GEOLOGINEN TUTKIMUSLAITOS

BULLETIN
DE LA
COMMISSION GÉOLOGIQUE
DE FINLANDE

N:o 196

SUOMEN GEOLOGISEN SEURAN JULKAISUJA
MEDDELANDEN FRÅN GEOLOGISKA SÄLLSKAPET I FINLAND
COMPTES RENDUS DE LA SOCIÉTÉ GÉOLOGIQUE DE FINLANDE

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GEOLOGICAL SURVEY OF FINLAND

OTANIEMI, FINLAND

HELSINKI 1961. VALTIONEUVOSTON KIRJAPAINO

SISÄLLYSLUETTELO — CONTENTS

SUOMEN GEOLOGISEN SEURAN HISTORIIKKI 1936—1961

KIRJOITTANUT

AARNE LAITAKARI

Kun Suomen Geologinen Seura vuonna 1936 vietti 50-vuotisjuhlaansa, julkaistiin seuran julkaisusarjan numerossa IX (ss. 1—64, 5 kuvaliitettä) professori Aarne Laitakarin kirjoittama seuran historiikki. Nyt on syytä tehdä selkoa tämänjälkeiseltä seuran 25-vuotistaipaleelta. Aikaisemmista tapahtumista puhuttaessa viitataan edellä mainittuun julkaisuun, mutta jotkut liitteet on tehty seuran koko toiminta-ajalta, ja jäsenluettelo on laadittu täydentämällä aikaisemmin tehty luettelo. Liitteessä 6 on esitetty vuosilta 1916—1923 Helsingin geologisen yhdistyksen tiedonannot, jotka sisältävät osittain esitelmien selostuksia.

1936—1945

Suomen Geologinen Seura vietti 50-vuotisjuhlaansa 22. 10. 1936. Juhlaa kunnioittivat läsnäolollaan silloinen Tasavallan presidentti P. E. Svinhufvud, pääministeri Kyösti Kallio sekä useat hallituksen jäsenet samoin kuin monien koti- ja ulkomaisten seurojen edustajat.

Juhlaa vietettiin tieteellisten seurojen juhlasalissa, joka oli kukin ja kivin koristettu, ja sen täytti runsas juhlapukuinen yleisö. Ylioppilaskunnan Soittajien esitettyä Sibeliuksen Andante festivon piti seuran puheenjohtaja, professori Aarne Laitakari tervehdyspuheen ja esitti seuran historiikin. Tämän jälkeen professori E. H. Kranck piti esitelmän: Suomen kallioperä ja kansainvälinen geologinen tutkimus, jonka jälkeen professori Pentti Eskola piti juhlaesitelmän aiheesta: Geologian merkitys elämän perustan tutkimuksessa ja ihmisen maailmankuvan perustana. Juhlakokouksen päätteeksi Ylioppilaskunnan Soittajat esitti Sibeliuksen Romanssin. Tämän jälkeen oli hotelli Seurahuoneessa ulkomaalaisten kutsuvieraiden ja seuran jäsenten keskeiset juhlaillalliset.

Seuran julkaisusarjan toimittaminen oli aloitettu edellisenä 10-vuotiskautena, ja sitä oli ehditty julkaista jo 8 numeroa. Tällä kaudella sarjan toimittamista jatkettiin, jolloin valmistuivat numerot IX—XVIII. Nämä sisälsivät 1 631 sivua sekä kuvaliitteitä ja karttoja. Julkaisutoiminta vaati paljon varoja, mutta kuten liitteestä 4 (rahatalous) käy selville, oli rahoituksesta aina selviydytty. Suurin osa rahoituksesta saatiin valtionapuna, mk 292 000, ja muita tuloja seuralla oli noin 78 000 mk. Tämän 10-vuotiskauden menot olivat noin 370 000 markkaa. Liitteessä 6 on selostettu seuran julkaisutoimintaa. On myöskin aiheellista mainita, että v. 1942 ilmestyneessä Geologische Rundschaun Suomi-niteessä oli 18 suomalaisen geologin kirjoittamaa tutkimusta.

Vuonna 1943 valittiin professori Pentti Eskola seuran ensimmäiseksi kunniajäseneksi.

Uusiksi kirjeenvaihtajajäseniksi valittiin tänä 10-vuotiskautena seuraavat: V. M. Goldschmidt 1936, Tadeusz Wojno 1936, Harry von Eckermann 1939, Richard Flint 1940, Richard J. Lougee 1940, Hans Cloos 1941. Aikaisemmin oli kirjeenvaihtajajäseniksi valittu seuraavat: Aladar Vendl 1931, Armin Öpik 1931, Eugene Wegman 1931, Nils Magnusson 1933. Vuonna 1945 seuraan siis kuului 10 kirjeenvaihtajajäsentä ja seuran jäsenmäärä oli kasvanut 93:sta 160:een. Seuran varsinaisista jäsenistä oli 32 ulkomaalaista.

Talvisota ja vv. 1941—45 sota vaikuttivat tietysti suuresti Suomen Geologisen Seuran toimintaan. Osallistuihan suurin osa sen jäsenistä eri tehtävissä sotatoimiin. Sodissa menettivät seuraavat seuran jäsenet henkensä: Erkki Mikkola 1940, Gunnar Brander 1940, Väinö Sihvonen 1940, Jalo Ant-Wuorinen 1940, Erkki Kuosmala 1940, Eero Malmivuo 1940, Reino Uusitalo 1940, Sampo Kilpi 1941, Onni Veikko Lumiala 1944, Erkki Austi 1941, Viljo Kanula 1941 ja Erkki Rosendal 1944. Lisäksi tänä kautena kuolivat seuran ansioituneet jäsenet W. W. Wilkman 1937, M. K. Palmunen 1938 ja V. Hackman 1941.

Seuran pääasiallisena toimintamuotona julkaisutoiminnan lisäksi ovat olleet kuukausikokoukset, joissa on pidetty esitelmiä geologian eri aloilta kaikkiaan 93. Kutsutut ulkomaalaiset ovat tänä kautena pitäneet 7 esitelmää. Osanotto kokouksiin on ollut vilkasta ja osanottajamäärä on keskimäärin ollut 35—40 jäsentä. Sotien aikana toiminta tietystikin oli vähäisempää. Liitteessä 5 on yksityiskohtainen luettelo pidetyistä esitelmistä. Kokoukset on yleensä pidetty yliopiston geologian laitoksen luentosalissa, ja yhdessäoloa ja keskustelua kokouksessa esillä olleista aiheista on jatkettu jossakin ravintolassa.

Suomen Geologisen Seuran kehityksessä heijastuu koko maan geologisen tutkimuksen voimakas kasvu. Samaan aikaan geologisen tutkimuslaitoksen henkilökunta on kasvanut ja sen toiminta on monella tavoin tehostunut ja myös vuoriteollisuuden piirissä on geologikunta saanut entistä enemmän tehtäviä ja geologinen henkilökunta on lisääntynyt.

1946—1955

Sodat ja niiden jälkeiset epävarmat olot maassamme eivät tietystikään voineet olla vaikuttamatta lamaannuttavasti seuran toimintaan, mutta vähitellen olot palautuivat entiselleen, ja seuran toiminta pääsi jatkumaan aikaisemmin viitoitettuun suuntaan, ja sen toiminnassa olikin tänä kymmenvuotiskautena havaittavissa ripeää kehitystä. Tänä aikana ilmestyivät julkaisusarjan numerot XIX—XXVIII, jotka sisälsivät 1 690 sivua. Täten useat tutkijat ovat saaneet tutkimustuloksensa julkisuuteen.

Kauan oli seuran piirissä keskusteltu oman tiedotuslehden aikaansaamisesta, ja vuonna 1949 alkoi ilmestyä Geologi-lehti, joka tästä lukien on ilmestynyt 10 kertaa vuodessa. Lehdessä on julkaistu lyhyehköjä kirjoituksia ja esitelmäselostuksia geologian eri aloilta, tiedotuksia sekä erityisesti geologeja kiinnostavia uutisia. Sivulukua kertyi näinä seitsemänä lehden ilmestymisvuotena yhteensä 452.

Julkaisutoiminnan aiheuttamat kulut on saatu rahoitetuksi valtionavun ja seuran muiden tulojen turvin. Valtionavun osuus oli tänä aikana 3 520 000 ja muut tulot 1 570 000 markkaa. Seuran vuosikertomukset ja tilit on julkaistu vuosittain Geologi-lehdessä. Samoin on siinä julkaistu lehden kirjoitusluettelo viisivuotiskausittain. Seuran kunniajäsenen, professori Pentti Eskolan 65-vuotispäivän johdosta ilmestyi vuonna 1947 seuran juhlajulkaisu numero XX ja vuonna 1949 professori Matti Sauramon 60-vuotispäiväksi juhlajulkaisu numero XXII.

Seuran kunniajäsen, professori Pentti Eskola valittiin vuonna 1953 hänen 70-vuotispäivänään seuran kunniapuheenjohtajaksi. Kunniajäseneksi kutsuttiin vuonna 1948 professori L. H. Borgström ja kirjeenvaihtajajäseniksi vuonna 1952 P. Niggli, B. Sander ja N. L. Bowen. Vuonna 1954 seuran kunniajäsen L. H. Borgström kuoli, ja kirjeenvaihtajajäsenistä kuolivat V. M. Goldschmidt 1947, H. Cloos 1951 ja P. Niggli 1953. Tämän kymmenvuotiskauden aikana seuran jäsenluku kasvoi 160:stä 304:ään, joista 11 ulkomaista kirjeenvaihtajajäsentä ja 73 ulkomaista jäsentä.

Vuonna 1950 seura toimeenpani geologisen retkeilyn Lappeenrantaan ja Vuoksenlaaksoon, ja vuonna 1954 seura järjesti VI Pohjoismaisen geologikongressin, johon liittyi retkeilyitä eri puolille Suomea. Kongressiin osallistui 186 henkilöä, joista 68 oli ulkomaalaisia, pääasiassa Pohjoismaista. Tämän kongressin järjestelyä ja valmistelua varten oli jo vuoden 1952 lopulla valittu toimikunta, johon kuuluivat professori Laitakari puheenjohtajana, tohtori Edelman sihteerinä, maisteri Repo rahastonhoitajana sekä muina jäseninä professorit Eskola, Sauramo ja Metzger, vuorineuvos Mäkinen, fil. tohtorit Härme, Salmi, Simonen ja Soveri, dipl. ins. Nieminen ja maisteri Lupander sekä seuran puheenjohtaja. Kongressin aiheuttamia kuluja varten saatiin valtionapua 300 000 markkaa, ja kun kulut kaikkiaan olivat

2 800 000 oli avustuksina hankittava 1 000 000 markkaa ja loput kerättiin lahjoituksina eri yhtiöiltä.

60-vuotisjuhlaansa seura vietti vaatimattomasti vain jäsentensä kesken vuonna 1946.

Seura menetti tänä kautena ansioituneista jäsenistään seuraavat: V. Tanner 1948, Th. Brenner 1949, H. Blankett 1949, B. Aarnio 1950, A. Hellaakoski 1952, I. Leiviskä 1953 ja E. Mäkinen 1953.

Seuran toimesta lyötettiin mitalit J. J. Sederholmista ja L. H. Borgströmistä.

1956—1961

Seura on toiminut verrattain edullisissa olosuhteissa, ja näin ollen se on voinut jatkaa ja tehostaa julkaisu- ym. toimintaansa. Tänä kautena ilmestyivät julkaisusarjan numerot XXIX—XXXII, jotka sisälsivät yhteensä 609 sivua. Geologi-lehti on jatkuvasti ilmestynyt ja sen sivumäärä on tänä 5-vuotiskautena kohonnut 516:een. Lehden sisältö on säilynyt samanlaisena kuin ennenkin. Seuran vuosikertomukset, tilit ja muut seuraa koskevat tiedotukset on julkaistu lehdessä samoin kuin lehden sisällysluettelo viisivuotiskaudelta. Julkaisutoimintansa seura on saanut rahoitetuksi valtionavun (3 800 000 markkaa) ja muiden tulojensa (829 998 markkaa) turvin.

Seuran kunniajäseniksi valittiin vuonna 1956 professori W. Wahl ja vuonna 1961 professorit C. E. Wegmann ja Aarne Laitakari. Uusiksi kirjeenvaihtajajäseniksi kutsuttiin vuonna 1960 professorit Tom Barth, Paul Ramdohr ja E. Szàdeczky-Kardoss. Kirjeenvaihtajajäsenistä kuolivat v. 1956 N. J. Bowen ja 1960 R. Lougee. Seuran jäsenistä kuolivat vuonna 1956 professorit J. G. Granö ja H. Väyrynen, v. 1958 professorit M. Sauramo ja H. Backlund ja v. 1959 professori A. Tammekann.

Seuran jäsenmäärä on kasvanut 304:stä 369:ään. Vuoden 1961 lopulla kuului seuraan 1 kunniapuheenjohtaja, 3 kunniajäsentä, 12 kirjeenvaihtajajäsentä, 91 ulkomaista jäsentä ja 264 suomalaista jäsentä eli siis yhteensä 369 jäsentä.

Seura teki v. 1957 paleontologisen ekskursion Gotlantiin. Matka suoritettiin lentäen ja siihen osallistui 24 jäsentä. Professori I. Hessland toimi asiantuntevana selostajana ja oppaana.

Vuonna 1958 seura maalautti kunniajäsenensä professori Pentti Eskolan muotokuvan, joka paljastettiin yliopiston geologisessa laitoksessa. Hänen 75-vuotispäivänään järjestettiin myös kansalaispäivälliset hotelli Kämpin peilisaliin. Tässä tilaisuudessa oli läsnä Tasavallan presidentti, useita ministereitä ja muita maan johtavia poliitikkoja, suuri joukko geologeja sekä lukuisia professori Eskolan henkilökohtaisia ystäviä.

Seuran toiminnasta on viimeisten 25 vuoden ajan ollut vastuussa kulloinenkin hallitus, mutta pääasiassa se on ollut vuosittain vaihtuvan puheenjohtajan ja joka toinen vuosi vaihtuvan sihteerin huolena. Liitteestä 2 käyvät ilmi, ketkä kulloinkin ovat olleet seuran johdossa. Vuodesta 1952 lähtien seuralla on ollut oma rahastonhoitajansa.

Seuran julkaisujen ja Geologi-lehden toimittajat ja toimituskunnat on mainittu liitteessä 6.

Seuran hallitus on huolehtinut myös suhteista ulkomaihin. Puheenjohtaja on huolehtinut edustustehtävistä ja hänen estyneenä ollessaan varapuheenjohtaja ja sihteeri.

Seura on eri tavoin kunnioittanut joukostaan poistuneitten muistoa, ja seuran puheenjohtaja on lähinnä seuraavassa kokouksessa esittänyt muistosanat.

Seuran suomalaisen jäsenistön ovat muodostaneet varsinaiset geologit, lähialojen tiedemiehet, geologiaa pääaineenaan opiskelevat ylioppilaat sekä muutamat harrastelijat. Seuran ulkomaiset jäsenet ovat ammattigeologeja, jotka tavalla tai toisella ovat olleet kiinnostuneita Suomen geologiasta. Kirjeenvaihtajajäseniksi on valittu ansioituneita ulkomaisia Suomen asioista kiinnostuneita tiedemiehiä. Kirjeenvaihtajajäsenistä ovat H. von Eckermann, N. H. Magnusson, H. Cloos, E. Wegmann ja P. Ramdohr esitelmöineet seurassamme.

Seura on säilyttänyt puhtaasti tieteellisen luonteensa, ja ammatillisten asioiden hoitamista varten perustettiin v. 1953 eri yhdistys, Geologiliitto.

Mineralogiaa harrastavilla seuran jäsenillä on oma yhdistyksensä, Suomen mineraloginen seura, joka perustettiin v. 1957, ja Vuorimiesyhdistyksen geologijaoston jäseniksi ovat liittyneet ne seuran jäsenistä, jotka ovat kiinnostuneita käytännöllisen geologian sovellutuksista.

V. 1958 perustettiin yhdistys Geologian naiset. Sen tarkoituksena on toimia geologian piiriin kuuluvien naisten kesken yhteenkuuluvaisuuden ja yhteisten henkisten harrastusten ylläpitämiseksi ja virkistyksen saamiseksi. Tarkoitustaan toteuttaakseen se toimeenpanee kokous-, esitelmä- ym. tilaisuuksia ja retkeilyjä.

SUMMARY

OF THE HISTORY OF THE GEOLOGICAL SOCIETY OF FINLAND 1936—1961

BY

AARNE LAITAKARI

The fiftieth anniversary of the Geological Society of Finland saw the publication of a history of the organization by Aarne Laitakari, including appendices and a summary in English. Now, on the occasion of the seventy-fifth anniversary of the society, I have brought the history up to date and revised the appendices.

The activities of the society during the past twenty-five year's have followed the same general pattern as before but undergone development in many ways. The membership now includes an honory president, honorary members, corresponding members, foreign members and native members. Honorary membership has been awarded members who have served the organization meritoriously, while deserving men of science abroad who have taken a kindly view of Finland in one way or another have been invited to become corresponding members.

The membership of the society has grown considerably during the period depicted. In 1936 it had six corresponding members, twenty-two foreign members and seventy-one native members, whereas now, in 1961, the corresponding figures are: one honorary president, three honorary members, twelve corresponding members, ninety-one foreign members and 264 native members.

The lecture program of the society is presented in Appendix 5, which lists all the lectures held each year. Appendix 6 deals with the publishing activities of the society. The series of publications brought out by the society now includes 32 volumes, totalling 3 837 pages. Since 1949 the society has also published a journal named *Geologi*, which contains short articles, reports on lectures and similar material. It has been appearing ten times a year for a total of 968 pages.

Appendix 3 lists the members throughout the seventy-five years of the society's existence, and Appendix 2 lists the presidents and secretaries. During the Finnish wars, activity was slight, for nearly all the members were in uniform; but at no time did activity cease altogether. Twelve members of the society were killed in the wars.

In 1954 the society organized the VI Nordic Geological Conference, as well as the excursions held in conjunction with the event, and in 1960 it took part in organizing the International Geological Congress.

The activities of the society have included meetings and lectures, the publication of literature and the organization of occasional excursions.

When Prof. Pentti Eskola, the society's honorary member, observed his sixty-fifth birthday in 1947, No. XX in the society's series of publications was dedicated to him, and when he reached the age of seventy in 1953, he was elected honorary president of the society.

The Geological Society of Finland has continued to be the unifying organ of all the geologists in Finland. It has remained a purely scientific association. In 1953 a separate organization, the *Geologiliitto* (Union of Geologists), was established to look after the economic and other material interests of Finnish geologists. In 1957 the Mineralogical Society of Finland was founded for the benefit of geologists particularly interested in mineralogy. Those geologists who are interested in the practical application of geology can join the Geological Section of the Society of Mining Engineers.

Liite 1
Appendix 1

SUOMEN GEOLOGISEN SEURAN SÄÄNNÖT

(Hyväksytty 7. 12. 1950 ja 29. 1. 1953)

Nimi ja tarkoitus

1 §. Suomen Geologinen Seura — Geologiska Sällskapet i Finland r. y. on rekisteröity yhdistys, jonka kotipaikka on Helsingin kaupunki ja jonka viralliset kielet ovat suomi ja ruotsi. Yhdistyksestä käytetään näissä säännöissä nimitystä seura.

2 §. Seuran tarkoituksena on edistää geologista ja mineralogista tutkimusta maassamme sekä lisätä harrastusta näihin tieteisiin.

Näiden tarkoitustensa toteuttamiseksi seura pitää kokouksia, toimittaa ja kustantaa tieteellisiä ja yleistajuisia julkaisuja, toimeenpanee retkeilyjä sekä käyttää muita samantapaisia keinoja. Seura voi harjoittaa julkaisutoimintaa, ottaa vastaan lahjoituksia, jälkisäädöksiä sekä omistaa kiinteistöjä. Tarvittaessa hankkii seura toimintaansa varten asianomaisen luvan.

Jäsenet

3 §. Seuran jäsenistön muodostavat kunniapuheenjohtaja, kunniajäsenet, kirjeenvaihtajajäsenet sekä kotimaiset ja ulkomaiset jäsenet.

4 §. Kunniapuheenjohtajakseen seura voi kutsua kotimaisen kunniajäsenensä. Kunniapuheenjohtajia seurassa voi olla vain yksi kerrallaan. Kunniapuheenjohtajaksi kutsumisesta on seuran hallituksen tehtävä kirjallinen ehdotus seuran kokoukselle. Kunniapuheenjohtajan valinta suoritetaan samoin kuin jälempänä on kunniajäsenten valitsemisesta määrätty.

Kunniajäseneksi voidaan kutsua kotimainen tai ulkomainen henkilö, jolla on harvinaisen suuret seuran toiminta-alaan kuuluvat tieteelliset ansiot tai joka muulla toiminnallaan on erikoisen tehokkaasti edistänyt seuran pyrkimyksiä. Kunniajäsenten luku on rajoitettu kymmeneksi. Kunniajäseneksi kutsumisesta on kahden kotimaisen jäsenen tehtävä perusteltu kirjallinen ehdotus hallitukselle. Jos hallitus puoltaa ehdotusta, on vaali toimitettava kahdessa perättäisessä kokouksessa, jolloin hyväksymiseen vaaditaan puoltavia ääniä vähintään kolme neljäsosaa sanotuista äänistä.

Kirjeenvaihtajajäseneksi voidaan kutsua ulkomailla asuva henkilö, joka on tieteellisesti ansioitunut jollakin suomalaisia geologeja kiinnostavalla alalla tai muuten edistänyt seuran tarkoitusperiä. Kirjeenvaihtajajäseneksi kutsuminen on tehtävä samalla tavalla kuin kunniajäsenen.

Ulkomaiseksi jäseneksi seura voi valita kahden kotimaisen jäsenen ehdotuksesta jokaisen sitä haluavan ulkomaisen geologin.

Henkilöitä, jotka eivät ole Suomen kansalaisia, saa seuran jäseninä olla enintään 1/3 seuran koko jäsenluvusta.

Kotimaiseksi jäseneksi seura voi valita kahden jäsenen ehdotuksesta Suomen kansalaisen, jonka tiedoillaan tai toiminnallaan katsotaan voivan edistää seuran tarkoitusperiä.

5 §. Kunniapuheenjohtajalla, kunniajäsenillä ja kirjeenvaihtajajäsenillä ei ole mitään taloudellisia velvollisuuksia. Ulkomaiset ja kotimaiset jäsenet suorittavat joko vuosimaksun tai kertakaikkisen jäsenmaksun, joka on voimassa oleva vuosimaksu 15 kertaisena. Jäsenen, joka 20 vuotena on suorittanut jäsenmaksunsa, seuran hallitus voi vapauttaa jäsenmaksun suorittamisesta. Vuosimaksun suuruudesta päätetään vuosikokouksessa.

6 §. Jäsen voi erota seurasta ilmoittamalla siitä suullisesti seuran kokouksen pöytäkirjaan tai kirjallisesti seuran hallitukselle tai sen puheenjohtajalle. Ellei jäsen huomautuksesta huolimatta ole suorittanut edellisen vuoden jäsenmaksua, seuran hallitus voi erottaa hänet seurasta.

Hallinto

7 §. Seuran asioita hoitaa yhdistyslain, näiden sääntöjen ja seuran kokousten päätösten määräämällä tavalla seuran hallitus. Hallitus on päätösvaltainen neljän sen jäsenen ottaessa osaa asiain käsittelyyn, puheenjohtaja tai varapuheenjohtaja mukaan luettuina.

Seuran hallituksen muodostavat puheenjohtaja, joka samalla on seuran puheenjohtaja, varapuheenjohtaja, sihteeri, rahastonhoitaja ja kolme lisäjäsentä, joiden kaikkien on oltava kotimaisia jäseniä.

8 §. Varapuheenjohtajan ja lisäjäsenet valitsevat kotimaiset jäsenet vaalikokouksessaan yhdeksi vuodeksi, eikä heitä voida valita samaan tehtävään lähinnä seuraavaksi vuodeksi. Kuluvan vuoden varapuheenjohtaja siirtyy seuraavan vuoden puheenjohtajaksi. Samassa kokouksessa valitaan vuorovuosina sihteeri ja rahastonhoitaja kahdeksi vuodeksi. Milloin seuran toimihenkilön tai hallituksen lisäjäsenen paikka vapautuu ennen toimikauden päättymistä tai asianomainen on jostakin syystä määräajan estetty hoitamasta tehtäviään, täytetään se toimikauden loppuun seuran seuraavassa kokouksessa.

Sihteerille ja rahastonhoitajalle maksetaan hallituksen määräämä palkkio.

9 §. Seuran toiminta- ja tilivuosi on kalenterivuosi.

10 §. Seuran nimen kirjoittaa puheenjohtaja tai varapuheenjohtaja, jompi kumpi yhdessä sihteerin kanssa.

11 §. Puheenjohtaja johtaa puhetta seuran ja hallituksen kokouksissa. Hän on oikeutettu ottamaan osaa asiain käsittelyyn kaikissa seuran toimikunnissa.

Puheenjohtajan ollessa estynyt varapuheenjohtaja hoitaa hänen tehtäviään.

Sihteeri valmistelee seuran ja hallituksen kokoukset sekä neuvoteltuaan ensin puheenjohtajan kanssa kutsuu ne koolle, pitää pöytäkirjaa kokouksissa, huolehtii näissä tehtyjen päätösten toimeenpanosta, laatii vuosikertomuksen, pitää luetteloa seuran jäsenistä sekä hyväksyy laskut ja valvoo seuran julkaisutoimintaa hallituksen päätösten mukaisesti. Hänen ollessaan estynyt toimii rahastonhoitaja hänen sijaisenaan.

Rahastonhoitaja hoitaa seuran varoja ja pitää niistä tiliä, sekä huolehtii seuran omaisuudesta ja julkaisujen jakelusta. Hänen on esitettävä hallituksen puolesta edellisen vuoden tilit tilintarkastajille tammikuun kuluessa.

Hallitus voi jättää sopiviksi katsomiaan kysymyksiä ja asioita toimikunnan tai yksityisen jäsenen käsiteltäväksi ja hoidettavaksi.

12 §. Vuosikokouksessa valitaan kaksi tilintarkastajaa ja heille kaksi varamiestä tarkastamaan kuluvan vuoden taloudenhoitoa ja tilejä. Heidän on annettava niistä lausunto hallitukselle seuraavalle vuosikokoukselle esitettäväksi enintään 3 päivää ennen kokousta.

Kokoukset

13 §. Seuran kokouksissa pidetään esitelmiä, keskustellaan ja käsitellään seuran työalaan kuuluvia asioita. Niissä on kaikilla jäsenillä puhevalta. Jäsenien, jotka haluavat pitää kokouksissa esitelmiä tai tiedonantoja, on sovittava siitä ennakolta sihteerin kanssa.

Myös seuraan kuulumattomat voivat puheenjohtajan tai sihteerin luvalla olla läsnä kokouksissa.

14 §. Seuran vaalikokous pidetään joulukuussa. Sihteeri voi kuitenkin puheenjohtajan kanssa neuvoteltuaan kutsua seuran tarpeen vaatiessa kokoukseen vaaleja varten muulloinkin.

15 §. Helmikuussa pidettävässä vuosikokouksessa

— esitetään edellisen vuoden toimintakertomus,

— esitetään edellisen vuoden tilinpäätös ja tilintarkastajien lausunto edellisen vuoden taloudenhoidosta ja tileistä,

— päätetään tilinpäätöksen vahvistamisesta ja vastuuvapauden myöntämisestä hallitukselle ja rahastonhoitajalle,

— valitaan tilintarkastajat ja heidän varamiehensä,
— määrätään kuluvan vuoden vuosimaksu.

16 §. Kaikilla kotimaisilla jäsenillä on äänioikeus ja äänestyksissä ratkaisee, ellei säännöissä ole muualla toisin määrätty yksinkertainen äänten enemmistö ja äänten mennessä tasan puheenjohtajan ääni.

17 §. Kutsut kokouksiin on toimitettava jäsenille joko viimeistään neljä päivää ennen kokousta postitse tai viimeistään neljä päivää ennen julkaistulla sanomalehti-ilmoituksella. Kutsuttaessa kokoukseen, jossa käsitellään yhdistyslain 14 §:ssä mainittuja asioita, on siitä kokouskutsussa erikseen mainittava.

Erityissäädöksiä

18 §. Sääntöjen muuttamista haluttaessa on siitä seuran jonkun tai joidenkin jäsenien tehtävä kirjallinen ehdotus hallitukselle, jonka on valmistettava asia ja esitettävä se seuran käsiteltäväksi kahdessa perättäisessä kokouksessa. Muutosehdotuksen hyväksymiseen vaaditaan, että vähintään kaksi kolmasosaa näissä kokouksissa annetuista äänistä sitä puoltaa.

19 §. Ehdotus seuran purkamisesta käsitellään samalla tavalla kuin sääntöjen muutosehdotus.

20 §. Jos seura purkautuu tai lakkautetaan, on sen jäljelle jääneet varat käytettävä sellaisiin tarkoituksiin, joiden hyväksi seura on toiminut.

THE BY-LAWS OF THE GEOLOGICAL SOCIETY OF FINLAND

(Revised Dec. 17, 1950 and Feb. 19, 1953)

Name and purpose

1 §. Suomen Geologinen Seura — Geologiska Sällskapet i Finland ry. is a registered corporation, the headquarters of which is situated in the city of Helsinki and the official languages of which are Finnish and Swedish. The corporation is called Society in these By-Laws.

2 §. The purpose of the Society shall be to promote geological and mineralogical research work in our country and to foster interest in these sciences.

In pursuance of its aims the Society shall hold meetings, edit and publish scientific and popular works, arrange excursions and employ similar methods. The Society may act as publisher, accept donations and legacies and possess real estate. When necessary the Society shall obtain permission from those concerned for its operations.

Membership

3 §. The membership of the Society shall be classified as Honorary President, Honorary Members, Correspondents, National Members and Foreign Members.

4 §. The Society may elect a National Honorary Member as its Honorary President. There shall not be more than one Honorary President at a time. The motion to elect an Honorary President shall be made by the Council. The election of the Honorary President shall be conducted in the same way as that of an Honorary Member, as prescribed in the following.

Honorary Membership may be extended to any person of Finnish or foreign nationality who is exceptionally distinguished for his scientific attainments in the field represented by the Society or who has effectively fourthered the aims of the Society through other activities. The number of Honorary Members shall be limited to ten. Nomination to Honorary Membership shall call for a well-founded proposition by two Finnish members submitted in writing to the Council. If the Council recommends the nomination, the election shall take place at two successive meetings, when three fourths of the votes are required in favor of the motion for final acceptance.

Eligible for election as Correspondent is any person resident abroad who is noted for scientific achievements of special interest to Finnish geologists or who has otherwise furthered the aims of the society. Invitation to membership of a Correspondent shall be carried out in the same way as of an Honorary Member.

Any foreign geologist desirous of membership may be elected to the Society as a Foreign Member if proposed by two Finnish members.

The number of Foreign Members shall not exceed a third of the total membership.

Any Finnish citizen whose knowledge or activity is deemed likely to further the ends of the Society may be elected a Member on the proposition of two Members.

5 §. The Honorary President, Honorary Members and Correspondents shall not be required to pay dues. Members and Foreign Members shall pay either the annual dues or the dues for a life membership, which is a single payment fifteen times the annual dues. The Council of the Society may free from further payments any Member who has payed the annual dues for 20 years. The amount of the annual dues shall be decided at each annual meeting.

6 §. A Member may announce withdrawal from the Society orally at a meeting for the minutes or by a written notification to the Council or to the President of the Society. If a Member, disregarding admonition, has not payed his dues for the previous year, his membership may be terminated by the Council of the Society.

Administration

7 §. The affairs of the Society shall be administered by the Council according to the By-Laws and the resolutios of the meetings. Four members of the Council, including the Chairman or the Vice-Chairman, shall constitute a quorum in dealing with affairs of the Society.

The Council of the Society shall be formed by the Chairman, who is at the same time the President of the Society, the Vice-Chairman, the Secretary, the Treasurer, and three additional members, all of whom shall be national members.

8 §. The Vice-Chairman and additional members of the Council shall be elected by national members for a term of one year at the electoral meeting,

and they shall not be eligible for the same office the following year. The Vice-Chairman of the current year succeeds to the chairmanship the following year. The Secretary and the Treasurer shall be elected at the same meeting every other year for a term of two years. Upon a vacancy's occurring in the Council or in any other office before the end of the term or when an officer is prevented from performing his duties for any length of time, the vacancy shall be filled by the next meeting for the remainder of the term.

The Secretary and the Treasurer shall receive compensation as fixed by the Council.

9 §. The Fiscal Year of the Society is the calendar year.

10 §. The President and the Vice-President, one of them together with the Secretary, shall be authorized to sign the name of the Society.

11 §. The President shall preside at the meetings of the Society and the Council. He shall be entitled to participate in the work of all the Committees of the Society.

The Vice-President shall assume the powers and duties of the President in the latter's absence.

The Secretary shall prepare the Meetings of the Council and the Society and he shall convene them after consultation with the President, keep the minutes at the meetings, see to the execution of resolutions passed at meetings, submit an Annual Report, keep a complete list of members, accept bills, and supervise the publishing activities of the Society according to the resolutions of the Council. In his absence the Treasurer shall act as his deputy.

The Treasurer shall be in charge of the funds of the Society, and keep records of them, look after all property belonging to the Society and manage the distribution of its publications. He shall submit the annual financial statement on behalf of the Council to the Auditors during the month of January.

The Council may entrust appropriate questions and tasks to a Committee or individual member to be dealt with.

12 §. At the Annual Meeting two Auditors and their deputies shall be elected to examine the accounts and financial operations of the current year. They shall deliver their report to the Council not later than three days before the Annual Meeting, to which it shall be submitted.

Meetings

13 §. At the meetings lectures shall be held, and matters belonging to the sphere of interest of the Society shall be discussed and dealt with. All members shall have the right to speak at meetings. Members who want to

give a talk or report shall make arrangements in advance with the Secretary.

Also non-members may be allowed by the President or the Secretary to be present at meetings.

14 §. The electoral meeting shall be held in December. The Secretary may, after consultation with the President, convene the Society for an electoral meeting at other times as well.

15 §. The order of business at the Annual Meetings in February shall be the following:

— Annual Report for the past year is submitted.

— The financial statement of the past year and the report of the Auditors upon the accounts and business management are submitted.

— Decision is made upon approving the accounts and releasing the Council and the Treasurer from responsibility.

— The Auditors and their deputies are elected.

— The Annual Dues are fixed for the current year.

16 §. All national members have the right to vote and a simple majority shall be decisive; in the event of a tie the vote of the President shall be decisive.

17 §. The members shall be called to meetings on not less than four day's notice by post or by an announcement in the newspapers published not less than four days in advance. If matters touched on in Section 14 of the Association Act are to be dealt with at any meeting mention of them shall be made in the invitation to attend.

Special Regulations

18 §. Any member or members of the Society proposing an amendment of the Rules shall submit a written proposal to the Council, who shall investigate and prepare it and submit it to the Society at two successive meeting. The proposal of an amendment shall be adopted if two thirds of the votes given at these meetings are in favor of it.

19 §. Any proposal to dissolve the Society shall be treated in the same way as a proposed amendment.

20 §. If the Society be dissolved or suppressed, the remaining funds shall be used toward the same ends as the Society pursued during its period of activity.

Liite 2
Appendix 2

GEOLOGISEN SEURAN PUHEENJOHTAJAT, SIHTEERIT JA RAHASTONHOITAJAT 1937—1961

THE PRESIDENTS, VICE-PRESIDENTS, SECRETARIES AND TREASURERS OF THE GEOLOGICAL SOCIETY 1937—1961

	Puheenjohtaja President	Varapuheenjohtaja Vice-President	Sihteeri Secretary	Rah.hoitaja Treasurer
1937	W. W. Wilkman	Erkki Mikkola	Esa Hyyppä	
1938	Erkki Mikkola	Heikki Väyrynen	Th. G. Sahama	
1939	Heikki Väyrynen	Erkki Kivinen	»	
1940	»	»	Sampo Kilpi	
1941	Erkki Kivinen	Aaro Hellaakoski	Sampo Kilpi ja Martti Salmi	
1942	Aaro Hellaakoski	Esa Hyyppä	Martti Salmi	
1943	Esa Hyyppä	Th. G. Sahama	»	
1944	Th. G. Sahama	Martti Salmi	V. Pääkkönen	
1945	Martti Salmi	Paavo Haapala	Kalervo Rankama	
1946	Paavo Haapala	Aarne Laitakari	» ja K. Virkkala	
1947	Aarne Laitakari	Erkki Aurola	K. Virkkala	
1948	Erkki Aurola	Ahti Simonen	»	
1949	Ahti Simonen	Martti Saksela	Mauno Härme	
1950	Martti Saksela	A. A. Th. Metzger	Simo Kaitaro	
1951	A. A. Th. Metzger	Kalervo Rankama	»	
1952	Kalervo Rankama	Maunu Härme	V. Okko	Reino Repo
1953	Maunu Härme	Urpu Soveri	»	»
1954	Urpu Soveri	Veikko Vähätalo	Arvo Matisto	»
1955	Veikko Vähätalo	Martti Salmi	»	I. Salli
1956	Martti Salmi	K. J. Neuvonen	Valto Veltheim	»
1957	K. J. Neuvonen	K. Virkkala	»	»
1958	K. Virkkala	Aarno Kahma	H. Ignatius	»
1959	Aarno Kahma	Juhani Seitsaari	»	»
1960	Juhani Seitsaari	Karl Mölder	Raimo Lauerma	»
1961	Karl Mölder	Aimo Mikkola	»	Kauko Meriläinen

Liite 3
Appendix 3

SUOMEN GEOLOGISEN SEURAN JÄSENLUETTELO 1886—1961

LIST OF MEMBERSHIP OF THE GEOLOGICAL SOCIETY OF FINLAND 1886—1961

Seuran jäsenluetteloita alkuajoilta ei ole säilynyt ollenkaan. Ensimmäinen sangen epätäydellinen on vuodelta 1897, ja sitä seuraava vuodelta 1906. Täydellistä ja täysin luotettavaa jäsenluetteloa ei siis ole voitu tehdä. Tähän luetteloon on alkuajoilta otettu ne pöytäkirjoissa mainitut henkilöt, jotka otaksuttavasti ovat olleet jäseninä. Liittymisvuodet ovat usein epävarmoja; tietojen puutteessa on täytynyt ottaa se vuosi, jolloin henkilö on ensi kerran pöytäkirjoissa mainittu. Vuodesta 1914 lähtien ovat tiedot luotettavampia; tarkat jäsenluettelot on vasta v:sta 1928 lähtien.

Kunniapuheenjohtaja — Honorary President

Eskola, Pentti, prof.	1953—

Kunniajäsenet — Honorary Members

Eskola, Pentti, prof.	1943— kunnia-puh.joht. 1953
Borgström, L. H., prof.	1948—†1954
Wahl, Walter, prof.	1956—
Laitakari, Aarne, prof.	1961—
Wegmann, Eugen, prof.	1961—

Kirjeenvaihtajajäsenet — Corresponding members

Vendl, A., prof., Hungary	1931—
Öpik, A., prof., Australia	1931—
Wegmann, Eugen, prof., Switzerland	1931—1961 Kunniajäsen v:sta 1961
Magnusson, Nils H., prof., Sweden	1933—
Goldschmidt, V. M., prof., Norway	1936—†1947
Wojno, Tadeusz, prof., Poland	1936—
von Eckermann, Harry, prof., Sweden	1939—
Lougee, Richard, prof., USA	1940—†1960
Flint, Richard F., prof., USA	1940—
Cloos, Hans, prof., Germany	1941—†1951

Cloos, Ernst, prof., USA	1947—
Niggli, Paul, prof., Switzerland	1952—†1953
Sander, Bruno, prof., Austria	1952—
Bowen, Norman S., prof., USA	1952—†1956
Barth, Tom. F. W., prof., Norway	1960—
Ramdohr, Paul, prof., Germany	1960—
Szàdeczky-Kardoss, Elemer, prof., Hungary	1960—

Ulkomaiset jäsenet — Foreign members

Tegengren, F. R., fil. toht.	1897—1905
Kranck, E. H., prof., Canada (Finland)	1916—
Backlund, Helge, prof., Sweden	1917—†1958
Wegmann, Eugen, prof., Switzerland	1927—1931 kirj. vaiht.jäsen v:sta 1931 kunnia » » 1961
Frauenfelder, H., toht. ins.	1924—1926
Barbour, G. E., prof., USA	1928—
Geol.-pal. Inst. d. Univ., Basel	1928—1931
Min.-petr. Inst. d. Univ., Basel	1928—
Collet, L. W., prof., Switzerland	1929—†1957
Argand, E., prof., Switzerland	1931—1940
Bruce, E. L., prof., Canada	1931—†1949
Bütler, H., fil. toht., geol., Switzerland	1931—
Dellwik, B., vuori-ins., Sweden	1931—1934
Fisch, E. W., fil. toht., Switzerland	1931—1934
Staub, R., prof., Switzerland	1931—1932
Ebert, H., prof., Germany	1933—
Gevers, T. W., prof., South Africa	1933—
Homma, F., prof., Japan	1933—
Grüssner, A. J., cand. rer. nat.	1934—1935
Noe-Nygaard, Arne, prof., Denmark	1934—
Wenk, Eduard, fil. toht., Switzerland	1934—
Malkowski, St., prof., Poland	1935—
Mellis, Otto., prof., Sweden	1935—
Zāns, Verners, fil. toht., Jamaica	1936—
Hietanen, Anna, fil. toht., USA (Finland)	1936—
Popoff, Boris, prof., Latvia	1936—†1950
Zeidler, Waldemar, vuori-ins., Sweden	1936—
Hall, A. Jean, M. A., Great Britain	1937—
Joensuu, Oiva, fil. maist., USA (Finland)	1937—
Mogensen, Fredrik, vuori-ins., Sweden	1938—1951
von Knorring, Oleg, fil. maist., Great Britain (Finland)	1939—
Adamson, Olge, fil. toht., Norway	1941—
Grip, Erland, fil. toht., Sweden	1942—
Dahlström, Elis, fil. lis., Sweden	1942—
Du Rietz, Torsten, fil. toht., Sweden	1942—
von Gaertner, Hans Rudolf, fil. toht., Germany	1942—
Eigenfeld, Rolf, fil. toht., Germany	1942—1945
Martna, Jüri, fil. lis., Sweden	1944—

Jaanusson, Valdar, fil. lis., Sweden	1944—
Gavelin, Sven, prof., Sweden	1945—
Hjelmqvist, Sven, prof., Sweden	1945—
Kulling, Oscar, fil. toht., Sweden	1945—
Landergren, Sture, fil. toht., Sweden	1945—
Meier, Otto, fil. toht., Sweden	1945—
Wickman, Frans E., prof., Sweden	1945—
Ödman, Olof H., prof., Sweden	1945—
Brognon, Georges, ins. geologi, Portugal	1945—
Ahrens, L. H., fil. toht., USA	1947—1956
Fromm, Erik, fil. toht., Sweden	1947—
van Straaten, L. M. J. U., fil. toht., Netherlands	1947—
Frederickson, A. F., prof., USA	1947—1956
Roubault, Marcel, prof., France	1948—
Hagner, Arthur, fil. toht., USA	1948—
Suisse, M., geologi, Ivory Coast	1949—
Järnefors, Björn, fil. lis., Sweden	1949—
Marklund, Nils, fil. lis., Sweden	1949—
Hejtman, Bohnslaw, fil. toht., Czecho-Slovakia	1949—
de Waard, D., fil. toht., Netherlands	1949—
Chen Kwan Yan, fil. lis., China	1949—
Graeter, Paul, fil. toht., Switzerland	1950—
Kullerud, Gunnar, fil. toht., Norway	1950—
Preston, John, fil. toht., Great Britain	1950—
Shaw, Dennis, fil. toht., Canada	1950—
Wells, Maurice, fil. toht., Great Britain	1950—
Sahlin, Anders, fil. maist., Sweden	1951—
Tisler, Janko, fil. toht., Jugoslavia	1952—
Eriksson, Tryggve, fil. lis., Sweden	1952—
Kautsky, Gunnar, fil. toht., Sweden	1952—
San Miguel, Alfredo, prof., Spain	1952—
Fuster, José, prof., Spain	1952—
Yoder, Hatten S., fil. toht., USA	1952—
Melcher, Geraldo, dipl. ins., Brasil	1952—
Sandrea, Andre, toht., France	1952—1954
Watanabe, Takeo, prof., Japan	1952—
MacLeod, John, fil. maist., Canada	1952—
Heurtebize, Georges, geologi, France	1953—
Lelubre, Maurice, fil. toht., Algeria	1953—
Dutt, A., geologist, India	1953—1954
Föyn, Sven, fil. toht., Norway	1953—
Gjelsvik, Tor, fil. toht., Norway	1953—
Schachner, Doris, fil. toht., Germany	1953—
Welin, Eric, fil. lis., Sweden	1953—
Shuaib, Saiyed Mohammed, fil. toht., Pakistan	1953—
Brotzen, Otto, fil. lis., Sweden	1953—
Disler, Jürg, fil. toht., Canada	1954—
Ryan, Donald J., fil. toht., USA	1954—
Linnman, Gunnel, fil. maist., Sweden	1955—
Carstens, Harald, fil. kand., Norway	1956—

Franco, Rui Ribeiro, prof., Brasil	1956—
Günthert, A., fil. toht., Switzerland	1956—
Hessland, Ivar, prof., Sweden	1956—
Moore-Levy, Georg, fil. toht., Ireland	1956—
Behr, Hansjürgen, fil. maist., Germany	1957—
Niemczynow, Grazyna, fil. maist., Poland	1957—
Schermerhorn, L. J. G., fil. toht., Angola	1957—
Warren, Harry, prof., Canada	1957—
Gault, Hugh R., fil. toht., USA	1958—†1961
Blake, Weston, fil. maist., USA	1958—
Kalliokoski, Jorma, prof., USA	1958—
Ramdohr, Paul, prof., Germany	1958— kirj. vaihtajajäsen 1960
Joplin, Germaine, fil. toht., Australia	1958—
Frietsch, Rudyard, fil. lis., Sweden	1958—
Misař, Zdeněk, fil. toht., Czecho-Slovakia	1958—
Neumann, Henrich, fil. toht., Norway	1959—
Filby, Royston, fil. maist., Great Britain	1959—
Cooray, Gerald, fil. toht., Ceylon	1960—
Jónsson, Jón, fil. lis., Iceland	1960—
Laruelle, Jacques, fil. toht., Belgium	1960—
Sarap, Hans, fil. lis., Sweden	1960—
Armstrong, Herbert, prof., Canada	1961—
Vera, Elpidio, M. Sc., Philippines	1961—
Ojakangas, Richard, M. A., USA	1961—
Dudek, Arnost, Dr., Czecho-Slovakia	1961—
Clark, Alan, B. Sc., Great Britain	1961—

Jäsenet — Members

Wiik, F. J., fil. toht., prof.	1886—1909
Moberg, K. Ad., fil. maist., geologisen toimikunnan johtaja	1886—1901
Tigerstedt, A. F., vuori-ins., tod. valtioneuvos, teoll. hallituksen yli-intendentti	1886—1918
Gylling, Hj., fil. maist., valtiongeologi	1886—1888
Solitander, C. P., vuori-ins., teoll. hallituksen johtaja	1886—1918
Sederholm, J. J., fil. toht., prof., geologisen toimikunnan johtaja	1886—1934
Arppe, N., fil. maist.	1886?
Warén, A. J., lääket. lis.	1886?
Frosterus, Benj., fil. toht., valtiongeologi, maatutkim. laitoksen johtaja	1886—1931
Stolpe, Th., ylim. geologi	1886?
Aartovaara (Abrahamsson), G., vuori-ins., lehtori	1886—1923
Stjernwall, H. J., fil. maist.	1886—1896
Furuhjelm, E. Hj., vuori-ins., vuorihallituksen intendentti	1886?
Ramsay, Wilhelm, fil. toht., prof.	1887—1928
Lisitzin, G., vuori-insinööri	1887—1914
Helaakoski, A. R., fil. toht., vanh. lehtori	1887—†1945
Plathan, A., fil. toht., vanh. lehtori	1891—1906
Berghell, Hugo, fil. toht., valtiongeologi	1891—1922

af Schultén, Aug., fil. toht., dosentti	1893—1899
Hackman, V., fil. toht., prof., valtiongeologi	1894—1941
Ailio (Ax), J. E., fil. toht., dosentti, kansallismuseon osastonjohtaja	1895—1933
Katila, E., toimittaja	1896—1897
Lindberg, H., fil. toht., yliopiston kustos	1896—1923
Åkerblom, V. L., ins., geologisen toimikunnan geodeetti	1897—1906
Wilkman, W. W., fil. toht., valtiongeologi	1897—1937
Ponsi (Pönnelin), Arvid, liikemies	? —1897
Kivilinna (Borg), V., fil. maist., vanh. lehtori	
Herlin, Rafael, fil. toht.	? —1897
Tappura, Aku, ylioppilas, ylimäär. geologi	? —1897
Skogström, A. M., lääket. lis.	? —1897
Otto, C. M., pääkonsuli	1897—1906
Borgström, L. H., fil. toht., prof.	1897—1948 kunniajäsen v:sta 1948
Segercrantz, W., fil. maist., pankinjohtaja	1897
Nyholm, E. T., metsänhoitaja	1897—1908
Sarlin, Emil, vuori-ins., vuorineuvos, Paraisten Kalkkivuori Oy:n johtaja	1897—1914
Nordenskiöld, Erik., fil. toht., dosentti	1897—1906
Sandelin, S., fil. toht.	? —1897
Rosberg, J. E., fil. toht., prof.	1897—1932
Mattsson, Gust., fil. toht., kirjailija	1897
Luther, A. F., fil. toht., prof.	1897—1906
Poppius, B. B.	1897?—1914
Wahl, W., fil. toht., prof.	1897—1956 kunniajäsen v:sta 1956
Hartwall, G., fil. toht., dosentti	1897—?†1945
Rindell, A., fil. toht., prof.	1897?—1919
Florin, E., ins.	1897?—1919
Aschan, O., fil. toht., prof.	1897—1914
Jantunen, P., ylioppilas, liikemies	1897—?
Lagerblad, S.	1897—†1928
Sundell, I. G., vuori-ins., rahapajanjohtaja	1897—
Sergelius (Sergejeff), M., fil. maist., teollisuuskoulun joht.	1897—1906
Trüstedt, Otto, vuori-ins., fil. toht., v. t. valtiongeologi	1904—1929
Aminoff, G., vapaaherra, vuori-ins.	1904
Granö, J. G., fil. toht., prof.	1904—†1956
Graeffe, F., ylioppilas, ylim. geologi	
Holm, J. R. G.	
Kopperi, E., fil. maist.	1904
Saarinen, J. H., ins.	1905
Tanner, Väinö, fil. toht., valtiongeologi, prof.	1906—†1948
Boldt, R., fil. toht.	1906
Buss., H., lääket. lis.	
von Christiersson, A., fil. maist., koulunjohtaja	
von Julin, A., vuori-ins., tilanomistaja	1906—1918
Franck, R., ins., tilanomistaja	1906—1919
Leiviskä, I., fil. toht., prof.	1906—1953
Krohn, L., ins.	

Blankett, H., vuori-ins., teollisuusneuvos, Suomen Kiviteollisuus Oy:n joht.	1906—1949
Kyrklund, H., ins., prof.	1906—
Streng, A. E.,	1906
Eskola, Pentti, fil. toht., prof.	1906—1943 kunniajäsen v:sta 1943
Petra, A. H., fil. maist., kemiall. lab. johtaja	1906—1928
Hausen, Hans, fil. toht., prof.	1906—
Mäkinen, Eero, fil. toht., vuorineuvos, Outokumpu Oy:n johtaja	1906—†1953
von Fieandt, A., fil. kand., vuori-ins.	1906—1912
Quist, W., ins.	1913—1917
Ehrnrooth, H., fil. maist.	1913—1928
Soikero, J. N., geologi	1913—†1946
Laitakari, Aarne, fil. toht., prof., geologisen tutkimuslaitoksen täysinpalv. joht.	1913— kunniajäsen v:sta 1961
Brenner, Th., fil. toht., prof.	1914—†1949
Sauramo, Matti, fil. toht., prof.	1914—†1958
Korvenkontio, V., fil. toht.	1914—†1944
Kyrklund, G., fil. maist.	1914—1915
Koskimies, M. R., fil. maist.	1914—1917
Krohn, V. J. S., fil. toht.	1914—?
Levander, G. V., lääket. lis.	1914 †1935
Aschan, M., ylioppilas	1914—1917
Wolff (Wolff-Nessler), Nelly, fil. toht., kemisti	1914—?
Aschan, J., fil. kand., vuori-ins.	1914—†1941
Väyrynen, Heikki, prof.	1915—†1956
Aarnio, B., fil. toht., prof., maatutkimuslaitoksen johtaja	1915—†1951
Forsman, W., fil. toht.	1915—1946
Lindberg, Birger, ins.	1915—1919
Toivonen, N. J., fil. toht., prof.	1915—
Lydén, R., fil. toht., prof.	1915—1919
Brofelt, M., fil. maist.	1916—1919
Järnefelt, H., fil. toht., prof.	1916—1931
Tomula, E. S., fil. toht., prof.	1916—1919
Auer, Väinö, fil. toht., prof.	1916—
Sjerfbeck, M., fil. maist.	1916—1929
Vuorinen, E., fil. maist.	1916—1923
Backman, A. L., fil. toht., dosentti	1916—
Metzger, Carl, prof.	1916—1930
Stening, I., dipl. ins.	1916—1923
Törnqvist, K. H., dipl. ins.	1916—1936
Sandelin, E., fil. toht.	
Stenberg, O. R., fil. maist.	1917—1919
Staudinger, R., tullinhoitaja	
Söderlund, P.	
Tallgren, H.	
Talvia, A., fil. maist., vanh. lehtori	1917—
Hasselström, T., tekn. toht.	1917—1919
Pehrman, G., fil. toht., prof.	1917—
Simberg, G., fil. maist.	1917—1919

Tennberg, I., dipl. ins.	1917—1924
Lokka, Lauri, fil. toht., prof.	1917—
Metzger, A. A. Th., fil. toht., prof.	1917—
Brenner, M. W., fil. toht., dosentti	1919—1932
Saksela (Saxén), Martti, fil. toht., prof.	1919—
Hellaakoski, Aaro, fil. toht., yliopettaja, dosentti	1919—†1952
Ojala, O. O., fil. maist.	1919—1923
Wuorinen, F.	1919
Grönlund, R., fil. maist.	1919—1923
Lönnroth, E., fil. toht., prof.	1919—1927
Lyytikäinen, V. T., fil. maist.	1921
Olsoni, O. B. A., fil. maist.	1919—1923
Palmunen, M. K., vuori-ins.	1919—†1938
Suomalainen, A. J., fil. maist.	1919—1931
Näätänen, Alb., fil. maist.	1919
Talvitie, A., tekn. toht.	1919—1929
Tolvanen, V., fil. toht., vanh. lehtori	1923—1929
Salminen, Antti, fil. toht., agrogeologi	1923—†1951
Grönros, Yrjö, dipl. ins.	1923—
Clopatt, A., fil. maist.	1923—1928
Nordenswan, E. A. O., fil. maist.	1923—
Bengelsdorff, G., fil. maist.	1927—1929
Brander, G., fil. toht.	1927—†1940
Forstén, R., dipl. ins., tekn. toht.	1927—1929
Lindberg, L., ylioppilas, liikemies	1927—1929
Linkola, Kaarlo, fil. toht., prof.	1927—1929
Lupander, Kurt, fil. maist.	1927—
Mikkola, Erkki, fil. toht.	1927—†1940
Ryselin, W., dipl. ins.	1927—1933
Sahama (Sahlstein), Th. G., fil. toht., prof.	1927—
Sihvonen, V., fil. toht., prof.	1927—1929†1940
Kokkonen, Pekka, fil. toht., prof.	1928—1945†1951
Erämetsä (Enwald), O., tekn. toht.	1928—
Aario, Leo, fil. toht., prof.	1929—
Haapala, Paavo, fil. toht., prof.	1929—
Hyyppä, Esa, fil. toht., dosentti	1929—
Renwall, Åge, fil. maist.	1929—
Savolainen, Eetu, fil. maist.	1929—
Suominen, Eero, fil. maist.	1929—
Kantele, Helvi, fil. maist.	1931—
Kalajoki, Reino, fil. maist.	1931—1949
Lehmann, W., prof.	1931—
Lukkala, O., maat. toht., prof.	1931—†1959
Wasastjerna, L., ylioppilas	1931—1932
Ant-Wuorinen, Jalo, fil. toht.	1933—†1940
Aurola, Erkki, fil. toht., dosentti	1933—
Kahlson, T., ylioppilas	1933—1935
Kilpi, Sampo, fil. toht.	1933—†1941
Kivekäs, E., fil. maist.	1933—†1946
Kivinen, Erkki, prof.	1933—

Nordin, W., dipl. ins.	1933—1948
Järnefelt, Olavi, fil. maist.	1933—1951†1961
Mainari (Swahn), B., tekn. ylioppilas	1933—?
Mikkola (Teittinen), Toini, fil. maist.	1933—
Koponen, J. S. W., fil. maist., seminaarin lehtori	1934—†1948
Rankama, Kalervo, fil. toht., prof.	1934—
Friberg, Sven, fil. maist.	1935—
Pääkkönen, Veikko, fil. lis.	1935—
Wessman, J., fil. maist.	1935—?
Zeidler, W., vuori-ins.	1936—
Wessman, Tor H., fil. maist.	1936—
Hall, W., dipl. ins.	1936—†1938
Huuskonen, Toivo S., fil. maist.	1936—
Kahma, Aarno, fil. toht.	1936—
Keränen, Jaakko, prof.	1936—
Lumiala, O., fil. maist.	1936—†1944
Maijala, Paavo, dipl. ins.	1936—
Nystén, Henrik, dipl. ins.	1936—
Okko, Veikko, apul. prof.	1936—
Simonen, Ahti, fil. toht., dosentti	1936—
Varma, Arno, fil. maist.	1936—
Virkkala, Kalevi, fil. toht., dosentti	1936—
Vähätalo, Veikko, fil. toht.	1936—
Mölder, Karl, fil. toht.	1937—
Sirén, Arne, fil. toht.	1937—
Eskola, Salli, fil. toht., apul. prof.	1937—
Levanto, L. A., dipl. ins.	1937—†1943
Helkavaara, Eero K., fil. maist.	1937—
Kanula, Aatto Viljo, ylioppilas	1937—†1944
Lehijärvi, Mauno, fil. toht.	1937—
Matisto, Arvo, fil. lis.	1937—
Simola, Kaarlo, kaivosmittaaja	1937—
Tuominen, Heikki, fil. toht.	1937—
Austi, E., ylioppilas	1937—†1944
Himmi, Reino, fil. maist.	1937—
Heiskanen, V. A., prof.	1938—
Salmi, Martti, fil. toht., dosentti	1938—
Mattila, Jorma, fil. maist.	1938—
Parras, Kauko, fil. toht.	1938—
Häyrynen, Matti, dipl. ins.	1939—1953
Ståhlberg, Elsa, fil. maist.	1939—†1954
Soininen, Antti, fil. maist.	1939—1945
Härme, Mauno, fil. toht., dosentti	1939—
Mikkola, Toivo, fil. maist.	1939—
Rosendahl, L. Erkki M., ylioppilas	1939—†1944
Salminen, Kyllikki, fil. lis.	1939—
Tammekann, August, prof.	1941—†1959
Aulanko, Heikki, dipl. ins.	1941—
Laakso, Perttu, fil. toht.	1941—1951
Nisonen, Eino, fil. maist.	1941—

Lunnasvaara, Väinö, fil. maist.	1941—1949
Soveri, Urpu, fil. toht., dosentti	1941—
Enkovaara, Antti, fil. maist.	1941—
Hämäläinen, Viljo, fil. maist.	1941—
Mikkola, Aimo, prof.	1941—
Seitsaari, Juhani, apul. prof.	1941—
Vahervuori, Teuvo, ylioppilas	1941—1947
Vaasjoki, Oke, fil. toht., dosentti	1941—
Saraste, Ahti, geologi	1941—
Linna, Antti, dipl. ins.	1942—
Kotilainen, Mauno J., prof.	1943—†1961
Halinen, Vihtori, geologi-insinööri	1943—1951
Koponen, Olavi, dipl. ins.	1943—
Valtasaari, Lea, rva	1943—1947
Laitakari, Lauri, pion. kapt. evp.	1943—
Marmo, Vladi, fil. toht., ylijohtaja	1943—
Mäntynen, Matti, fil. maist.	1943—
Stigzelius, Herman, tekn. toht.	1944—
Ohlson, Birger, fil. lis.	1944—
Puranen, Maunu, fil. lis.	1944—
Kaitaro, Simo, fil. toht.	1944—†1957
Neuvonen, K. J., prof.	1944—
Veltheim, Valto, fil. lis.	1944—
Kanerva, Reino, fil. toht.	1945—
Simola, Torsti, dipl. ins.	1945—
Holm, Caj, dipl. ins.	1945—
Viento, Aimo, johtaja	1945—
Alenius, P., dipl. ins.	1946—
Jalander, Holger, dipl. ins.	1946—
Ratu, Martti, dipl. ins.	1946—1952
Heino, Eino, dipl. ins.	1946—
Heinonen, Leo, fil. toht.	1946—†1959
Vuorjoki, Aarre, fil. maist.	1946—
Wiik, Birger, fil. toht.	1946
Peltola, Esko, fil. toht.	1946—
Kulonpalo, Max, fil. maist.	1946—
Repo, Reino, fil. toht.	1946—
Hyppönen, Viljami, fil. maist.	1946—
Schröder, Henning, fil. maist.	1946—
Toivonen, A. V. P., fil. maist.	1946—
Valovirta, Veikko, fil. maist.	1946—
Järvinen, Kauko, prof.	1947—
Edelman, Nils, fil. toht.	1947—
Arvela, August, dipl. ins.	1947—
Vanninen, Mikko, fil. maist.	1947—
Oksanen, Oiva, ylioppilas	1947—1954
Siikarla, Toivo, dipl. ins.	1947—
Grönlund, Rolf, fil. maist.	1947—
Jurva, Risto, prof.	1947—†1953
Huhma, Aarto, fil. maist.	1947—

von Timroth, M., dipl. ins.	1947—
Paarma, Heikki, fil. maist.	1947—
Laatio, Gunnar, dipl. ins.	1948—
Okkonen, Ilmo, dipl. ins.	1948—†1953
Tanner, Heikki, dipl. ins.	1948—
Lindholm, Ole, fil. maist.	1948—
Strandström, George, fil. maist.	1948—
Hukki, Risto, prof.	1949—
Salli, Ilmari, fil. maist.	1949—
Salemaa, J., fil. maist.	1949—1953
Rancken, Ragnar, fil. lis.	1949—
Borg, Torvald, fil. maist.	1949—
Donner, Joakim, fil. toht., dosentti	1949—
Ervamaa, Pentti, fil. lis.	1949—
Halonen, Olli J., ylioppilas	1949—
Halonen, Olavi T., fil. maist.	1949—
Heiskanen, Erkki, fil. maist.	1949—
Helovuori, Olavi, fil. maist.	1949—
Huhta, Jussi, fil. maist.	1949—
Hytönen, Kai, fil. toht.	1949—
Hyvärinen, Lauri, fil. maist.	1949—
von Volborth, Aleksis, fil. toht.	1949—
Ignatius, Heikki, fil. toht.	1949—
Isokangas, Pauli, fil. maist.	1949—
Kalla, Juha, fil. maist.	1949—
Korpela, Kauko, fil. maist.	1949—
Koskinen, Juhani, fil. maist.	1949—
Kouvo, Olavi, fil. toht.	1949—
Laiti, Ilpo, fil. maist.	1949—
Laitala, Matti, fil. kand.	1949—
Marttila, Erkki, fil. maist.	1949—
Mikkonen, Antti, fil. maist.	1949—
Hämäläinen (Nieminen), Kaisa, fil. maist.	1949—
Nortio, Jaakko, fil. maist.	1949—
Palin, Urpo, fil. maist.	1949—
Pehkonen, Eero, fil. maist.	1949—
Räsänen, Veikko, fil. maist.	1949—
Savolahti, Antti, fil. toht.	1949—
Stenberg, Aarre, fil. maist.	1949—
Suila, Matti, fil. maist.	1949—
Okko (Syvänen), Marjatta, fil. lis.	1949—
Tynni, Risto, fil. lis.	1949—
Vaasjoki, Pirjo, rouva	1949—
Waldén, Olavi, fil. maist.	1949—
Vesasalo, Arvo, fil. maist.	1949—
Viluksela, Erkki, fil. maist.	1949—
Yletyinen, Veijo, fil. maist.	1949—
Näykki, Ossi, fil. maist.	1949—
Laitakari, Ilkka, fil. maist.	1949—
Ahlfors, Bruce, dipl. ins.	1950—

Alanko, Risto, dipl. ins.	1950—
Hyyppä, Jussi, fil. maist.	1950—
Kauranne, Kalevi, fil. lis.	1950—
Koskela, Erkki	1950—
Laitakari, Aatto J., fil. maist.	1950—
Lauerma, Raimo, fil. lis.	1950—
Lindberg, Eric, fil. maist.	1950—
Niini, Risto, prof.	1950—
Nousiainen, Erkki, ekonomi	1950—
Nuutilainen, Juhani, fil. maist.	1950—
Oivanen, Paunu, fil. maist.	1950—
Oksanen, Erkki, johtaja	1950—
Penttilä, Aarne, fil. maist.	1950—
Pesola, Pentti, dipl. ins.	1950—
Runolinna, Urmas, tekn. toht.	1950—
Stolpe, Tor-Björn, fil. maist.	1950—
Tavela, Matti, fil. lis.	1950—
Turunen, Eero, dipl. ins.	1950—
Wennervirta, Heikki, fil. maist.	1950—
Juurinen, Aarno, fil. toht.	1951—
Meriläinen, Kauko, fil. lis.	1951—
Pohjanlehto, V. O., autoilija	1951—
Springert, Hans, fil. maist.	1951—†1952
Suvenmaa, L., malminetsijä	1951—
Nieminen, Kalle, dipl. ins.	1952—
Penttilä, Seppo, fil. lis.	1952—
Heloma, Erkki, arkkitehti	1952—
Makkonen, Väinö, fil. maist.	1952—
Rosblom, Heikki, fil. maist.	1952—
Similä, Pentti, fil. maist., dipl. ins.	1952—
Säynäjärvi, Klaus, fil. maist.	1952—
Vanhala, Risto, fil. maist.	1952—
Pipping, Fredrik, fil. maist.	1952—
Konttinen, Lauri, fil. maist.	1952—
Järvinen, Matti, fil. maist.	1952—
Laurila, Matti, fil. maist.	1952—
Haapala, Olavi, dipl. ins.	1953—
Heikkinen, Aulis, fil. maist.	1953—
Veijola, Erkki, fil. maist.	1953—
Nykänen, Osmo, fil. maist.	1953—
Halme, Erkki, tutk. assist.	1953—
Öhman, Börje, fil. maist.	1953—
Vormisto, Kauno, fil. maist.	1953—
Pipping (Stjernvall), Gisela, fil. maist.	1954—
Huhma, Maija, fil. maist.	1954—
Granö, Olavi, apul. prof.	1954—
Wessman, Curt, fil. maist.	1954—
Kurppa, Reino, dipl. ins.	1954—
Virkkunen, Marjatta, fil. maist.	1954—
Aho, Lea, fil. maist.	1954—

Lyytikäinen, Erkki, fil. maist.	1954—
Vorma, Atso, fil. lis.	1954—
Rouhunkoski, Pentti, fil. lis.	1954—
Huopaniemi, Pertti, fil. maist.	1954—
Suominen, Paavo, fil. maist.	1954—
Saastamoinen, Jyry, fil. maist.	1954—
Boström, Rolf, fil. maist.	1955—
Väänänen, Paavo, fil. maist.	1955—
Maunula, T. E., maanvilj. teknikko	1955—
Piirainen, Tauno, fil. maist.	1955—
Häkli, Tauno, fil. lis.	1955—
Sipilä, Esko, fil. maist.	1955—
Lonka, Anssi, fil. lis.	1956—
Leveinen, Antero	1956—
Purokoski, Paavo, prof.	1956—
Nyberg, Harald, laboratoriomestari	1956—
Ruhanen, Veikko, kapteeni	1956—
Carlson, C. E., dipl. ins.	1956—
Markkanen, Pentti, ylioppilas	1956—
Kuhmonen, Heikki, fil. maist.	1956—
Hacklin, Bertel, johtaja	1956—
Penttilä, Esko, fil. maist.	1956—
Tyrväinen, Aimo, fil. maist.	1956—
Virtanen, Pentti, fil. maist.	1956—
Lammela, Torsti, fil. maist.	1957—
Keto, Leijo, fil. maist.	1957—
Talvitie, Jouko, fil. maist.	1957—
Tyni, Matti, fil. maist.	1957—
Hirvonen, Yrjö, fil. maist.	1957—
Löfgren, Arvo, fil. maist.	1957—
Nurmi, Aimo, fil. maist.	1957—
Åberg, Ragnar, fil. maist.	1957—
Mattila, Tatu, fil. maist.	1957—
Palosuo, Erkki, fil. toht.	1957—
Sallinen, Kalevi, fil. maist.	1957—
Villman, Jussi, autoilija	1957—
Hoffren, Väinö, fil. maist.	1957—
Kurtén, Björn, fil. toht., dosentti	1958—
Vasari, Yrjö, fil. lis.	1958—
Kostiainen, Nils, arkkitehti	1958—
Kujanpää, Jorma, fil. maist.	1958—
Sippola, Martti, fil. maist.	1958—
Laurell, Nils-Olof, tutk. assist.	1958—
Lappalainen, Veikko, fil. maist.	1958—
Papunen, Heikki, fil. maist.	1958—
Lammi, Paavo	1958—
Leino, Antti, tutk. assist.	1959—
Niini, Heikki, luonnont. kand.	1959—
Paronen, Tauno, johtaja	1959—
Peltonen, Pietari, dipl. ins.	1959—

Pouttu, Topi, ylioppilas	1959—
Säde, Railo, kultaseppä	1959—
Alhonen, Pentti, ylioppilas	1960—
Hannikainen, Raili, ylioppilas	1960—
Hämäläinen, Uljas, ins.	1960—
Järvimäki, Paavo, tutk. ssist.	1960—
Kae, Eero, ylioppilas	1960—
Kaskimies, Keijo, ekon.	1960—
Lehtinen, Asko, ylioppilas	1960—
Peltonen, Anna-Liisa, luonnont. kand.	1960—
Punakivi, Kalevi, ylioppilas	1960—
Salmi, Martti H., ylioppilas	1960—
Simola, Liisa, luonnont. kand.	1960—
Suomalainen, Mikko, luonnont. kand.	1960—
Tolonen, Kimmo, luonnont. kand.	1960—
Westerlund, Keijo, tutk. assist.	1960—
Ylinen, Mauno, fil. maist.	1960—
Luho, Ville, fil. toht.	1961—
Aarne, Uuno V., kultaseppä	1961—
Paavilainen, Eero, maat. ja metsät. kand.	1961—
Haapala, Ilmari, luonnont. kand.	1961—
Mälkki, Esko, ylioppilas	1961—
Paakkola, Juhani, ylioppilas	1961—
Piispanen, Risto, ylioppilas	1961—
Rantala, Erkki, ylioppilas	1961—
Sarikkola, Risto, ylioppilas	1961—
Hiltunen, Aimo, ylioppilas	1961—
Turkka, Seppo, ylioppilas	1961—
Heikkilä, Heikki, insinööri	1961—

5 2192—61

Liite 4

Appendix 4

SUOMEN GEOLOGISEN SEURAN RAHA-ASIAT 1924—1960

SUMMARY OF FINANCES OF THE GEOLOGICAL SOCIETY 1924—1960

Vuosi	Säästö edell. v:lta	Jäsen-maksuja	Vakituinen valtionapu	Yli-määräinen valtionapu	Korkoja	Ylipainoksista ym.	Menoja
1924	120	135	1 000	—	—	500	1 385
1925	370	1 230	2 000	—	—	—	415
1926	3 185	405	—	—	—	—	485
1927	3 105	840	10 000	—	—	—	530
1928	13 415	1 460	10 000	53 000	117	—	50 752
1929	27 241	2 665	10 000	3 000	785	689	21 540
1930	22 840	1 050	13 000	2 000	1 495	228	24 921
1931	15 693	4 355	13 000	7 000	177	2 605	33 418
1932	9 413	3 300	9 300	3 700	887	260	8 604
1933	18 256	3 125	7 000	10 000	1 041	613	23 195
1934	16 841	3 325	7 000	6 000	730	1 689	21 641
1935	13 945	4 075	7 000	9 000	892	940	1 944
1936	34 078	3 424	7 000	5 000	1 325	100	51 273
1937	5 655	4 277	8 000	26 000	248	—	15 055
1938	23 126	1 750	22 000	15 000	522	—	41 137
1939	21 262	—	16 500	7 500	729	—	21 144
1940	24 847	—	12 500	14 000	1 117	4 625	2 798
1941	56 602	8 800	4 500	28 000	1 266	1 342	12 564
1942	87 946	4 825	4 500	35 250	2 915	265	51 626
1943	84 066	3 875	4 500	25 000	3 167	345	42 497
1944	93 457	—	4 500	24 000	4 037	—	35 284
1945	108 994	19 575	4 500	30 000	4 333	5 485	102 965
1946	70 070	19 940	4 500	52 000	1 905	—	124 085
1947	24 330	19 849	4 500	160 000	4 773	232 049	418 992
1948	26 513	39 098	144 000		39 398	246 721	371 838
1949	123 890	37 703	80 000	295 000	4 554	117 805	324 919
1950	341 843	53 513	100 000	200 000	8 096	19 950	325 011
1951	398 391	49 573	120 000	280 000	11 256	18 920	708 114
1952	170 026	64 300	180 000	300 000	6 604	35 170	658 393
1953	97 707	132 388	180 000	320 000	5 689	32 500	603 307
1954	164 977	119 494	600 000		7 402	128 727	864 751
1955	165 839	69 900	500 000		11 247	55 270	620 899
1956	181 357	61 500	500 000		5 104	116 840	698 660
1957	166 301	69 100	500 000		2 304	122 829	721 369
1958	159 805	58 400	900 000		2 603	110 853	1 042 985
1959	188 676	67 900	1 050 000		3 098	121 645	1 249 817
1960	182 082	98 194	850 000		3 718	199 008	1 633 022

Liite 5

Appendix 5

SUOMEN GEOLOGISEN SEURAN KOKOUSTEN ESITELMÄT JA TIEDONANNOT 1936—1961

LECTURES AND COMMUNICATIONS DELIVERED AT THE MEETINGS OF THE GEOLOGICAL SOCIETY 1936—1961

1936

E. H. Kranck: Om intrusion och tektonik i Västra Skärgården.
E. Aurola: Postglasiaalinen rannansiirtyminen Varsinais-Suomessa.
P. Eskola: Ruokolahden Suur-Lintusaaren pallograniitti.
S. Kilpi: Sotkamon reitin Kainuun myöhäisglasiaalisesta kehityksestä.
M. Saksela: Havaintoja Pohjois-Ruotsin ja Norjan kiisumalmialueilta.
P. Eskola: Geologian merkitys elämän perustan tutkimuksena ja ihmisen maailmankuvan perustana.
E. H. Kranck: Finlands urberg och den internationella geologiska forskningen.
Aarne Laitakari: Suomen geologisen seuran 50-vuotishistoriikki.

1937

Antti Salminen: Savien raekokoomuksen alueellisesta vaihtelusta.
E. Hyyppä: Terijoen pohjavesipurkaus.
Aarne Laitakari: Uusi suomalainen geologikompassi.
E. H. Kranck: Om svecofenniderna i Sverige och Finland.
P. Eskola: Rapakiilteen rapautumisesta.
M. Sauramo: Hiili aineksena kerrallisessa savessa.
H. Väyrynen: Petsamon nikkelimalmin synnystä ja sijoittumisesta.
M. Sauramo: Muistelmia kansainvälisen kvartäärigeologiyhtymän kolmannesta kongressista Wienissä syyskuun 1—6. päivinä 1936.
P. Eskola: S. Hitchenin diagrammista, joka osoittaa kvartsin liukenevaisuutta vedessä eri lämpötiloissa.
Aarne Laitakari: Näytteitä ja kuvia Norjan kiisumalmeista.
H. Väyrynen: Petsamon alkalirikkaista kivilajeista.
Anna Hietanen: Lapuan Simsiönvuoren kvartsiitista.
A. Hellaakoski: Vuoksen puhkeamisdeltta.
Th. Brenner: Harjuaineksen sisästä löydetty savikimpale.
P. Eskola: Moskovan geologikongressin 1937 Kuollan-retkestä.
M. Saksela: Kesän 1937 malmitutkimuksista.
H. Väyrynen: Moskovan geologikongressin 1937 Karjalan ja Aunuksen retkiltä.

E. Hyyppä: Etelä-Suomen postglasiaalisesta rannansiirtymisestä.
A. Simonen: Kemijärven pallograniitti.
E. Savolainen: Jänisjärven Leppäniemen dasiittikallio.
A. Laitakari: Lammastenkosken hiekkakivileikkaus.

1938

P. Haapala: Vaikutelmia Sudburyn nikkelimalmialueelta.
E. Savolainen: Fossiileja sisältävä kalkkikivilohkare Koitonjärveltä Suistamosta.
E. Mikkola: Itä-Lapin kallioperästä.
—»— Lapin kallioperää granuliittialueelta Länsi-Lappiin.
A. Salminen: Maalajiemme raekokoomus pystysuorassa suunnassa.
H. Väyrynen: Mikrokuvia Petsamon nikkelimalmeista.
M. Sauramo: Pohjamoreenin suuntaisrakenne.
L. H. Borgström: Om mineralernas kristallstruktur.
M. Saksela: Laver-Kaivos.
P. Eskola: Liuskeisuuden synty.
V. Pääkkönen: Kalsiumkarbonaatin vaikutuksesta eräisiin alkalimineraaleihin.
E. Mikkola: Etu-Lapin vanhemmat kvartsiitti-liuske- ja vihreäkivimuodostumat.
Aarne Laitakari: Tafonirapautuminen ja Karhunpesäkivi.
—»— Kemiön tantaliitit.
O. Erämetsä: Suomalaisten flogopiittien korkea rubidiumpitoisuus.
M. Sauramo: Kovelliittia sisältävä lohkare.
—»— Fulguriittia Kuhrische Nehrungilta.
E. Mikkola: Kuopion kallioperäkarttalehti Wilkmanin kuvaamana.
E. Aurola: Outokumpu Oy:n suorittamat malmitutkimukset Pohjois-Karjalassa 1935—38.
E. Mikkola: Uusi kallioperäkartoitus Etelä-Suomessa.
—»— Kumpu-Oraniemen muodostuma.

1939

E. H. Kranck: Labradorkustens betydelse för den geologiska konnektionen mellan Nordamerika och Grönland.
P. Eskola: Hiilen kiertokulku.
E. Kivinen: Moreenimaiden ominaisuuksista vaara-alueilla.
Th. Brenner: Skred i Ilmajoki.
H. Väyrynen: Outokummun alueen tektoniikasta.
Leo Aario: Kolosjoelta löydetyistä pyöriäisen luista.
Harry von Eckermann: De alkalina bergarternas genesis i belysning av nya forskningsrön från Alnön.
M. Sauramo: Suomen metsien historia.
Th. G. Sahama: Wiikiitin maametallikokoomuksesta.
Paul Thomson: Interglasiaalinen suo Virossa.
P. Eskola: Yliopiston mineralogis-geologisella laitoksella rakennetut kidemallit.
H. Väyrynen: Pohjois-Karjalan liuskemuodostumien stratigrafiasta.
K. Rankama: Silikaattianalyysin piihapposakan sisältämästä epäpuhtauksista.
P. Eskola: Honkilahden rapakivestä tavatusta kidekellarista ja sen mineraaleista.

1940

Martti Salmi: Patagonian postglasiaalisista purkauskerroksista.
Kalervo Rankama: Geokemiallisesta prospektauksesta.
A. A. Th. Metzger: Om cementpetrografi.
P. Eskola: Uusia yliopiston geologian laitoksella valmistettuja kiderakennemalleja.
Anna Hietanen: Appalakien itäisen osan geologiasta.
Thord Brenner: Slamstenssedimentet i Muhos.

1941

Veikko Okko: Rantaviivan siirtymisesta Oulujärven ympäristössä ja Oulujoen eteläpuolella.
Kalervo Rankama: Niobin ja tantalin esiintymisestä muutamissa Suomen peruskallion graniiteissa.
—»— Lausunto kauppa- ja teollisuusministeriölle geologisen toimikunnan perustellusta ehdotuksesta niistä muutoksista, joita toimikunnan käytännöllisgeologisen tutkimustyön tehostaminen aiheuttaa geologisesta toimikunnasta voimassa olevaan lakiin ja asetukseen.
Toini Mikkola: Neulakvartsista.
Matti Sauramo: Jään ja vesipeitteen peräytyminen Joensuun itäpuolella.
Kauko Parras: Länsi-Uudenmaan kallioperästä.
Hans Cloos: Über Struktur und Funktion von vulkanischen Tuffschloten.
Aarne Laitakari: Ylämaan labradorikivikallio.
August Tammekann: Mannerjäätikön viimeinen perääntyminen Suomenlahden eteläpuolella.
Aarno Kahma: Satakunnan oliviinidiabaasin kontakteista.

1942

Matti Sauramo: Itä-Karjalan kvartääristä.
Pentti Eskola: Hugo Struntzin Mineralogische Tabellen.
Anna Hietanen: Kalannin alueen geologiasta.
Thord Brenner: Geologisia havaintoja Oulujoen Pyhäkosken ympäristössä.
Pentti Eskola: Itä-Karjalan geologiasta.

1943

Karl Mölder: Resenttisten piilevien merkitys kvartäärigeologisessa tutkimuksessa.
W. Wahl: Om möjligheterna att bestämma orogenesernas ålder.
Aarne Laitakari: Unkarin kaivannaisista.
Heikki Väyrynen: Utajärven-Kiimingin liuskemuodostumasta.
Aarne Laitakari: Viron palavakivikaivoksen trilobiitti.
—»— Viron Jöhvin luona suoritetuista syväkairauksista.
O. V. Lumiala: Nykyisistä eloperäisistä sedimenteistä ja niiden synnystä.
Matti Sauramo: Karjalan kannaksen tuulenhiomista kivistä ja maalajeista.

1944

H. Stigzelius: Geologisk beskrivning av Haveri gruvan och dess omgivningar.
V. Pääkkönen: Reposaarella tapaamani rapakiveä muistuttava graniittiesiintymä.
Pentti Eskola: Kuhn-Rittmannin teoria maapallon sisuksesta.
Heikki Väyrynen: Lapin kallioperän tektoniikasta.

1945

Anna Hietanen: Naantalin seudun geologia.
Esa Hyyppä: Vihannin rikkikiisulohkareiden emäkallion etsinnästä glasiaaligeologisilla perusteilla.
E. H. Kranck: Om Mätäsvaara molybdenförekomst.
Mauno J. Kotilainen: Yhteistehtävistä ja yhteistyöstä kasvimaantieteen ja geologian alalla.
Heikki Väyrynen: Pohjois-Suomen kallioperän tektonisesta rakenteesta.
Sture Landergren: Några drag i järnmalmernas geokemi.
Frans E. Wickman: Några atomistiska synpunkter på den magmatiska differentiationen.
Thord Brenner: Lerornas vattenhalt.
Aarne Laitakari: Geologisista hajahavainnoista.
Heikki Väyrynen: Havaintoja Etelä-Suomen kallioperän tektoniikasta.
Matti Sauramo: Rannansiirtymiset maankohoamisalueen reunavyöhykkeellä.

1946

Aarne Laitakari: Helsingin ympäristön kallioperän kaivannaiset.
M. Kulonpalo: Pakilan lyijymalmi.
Martti Salmi: Soiden teknillinen käyttö ja käyttömahdollisuudet Helsingin lähiympäristössä.
Urpu Soveri: Helsingin lähiympäristön savien teknillinen käyttö ja tutkimus.
K. Virkkala: Teknillisesti käyttökelpoisista sora- ja hiekkaesiintymistä Helsingin seudulla.
Ahti Simonen: Hämeenlinnan alueen kallioperästä.
Aarne Laitakari: Geologista tutkimuslaitosta koskeva uusi asetus.
V. A. Heiskanen: Uutta amerikkalaista kirjallisuutta geotieteiden alalta.
K. Virkkala: Havaintoja Pohjois-Satakunnan myöhäisglasiaalisista vaiheista.
Veikko Okko: Oulunlaakson kvartäärigeologiasta.
Heikki Väyrynen: Kiteiden symmetrialuokitus ja kiderakenne.
Veikko Pääkkönen: Otanmäen löytö ja tähänastiset tutkimusvaiheet.
L. H. Borgström: Peridotiternas ombildning till serpentin- och talkmagnesitstenar.
Aarne Laitakari: Eräitä uutuuksia maamme kaivannaisten alalta.
Esa Hyyppä: Harjujen synty.
—»— R. Flintin kvartäärigeologinen kartta Pohjois-Amerikasta.
Heikki Väyrynen: Valon eteneminen anisotrooppisessa aineessa.
Martti Salmi: Ancylustransgressio Sippolassa.
Aarne Laitakari: Säkkijärven labradorisoiva labradoriitti.
Aimo Mikkola: Havaintoja Pohjois-Ruotsin kallioperäkartoituksesta.

1947

Aarne Laitakari: Someron petaliitista.
Toini Mikkola: —»—
Pentti Eskola: Amerikan matka.
A. Metzger: Puolangan Pihlajan kaoliinista.
Veikko Okko: Puolangan Pihlajan kaoliinin subfossiileista.
Martti Salmi: Turpeiden tuhkapitoisuudesta ja lämpöarvoista.
E. Edelman: Om reaktionen kvartsglas-kristobalit.
—»— Rapakiviesiintymästä Paraisilla.
Veikko Okko: Alavieskan rikkikiisulohkare.
Veikko Pääkkönen: Otanmäen malmin ja sen ympäristön geologisesta rakenteesta.
M. Kulonpalo: Inkeroisten rapakivessä tavatusta lyijyhohdejuoniseurueesta.
Martti Salmi: Heklan tuhkasateesta.
Ahti Simonen: Geologisen tutkimuslaitoksen kallioperäkartoituksesta 1946.
Esa Hyyppä: Geologisen tutkimuslaitoksen maalajitutkimuksen uudelleenjärjestelystä.
Martti Saksela: Piirteitä malmimuodostuksesta Tampereen liuskealueella.
N. Edelman: Kemiön saaristosta tavatusta todennäköisesti postglasiaalisesta siirroksesta.
L. M. J. U. van Straaten: Subaquatische Abrutschungserscheinungen.
S. Thorarinsson: Hekla utbrottet 1947.
Martti Saksela: Outokummun löytö.
Martti Salmi: Turpeiden bitumipitoisuudesta.
Harry von Eckermann: Från Alnö området.
Heikki Tuominen: Bergenin alueen geologiasta.
Thord Brenner: Jostedalsbrä.
Aarne Laitakari: Sulitjelmasta ja Blaamandsiseniltä.

1948

Iivari Leiviskä: Kerrallisen saven hiilipitoisuudesta.
Mauno Härme: Preglasiaalisesta rapautumisesta Tyrväällä.
Nils Edelman: Tektoniken vid Gullkronafjärden.
Aarne Laitakari: Vihannin tuulenhiomista kivistä.
Matti Sauramo: Näytetiheys siitepölyanalyysissä.
V. Marmo: Suojun vulkaanisen kompleksin tulivuoritoiminnasta ja siinä syntyneistä laavoista.
Erkki Aurola: Eräitä glasiaaligeologisia havaintoja Karjalasta.
Iivari Leiviskä: Satakunnan hiekkakivestä.
Mauno Härme: Kemin ympäristön kallioperä.
Ahti Simonen: Geologisen tutkimuslaitoksen kallioperäkartoituksesta kesällä 1947.
Iivari Leiviskä: Maankohoamisesta ja vajoamisesta.
Martti Saksela: Lappajärven kärnäiitti ja sen leviäminen irtolohkareina.
Nils Edelman: En porfyrgranit i västra Hitissocknen.
Heikki Tuominen: Orijärven Mg-Al-rikkaitten kivien rakenteellinen sijainti ja kysymys magnesiametasomatoosista.
Pentti Eskola: Lontoon geologikongressin retkeilyiltä Cornwalliin ja Devonshireen.
Ahti Simonen: Lontoon geologikongressin retkeilystä Skotlantiin.
Oke Vaasjoki: Piirteitä Tšekkoslovakian geologisesta tutkimuksesta ja malmiesiintymistä.

1949

Veikko Okko: Geologi-lehden ilmestymisen alkuvaiheista.
Urpu Soveri: Terminen analyysi savitutkimuksemme apuna.
Th. G. Sahama: Termokemia teoreettisen petrologian aseena.
Simo Kaitaro: Kalkografisten ja röntgenografisten tutkimusten tuloksena todettu Suomelle uusi Pb-Sb-sulfosuolamineraali menegheniitti.
H. B. Wiik: Pyrallolit, ett mineral med dåligt rykte.
A. A. Th. Metzger: En resa genom Englands bergsindustri.
Aarne Laitakari: Ruokolahden dumortieriitti.
—»— Nordsjön kartanon alueella oli todettu sähkövirran sulattamassa hiekassa syntyneen fulguriittia.
—»— Ivigtutin kryoliitista ja sen rikastamisesta.
Veikko Pääkkönen: Havaintoja ja päätelmiä lohkarekuljetuksesta.
Nils Edelman: Några morfologiska drag i Gullkrona området.
A. A. Th. Metzger: Syvällä olevien malmien etsintä.
Veikko Okko: Kivinäytteitä Islannista.
Erkki Aurola: Havaintoja eräiltä Ruotsin kiviteollisuusalueilta.
Ahti Simonen: Geologisen tutkimuslaitoksen kallioperäosaston viimeaikaisista töistä.

1950

Oke Vaasjoki: Näkökohtia eruptiivikivianalyysien tulkinnasta ja magmadifferentiatiosta.
Aarne Laitakari: Suomalaisia korukiviä.
Heikki Väyrynen: Kantagraniitista ja sen muodostumisesta.
Veikko Okko: Glasiaaligeologisia havaintoja Islannista.
V. Marmo: Kasvihavainnoista geologien apuna.
Martti Saksela: Lisiä Pitkärannan malmialueen mineralogiaan.
A. A. Th. Metzger: Lappeenrannan ympäristön kallioperästä.
Matti Sauramo: Lappeenrannan seudun kvartäärigeologisesta kehityksestä.
M. K. Wells: Some Aspects of Scottish Geology.
Toivo Mikkola: Orijärven alueen rakennetta ja stratigrafiaa.
Martti Saksela: Pitkänrannan malmien synnystä.
Aarne Laitakari: Geologisilta retkeilyiltä Sveitsissä.
Esa Hyyppä: Kuvia Salpausselän rakenteesta.
A. A. Th. Metzger: Uusi kärnäiittiesiintymä Lappajärven kaakkoispuolella Vimpelissä.
Kalervo Rankama: Terveisiä Chicagosta.

1951

Oke Vaasjoki: Iilijärven lillianiitista, joka on todettu seleenipitoiseksi kosaliitiksi.
V. Marmo: Kiisuliuskeista.
A. A. Th. Metzger: Glangeudin teoria magmakivilajien synnystä.
A. A. Th. Metzger: Stereogrammeista.
Heikki Paarma: Avaruusmalleista.
Heikki Väyrynen: Rikkikiisu-magneettikiisuyhtymän tasapainosuhteesta.

E. H. Kranck: Om geologin i Kanada.
Pentti Eskola: Chubb-meteorikraaterista Labradorissa.
Juhani Seitsaari: Metamorfoosi ja syväkivet Tampereen alueen itäosassa.
Veikko Vähätalo: Outokummun malmin hivenmineraaleista.
Heikki Väyrynen: Salonsaaren pyrokseenigneissistä Ruokolahdella.

1952

Aarne Laitakari: Posion vermikuliitti.
Frans E. Wickman: Några undersökningar av variationen i kolisotopernas halt hos olika material.
Pentti Eskola: Fluoboriitti Pitkärannasta.
Kalervo Rankama: Rikin isotooppigeologia.
Heikki Väyrynen: Teknillisen korkeakoulun uusittu geologian laitos.
Yrjö Huuskonen: Öljyfilmejä Texasista.
Urpo Soveri: Routa ja routavauriot ruotsalaisilla teillä.
Ahti Simonen: Jokioisten aksiniitista.
Simo Kaitaro: Åvan graniitin ja siihen liittyvien intrusioiden rakenteesta ja mekanismista.
Martti Saksela: Kiisujen rapautumisesta.
Nils Edelman: Om migmatittektoniken.
A. von Volborth: Ihalaisten apofylliitista.
Eugen Wegmann: Stockwerk-tektonik.
Aarno Kahma: Matkavahaintoja Ruotsista.
Martti Salmi: —»— Länsi-Euroopasta.
Mauno Härme: —»— Sveitsistä.
Toini Mikkola: —»— Algeriasta ja Marokosta.
Aarne Laitakari: —»— —»—
V. Marmo: —»— —»—
J. Donner: —»— Brittien saarilta.
R. Lauerma: —»— Grönlannista.
Urpu Soveri: —»— Yhdysvalloista ja Kanadasta.
Oke Vaasjoki: —»— —»—
Olof Ödman: Urbergsgeologiska problem i Norrbotten.
Erik Fromm: Kvartärgeologiska undersökningar i Norrbotten.
Matti Mäntynen: Suomen tiiliteollisuudesta, sen tuotteiden ja raaka-aineiden tutkimustoiminnasta.
Nils Edelman: Kontakter och graniter i Gullkronaområdet.

1953

V. Marmo: Geo- ja biokemiallisesta malminetsinnästä.
Maunu Härme: Mustion alueen rakenteesta ja stratigrafiasta.
Th. G. Sahama: Brittiläisen Itä-Afrikan nuorista vulkaniiteista.
Martti Saksela: Migmatiittien asema orogeenisessa kehityksessä.
Erkki Aurola: Stansvikin sepioliitista.
Georges Heurtebize: Adelien maan geologiasta.

Pentti Eskola: Sörnäisten keskusvankilan alueen lamprofyyrijuonesta.
A. A. Th. Metzger: Högtemperaturkalksilikater i naturen och i klinker.
Veikko Okko: Inqua ja sen IV kongressi.
Valto Veltheim: Geologisia matkahavaintoja Brasiliasta.
Mauno Kotilainen: Hiekkakerrosten alle hautautunut suo Kuopion yläkaupungilla.
Heikki Ignatius: Itä-Kanadan kvartäärigeologiasta.
K. J. Neuvonen: Maasälvistä.

1954

Aarne Laitakari: Göteborgin geologikokouksesta.
Pentti Eskola: Kivianalyysien laskemisesta ioniprosenteissa.
—»— Uusi geologinen aikataulu.
A. von Volborth: Eräjärven Viitaniemen Li-pegmatiitin mineraaleista.
R. J. Lougee: Glacial Shoreline History in Eastern North America.
A. von Volborth: Väyryneniitistä.
H. Stigzelius: Kultaesiintymät Pohjois-Lapissa.
Paavo Haapala: Colquijircan kaivoksen geologiasta.
S. Deevey: Radiocarbon Dating.
Kauko Parras: Charnokiitti — itsenäinen ongelma vaiko osa kokonaisesta.
Martti Salmi: Turvekemiallinen malminetsintämenetelmä.
A. A. Th. Metzger: Om några grundproblem i geologin.
V. Marmo: Sierra Leonen geologiasta.
Simo Kaitaro: Keski-Saharan Hoggarista.

1955

Aimo Mikkola: Elisabeth-kupariesiintymän rakenteesta, Vermont USA.
Oke Vaasjoki: Eräs silikaatti- ja oksidimineraalien yhteenkasvettumisrakenne.
Toivo Mikkola: Primääri- ja pseudorakenteista Karjalan kvartsiiteissa.
Heikki Väyrynen: Niinin jäähtymiskaavaan liittyvistä virhelähteistä.
Ilmari Salli: Vieskan—Himangan liuskemuodostuman rakennetta ja stratigrafiaa.
Urpu Soveri: Saviemme koostumuksesta.
Anders Kvale: Strukturundersökelser i metamorfe bergarter.
Oke Vaasjoki: Geologia Geneven atomivoimakonferenssissa.
Walter Schütt: Glaciologiska problem, specifika för polära glaciärer.
Juhani Seitsaari: Perniön—Kemiön alueen geologiasta.

1956

Paavo Haapala: Havaintoja Espanjan ja Portugalin rikkikiisumalmeista.
Pentti Eskola: Pascual Jordanin maailman syntyhistoriasta.
Hans Schneiderhöhn: Allgemeines räumlich-zeitliches Vorkommen der Erzlagerstätten in der obersten Erdrinde.
—»— Magmatische Abfolge.
Sven Gavelin: Om Västerbottens pre-kaledoniska geologi.

Heikki Ignatius: M/S Arandan tutkimuksista kesällä 1956.
Pentti Eskola: Peruskallion ikä.
Martti Saksela: Outokummun malmin synty tektonis-metamorfisen aineen mobilisation valossa.
Martti Salmi: Selostus Meksikon kansainvälisestä geologikongressista 1956.
V. Marmo: Länsi-Afrikan timanteista.

1957

I. P. Shumilin: Uraanimalmien prospektauksesta (tulkkina V. Marmo).
Marjatta Syvänen: Rapautumisesta ja kulumisesta Lounais-Yhdysvaltojen puoliaavikolla.
Harry von Eckermann: Alnö alkalina intrusion i belysning av gångarna i Bergeforsens kraftstationstunnlar.
Otto Mellis: Några resultat av petrografiska undersökningar av djuphavssediment från Atlantiska oceanen.
Nils Edelman: Stratigrafin i Åbo skärgård.
Ahti Simonen: Itä-Karjalan ja Kuolan kallioperästä.
Aarno Kahma: Malminetsinnästä Itä-Karjalassa.
V. Marmo: Uralin geologiasta ja Keski-Uralin malmeista.
Maunu Puranen: Geofysikaalisista malminetsintämenetelmistä Neuvostoliitossa.
L. K. Kauranne: Pedogeokemiallisesta malminetsinnästä.
K. J. Neuvonen: Kalimaasälvästä.

1958

Paul Ramdohr: Die Gold- und Uranlagerstätten des Witwatersrandes in Vergleich mit den neuen Uranvorkommen in Blind River District.
Esa Hyyppä: Inqua-kongressista.
Olavi Kouvo: Mineraalien iän määräämisestä.
V. Marmo: Graniiteista ja malmeista.
K. Virkkala: Tampereen seudun maaperägeologiasta.
Raimo Lauerma: Grönlannista ja sen geologisesta tutkimuksesta.
Toini Mikkola: Lohkodiagrammien laatimisesta.
Martti Salmi: Ruotsin länsirannikolta ja Värmlannista.
Heikki Ignatius ja Valto Veltheim: Havaintoja Selkämeren pohjan geomorfologiasta ja mahdollisuuksista litologian kontrolloimiseen pohjanäytteiden avulla.

1959

Pentti Eskola: Svekofenniidit ja kareliidit.
Aarno Kahma: Waite Amuletin kirsimalmeista.
V. Marmo: Graniittisten kivien K-A-ikien tulkinnasta.
Walter Schütt: Några synpunkter på landisar.
Tom. F. W. Barth: Geologisk termometri.
Reino Himmi: Parosten geologiasta.
Esa Hyyppä: Näkymiä Pohjois-Amerikan matkalta.

1960

Pentti Eskola: Unkarin matkasta.
Erkki Viluksela: Ekskursiovaikutelmia Reininlaakson peruskallioalueilta.
Heikki Paarma: Magneettisten anomalioitten visuaalisesta tulkinnasta.
Arto Levanto: Magneettisten anomalioitten matemaattisesta tulkinnasta.
Juhani Seitsaari: Ahlaisten—Merikarvian rannikkoalueen geologiasta.
E. H. Kranck: Peruskallioprobleemeja Kanadassa.
Toivo Mikkola: Kolarin alueen geologiasta.
Aimo Mikkola: Vihannin malmiesiintymän mineralisaatiosta.
Esa Hyyppä: Itämeren historiasta.

1961

Oke Vaasjoki: Kemin kromiittimalmin mineralogiasta.
Veikko Pääkkönen: Wolframiittia irtokivessä.
K. Mölder: Eestin uusimmista geologisista tutkimuksista.
Aarre Laitakari: Kuvia Adrianmeren rannikolta.
Veikko Pääkkönen: Seinäjoen antimoniesiintymän geologiasta.

Liite 6

Appendix 6

SEURAN JULKAISUTOIMINTA

THE PUBLISHING ACTIVITY OF THE SOCIETY

Vuosina 1916—1923 julkaistiin seuran toiminnasta lyhyet otteet, osittain myös esitelmäselostukset, aluksi ylipainoksina Teknikern-lehdestä ja vv. 1921—23 erillisinä. Nämä Helsingin geologisen yhdistyksen tiedonannot olivat pääasiallisesti ruotsinkielisiä, ja nämä vaatimattomat tiedonannot päättyivät vuoden 1923 lopussa.

Vuonna 1925 valittiin julkaisutoimikunta, johon kuuluivat A. Laitakari, W. Ramsay, P. Eskola, V. Tanner ja M. Saxén, harkitsemaan omaa julkaisutoimintaa. Vuonna 1927 hyväksyttiin toimikunnan esitys oman julkaisusarjan perustamisesta, ja samana vuonna julkaisusarja alkoi ilmestyä »Suomen geologisen seuran julkaisuja»-nimisenä omina numeroinaan geologisen tutkimuslaitoksen Bulletin-sarjassa. Seuran julkaisujen ensimmäinen numero ilmestyi Bulletin-sarjan numerona 85. Aina vuoteen 1951 saakka seuran julkaisujen toimittajana oli seuran kulloinenkin sihteeri. Koska julkaisut liittyivät geologisen tutkimuslaitoksen Bulletin-sarjaan, oli tutkimuslaitoksen johtajan ne hyväksyttävä, eivätkä tutkimuslaitoksen johtajat (professorit J. J. Sederholm ja Aarne Laitakari) kertaakaan ole katsoneet olevan syytä evätä julkaisujen liittämistä Bulletin-sarjaan.

50-vuotisjuhlajulkaisua varten seura valitsi toimikunnan, johon kuuluivat Aarne Laitakari, Pentti Eskola, E. H. Kranck, W. W. Wilkman ja Erkki Mikkola. Komea 500-sivuinen juhlajulkaisu ilmestyikin sitten v. 1936. Suurimman taakan julkaisun julkisuuteen saattamisesta kantoi tohtori Erkki Mikkola, ja hänelle kuuluu myös kunnia sen täysipainoisesta sisällöstä ja painoasuun saattamisesta.

Professori Pentti Eskolan 65-vuotisjuhlajulkaisua varten seura valitsi v. 1945 toimikunnan, johon kuuluivat Aarne Laitakari sekä Hyyppä, Okko, Sahama, Simonen ja Virkkala. Tämä yli 300-sivuinen juhlajulkaisu ojennettiin professori Eskolalle hänen 8. 1. 1948 täyttäessään 65 vuotta. Mainittakoon, että tätä juhlajulkaisua varten amerikkalaiset geologit professori Ernst Cloosin aloitteesta lähettivät noin 750 dollarin suuruisen avustuksen.

Vuodesta 1951 on seura valinnut vuosittain julkaisutoimikunnan ja julkaisutoimittajan. Tähän toimikuntaan on geologisen tutkimuslaitoksen johtajan lisäksi kuulunut muutamia seuran jäseniä. Julkaisujen toimittajina ovat olleet: K. Rankama vv. 1951—53, S. Kaitaro 1954—55, J. Seitsaari 1956—57, O. Vaasjoki 1958—60 ja K. Virkkala 1961.

Näiden julkaisujen rahoitus onkin niellyt pääosan seuran varoista, ja jatkuvasti on seuralla ollut rahoitusvaikeuksia, sillä valtionapu on ollut riittämätön. Sangen usein on Outokumpu Oy:n säätiö tukenut julkaisutoimintaa raha-avulla, joka tässäkin kiitollisuudella todettakoon.

Seuran julkaisusarjan niteiden I—XXXI sisällysluettelot on esitetty geologisen tutkimuslaitoksen toimittamassa erillisjulkaisussa »Guide to the publications of the Geological Survey of Finland, 1879—1960».

GEOLOGI-LEHTI

COMMUNICATIONAL PAPER »GEOLOGI»

Joulukuussa 1945 professori Aarne Laitakari teki esityksen geologisen tiedonantolehden perustamisesta.

Seuran jäsenet Aarne Laitakari, V. Okko ja V. Marmo jättivät seuralle kirjallisen, perustellun esityksen tiedotuslehden aikaansaamisesta, ja seura asetti toimikunnan asiaa valmistelemaan. Tähän toimikuntaan valittiin aloitteen tekijät.

Kokouksessaan 9. 12. 1948 seura päätti tämän toimikunnan esityksestä ruveta vuoden 1949 alusta julkaisemaan (aluksi kokeilutarkoituksessa) Geologi-nimistä tiedonantolehteä ja jätti taas asian aikaisemmin valitun toimikunnan tehtäväksi.

Tammikuussa 1949 »Geologi» sitten ilmestyikin ensimmäisen kerran. Lehden toimittajana oli V. Okko ja sen alkukirjoituksena oli professori Pentti Eskolan kirjoitus: »Geologin» tehtävä. Tässä kirjoituksessa mainitaan: »Käyttäkööt geologit sen palstoja antamaan uutisia pienen erikoismaailman asioista, esittämään tarpeitaan ja toiveitaan, esittämään edeltävät tiedot tutkimuksistaan ja löydöistään. — Kuvittelen nyt, että tämä lehti olisi ollut olemassa jo silloin kun 45 vuotta sitten aloin käydä Geologisen seuran kokouksissa ja että se olisi ilmestynyt koko ajan nelisivuisena 10 kertaa vuodessa. Minulla olisi nyt 1 800-sivuinen muistojen ja tiedon aarre.»

Lehti on sitten ilmestynyt säännöllisesti 10 kertaa vuodessa, mutta harvoin on voitu tyytyä vain nelisivuiseen. Kaikkiaan on näiden 12 ilmestymisvuoden aikana »Geologiin» karttunut 870 sivua. Lehden toimittajina ovat olleet Veikko Okko, Erkki Aurola, Simo Kaitaro, V. Marmo, Marjatta Syvänen, A. Matisto ja M. Lehijärvi. Geologi-lehti on osoittautunut tarpeelliseksi ja hyödylliseksi koko geologikunnalle.

Koska lehden sisällysluettelo on julkaistu lehdessä 5-vuotiskausittain, ei sitä ole syytä tässä historiikissa julkaista.

HELSINGIN GEOLOGISEN YHDISTYKSEN TIEDONANTOJA

MEDDELANDEN FRÅN GEOLOGISKA FÖRENINGEN I HELSINGFORS

Helsingin geologisen seuran pöytäkirjat esitelmäselostuksineen julkaistiin vuosina 1916—1920 eripainoksina Teknikern-lehdestä. Selostuksissa ei useastikaan ole esitelmien tarkkaa nimeä mainittu, joten ne on osittain kirjoittajan muovaamia. Suomenkielisten esitelmien nimet oli selostuksissa yleensä käännetty ruotsiksi. Vuosien 1921—1923 selostukset ovat eksaktisempia. Niissä on sivuilla 28—32 julkaistuna myös Helsingin geologisen yhdistyksen r. y. säännöt.

1916

1917

BORGSTRÖM, L. H. Några metoder för bestämning av radioaktiva mineral, s. 1.
RAMSAY, W. Geokronologiska bestämningar gjorda av prof. De Geer vid jämförelse av några av herr Sauramos lerprofiler med diagram från lokaliteter i Sverige, s. 2.
—»— C. A. F. Molengraaffs teori om korallbildning, s. 2.
AILIO, J. Polemiikki Berghellin käsityskantaa vastaan Laatokkakysymyksestä, s. 2.
LAITAKARI, AARNE Paraisilta vasta tavattu klintoniitti, s. 2.
ESKOLA, P. Stilpnomelaani Nadezdan kaivoksesta Aunuksen Pertniemestä, s. 2.
BORGSTRÖM, L. H. Ett försök att beräkna det djup under jordytan, vid vilket vissa mineral med relativt låg kokpunkt eller dissosiations punkt kunna existera, s. 2.
RAMSAY, W. Marina gränserna i södra Finland, s. 2.
—»— Om några detraktionsfenomen, s. 3.
ESKOLA, P. Hiljattain löytynyt pallodioriitti Pöytyällä, s. 4—5.
BRENNER, TH. En synnerligen kalkhaltig morän från Pälkjärvi, Tyytynsaari, Jänisjärvi, s. 5.
SEDERHOLM, J. J. Om expeditionen till Urienhai i gränstrakterna mot norra Mongoliet, s. 5.
VÄYRYNEN, H. Hämeenlinnan seudun kivilajit, s. 5—8.
LAITAKARI, AARNE Etelä-Suomen albiittiepidoottikivistä, s. 8.

1918

BORGSTRÖM, L. H. Till minnet av berging. A. von Julin, s. 9—10.
WAHL, W. Intendenten P. E. Solitander, s. 10.
—»— Bergartsombildning genom inverkan av olikformigt tryck, s. 10—13.
ESKOLA, P. Aunuksen Karjalan geologian pääpiirteet, s. 13—18.

1919

BORGSTRÖM, L. H. Om kvartssand i Finland, s. 1.
SEDERHOLM, J. J. Om migmatit-problemet, s. 1—2.
ESKOLA, P. Malmien systematiikasta, s. 2—3.
BRENNER, TH. Om kvartärgeologiska förhållanden på Kolahalvön, s. 3—7.
SAURAMO, M. Jäänreunan perääntymisesta Keski-Suomessa ja Pohjanmaalla, s. 7.
SEDERHOLM, J. J. Mellersta Fennoskandias geologi, s. 7.

1920

HELLAAKOSKI, A. Halikon vuoden 1919 maanvyörymästä, s. 8—9.
METZGER, A. Om krigsgeologin, s. 9.
RAMSAY, W. Om litorina gränsen i Finland, s. 9.
MÄKINEN, EERO Outokummun alueen geologiasta, s. 10—17.
ESKOLA, P. Tutkimuksia eklogiiteista ja niiden merkityksestä maankuoren rakenteeseen, s. 17.
HACKMAN, V. Kritik av Iddings klassifikation, s. 17.
SAXÉN, M. Otravaaran malmikentän geologiasta, s. 17—21.

METZGER, A. Om den bionomiska paleontologien och de fossila sedimenten I. Faciesförhållandena och deras systematik, s. 21—22.
ESKOLA, P. Jänisjärven tulivuoren jäännöksistä, s. 22—24.
RAMSAY, W. Om metakrona och synkrona isobaser, s. 24.
AUER, V. Meriraja Tampereen eteläpuolella, s. 24.
ESKOLA, P. Aggregatio-olomuodon määritys ja nimitys, s. 24—25.
SAURAMO, M. Mihin perustuu esittämäni Salpausselkä-kronologia, s. 25.
RAMSAY, W. Om språnget mellan två serier av de högsta strandgränserna i Södra Finland, s. 26.

1921

SAURAMO, M. Kannusjärven keltamullasta, s. 1—2.
JÄRNEFELT, H. Suolattoman veden sedimenttien synnystä, s. 2—3.
LAITAKARI, AARNE Suomen geologisen kartan kehityksestä, s. 4—7.
AARNIO, B. Litorinasaven tumman värin syistä, s. 7—11.
LAITAKARI, AARNE Paraisten kalkkikiviesiintymän mineraaleja ja kivilajeja, s. 11.
HACKMAN, V. Hommels petrografiska system, s. 11—12.
METZGER, A. A. TH. Om fossila fiskdjur, s. 13.
SEDERHOLM, J. J. Om den jotniska sandstenens åldersförhållanden, s. 13—14.
LAITAKARI, AARNE Eräs kvartsiittilohkare Kontiolahdelta, s. 14.
TRÜSTEDT, O. Den magnetiska anomalien i guvernementet Kursk (Ryssland), s. 14—15.
METZGER, A. A. TH. Ett fynd av en mammuttand i Esbo, s. 15.
BORGSTRÖM, L. H. Om mineralens hårdhet, s. 15—16.
VÄYRYNEN, H. Puolangan kaoliniesiintymästä, s. 16—18.

1922

TRÜSTEDT, O. Rammelsberg-malmfyndighetens genesis, s. 19—20.
BRENNER, W. Olika jordarters och jordmåners surhetsgrad, s. 20.
TRÜSTEDT, O. Kvartsit-pegmatitkontakter från Outokumpu, s. 20.
BACKLUND, H. Om sambandet mellan vulkanism och tektonik, s. 20.
HAUSEN, H. Jenisej-källområdets geologi, s. 20—21.
PEHRMAN, G. Om ett nickeljärn från Tschinga, s. 21.
VÄYRYNEN, H. Vuorilajien fakiesluokittelusta, s. 21.
HAUSEN, H. Petsamos geologi och malmer, s. 22.
WILKMAN, W. W. Kuopionseudun kivilajeista, s. 22.
KRANCK, E. H. Om tvänne gabbro-områden i Kangasniemi, s. 22—24.

1923

WILKMAN, W. W. Till minne av statsgeologen Hugo Berghell, s. 25—26.
SAURAMO, M. Lohkareen emäkallion etsimisestä, s. 26—27.
ESKOLA, P. Korkean paineen mineraaleista, s. 27.
RAMSAY, W. Nyare arbeten om nivåförändringar, s. 32.
—»— Den baltiska issjöns tappning, s. 33—38.

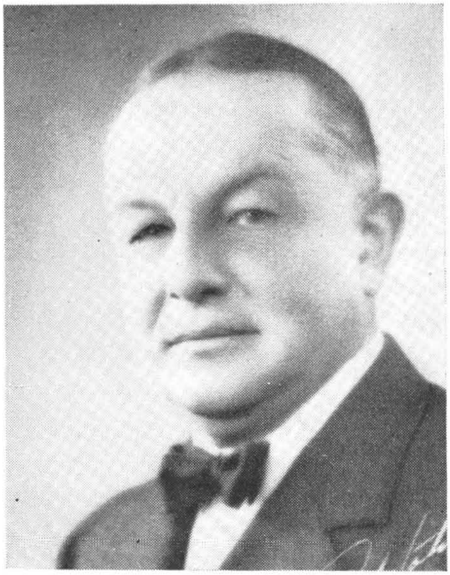
Suomen Geologisen Seuran 50-vuotisjuhlakokous 22. 10. 1936.
The 50th anniversary meeting of the Geological Society of Finland, October 22, 1936.

Pentti Eskola

Kunniapuheenjohtaja v:sta 1953.
Honorary President from 1953.

L. H. Borgström 1876—1954

Kunniajäsen v:sta 1948.
Honorary Member from 1948.

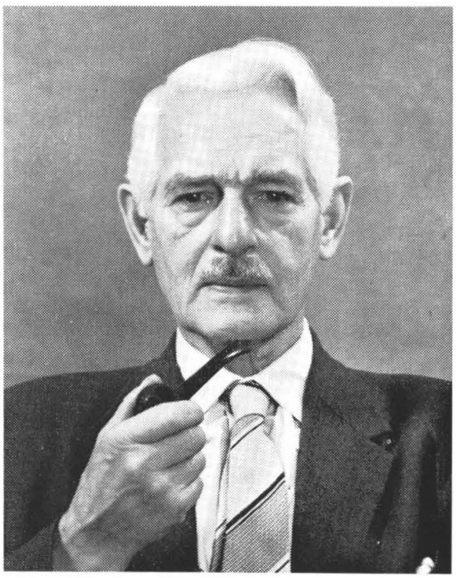


Walter Wahl

Kunniajäsen v:sta 1956.
Honorary Member from 1956.

Eugen Wegmann

Kunniajäsen v:sta 1961.
Honorary Member from 1961.



Aarne Laitakari

Kunniajäsen v:sta 1961.
Honorary Member from 1961.

1

SEDIMENT GROUPS, PARTICULARLY FLYSCH, OF THE PRECAMBRIAN IN FINLAND [1]

BY

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ABSTRACT

The stratigraphic scheme of the Finnish Precambrian presented here has been based on the concept of the cycle of sedimentation. Only one orogeny existed, called Fennian in this text. The former Karelian and Svecofennian sediments have been divided into sediment facies and groups according to the cycle mentioned. There are no important changes in classifying the Karelian sediments anew. The marine and continental flysch facies have been added to the Svecofennian. The sediment area near Oulu has been moved from the Karelian sediments to the continental flysch, Svionian group.

CONTENTS

INTRODUCTION

The stratigraphic scheme of the Finnish Precambrian has temporarily changed during the last half century. The methods of investigation have varied, but the granites and their contacts have been the most important

[1]) Received January 14, 1961.

evidences in comparing the age of the sediment groups. If no suitable contacts existed, the degree of metamorphism mostly served as a criterion.

Until the year 1950, the ideas about the stratigraphy of Finland had been crystallized and simplified, at first by J. J. Sederholm and then in the main by P. Eskola. It was generally thought that there had been two major orogenies, the older Svecofennian and the younger Karelian. In addition, there is a sandstone formation, »oldest red», which has been presumed to be younger than the orogenies mentioned.

Wegmann (1928) introduced the concept of diastrophism and the sedimentation cycle to the study of the Karelian sediments. Recently the same method has also been applied to the stratigraphy of all the Archaean sediments of Finland (Mikkola, 1953). It is evident that in the Karelian belt there are only foreland sediments, whereas in the Svecofennian geosynclinal sedimentation has taken place. There is no evidence against the connection of these two orogenies. Two different cycles had been earlier supposed, principally for the reasons listed below:

1. There is a basal conglomerate in the Karelian but not in the Svecofennian.
2. The sedimentation facies is different.
3. There is a difference in the manner and in the direction of the deformation.

It is easy to see that these differences are dependent upon differences in the environment of the sedimentation and on the stage of the deformation, but not on the age of the orogeny. The Karelian sediments grade into the Svecofennian in several places, and no contact zone can be pointed out. Further, two different orogenies with incomplete cycles of sedimentation, which together form a complete one, are not probable. Therefore only one orogeny has been supposed instead of the former two, and the name Fennian orogeny will here be suggested for this orogeny.

Because there were no absolute age determinations available at the time, the ideas presented were not acceptable for most of the Finnish geologists. The recent publication of some age determinations, made for example by Kouvo (1958), has changed the situation completely. Three papers starting from the idea that only one orogenic cycle existed, have been published to date (Metzger, 1959; Simonen, 1960 a and b). A long discussion preceded these papers in the Finnish periodical »Geologi». Therefore it is quite unbelievable that Metzger arrived at his ideas independently. The development of the stratigraphic scheme of the Finnish Precambrian until the year 1940 has been discussed in greater detail by Eskola (1942). The opinion of an »outsider» is seen in a paper of Schmidt (1960).

The method of absolute age determination is questionable, not the method itself but merely the sampling and, on the whole, the possibility of finding a mineral with an age of the sediment and not of the orogeny or of the source area. On the other hand, the common stratigraphic methods also fail in the unfossiliferous Precambrian. No common agreement exists about the cycle of sedimentation and even the terminology is confusing for a student of metamorphic geology, »a hard rock man». Now that the first summaries have been made, it seems probable that the new methods give results comparable with each other. Therefore, there is no reason to abandon either of them. The usage of the method of absolute ages is simple, and requires only plotting the values on the map and drawing in the zones of the same age. Of course there is difficulty, but nothing can be done to avoid it. These zones concern orogenies only and don't have much to do with the stratigraphy. The sedimentary method, using the idea of diastrophism, gives the sequence but the possibilities of bad errors are many.

There is no use to generalize too much in investigating the Finnish Precambrian; there is no possibility of mixing up different orogenies, as only one has been indicated by both new methods. Still there is difficulty enough in determining the different groups of sediments without any possibility of seeing the form of the geosyncline, with very little known about the environment, and with even the top of the sequence seldom seen. Little can be done to find out the environment of sedimentation, but it is possible to decipher the course of orogeny by using a broad generalization of the concept of the geologic cycle.

THE CYCLE OF SEDIMENTATION

It is well known that no common cycle of sedimentation, but only an average one, exists. It might be possible to assume that the orogeny is quite similar from place to place, but even then the environments of sedimentation differ enormously, and thus the sediments deposited at the same stage of deformation cannot be similar. However, many geologists have found it possible to distinguish between sediments deposited *e. g.*, in a geosyncline or somewhere else, and thus obtain important knowledge of stratigraphy (Krynine, 1941; Pettijohn, 1948; Krumbein & Sloss, 1951). On the other hand, there are geologists who stress the opinion that the type of sedimentation is much more dependent upon the environment than on the stage of orogeny (Twenhofel, 1950; Kay, 1951). Certainly some sediment type, for example grauwacke, can be found outside its common association (geosynclinal), but when larger areas have been investigated, the sediment association in every particular stage of sedimentation seems always to be quite the same. Be-

Table 1. The generalized cycle of sedimentation.

Degree of deformation	Stage of orogeny	Stage of sed. cycle	Sediment facies	Characteristic sediments
None	Quiescence	Peneplanation	Shelly facies	Orthoquartzite-dolomite
Weak	Weak down-warping	Geosynclinal	Miogeosynclinal	Limestone-argillite
Moderate	Downwarping		Eugeosynclinal	Grauwacke-slate, spilite, chert
High	Elevation	Orogenic	Flysch	Marl-grauwacke, breccias
Moderate	Weak elevation	Post-orogenic	Molasse	Subgrauwacke, arkose

cause this seems to be almost the only way to study the stratigraphy of the Precambrian, when strongly metamorphic, it is best to assume an ideal cycle and modify it from place to place. Ideas about that cycle vary widely (Pettijohn, Twenhofel, Kay, *op. cit.;* Dunbar and Rodgers, 1958), but for Precambrian investigation a cycle as detailed as possible would be best (Quennell and Haldemann, 1960; Harpum, 1960). The controversy of speaking about sediments in a geosyncline (Kay, 1951) or geosynclinal sediments (Pettijohn, 1948) is not of interest now. The comparison of the foreland sediments of Pettijohn with the sediments deposited in miogeosyncline of Kay is of the same order of importance, and it seems possible to divide the foreland facies into two groups: Orthoquartzite-dolomite suite (Shelf) and limestone-argillite (Miogeosyncline). The terminology of Pettijohn is the most familiar in Finland, but even though this discussion is about Finnish rocks, his definitions are not detailed enough this time. Keeping in mind that every known sedimentation cycle is fit for one orogeny only, and perhaps for one section of it, and that every cycle intended to explain all types of orogenies must be a very broad generalization, the following scheme (Table 1) will be given as a basis for the stratigraphic study of the Finnish rocks. It cannot be a final one, and is subject to change at any time when necessary. It is not given in a form of a circle because a process of this kind is not a circle, neither in space nor in time. An area, once having been a geosyncline, cannot become a geosyncline again. Contrary ideas have been presented, too (Gastil, 1960).

Concerning this scheme, it has often been thought that the deformation begins along the shore line in a zone of reverse faults, with overthrusting of the continental side. It is also assumed that two parallel geosynclines, miogeosyncline and eugeosyncline (orthogeosyncline), do not exist but that the shore line with sandstones will gradually change into a miogeosyncline with carbonate and argillite sedimentation. This will in turn deepen to a eugeosyncline with a volcanic island arc, volcanism being caused by overthrusting. During the upwarping and strongest deformation of the geosyncline, the flysch facies will form, and after the culmination of the deformation the molasse will be deposited, with these grading into each other. Molasse will complete the cycle.

Only the main features of these sediment groups can be described here: The foreland facies, which is similar to the sediments of the shelf and miogeosyncline, is quite simple. Orthoquartzite with dolomitic carbonate rock is the most characteristic association. Often the group will continue with black schists, limestones, and argillites. Only the proximal areas may show some volcanism. Iron formations are as a rule connected with these sediments, if Precambrian.

The most important and typical sediment association in a eugeosyncline is grauwacke-slate. It is a very monotonous, thick, graded-bedded sequence. Massive grauwackes and grauwacke conglomerates in the upper portions of the sequence are common, as are volcanics and cherts. Also characteristic of the geosynclinal suite is the thickness of the sediments.

Flysch was originally a local name for Alpine sediments, but it has also been used for sediments formed during the elevation of the geosyncline. In Pettijohn's terminology, flysch is a part of the geosynclinal facies. Typical rocks are thin layered grauwacke-slates with graded bedding, marly limestones, grauwackes, and grauwacke conglomerates in the upper parts, and breccias. Volcanics are not common. Some part of the flysch may be continental delta deposits with cross beds and ripple marks, mostly water laid but perhaps partly aeolian, too. In the marine flysch, ripple marks and cross beds are absent. Usually the sediments become coarser upwards.

Molasse is an Alpine name, too. There is no exact limit between flysch and molasse and they can overlap. However, molasse in this text is a general name for the postgeosynclinal, orogenic facies. According to Pettijohn, the characteristic sediment of molasse is subgrauwacke. This is true of the molasse which has been deposited near the source area and is often deformed. The molasse of this kind does not, however, include the arkosic rocks, which are the most typical orogenic sediments and have been deposited near the granitic source area. They are only little deformed by epeirogenic movements. As to the age relations of these different types of molasse, nothing certain can be said in every particular case but probably the less-deformed arkosic rock is usually younger than the molasse deformed by the orogenic movements of the same orogeny. The arkosic molasses have been classified by Dunbar and Rodgers (1958) as rocks of post-orogenic basins. The molasse will commonly become finer upwards. The color may or may not be red or brownish.

This description is not enough for identification of groups or investigation of the stratigraphy of the Precambrian. A sedimentologist cannot find anything new in it. However, this paper can perhaps serve, for a student of metamorphic rocks, as a guide to textbooks and original papers or, most important, to field work. The idea of diastrophism does not work as simply as it has sometimes been applied *e. g.*, in the Finnish Precambrian. The study of sediments includes three main subjects: Sedimentary petrography, sedimentation, and stratigraphy. Of course the stratigraphy is the most interesting, because it is the history of the sediments and the orogeny, but there is no short cut to the stratigraphy without the study of sedimentation and sedimentary petrography. Therefore, no sign of the manner of sedimentation is too insignificant, particularly since there are not too many left. Cross beds and current ripples should be measured. The top of the sequence must be observed where possible, and very often it is. The cross beds and graded beds may serve as criteria but there are others, too, as described by Shrock (1948). The coarseness of the detrital material and its grading may help investigation. Wave ripples could give some information on the site of deposition, and the grading in the thickness of layers may show the direction of the sediment transport, etc.

THE RÔLE OF METAMORPHISM

In the highly metamorphosed areas, the best knowledge of a stratigrapher cannot help us if we do not know the rocks. The study of metamorphism can help now. Different processes may have taken place. Everybody knows that material can change and be differentiated in sedimentary or magmatic processes. Differentiation by metamorphic processes seems to be less familiar, though it can change rocks enormously. Nodules, concretions, and layers formed by metamorphic differentiation have been described *e. g.*, by Eskola (1932). Tuominen and Mikkola (1950) have explained the cordierite rocks as a product of metamorphic differentiation, and have suggested that granitic rocks and sulphide ore bodies may form in the same process. Tuominen (1951) has interpreted the skarns and Mikkola (1955) the ultrabasics of the Orijärvi area in the same way. The diopside-gneisses of the Precambrian have generally been assumed to be metamorphic derivatives of marls (Eskola, 1914).

On the other hand, breccias have sometimes been described as agglomerates, sedimentogenic leptites rich in feldspar as arkoses, evengrained amphibolites as basalts, *etc.* Those errors may happen and particularly so in the flysch areas which have undergone the strongest deformation. If every breccia is taken as an agglomerate, and every amphibolitic layer as a basalt, the degree of volcanism would appear much higher than it really was. The high degree of volcanism would then point to a wrong environment of sedimentation. As to the leptites and amphibolites, the process of metamorphic differentiation as suggested by Tuominen and Mikkola *(op. cit)* has two end members: The mica-feldspar-rich, massive, dark rock, and the light colored leptite. Strongly simplified, the light and dark layers of an argillite have been separated from each other by sliding, and in the continuous process the shear movements have changed the composition of these two rocks. The light layers now form a leptite, its composition but not origin, being very near to an arkose. Every leptite may change, loose all the sedimentary features, and finally form a granitic rock. The origin of the fine grained leptites of the strongest deformed areas is not easy to find out. Therefore the cycle of sedimentation, if it is known in the vicinity, should be used as a guide. The dark layers, which at first form a mica-feldspar rock, can be further altered in at least two different ways. As described in the paper mentioned (Tuominen and Mikkola, 1950) the end member can be a cordierite-anthophyllite rock. All the excess constituents have migrated out. Another possible end member is an amphibolitic rock. An addition of lime could convert the rock formed from dark layers into amphibolite. It is evident that this process has happened, too, in some way or other. This explanation is probably fit for a part of amphibolites, the other part having been marly clays, grauwackes, *etc.* How much the interpreted processes have changed the sediment suite, we can only imagine. However, we must take them into consideration when investigating the Precambrian stratigraphy.

SEQUENCE OF THE FINNISH PRECAMBRIAN

The basement of the Finnish Precambrian sediments is a pre-Karelian granite-gneiss complex, as is now generally assumed. The name Karelian would be better to abandon, since it seems evident as described earlier, that a separate Karelian orogeny does not exist. It is a pity, because the name is a good one. Perhaps it could be used as a group name or better yet as a sediment facies name (Foreland). Instead of Karelian, the older names Kalevian and Jatulian have been used in this text. The sedimentation of the Jatulian begins with a thin basal arkose conglomerate or breccia. It continues with an orthoquartzite, which has commonly been described as continental desert sand sediment (Väyrynen, 1928). The shelly (marine) environment of sedimentation, however, is evident (Mikkola, 1960). The sequence continues with dolomite, black schist, and iron formation. The argillites, and amphibolites of unknown origin are next in the sequence and they belong to the Kalevian group. The unconformity between Jatulian and Kalevian is by no means clear. According to the author's investigations during prospecting work some years ago, the thickness of the Jatulian group is not great. The quartzite does not exceed some hundred metres, the dolomite will seldom exceed a hundred metres, the black schist formation is thinner than the dolomite, and the iron formation is mostly nonproductive, being only some metres thick. Thus the total thickness of the Jatulian is usually less than half a km. The exact determination of the thickness of Kalevian sediments is difficult because of the strong deformation, but a thickness of less than two kilometers is most probable. There are breaks or diastems or discontinuities everywhere. These groups could serve as an example of foreland facies. They would not require any generalizations.

In the stratigraphic sense, the areas right outside of the Kalevian and Jatulian are little known. In central Finland, the grauwacke-slate group of the classic Tampere area is well investigated. The pioneer work in this area was done by J. J. Sederholm (1897). He called the area and group the Bothnian as it is called in this text, too. Seitsaari (1951) has described the petrography, metamorphism, and the structure of the Tampere schist belt. From the stratigraphic point of view, Simonen (1953) has made a useful study in this area. He has estimated the thickness of sediments to be at least 8 km and the sequence from top to base is as follows:

Basic volcanics	>1 000 m
Conglomerates, grauwackes and arkoses	700— 800 m
Basic and intermediate volcanics	800—1 500 m
Arkoses, grauwackes and pyroclastics	>3 000 m

The associations of grauwacke-slates and basic volcanics are typical of the geosynclinal facies. Other rocks existing here belong to the same geosynclinal suite. The only main types missing are chert and spilite, but there can be metamorphic derivatives of them not discovered or identified as yet. The thickness of the sequence is one of the most characteristic features of the geosyncline. On the other hand, there is something strange in the direction of the geosyncline as compared to the foreland area, because it is difficult to see how these two areas could belong into the same orthogeosyncline (Kay, *op. cit.*). This must be a subject of further investigation.

The next member in the sequence, as far as it is known, is the Svionian group. On the map it is seen between Helsinki and Turku and a part of it is best known as the Orijärvi region. The old name Svecofennian formation has been abandoned here for the same reason that Karelian has been abandoned. Further, the term »formation» has, in the American literature, a completely different meaning than it has had in Finland, and the American terminology has been preferred in this text.

The petrography of the Orijärvi area has been described by Eskola (1914), the structure has been described by Tuominen (1957), and some efforts to determine the stratigraphy have been made by Mikkola (1950). According to later investigations of the author, the most striking feature in the sediments of the Orijärvi region is the discontinuity. It is much due to the faulting but at least as much due to the manner of sedimentation. The deformation has been very strong and therefore determination of the sequence has been extremely difficult. The primary sedimentary structures have been destroyed in most places. In a few cases the graded bedding has been preserved and it has been used to determine the tops of the beds. Because of the strong cross folding and often vertical or overturned dip of the strata, the top would have been too difficult to determine without sedimentary structures. Assuming that the repetition of the sequence is due only to the folding and faulting, the thickness of this group is about one kilometer. But if the cycle is due to rhythmic deposition, the thickness may be twice or three times as much. The lowermost rock in the Orijärvi area is an amphibole schist. Formerly, this rock was supposed to be the same as the topmost amphibolite, due to repetition, but it is so much different that it cannot be the same. The cycle mentioned above begins with an argillite, now mostly in the form of mica cordierite schist. Some of the coarser parts of this argillite are merely grauwacke-slates with graded bedding. The topmost part of this rock is rich in lime, in places it has been a marl, and it now occurs in the form of a diopside-amphibolite. Thin limestone beds, some metres in the whole, have also been deposited, but they have mostly glided to the crests of folds forming thick lenses. Next in the sequence is a black, sulphide bearing schist, but its occurrence is very irregular and it is always a thin one, never exceeding some ten metres. Whether it represents a black schist-forming environment or only a reworked older black sediment is not exactly known, but because it occurs in close connection with an ironstone formation (uneconomic), reworking does not seem probable. After the black schist-ironstone horizon, there probably is a break in the sedimentation, and the sequence begins anew with a polymict grauwacke conglomerate. Sometimes there are graded bedded slates associated with the conglomerate, and on the top of the conglomerate there is in places a massive bed of grauwacke only some metres thick. The topmost member is a bed of amphibolite. It could have been a volcanic rock, but most probably it is a metamorphic variant of some lime-rich sediment, because there are remnants of rock resembling diopside-amphibolite. The sequence, as determined from limited areas, is from the top down:

Amphibolite
Grauwacke (Subgrauwacke?)
Grauwacke conglomerate
Banded iron ore
Black schist
Marble (Limestone)
Diopside-amphibolite (Marl)
Argillite with grauwacke-slate
Amphibole schist
Basement unknown

It is evident that the group described above belongs to the geosynclinal suite. The characteristic association of these sediments is that of marine flysch: The beds and the sequence are discontinuous, the deformation is strong, and it is possible that the deformation has begun with subaquaceous sliding in an early stage of sedimentation (Tuominen and Mikkola, *op. cit.*). The coarser rocks occur in the uppermost parts of the sequence. Volcanics are rare or absent. Quartzites are very rare (see Pettijohn, 1943). Grauwackes are not common as in true geosynclines, but they do occur. Current beds and ripples do not exist. As to the upper parts of the sequence, the possibility of molasse is not out of question.

From the investigations of Salli (1955), from the personal communication of Mr. J. Koskinen, M. A., from explanations of geologic maps, and from what has been seen during excursions and minor field works, the conclusion could be made that at least a part of the schist area between Tampere and Oulu belongs to the continental flysch facies, as delta deposits with cross beds and muddy sandstones are characteristic features of the sedimentation. The sediment area of Oulu has always been connected with Karelian sediments, but the sediments of the area do not belong to the foreland facies. There are very few true orthoquartzites or dolomites. Conglomerate areas are large and the rocks are not basal conglomerates typical of the Karelian zone. The sediments closely resemble the Svionian sediments, and have here been classified in the upper flysch. A part of these sediments may belong to the molasse. The stratigraphy of these areas may be the most complicated in Finland, but because many of the sedimentary structures have been preserved, knowledge of the environment of sedimentation and of the source areas and transport is possible to determine through detailed investigation and measurements.

There are also some minor areas of molasse in Finland. Beginning from the north, the Kumpu formation has been thought to be a molasse (Väyrynen, 1954). It consists of coarse, brownish sediments, the source rocks probably having been Kalevian and Jatulian sediments. The Fennian orogeny, which has deformed the Kalevian and Jatulian sediments, has also deformed the Kumpu formation. The formation has been here classified in the lower molasse. The next area to the south is the Muhos formation near Oulu. There are undeformed redbeds which may or may not belong to the Precambrian. There is no evidence, but the area has been tentatively classified as Jotnian (Simonen, 1960 b). The area near Pori is a typical arkose sandstone, coarse-grained, red-colored rock (Simonen and Kouvo, 1955). It is called Jotnian sandstone or the »oldest red». This rock belongs to the typical orogenic suite, and it is difficult to understand that there could be a question about an orogeny other than Fennian. As seen in Table 2, the age of the Jotnian sediments should be 1.3 bill. years (Simonen, 1960 b)

Table 2. Development of the stratigraphic concept of the Finnish Precambrian.

Eskola, 1921	Sederholm, 1932	Mikkola, 1953	Väyrynen, 1954	Metzger, 1959	
Jotnian	Jotnian Hoglandian	Jotnian	Jotnian	Jotnian	
Karelian	Kalevian Jatulian	Svecofennian	Kalevian Jatulian		
Svecofennian	Bothnian Ladogian Svionian	Karelian	Fennonian Svionian	Sveco- fennian	Kare- lian
Basement unknown		Pre-Karelian basement	Basement unknown	Pre-Karelian basement	

Simonen, 1960

Svecofennian belt	Karelian belt: East Finland	Karelian belt: North Finland	Age
Jotnian sediments, unmetamorphic			Age: 1 300 m. y., diagenesis.
Anorogenic rapakivi granites			1 620 m. y.
Orogenic plutonic rocks			1 750—1 850 m. y.
Upper Sveco-fennian		Kumpu formation	
Middle Sveco-fennian	Kale-vian	Lappo-nian	
Lower Sveco-fennian	Jatu-lian		
	Belomorides: Tuntsa-Savukoski formation		1 900—2 000 m. y.
Basement unknown	Pre-Karelian basement		2 600 m.y.

This text, 1961

CYCLE: FENNIAN

Group:	Sediment facies:
Jotnian	Upper molasse
Kumpu	Lower molasse
Svionian, continental	Upper flysch
Svionian, marine	Lower flysch
Bothnian	Eugeosynclinal
Kalevian	Miogeosynclinal
Jatulian	Shelf

Basement: Gneisses of the earlier orogeny (Pre-Fennian)

but it is the age of diagenesis. On the other hand, because no younger orogeny can be pointed out, the Jotnian formation has been classified in the upper molasse of the Fennian orogeny.

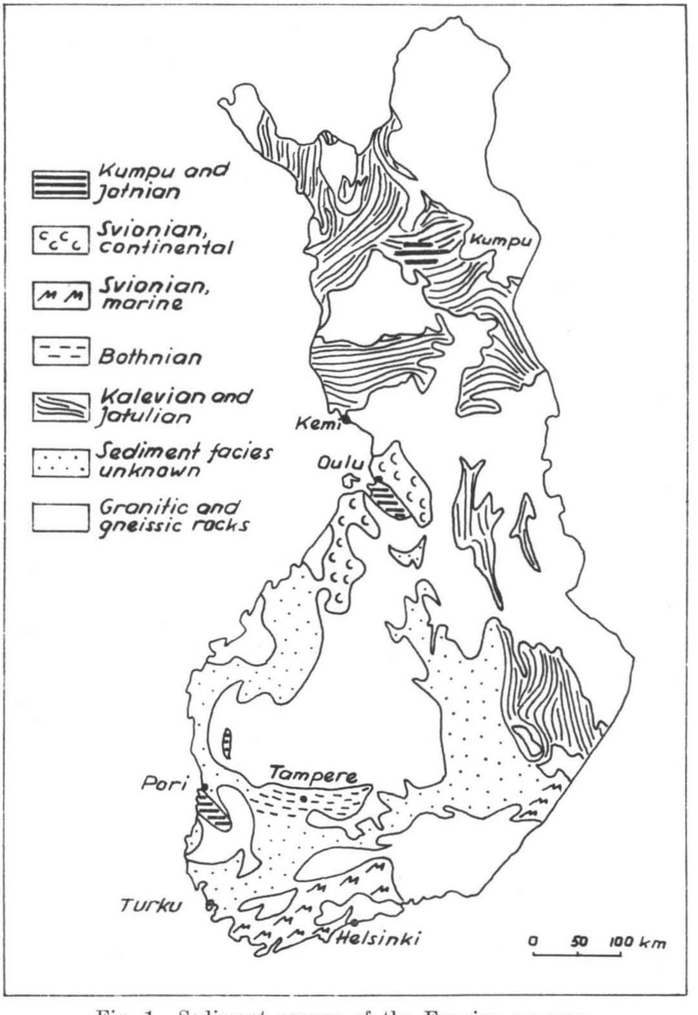

Fig. 1. Sediment groups of the Fennian orogeny.

SUMMARY

In the foregoing, the Finnish Precambrian has been divided as follows: The former Karelian formation into Jatulian and Kalevian, as has been done earlier, too. It is done here only so that the cycle of sedimentation would be easier to apply. The former Svecofennian has been divided into Bothnian and marine and continental Svionian groups, beginning from the lowermost members. The youngest Precambrian sediments are the Jotnian sandstones and slates. All these sediments have been deformed more or less by one orogeny, and the Jotnian has been deformed only very little. This orogeny has no name as yet, but it has been called »Fennian» in this text. Thus all the sediments mentioned here belong to the Fennian system or cycle. It would be better to abandon the older formation names of Karelian and Svecofennian.

As seen from the map, a large part of the Finnish Precambrian has not been divided into the groups mentioned, because they are very little known from the stratigraphic point of view. Probably some more subdivision is necessary, too. Errors in the groups already defined are very possible.

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2

LAUMONTITE FROM VIITASAARI, CENTRAL FINLAND [1]

BY

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ABSTRACT

Laumontite (leonhardite) has been found in a hydrothermal vein in Archaean gneisses, in Central Finland. A chemical analysis, the physical properties, and the dimensions of the leonhardite monoclinic unit cell are listed. A list of d-spacings is included.

INTRODUCTION

Zeolites have earlier been found in several connections in Finland, though only a few have been thoroughly investigated and described in the literature. The most comprehensive study on a zeolite from Finland is a recently published paper on laumontite by Eskola (1960). A short note on the same subject was already published 25 years ago (Eskola, 1935). Laumontite and analcite have later been identified from a Li-pegmatite by Neuvonen and Vesasalo (1960). By oral communication the author is aware of several other zeolite occurrences connected with ore bodies (Toivo Mikkola: Kolari, Lapland; O. Kouvo and Y. Vuorelainen: Vihanti and Korsnäs mines). Eskola further mentions that laumontite has been found in loose boulders in Central Finland (Eskola, 1960). A similar occurrence has been reported by O. Näykki, M. A. (oral communication).

OCCURRENCE

In connection with ore-drillings in Viitasaari, Central Finland, in the years 1957—1959, the author had an opportunity to investigate the core

[1] Received February 15, 1961.

material of some 30 diamond-drillings. The geological setting is: Archaean amphibolitic and micaceous gneisses, chiefly of sedimentary origin, with garnetiferous zones. The ore mineralization is represented by the minerals sphalerite, chalcopyrite and galena in rather uneven dissemination. A still later phase of mineralization is represented by sparse and thin veinlets of zeolites, mainly laumontite. In some instances even a narrow zone of the surrounding rocks has been partly or almost wholly altered to laumontite.

The material for this study was taken from a 0.5 cm thick veinlet occurring about 50 m below ground level (borehole R 36/Vs—58/77.60—77.61). The mineral was first identified by optical means in thin section, and later by X-ray methods. Accessory analcime, prehnite and natrolite were identified in thin section and later confirmed by X-ray powder photographs.

CHEMICAL COMPOSITION

A sample of 200 mg of carefully handpicked material from the vein mentioned was analyzed by H. B. Wiik (Table 1, I). Later he analyzed a similar sample of 57 mg for the alkalies (Na_2O and K_2O), using a Beckman-type flamephotometer.

The analysis conforms well with earlier published ones. For comparison three analyses from recent publications have been cited (Coombs 1952; Koch 1959; and Eskola 1960, see Table 1). The cation proportions and cation numbers on the basis of 48 Oxygen atoms have been calculated by the present author.

The unusually high Ca content in the laumontite from Viitasaari is probably due to slight impurities of prehnite occurring with the laumontite. The high Al and low Si suggest the same source of deviation. Taking in consideration, however, the very small sample analyzed, it is not surprizing that only a few grains of prehnite might cause such a deviation. As any treatment with liquids was avoided before analyzing, a definite control of possible impurities in the sample was difficult.

The low water content indicates that the sample, stored dry at room temperature for several months, has given off some water and formed the variety known as leonhardite (see *e. g.* Coombs, 1952). As will be seen this is confirmed by the optical and X-ray investigations.

PHYSICAL PROPERTIES AND X-RAY DATA

The specific gravity of the laumontite from Viitasaari was determined by the floating method with the aid of a Westphal balance. The optical properties were determined on a crystal fragment with the aid of a Wald-

Table 1.

	I	II	III	IV
Weight percentage				
SiO_2	49.95	52.04	51.57	50.70
Al_2O_3	23.36	21.46	22.80	22.53
TiO_2	n.d.	—	—	—
Fe_2O_3	n.d.	0.12	0.24	0.04
MgO	0.00	tr.	0.11	—
CaO	12.94	11.41	10.43	11.54
Na_2O	0.4	0.20	0.92	0.40
K_2O	0.3	0.66	1.09	0.30
H_2O tot.	12.82	13.80	12.98	14.41
	99.77	99.69	100.14	99.92
Cation proportions				
Si	8 312	8 660	8 582	8 437
Al	4 582	4 209	4 472	4 419
Fe	—	15	30	5
Mg	—	—	27	—
Ca	2 307	2 035	1 860	2 058
Na	129	65	297	129
K	64	140	231	64
OH	14 232	15 318	14 408	15 996
Cation numbers on basis of 48 Oxygen atoms				
Ca	4.27	3.79	3.43	3.85
Na	0.24	0.12	0.55	0.24
K	0.12	0.26	0.43	0.12
Al	8.49	7.84	8.26	8.26
Si	15.41	16.13	15.85	15.78
H_2O	13.19	14.23	13.31	14.96
Ca + (Na + K)/2	4.45	3.98	3.91	4.03
Ca + Na + K	4.63	4.17	4.41	4.21
Al + Si	23.90	23.97	24.11	24.04
Physical properties				
Sp. gr.	2.26 ($\pm$0.01)	2.29 ($\pm$0.01)	2.28 ($\pm$0.01)	2.26—2.29
α	1.504 ($\pm$0.001)	1.507 ($\pm$0.002)	1.509	1.505—1.513
β	1.512 ($\pm$0.001)	1.516 ($\pm$0.002)	—	—
γ	1.516 ($\pm$0.001)	1.518 ($\pm$0.002)	1.521	1.516—1.521
$2V_\alpha$	44° ($\pm$3°)	26° ($\pm$4°)	30° (var.)	—
$c \Lambda \gamma$	30° ($\pm$2)	32 ($\pm$2°)	—	38°

Explanation to Table 1.

I. Laumontite from hydrothermal vein, Viitasaari, Finland. Analysis made on 200 mg sample by H. B. Wiik; Na_2O and K_2O determinations were made later on a separate 57 mg sample, also by H. B. Wiik. The optical and specific gravity determinations were made by F. Pipping.

II. »Laumontite on tuffa, Hungary, No. 192». Mineralogical Museum, Cambridge. Analysis and optical determinations by D. S. Coombs (Coombs, 1952, Tables I and III). Leonhardite form optics.

III. Laumontite from Kuhmoinen, Finland. Laumontitized migmatite. Analyst Y. Hentola (Eskola, 1935, 1960). »Fresh-mineral» optics. Sp. gr. determined by the present author.

IV. Laumontite from Petersberg, Halle a. d. Saale, Germany. Hydrothermal miarolitic mineral. Analyst G. Schneidereit (Koch, 1959).

Table 2. Cell dimensions for laumontite (leonhardite) from Viitasaari, Finland obtained by single crystal and powder photograph methods.

	1	2	3	4	5
a_o		14.78 Å		14.72 Å	14.79 Å
b_o	12.91 Å			13.02 Å	13.00 Å
c_o		7.52 Å	7.57 Å		7.57 Å
β	112°.0	112°.0 (± 0.3)			111°.9 (±0.7)
(110) Λ (1$\bar{1}$0)				92°.74	

Explanation to Table 2.

1. b-axis oscillation photograph.
2. b-axis Weissenberg zero-layer photograph.
3. c-axis oscillation photograph.
4. c-axis Weissenberg zero-layer photograph.
5. Calculated from powder-photograph reflexions 200 (6.86 Å), 040 (3.25 Å), 002 (3.51 Å), 201 (4.17 Å) and $\bar{2}$01 (6.15 Å)

For the calculations the following wavelengths have been used:

Cu Kα = 1.5418 Å

Cu Kβ = 1.3922 Å

Table 3. The dimensions and content of the unit cell of laumonite (leonhardite) from Viitasaari, Finland.

a_o	= 14.76 Å (±0.02)	Si	15.17
b_o	= 12.98 Å (±0.03)	Al	8.36
c_o	= 7.55 Å (±0.02)	Ca	4.21
β	= 112°.0 (±0.3)	Na	0.24
V_o	= 1341.2 × 10-24 cm³	K	0.12
$D^{4°}_{20°}$	= 2.26 (±0.01)	H_2O	12.99
$V_o \cdot D$	= 3031.1 × 10-24 gr.	O	34.74

mann-sphere in a U-stage. The values listed in Table 1 indicate the leonhardite form, having low indices of refraction and a rather great negative axial angle. The specific gravity and optical properties of the earlier cited laumontites are also given in Table 1.

Thorough X-ray studies of laumontite have been published by Coombs (1952) and Heritsch (1956). The former author established clearly measurable differences in cell-size between laumontite and its water-poor equivalent leonhardite. The corresponding changes in optical properties were also recorded by Coombs. The study by Heritsch is a purely X-ray crystallographic investigation.

Table 4. D-spacings for laumontite (leonhardite) from Viitasaari, Finland obtained with 114.6 mm diameter camera, Cu K_{α}-radiation, wavelength = 1.5418 Å. Silicon used as internal standard.

d		d	I
9.4	vs	2.788	vw
6.86	vs	2.571	vw
6.15	vw	2.436	w
5.04	vw	2.367	vw
4.72	s	2.275	vw
4.50	w	2.151	w
4.17	vs	1.997	vw
3.51	s	1.957	vw
3.25	ms	1.704	vw
3.18	w	1.528	vw
3.04	w	1.447	vw
2.875	vw	1.310	vw

The present X-ray data (listed in Tables 2 and 4) are based on single crystal oscillation and Weissenberg photographs with b and c axes rotation, and also on powder photographs taken with a 114.6 mm diameter camera. (Buerger, 1942; Klug and Alexander, 1954). D-spacings for 2 Θ angles up to 75° are listed in Table 4, with ocular estimations of intensities.

The relative intensities of the reflections of the Weissenberg photographs agree exactly with those listed by Heritsch (1956). Only some of the very weak reflections listed by him are absent from the photographs taken by the present author, obviously due to differences in time of exposure. The unit cell dimensions given here (Table 3) are averages of the values contained in Table 2. The unit cell content has been calculated on the basis of these average numbers.

The unit cell constants of leonhardite now presented agree with the values given by Coombs (1952) and Heritsch (1956), within the limits of error as indicated by them, except for the b_0 which is about 1 % smaller than their values.

Acknowledgements — The author is indebted to H. B. Wiik, Ph. D. for making the analyse. Sincere thanks of the author are also due to professor Pentti Eskola for kind encouragement to publish these data, and for placing his material at the authors disposal for comparison. He also wants to express his thanks to prof. K. J. Neuvonen, University of Turku, and Atso Vorma, M. A. of the Geological Survey of Finland, for their abundant advice during this work.

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3

METAMORPHIC FACIES AND STYLE OF FOLDING IN THE BELT SERIES NORTHWEST OF THE IDAHO BATHOLITH [1]

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ABSTRACT

The sedimentary rocks of the Belt series of Precambrian age northwest of the Idaho batholith were folded and metamorphosed during Jurassic time. The grade of metamorphism increases from the greenschist facies to the amphibolite facies in a distance of about 60 km. The area studied lies within the PT-field of the epidote-amphibolite and amphibolite facies, which are divisible into several subfacies. Index minerals such as staurolite, almandite, cordierite, kyanite, andalusite, and sillimanite always accompanied by muscovite, together with the occurrence of the three aluminium silicates, make it possible to estimate the relative temperature and pressure range of recrystallization for each of the subfacies. The areal distribution of typical mineral assemblages is related to the intensity and type of folding. High-pressure assemblages are more common, and the folding tighter, in the western part of the area.

In most of the rocks of the epidote-amphibolite facies the folds are large and open, but toward the batholith they become increasingly tighter, until in the zone next to the batholith isoclinal flow folds prevail. The distribution of axial-plane cleavage in the middle zones depends largely on the material folded.

CONTENTS

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INTRODUCTION

The area studied lies in the southern part of the Idaho »panhandle» and northwest of the Idaho batholith (Fig. 1). The metamorphic country rocks of the batholith belong to the lower part of the Belt series of Precambrian age. They consist of interbedded pelitic schists, quartzites, granofelses, and diopside- or biotite-bearing gneisses. The gneisses and their granular low-metamorphic equivalents, the granofelses [1]), contain calcium-rich layers that, together with aluminium-rich pelitic schists, form an ideal material for the study of metamorphism. The alternation of fairly thin units of schist, quartzite, and gneiss or granofels, and the presence of distinct bedding within each unit, make it possible to observe structural elements in detail and to determine their attitudes.

Details of stratigraphy, petrology, and structure of highly and moderately metamorphosed rocks in zones 0 to 15 and 15 to 50 km north of the batholith, are described in longer reports (Hietanen, in press). The much less metamorphosed equivalents of these rocks farther north have been described by Ransome and Calkins (1908), Pardee (1911), Calkins and Jones (1911), Umpleby and Jones (1923), and Wagner (1949).

The metamorphism increases steadily from north to south over a distance of about 60 km. In the north the rocks were metamorphosed to the greenschist facies, and in the south, near the batholith, to the higher-temperature amphibolite facies. Several subfacies can be recognized and mapped; each

[1] A rock name coined by Goldsmith (1959).

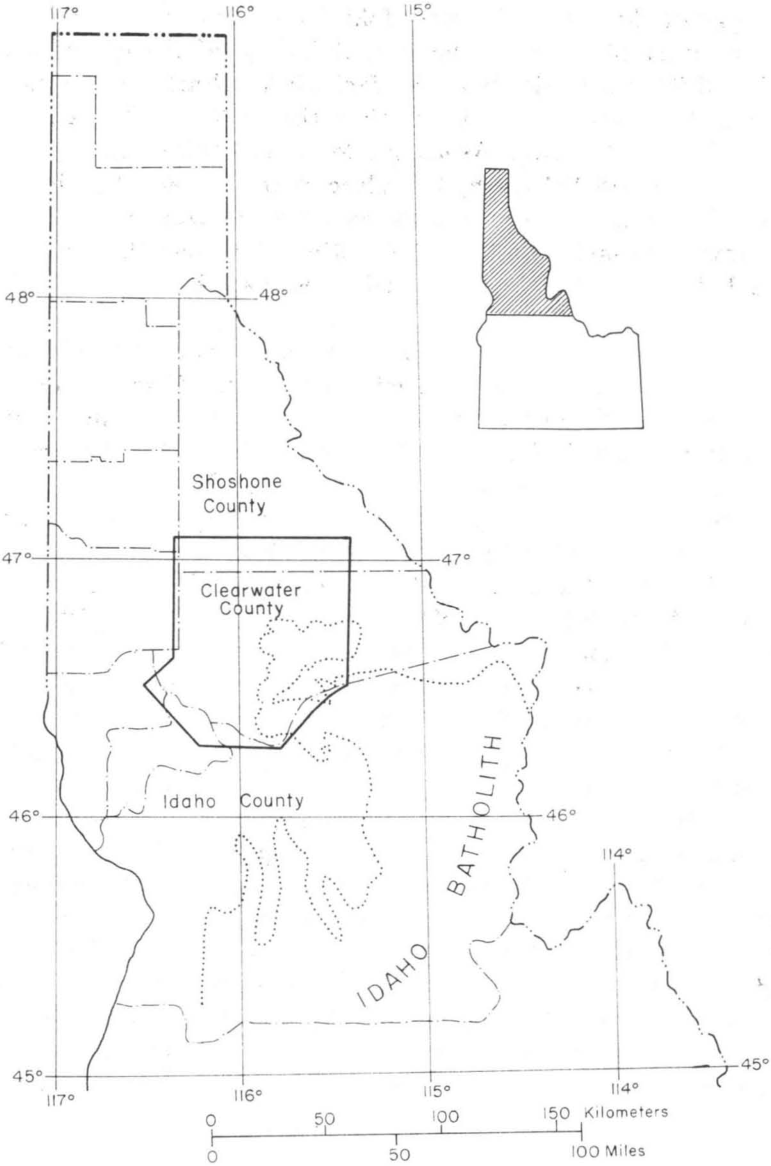

Fig. 1. Location of area studied (encircled) in northern Idaho. Outline of the northern part of the Idaho batholith is shown by a dotted line.

is characterized by one or more index minerals, such as staurolite, andalusite, kyanite, sillimanite, and cordierite, and by typical mineral assemblages, each stable only within a certain temperature-pressure field (Fig. 2, Plate I).

The rocks of the Belt series were folded and metamorphosed during the Nevadan orogeny (Jurassic). The lead-alpha age of the quartz monzonite batholith and its quartz diorite border facies is late Early Cretaceous (Larsen and others, 1958), and thus younger than the folding and attendant metamorphism of the Belt series. As the grade of even this early metamorphism increases toward the batholith, the place now occupied by the batholith must have been a heat center at least as early as Jurassic time. Structural studies show that many of the migmatitic veins near the batholith were emplaced before the folding, which indicates that the granitic rocks began to form even earlier.

In the following pages the rocks and their mineral assemblages are described first, then some of their structural features. Comparison of the intensity of metamorphism with the structures shows that the course of recrystallization in the field of changing temperature and pressure, and the intensity and type of folding are closely related.

It has been pointed out by de Sitter (1956) that two different types of folding, concentric folding and cleavage folding, are frequently found in alternating layers of the same strata. Concentric folds, called flexural slip folds by Knopf and Ingerson (1938) and flexure folds by Fairbairn (1949), are formed in distinctly stratified rocks by bending of compositional layers. In many thick homogeneous layers of the same strata, a new shear plane may be developed as a result of deformation, and folds are then formed by slip along these rather than along the compositional layers. These folds, called cleavage folds by de Sitter (1956), flexuralslip folds by Knopf and Ingerson (1938), and shear folds by Fairbairn (1949), are most common in rocks deformed at moderate depths, but become increasingly rare where the deformation occurred at higher and lower levels in the earth's crust.

When the same strata are studied throughout a large area within which the degree of metamorphism varies widely, it is possible to eliminate the effects of differences in stress level and in homogeneity of the folded material, and to observe the relations between style of folding and degree of metamorphism. These relations are well demonstrated in the various metamorphic zones northwest of the Idaho batholith.

OUTLINE OF LITHOLOGY AND STRATIGRAPHIC SEQUENCE IN THE BELT SERIES

Metamorphosed equivalents of the rocks of the Belt series northwest of the Idaho batholith consist of several units of quartzite and diopside- and biotite-bearing gneisses interbedded with pelitic schists. Comparison of the lithology, stratigraphic sequence, and thicknesses of the units indicates that equivalents of all but the youngest formation exposed in the Coeur d'Alene district — Prichard, Burke, Revett, St. Regis, and Wallace (Ransome and Calkins, 1908) — are represented in various parts of the area. The lowest formation, the Prichard, which covers the north-central part of the area, consists mainly of garnet-mica schist with at least two quartzite units interbedded. In most sections the two quartzite units range in thickness from 30 to 100 m, and are separated by more than 700 m of garnet-mica schist. About 400 m of garnet-mica schist is exposed above the upper quartzite unit; anorthosite, lime-silicate rocks, and about 600 m of aluminium-silicate-bearing schist underlie the lower quartzite unit. The bottom of the Prichard formation is not exposed.

The Burke formation, which in northern Idaho consists of impure quartzite, could be identified only in the northeastern part of the mapped area, where it consists of about 300 m of thin-bedded micaceous granular quartzite. The Revett quartzite is about 500 m thick; it consists of thick-bedded coarse-grained pure quartzite that is easy to identify in the field. The St. Regis formation, which overlies the Revett quartzite, consists, in this area, of garnet-mica schist. Above the St. Regis formation are at least two lithologic units of diopside- and biotite-bearing gneisses interbedded with aluminous mica schist. These rocks form a heterogeneous sequence that, allowing for changes due to metamorphism, is lithologically similar to the Wallace formation and is correlated with it. As this formation consists of calcareous layers interbedded with aluminium-rich pelitic schist, it is better suited than any of the others for study of the grade of metamorphism. The lithologic units within it, moreover, can be followed for a distance of about 50 km from northwest to southeast, and thus from the field of moderate metamorphism to that of high-grade metamorphism.

METAMORPHIC FACIES

The distribution of metamorphic facies in the Belt series was determined by study of the mineral assemblages in the aluminium-rich pelitic layers and the interbedded calcareous layers. The composition of plagioclase in the gneissic layers that contain epidote minerals was also helpful as an indicator of facies, especially where no aluminium-rich silicates were found. The distribution of some of the critical minerals in the pelitic schist layers is shown in Figure 2 (Pl. I). This needs, however, to be supplemented by describing the mineral assemblages in layers of differing composition within each metamorphic zone.

EPIDOTE-AMPHIBOLITE FACIES

The schist in the northwestern and northeastern parts of the area is fine to medium grained, and contains some plagioclase and a considerable amount of muscovite, biotite, and garnet. In addition to these minerals, staurolite and kyanite are common in the schist near Clarkia and in the northeastern corner of the mapped area.

The gneissic layers interbedded with the schist consist of quartz, plagioclase, microcline, and one or more of the minerals diopside, biotite, and hornblende. These layers are fine grained and can be seen in the field to be granular, but under the microscope the grains have rather irregular, slightly interlocking borders.

No aluminium silicates are in the schist in the extreme northwestern part of the area, nor in the schist 5 km or more north of the mapped area. In this outer zone, biotite and garnet crystallized late, and the plagioclase of the gneissic layers is more sodic than that in the gneissic layers that are interbedded with kyanite- and staurolite-bearing schist near Clarkia. Hornblende instead of diopside crystallized in many calcareous layers north of Clarkia. Thus it seems that the rocks north of Clarkia and those 5 km or more north of Bathtub Mountain recrystallized at lower temperature and pressure than those near Clarkia; accordingly two subfacies are recognized, one characterized by biotite, almandite, epidote, and albite and the other by kyanite and staurolite. The third subfacies, characterized by the occurrence of andalusite instead of kyanite, is found only in the east-central part of the area.

BIOTITE-ALMANDITE SUBFACIES

The common mineral assemblages in the various layers of the Wallace formation in the extreme northwestern part of the area are as follows:
Pelitic assemblages:

1) Almandite-biotite-muscovite-plagioclase (An_{5-15})-(zoisite)-quartz.
2) Muscovite-biotite-microcline-plagioclase (An_{8-27})-(clinozoisite)-quartz (Table 1, No. 1908).

Quartzo-feldspathic assemblages:

3) Biotite-muscovite-plagioclase (An_{8-15})-(clinozoisite)-scapolite-quartz.
4) Biotite-microcline-plagioclase (An_{8-10})-(quartz (Table 1, No. 1906).

Calcareous assemblages:

5) Hornblende-biotite-plagioclase (An_{10-15})-epidote-quartz, with or without microcline.
6) Hornblende-biotite-microcline-calcite-quartz.

7) Hornblende-microcline-plagioclase (An_{5-15})-scapolite-quartz, with or without calcite and epidote (Table 1, 1915).
8) Biotite-calcite-epidote-albite-quartz, with or without microcline or scapolite or both.

The anorthite content of the plagioclase that occurs with the epidote minerals is highly variable and depends in part on the calcium-aluminium ratio of the original sediment. Where enough aluminium was available, the plagioclase is richer in anorthite; the highest An-content measured in the thin sections is An_{27}. If this value is compared with the equilibrium diagram of epidote and plagioclase given by Ramberg (1949), the indicated temperature of recrystallization in this part of the area would have been about 350°C, and thus within the range of the epidote-amphibolite facies but below the upper limit of this facies which, according to Barth (1952, p. 338), should be placed at 400° C, where plagioclase An_{37} crystallizes in equilibrium with epidote.

STAUROLITE-KYANITE SUBFACIES

Small brown crystals of staurolite, and white clusters that are ½ to 2 cm long and consist of small crystals of kyanite and muscovite are common in the fine-grained garnet-mica schist east of Clarkia. Near Anthony Peak and in the northeastern part of the mapped area, the rocks contain aggregates and individual crystals of kyanite 2 to 8 cm long and garnet crystals 3 to 10 mm in diameter. Staurolite crystals 0.5 to 2 cm long are common in the kyanite-bearing garnet-mica schist in the northeastern part of the area. Staurolite without kyanite was found only in a few isolated localities, such as the ridge extending south of Bathtub Mountain, where staurolite crystals 3 to 6 cm long form about 25 percent of the rock. Since most outcrops around this locality also contain kyanite, it seems likely that the crystallization of abundant staurolite was due to a high content of iron rather than to a locally lower temperature and pressure. This view is supported by an occurrence of minute needles of sillimanite in a few thin sections of rocks collected at Bathtub Mountain.

Staurolite crystals, ranging from 1 to 2 cm in length, are abundant in the garnet-mica schist north of Surveyors Ridge, in the northeastern corner of the mapped area. Pseudomorphs of muscovite after either andalusite or kyanite are common here, and indicate a later metasomatic addition of potassium. The anorthite content of the plagioclase in the epidote-bearing layers interbedded with these schists is An_{30-32}, which indicates that the

Table 1. Chemical composition in weight and ionic percentages,
series metamorphosed to the epidote-

	1558 Revett quartzite Surveyors Ridge	1615 Kyanite-almandite schist 5 km east of Clarkia	1613 Kyanite-almandite schist 6 km east of Clarkia	1579 Kyanite-staurolite schist Bathtub Mountain	1578 Biotite quartzite 7 km SE. of Bathtub Mtn.	1908 Biotite gneiss Merry Creek, 7 km NE. of Clarkia
Constituent						Weight
SiO_2	95.90	62.51	75.14	66.28	83.55	73.21
Al_2O_3	2.25	19.78	12.51	16.01	7.53	12.87
Fe_2O_3	0.17	1.24	0.61	1.31	0.52	0.86
FeO	0.00	5.86	2.97	4.86	0.70	1.39
MnO	0.00	0.09	0.04	0.05	0.04	0.02
MgO	0.09	2.09	1.76	2.78	1.49	2.17
CaO	0.00	0.54	0.57	0.36	1.66	1.48
Na_2O	0.06	0.75	0.81	0.89	1.26	3.32
K_2O	0.66	3.78	3.17	3.88	1.51	2.81
TiO_2	0.17	0.56	0.49	0.55	0.26	0.49
P_2O_5	0.01	0.20	0.13	0.16	0.05	0.06
CO_2	0.02	0.02	0.01	0.01	0.01	0.01
Cl [1]	—	—	—	—	—	—
H_2O+	0.34	2.15	1.43	2.37	0.82	0.72
H_2O-	0.02	0.12	0.06	0.13	0.15	0.17
Total	99.69	99.69	99.70	99.64	99.55	99.58
						Cation
SiO_2	95.95	61.10	73.51	64.97	82.70	69.41
$AlO_{3/2}$	2.66	22.79	14.43	18.50	8.30	14.38
$FeO_{3/2}$	0.13	0.92	0.45	0.97	0.37	0.61
FeO	0.00	4.79	2.43	3.98	0.54	1.10
MnO	0.00	0.08	0.02	0.04	0.03	0.02
MgO	0.13	3.04	2.57	4.06	2.08	3.06
CaO	0.00	0.56	0.60	0.38	1.66	1.50
$NaO_{1/2}$	0.12	1.42	1.54	1.70	2.28	6.11
$KO_{1/2}$	0.84	4.71	3.97	4.85	1.80	3.40
TiO_2	0.13	0.41	0.36	0.41	0.19	0.35
$PO_{5/2}$	0.01	0.16	0.11	0.13	0.04	0.05
CO_2	0.03	0.02	0.01	0.01	0.01	0.01
H_2O	(1.14)	(7.01)	(4.67)	(7.75)	(2.55)	(2.28)
Total	100.00	100.00	100.00	100.00	100.00	100.00
O	195.90	163.55	174.06	164.29	182.70	170.30
OH	2.28	14.02	9.34	15.50	5.10	4.56
Cl [1]	—	—	—	—	—	—
Anions	198.18	177.57	183.40	179.79	187.80	174.86

Analysts: Paula Montalto, U. S. Geological Survey, Nos. 1558, 1579, 1578, 1580, 1541.
Dorothy Powers, U. S. Geological Survey, Nos. 1615, 1613, 1908, 1906, 1527, 1915, 1260, 1262.

[1] Amount needed to form scapolite.

molecular norms and modes of the equivalents of the Belt amphibolite and amphibolite facies.

1906	1580	1527	1915	1541	1260	1262
Biotite gneiss Merry Creek, 3 km NE. of Clarkia	Biotite-scapolite gneiss 1 ½ km NE. of Bathtub Mountain	Hornblende gneiss St. Maries River, 7 km E. of Clarkia	Hornblende-scapolite gneiss Merry Creek, 8 km NE. of Clarkia	Diopside-plagioclase gneiss 8 km ENE. of Bathtub Mtn.	Diopside-biotite gneiss Potlatch Creek	Diopside-scapolite gneiss Potlatch Creek
percent						
77.84	62.56	73.07	60.74	71.35	70.28	65.51
9.36	14.36	11.22	12.37	9.64	9.65	10.18
0.67	0.66	0.64	1.36	0.52	0.45	0.48
1.56	5.35	2.17	2.70	2.18	1.82	2.25
0.03	0.06	0.04	0.09	0.10	0.05	0.07
1.68	6.20	2.63	6.06	2.79	2.98	4.12
2.44	1.98	5.83	8.72	7.40	7.52	11.83
2.85	1.59	2.65	2.43	2.37	2.08	1.72
1.43	4.25	0.21	2.46	1.98	2.57	1.57
0.37	0.53	0.43	0.46	0.36	0.50	0.45
0.06	0.10	0.06	0.12	0.10	0.07	0.07
0.81	0.13	0.38	0.01	0.06	0.91	0.31
—	(0.13)	—	(1.14)	—	—	(0.56)
0.54	1.90	0.48	0.86	0.47	0.26	0.36
0.08	0.11	0.04	0.10	0.21	0.10	0.05
99.72	99.78	99.85	98.48	99.53	99.24	98.97
percent						
74.41	59.56	69.54	57.65	68.18	66.90	62.52
10.55	16.11	12.58	13.84	10.85	10.82	11.45
0.48	0.47	0.46	0.97	0.38	0.32	0.34
1.25	4.26	1.73	2.15	1.74	1.45	1.79
0.02	0.05	0.03	0.07	0.08	0.04	0.06
2.39	8.80	3.73	8.57	3.97	4.23	5.86
2.50	2.02	5.94	8.87	7.58	7.67	12.10
5.28	2.94	4.89	4.47	4.39	3.85	3.18
1.75	5.16	0.25	2.98	2.41	3.12	1.92
0.26	0.38	0.31	0.33	0.26	0.36	0.32
0.05	0.08	0.05	0.09	0.08	0.06	0.06
1.06	0.17	0.49	0.01	0.08	1.18	0.40
(1.72)	(6.04)	(1.52)	(2.72)	(1.50)	(0.82)	(1.15)
100.00	100.00	100.00	100.00	100.00	100.00	100.00
176.08	158.43	172.84	159.08	169.36	167.80	165.52
3.44	12.08	3.04	5.44	3.00	1.64	2.30
—	(0.21)	—	(1.83)	—	—	(0.90)
179.52	170.51	175.88	164.52	172.36	169.44	167.82

Table 1.

	1558 Revett quartz-ite Sur-veyors Ridge	1615 Kyanite-almandite schist 5 km east of Clarkia	1613 Kyanite-almandite schist 6 km east of Clarkia	1579 Kyanite-staurolite schits Bathtub Mountain	1578 Biotite quartzite 7 km SE. of Bathtub Mtn.	1908 Biotite gneiss Merry Creek, 7 km NE. of Clarkia
						Molecular
Q	92.98	35.13	51.72	37.83	65.02	34.53
Or	4.20	23.55	19.85	24.25	9.00	17.00
Ab	0.60	7.10	7.70	8.50	11.40	30.55
An	—	1.35	2.05	0.75	7.90	7.05
C	1.70	16.12	8.10	11.65	1.06	2.05
Wo	—	—	—	—	—	—
En	0.18	6.08	5.14	8.12	4.16	6.12
Fs	—	8.00	3.74	6.26	0.40	0.94
Ap	0.02	0.43	0.29	0.35	0.11	0.13
Il	0.13	0.82	0.72	0.82	0.38	0.70
Mt	0.13	1.38	0.67	1.45	0.55	0.91
Cc	0.06	0.04	0.02	0.02	0.02	0.02
Total	100.00	100.00	100.00	100.00	100.00	100.00
						Calculated
Quartz	93.02	35.25	54.34	41.85	66.54	38.39
Plagioclase ..	0.60	An_{12} 7.30	An_{13} 8.15	An_{9} 8.25	An_{30} 15.55	An_{10} 33.45
Scapolite	—	—	—	—	—	—
Microcline ...	0.20	0.65	0.75	2.55	5.20	6.95
Muscovite ...	5.60	25.76	20.80	18.72	—	5.20
Biotite	—	12.46	10.96	17.98	8.27	14.82
Chlorite	0.35	—	—	—	—	—
Almandite ..	—	10.16	4.02	7.16	—	—
Staurolite ...	—	—	—	2.10	—	—
Kyanite	—	11.10	3.25	3.47	—	—
Diopside	—	—	—	—	—	—
Tremolite ...	—	—	—	—	—	—
Hornblende .	—	—	—	—	—	—
Epidote	—	—	—	—	—	
Zoisite	—	—	—	—	4.26	1.70
Clinozoisite ..	—	—	—	—		—
Apatite	—	0.43	0.29	0.35	0.11	0.13
Magnetite ...	—	0.81	0.57	1.23	0.55	0.81
Hematite ...	0.13	—	—	—	—	—
Ilmenite	—	0.74	0.68	0.56	—	—
Sphene	—	—	—	—	0.57	0.99
Rutile	0.13	—	—	0.10	—	—
Calcite	—	0.04	—	0.02	0.02	—
Magnesite ...	0.06	—	—	—	—	—
Xenotime ...	0.02	—	—	—	—	—
Tourmaline ..	—	—	—	—	—	—
	100.11	104.70	103.81	104.34	101.07	102.44

Continuation.

1906 Biotite gneiss Merry Creek, 3 km NE. of Clarkia	1580 Biotite-scapolite gneiss 1 ½ km NE. of Bathtub Mountain	1527 Hornblende gneiss St. Maries River, 7 km E. of Clarkia	1915 Hornblende-scapolite gneiss Merry Creek, 8 km NE, of Clarkia	1541 Diopside-plagioclase gneiss 8 km ENE. of Bathtub Mtn.	1260 Diopside-biotite gneiss Potlatch Creek	1262 Diopside-scapolite gneiss Potlatch Creek
norm						
47.44	19.32	40.08	13.42	33.05	32.48	25.22
8.75	25.80	1.25	14.90	12.05	15.60	9.60
26.40	14.70	24.45	22.35	21.95	19.25	15.90
6.80	8.60	18.60	15.96	10.13	9.61	15.87
0.80	4.57	—	—	—	—	—
—	—	3.30	11.04	10.68	8.94	16.86
4.78	17.60	7.46	17.14	7.94	8.46	11.72
1.54	7.40	2.44	2.82	2.74	1.94	2.72
0.13	0.21	0.13	0.24	0.21	0.16	0.16
0.52	0.76	0.62	0.66	0.52	0.72	0.64
0.72	0.70	0.69	1.45	0.57	0.48	0.51
2.12	0.34	0.98	0.02	0.16	2.36	0.80
100.00	100.00	100.00	100.00	100.00	100.00	100.00
molecular mode						
50.98	33.80	41.49	17.85	32.74	33.47	26.22
An_8 28.00	An_{28} 11.25	An_{30} 33.65	—	An_{31} 31.90	An_{32} 27.85	An_{58} 14.05
—	Me_{45} 6.46	—	Me_{46} 31.28	—	—	Me_{55} 18.90
2.60	—	—	14.90	11.80	12.95	9.40
0.64	3.34	—	—	—	—	—
12.56	48.33	2.73	—	0.58	5.77	0.40
—	—	1.52	—	—	—	—
—	—	—	—	—	—	—
—	—	—	—	—	—	—
—	—	—	—	—	—	—
—	—	—	—	19.31	16.96	30.59
—	—	—	—	2.18	—	—
—	—	15.63	37.23	—	—	—
} 2.84	—	—	1.71	—	—	—
	} 1.28	4.93	—	—	—	—
—		—	—	—	—	—
0.13	0.21	0.13	0.24	0.21	0.16	0.16
0.63	0.48	0.03	—	0.57	0.06	0.30
—	—	—	—	—	—	—
—	0.20	—	—	—	—	—
0.72	0.75	0.81	0.99	0.78	1.08	0.96
—	—	—	—	—	—	—
2.12	—	0.98	—	0.16	2.36	—
—	—	—	—	—	—	—
—	—	—	—	—	—	—
0.57	—	—	—	—	—	—
101.79	106.10	101.90	104.20	100.23	100.66	100.98

temperature of recrystallization in this zone was somewhat higher than in the extreme northwestern corner of the mapped area. Common mineral assemblages are as follows:

Pelitic assemblages:

1) Kyanite-staurolite-almandite-biotite-muscovite-plagioclase (An_{9-15})-quartz (Table 1, Nos. 1579, 1613, 1615).
2) Almandite-biotite-muscovite-(plagioclase)-quartz.

Quartzo-feldspathic assemblages:

3) Biotite-microcline-plagioclase (An_{15-20})-quartz, with or without muscovite and epidote minerals (Table 1, Nos. 1558, 1578).
4) Biotite-plagioclase (An_{30-32})-zoisite-scapolite-quartz.
5) Biotite-muscovite-plagioclase (An_{28-32})-zoisite-quartz, with or without scapolite (Table 1, No. 1580).

Calcareous assemblages:

6) Diopside-hornblende-biotite-plagioclase (An_{12-32})-quartz.
7) Hornblende-biotite-(microcline)-plagioclase (An_{30})-zoisite-quartz (Table 1, No. 1527).
8) Diopside-tremolite-biotite-plagioclase (An_{15-32})-quartz, with or without microcline (Table 1, No. 1541).
9) Diopside-biotite-microcline-plagioclase (An_{15-30})-quartz, with or without scapolite and epidote.
10) Diopside-calcite-microcline-plagioclase (An_{30-32})-quartz.

The mineral assemblages listed above indicate that the rocks near Clarkia and near Bathtub Mountain were metamorphosed to a subfacies that would correspond to the staurolite subfacies of Turner (in Fyfe, Turner, and Verhoogen 1958, p. 229), except that the rocks nearly all contain kyanite as well as staurolite. This subfacies is therefore called the staurolite-kyanite subfacies; it is regarded here as distinct from the andalusite-staurolite subfacies discussed in the following section. It is not known whether staurolite without kyanite occurs north of the kyanite-staurolite schists. Reconnaissance showed diopside gneiss and quartzite just north of the kyanite-staurolite schists, and farther north almandite-biotite schist.

The most calcic plagioclase in the epidote-bearing gneisses interbedded with the kyanite-staurolite schists is An_{32}. This An content is too low for rocks of the amphibolite facies as defined by Barth (1952), and it supports Francis' (1956, p. 356) conclusion that staurolite crystallizes in the temperature-pressure field of the epidote-amphibolite facies. Turner (in Fyfe, Turner, and Verhoogen, 1958, p. 229) includes the staurolite subfacies in the almandite-amphibolite facies, and thus makes it a wider range than Francis' amphibolite facies. As epidote minerals are fairly common in the gneissic layers interbedded with the staurolite-(kyanite)-bearing schists but rare in the higher-temperature zones, it seems useful, in describing the rocks of this part of Idaho, to distinguish between the amphibolite facies proper, and an epidote-amphibolite facies, which would include the staurolite-kyanite subfacies.

ANDALUSITE-STAUROLITE SUBFACIES

Andalusite instead of kyanite crystallized in the eastern part of the area south of The Nub, and there the mineral assemblages in the pelitic layers are as follows:

1) Andalusite-staurolite-(sillimanite)-biotite-muscovite-plagioclase (An_{10-15})-quartz.
2) Andalusite-almandite-biotite-muscovite-(plagioclase)-quartz.
3) Staurolite-almandite-biotite-muscovite-(plagioclase, An_{10-15})-quartz.

The mineral assemblages in the quartzo-feldspathic and calcareous layers are similar to those in the staurolite-kyanite subfacies.

The presence of andalusite instead of kyanite indicates a lower pressure and probably a somewhat higher temperature. These stability relations between the critical minerals can be illustrated by a PT-diagram similar to those used by Thompson (1955) and Francis (1956), but certain marked differences from their diagrams were found. Staurolite was found in some samples of kyanite-sillimanite schist, in andalusite-sillimanite schist, and also where all three aluminium silicates occur together (Hietanen, 1956); the stability field of staurolite may therefore include the triple point of the three aluminium silicates (Fig. 3). The common occurrence of andalusite in the staurolite schists may indicate that in the andalusite field the stability curve of staurolite is closer to that of the andalusite $\rightarrow$ sillimanite inversion curve than shown in Figure 3. The mineral associations in the Boehls Butte quadrangle (Hietanen, 1956) raise the question whether cordierite is stable at the temperature and pressure of the triple point. Since cordierite was not found in either the kyanite rocks or in the andalusite-kyanite rocks, the stability field of cordierite does not overlap that of kyanite. Thus the stability curve for cordierite may pass through the triple point, or deviate from it only a little toward a lower pressure, and may divide the andalusite field in such a manner that the stability field of cordierite overlaps that of staurolite. In that part of the andalusite field where both staurolite and cordierite would be stable, the iron-magnesium ratio of the sediment would determine which of these two minerals crystallized. This is shown, for instance, by the occurrence of cordierite in some layers that are interbedded with staurolite-bearing layers in the north-central part of the mapped area. The occurrence of sillimanite in the kyanite-garnet gedritite near Orofino (Hietanen, 1959)

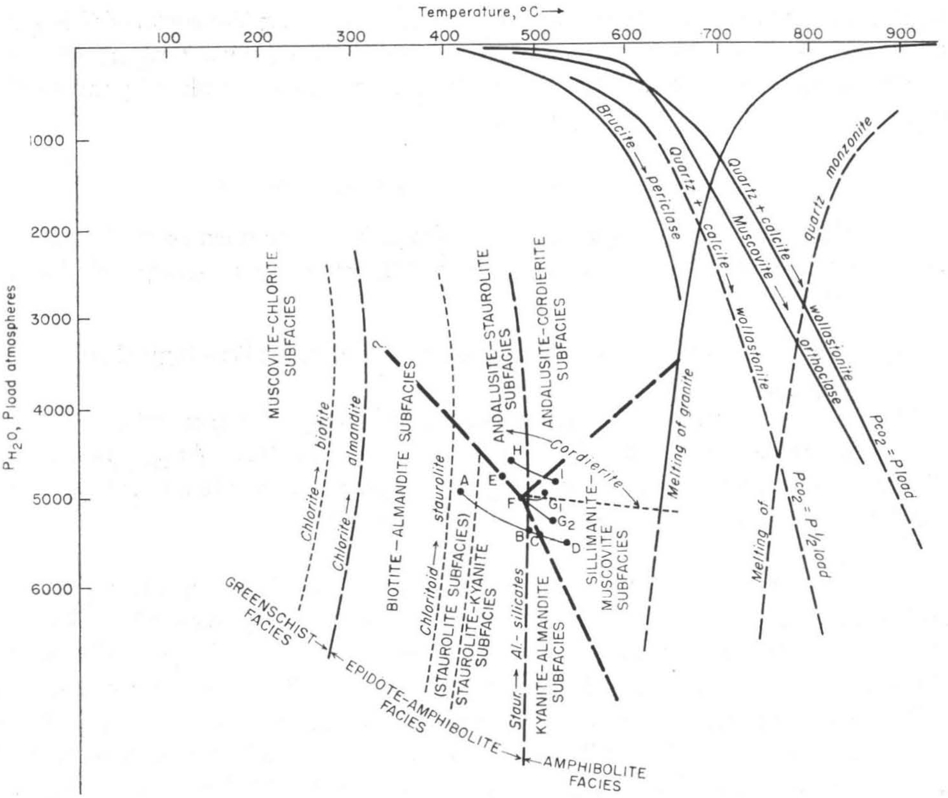

Fig. 3. PT-diagram showing fields of stability of metamorphic facies. Hypothetical stability curves for the index minerals are superimposed on the inversion curves of kyanite, andalusite, and sillimanite and divide the PT field into subfields each of which represents the stability field of a certain subfacies. For comparison are plotted the minimum melting curve for granite and the stability curve for muscovite after Yoder and Eugster (1955), the wollastonite curve after Harker and Tuttle (1956), and the brucite-periclase curve after McDonald (1955). The minimum melting curve for quartz monzonite is hypothetical. The inversion curves for the three Al-silicates are similar to those suggested by Miyashiro (1949).

would place the PT-conditions of this rock on the kyanite → sillimanite inversion curve, and thus into a higher pressure than that in which cordierite is stable.

AMPHIBOLITE FACIES

The transition from the epidote-amphibolite to the amphibolite facies is placed along the zone where staurolite disappears and minute needles of sillimanite appear between the muscovite flakes, within the biotite flakes, and in some places around the kyanite and andalusite crystals.

In the western part of the mapped area, kyanite and almanite crystallized instead of staurolite, according to the following reaction:

$$\underset{\text{6 staurolite}}{6HFe_2Al_9Si_4O_{24}} + \underset{\text{11 quartz}}{11SiO_2} \rightarrow \underset{\text{23 kyanite}}{23Al_2SiO_5} + \underset{\text{4 almandite}}{4Fe_3Al_2Si_3O_{12}} + \underset{\text{3 water}}{3H_2O}$$

Pseudomorphs consisting of small crystals of kyanite, staurolite, garnet, and muscovite were found east of Papoose Mountain. These pseudomorphs are 4 to 5 cm long, and their shapes suggest that they replaced staurolite crystals. The recrystallization took place on the borderline between the staurolite-kyanite and kyanite-almandite subfacies and involved addition of potassium. In the east-central part of the area, east and west of Eagle Point, andalusite and cordierite are common, and in the north-central part all three polymorphs of Al_2SiO_5 crystallized with staurolite or cordierite.

In the gneissic layers anorthite and microcline crystallized instead of zoisite and muscovite as follows:

$$\underset{\text{4 zoisite}}{4Ca_2Al_3Si_3(OH)O_{12}} + \underset{\text{muscovite}}{K_2Al_6Si_6(OH)_4O_{20}} + \underset{\text{4 quartz}}{4SiO_2} \rightarrow \underset{\text{8 anorthite}}{8CaAl_2Si_2O_8} + \underset{\text{2 microcline}}{2KAlSi_3O_8} + \underset{\text{4 water}}{4H_2O}$$

All reactions along the facies boundaries involve water, and the stability curves for the index minerals are therefore assumed to be of the type calculated for the equilibrium of hydrates (see Fyfe, Turner, and Verhoogen, 1958, p. 119). Some of the reactions, such as biotite $\rightarrow$ sillimanite, involve metasomatic removal of certain elements (*e.g.* Mg, K), which may form minerals rich in these elements in the adjoining layers. Some of the cordierite and segregations of biotite were probably formed in this manner. A slight change of pressure in the muscovite-sillimanite field near the triple point of the three aluminium silicates would determine whether or not cordierite would crystallize with sillimanite (see Fig. 3, G_1 and G_2).

The rocks in which staurolite is found with andalusite, kyanite, and sillimanite occur around the anorthosite bodies in the north-central part of the area. These rocks probably crystallized at or near the temperature and pressure of the triple point (Hietanen, 1956). Staurolite with andalusite and sillimanite was found in the schist on The Nub, and staurolite with kyanite and sillimanite on Bathtub Mountain. No signs of disequilibrium were found in any of the thin sections studied. Thus it seems that in this area the stability field of staurolite may overlap that of sillimanite (Fig. 3), making it difficult to define the lower limit of the amphibolite facies. Sillimanite with kyanite is common in the rocks near Hemlock Butte. This association occurs south of the staurolite-bearing rocks, and thus in a higher temperature zone closer to the batholith. The temperature and pressure of recrystallization in this zone must have been higher than those at the triple point (Fig. 3, point C). Andalusite may have crystallized at the same temperature but under lower pressure. The mineral association andalusite-cordierite is common in rocks formed at high temperatures and low pressures; as shown schematically in Figure 3, it would occupy the higher-temperature part of the andalusite field. The occurrence of muscovite with sillimanite in all high-grade rocks of the mapped area indicates, according to the experimental work by Yoder and Eugster (1955), that the temperature did not exceed 650°C. The amphibolite facies below this temperature can be subdivided into three subfacies, all of which may have been formed in the same temperature range but at different pressures. These subfacies are, in order of increasing pressure: andalusite-cordierite, sillimanite-muscovite, and kyanite-almandite. Most of the rocks of the sillimanite-muscovite subfacies, however, crystallized in a higher temperature because they are next to the batholith.

KYANITE-ALMANDITE SUBFACIES

The schist of this subfacies is a little coarser grained than that of the staurolite-kyanite subfacies. The mica flakes range from ½ to 2 mm in length, and the garnet crystals from 3 to 10 mm in diameter. The kyanite crystals are from ½ to 2 cm long and in many places are surrounded by sillimanite. The common mineral assemblages are as follows:
Pelitic assemblages:

1) Kyanite-sillimanite-biotite-muscovite-(plagioclase)-quartz.
2) Kyanite-almandite-biotite-muscovite-(plagioclase)-quartz.
3) Almandite-biotite-muscovite-plagioclase-quartz.

Most of the calcareous layers contain both diopside and biotite. Megascopically these layers are similar to those of similar composition in the epidote-amphibolite facies. The common mineral assemblages are:

4) Diopside-(biotite)-microcline-plagioclase (An_{30-58})-scapolite (Me_{55})-quartz (Table 1, No. 1262).

5) Diopside-(biotite-microcline)-plagioclase (An_{30-50})-clinozoisite-quartz (Table 1, No. 1260).

Plagioclase near An_{50} is associated with clinozoisite in a sample collected 10 km west of Hemlock Butte, which according to Ramberg's diagram (1949) would indicate a temperature close to 430° C.

ANDALUSITE-CORDIERITE SUBFACIES

The stability relations between the three aluminium silicates indicate that andalusite, instead of kyanite, crystallizes during thermal metamorphism under low pressures. Andalusite is common near the contacts of igneous bodies that were emplaced after the dynamic phase. Since cordierite occurs with all three aluminium sillicates (Hietanen, 1956), the stability curve for cordierite in this area probably lies close to or passes through the triple point (Fig. 3).

The rocks belonging to the andalusite-cordierite subfacies occur in the eastern part of the area, where folding is gentler than in the western part. Reconnaissance shows that andalusite with or without sillimanite is common near the North Fork of the Clearwater River, southeast of the Canyon Ranger Station. The pelitic assemblages are:

1) Andalusite-(sillimanite)-biotite-muscovite-plagioclase (An_{28})-quartz, with or without cordierite.

2) Almandite-biotite-muscovite-plagioclase-quartz.

Some andalusite-bearing layers contain irregularly shaped pseudomorphs of chlorite. Similar pseudomorphs of chlorite are common in the quartz monzonite near Beaver Butte, where they have replaced cordierite; this rock exhibits every stage of alteration of cordierite — first to pinite and then to aggregates consisting of chlorite or of chlorite and muscovite. It therefore seems very likely that the chlorite aggregates in the andalusite schist also were formerly cordierite, and that these rocks were crystallized to the andalusite-cordierite subfacies. Sillimanite occurs in some layers, which indicates proximity to the inversion curve andalusite → sillimanite.

The calcareous layers interbedded with the schist consist mainly of quartz, plagioclase, diopside, and biotite. In this respect and in megascopic appearance they are similar to the calcareous layers in the kyanite-almandite subfacies, but in thin sections it is seen that they are generally more granular and fine grained. They are granofelses (Goldsmith, 1959) rather than gneisses.

SILLIMANITE-MUSCOVITE SUBFACIES

The wide areal distribution of the sillimanite-bearing rocks and the increase of grain size within them toward the batholith indicate that there was a wider range of temperature in the sillimanite-muscovite subfacies than in the kyanite-almandite or andalusite-cordierite subfacies. The stability relations shown in Figure 3 illustrate the situation. The assumed course of crystallization in various parts of the area is illustrated by curves A to D, E to G, and H to I. Since all the curves are fairly close to the triple point of the three aluminum silicates, they pass to the sillimanite field at a temperature of around 500° C. In rocks that recrystallized at higher or lower pressures than these curves indicate, the transitions kyanite → sillimanite and andalusite → sillimanite would take place at higher temperatures, perhaps close to 600° C.

The mineral assemblages in the sillimanite-muscovite subfacies are as follows:

Pelitic assemblages:

1) Sillimanite-almandite-biotite-muscovite-plagioclase (An_{20-35})-quartz.
2) Almandite-biotite-(muscovite)-plagioclase (An_{20-35})-quartz.

Quartzo-feldspathic assemblages:

3) Biotite-(muscovite-microcline)-plagioclase (An_{25-45})-quartz.

Calcareous assemblages:

4) Diopside-actinolite-biotite-(microcline)-plagioclase (An_{12-85})-quartz.
5) Diopside-(biotite)-plagioclase (An_{30-35})-quartz.
6) Diopside-grossularite-plagioclase (An_{85})-quartz.
7) Diopside-calcite-grossularite-quartz.

All gneissic layers in the sillimanite-muscovite subfacies are coarser grained than the corresponding layers interbedded with kyanite- and andalusite-bearing schists.

No wollastonite was found in any of the thin sections examined.

THE OCCURRENCE OF KYANITE OR OF ANDALUSITE AND THE INTENSITY OF DEFORMATION

The sequence of mineral assemblages in the pelitic layers in the western, central, and eastern parts of the area shows certain trends that are illustrated with curves in the PT-diagram (Fig. 3). In the western part the trend of crystallization followed approximately the line A-D, in the central part E-G, and in the east-central part H-I. The temperature range in all parts of the area may have been much the same, but the pressures may have differed to some extents. As the structures and textures show that in general the rocks in the eastern part are less intensely folded and deformed than those in the western part, these pressures may have been directed instead of hydrostatic ones.

CHEMICAL COMPOSITION AND EQUILIBRIUM DIAGRAMS

Thirteen new chemical analyses of various layers of the Wallace formation in the PT-field of the epidote-amphibolite facies are shown in Table 1. The cation percentages and molecular norms were calculated according to the method proposed by Barth (1952) and Eskola (1954). The molecular modes were calculated from the molecular norms as follows: for biotite and hornblende the formulas worked out on the basis of optical properties, Winchell's tables, and available analyses were used. The amount of each

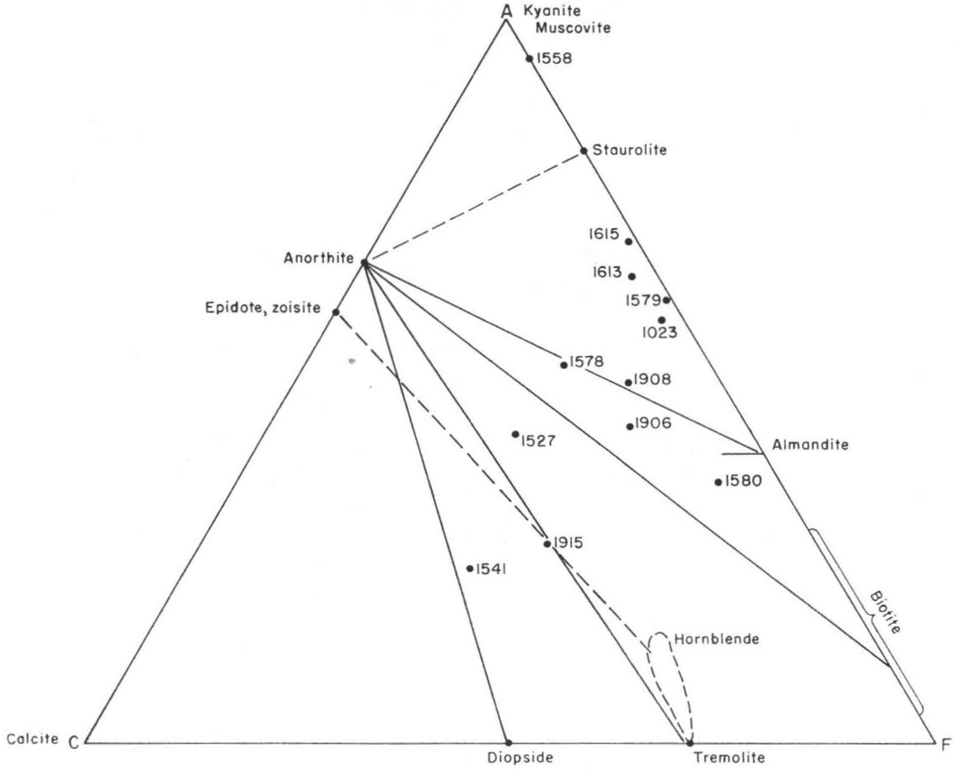

Fig. 4. ACF-diagram (plotted in terms of molecular percent) for the rocks metamorphosed to the epidote-amphibolite facies. The plotted points refer to the following analyses and mineral assemblages (Table 1):

1558. (Biotite)-chlorite-muscovite-albite-quartz.
1615. Kyanite-almandite-biotite-muscovite-plagioclase (An_{12})-quartz.
1613. Kyanite-almandite-biotite-muscovite-plagioclase (An_{13})-quartz.
1023. Almandite-biotite-muscovite-quartz (Hietanen, in press, B).
1579. Kyanite-almandite-staurolite-biotite-muscovite-albite (An_9)-quartz.
1578. Biotite-zoisite-clinozoisite-microcline-plagioclase (An_{30})-quartz.
1908. Biotite-muscovite-zoisite-microcline-albite (An_{10})-quartz.
1906. Biotite-muscovite-zoisite-microcline-albite (An_8)-quartz.
1580. Biotite-muscovite-zoisite-plagioclase (An_{28})-quartz.
1527. Hornblende-biotite-chlorite-microcline-zoisite-plagioclase (An_{30})-quartz.
1915. Hornblende-microcline-epidote-scapolite (Me_{46})-quartz.
1541. Diopside-(biotite)-tremolite-microcline-plagioclase (An_{31})-quartz.

accessory — sphene, ilmenite, and magnetite, was estimated in the thin sections and checked against the analyses. Where only one iron-magnesium mineral was present, all the magnesium, and the iron that remained after subtracting the part contained in ilmenite and magnetite, was allotted to this mineral. All sodium was assigned to albite, and the amount of anorthite was calculated from this and from the anorthite content of plagioclase as deter-

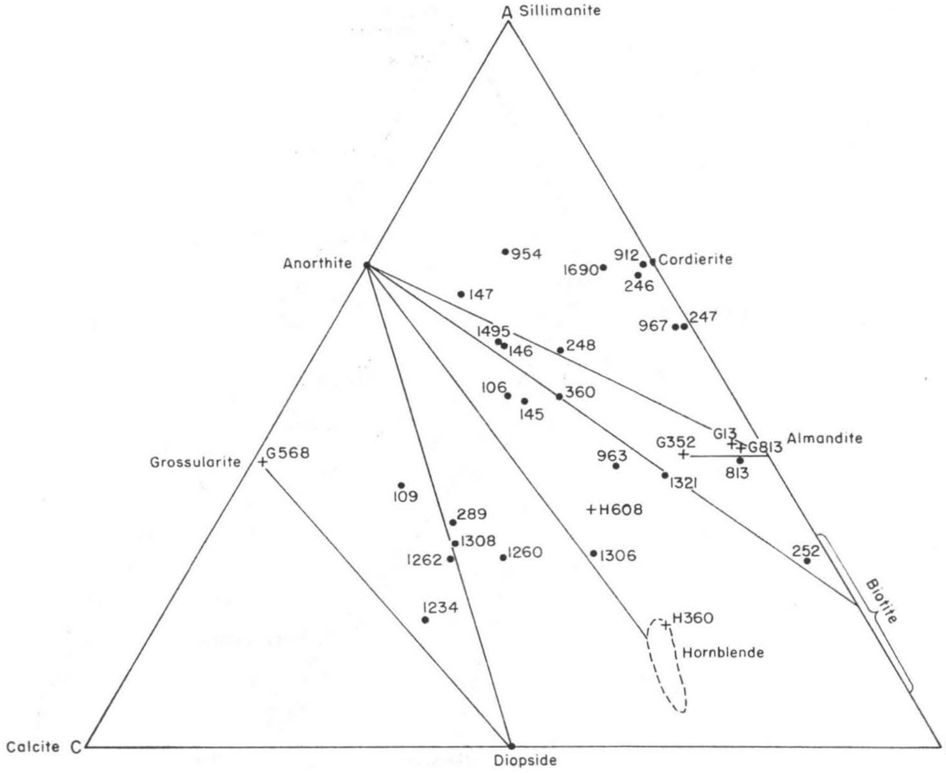

Fig. 5. ACF-diagram (plotted in terms of molecular percent) for the rocks of the amphibolite facies. The analyses points and mineral assemblages are as follows:

247. Sillimanite-almandite-biotite-muscovite-quartz.
246. Sillimanite-almandite-biotite-muscovite-plagioclase (An_{20})-quartz.
1690. Sillimanite-almandite-biotite-muscovite-orthoclase-plagioclase (An_{46})-quartz.
912. Kyanite-andalusite-sillimanite-cordierite-biotite-plagioclase (An_{17})-quartz (Hietanen, 1956).
967. Kyanite-andalusite-sillimanite-cordierite-biotite-plagioclase (An_{11})-quartz.
954. Kyanite-andalusite-sillimanite-biotite-(orthoclase)-plagioclase (An_{38})-quartz.
813. Garnet-kyanite-sillimanite-gedrite-biotite-plagioclase (An_{20})-quartz (Hietanen, 1959).
147. Biotite-muscovite-microcline-plagioclase (An_{11})-quartz.
248. Biotite-muscovite-microcline-plagioclase (An_{35})-quartz.
1495. Biotite-orthoclase-plagioclase (An_{44})-quartz.
146. Biotite-epidote-microcline-plagioclase (An_{24})-quartz.
252. Biotite-muscovite-microcline-plagioclase (An_{3})-quartz.
106. Hornblende-microcline-plagioclase (An_{27})-quartz.
145. Hornblende-biotite-epidote-microcline-plagioclase (An_{22})-quartz.
360. Garnet-hornblende-biotite-plagioclase (An_{40})-quartz.
1321. Garnet-hornblende-epidote-plagioclase (An_{36})-quartz.
963. Garnet-hornblende-plagioclase (An_{54})-quartz.
1306. Tremolitic hornblende-plagioclase (An_{79})-quartz.
109. Diopside-epidote-plagioclase (An_{38})-quartz.
289. Diopside-plagioclase (An_{85})-quartz.
1308. Diopside-plagioclase (An_{86})-quartz.
1234. Diopside-calcite-orthoclase-plagioclase (An_{72})-(scapolite Me_{60})-quartz.
1260. Diopside-biotite-epidote-microcline-plagioclase (An_{32})-quartz (Table 1).
1262. Diopside-(biotite)-microcline-plagioclase (An_{58})-scapolite (Me_{55})-quartz (Table 1).

Nos. 247, 246, 147, 248, 146, 252, 106, 145, 360, 109, 289 are from Hietanen (in press, A); Nos. 1690, 967, 954, 1495, 1321, 963, 1306, 1308, 1234 are from Hietanen (in press, B). Crosses refer to analyzed minerals.

mined in the thin section. Potassium and the aluminium not contained in plagioclase were divided between micas and microcline in the way proposed by Barth (1959, p. 139) for calculating the standard minerals of mesonorms. This method was also used to divide calcium between scapolite, plagioclase, and diopside, and for dividing iron and magnesium between hornblende, diopside, and biotite in the gneissic layers and between garnet and biotite in the schist.

The analyses of rocks and minerals used in the calculation of molecular modes have been plotted on an ACF diagram (Fig. 4). Lines connecting these minerals divide the ACF triangle into sub-triangles. The minerals at the corners of each sub-triangle crystallize in stable equilibrium in rocks that plot within the sub-triangle. For instance, the rocks that plot in the kyanite-almandite-anorthite triangle contain kyanite and staurolite or a large amount of muscovite. Such rocks have the mineral assemblages of the pelitic schists. The analyses of the quartzo-feldspathic layers plot within the anorthite-almandite-biotite triangle, those of the hornblende gneisses within the biotite-hornblende-anorthite (zoisite) triangle, and those of the diopside-bearing gneisses within the tremolite-diopside-anorthite triangle. The mineral assemblages of the analyzed samples, as listed under the ACF diagram and in Table 1, and the position of the analyses points in the ACF diagram, indicate that stable equilibrium was probably attained during the recrystallization.

For comparison the analyses of rocks of the amphibolite facies, described earlier (Hietanen, in press), have also been plotted on an ACF diagram (Fig. 5). Sillimanite instead of kyanite appears in the A corner, and cordierite crystallizes with it. The mineral assemblages listed under Figure 5, and the position of the corresponding points, indicate that equilibrium also was generally reached in the zone next to the batholith.

TEMPERATURE AND PRESSURE DURING RECRYSTALLIZATION

All the stability curves shown in Figure 3 are hypothetical, but they show the relations as well as they can be deduced from our present theoretical knowledge and the field evidence. Experimental work as well as geologic thermometers should be used to establish the true position and shape of the curves. Because the stabilities of many minerals are influenced by the presence or absence of water or carbon dioxide, and even by some trace elements, the shape and position of the curves may change considerably from place to place. This is well demonstrated, for instance, by the shift of the curves for staurolite and cordierite (compare Francis, 1956). The temperature of the triple point for andalusite, kyanite, and sillimanite can be estimated to lie at various points between 400° and 600° C, depending on what indirect laboratory and field evidence is used. Crystallization proceeds, in most areas, either from the andalusite field to the sillimanite field or from the kyanite field to the sillimanite field. The inversion curves of Figure 3, if approximately correct, show that in such areas the border of the sillimanite fields is reached at a higher temperature than that of the triple point. Thus the temperature at the triple point should be the minimum for all recorded temperatures in which sillimanite is stable.

Crystallization of orthoclase $Or_{78}Ab_{22}$ in some dikes near Boehls, where all three aluminium silicates occur in the country rock, indicates, according to Barth (1956), a temperature of about 420° C, and the occurrence of An_{55} with epidote in the anorthosite bodies gives, according to Ramberg's curve (1949), a temperature of 430° C. The recent work by Wyllie and Tuttle (1960) shows that volatile materials and certain minor elements may considerably lower the melting temperature of granite. In the light of their work it seems at least reasonable to assume that the sillimanite-muscovite subfacies crystallized at or below 570° C, or not much higher, for the rocks of this subfacies had not generally begun to melt. This places the amphibolite facies in the temperature range 500°—600° C, an estimate which is compatible with that made by Engel and Engel (1958) for similar rocks of the Adirondack region. The rocks that contain all three aluminium silicates are farther from the batholith, and hence in a zone of lower temperature. Since cordierite and staurolite are common constituents of aluminium-silicate-bearing rocks, the triple point should lie within or close to the stability fields of these minerals. For illustrative purposes the triple point was assumed to be near 500° C.

The pressure of recrystallization is more difficult to estimate than the temperature. The experimental work on stability of muscovite by Yoder and Eugster (1955) and on the wollastonite reaction by Harker and Tuttle (1956) throws some light on this problem. If it be assumed that the water pressure in the field of middle grades of metamorphism is equal to the load pressure, the stability curve for muscovite and the minimum melting curve for granite are about as shown in Figure 3. The minimum melting temperature would be higher for quartz monzonite and for quartz diorite, which are the two major rock types of the Idaho batholith, than it is for granite. The stability curve for wollastonite is plotted for two pressures: $P_{CO_2} = P_{load}$ and $P_{CO_2} = \frac{1}{2} P_{load}$. As muscovite is stable with sillimanite up to the contact of the batholith and no wollastonite was formed, the melting curve for the igneous rocks must intersect the muscovite and wollastonite curves at lower pressures than prevailed near the contact of the batholith. If it be assumed that the quartz monzonite started to melt between 750° and 800°C, and that the carbon dioxide pressure was about half the load pressure, the wollastonite curve would intersect the melting curve of the igneous rocks at pressures around 4 500 atm., and this suggests a still higher pressure for the triple point of andalusite-kyanite-sillimanite. It was assumed, for illustrative purposes, that the pressure at this point was about 5 000 atm.

CHANGES IN STRUCTURE OF THE BELT SERIES NEAR THE BATHOLITH

In the area considered here, two or three sets of folds can be observed at many single outcrops. In some places three or four sets of fold axes can be identified. Two of these, trending northwestward and northeastward, are parallel to the trends of the Jurassic folding, but the folding around the eastward-plunging axes began much earlier, perhaps in Precambrian time, and the folds were rejuvenated in Jurassic time. Lineation is well developed along the flanks of the folds; it is always a wrinkling or micro folding around a second fold axis.

In the course of the study of metamorphism it was observed that the distribution of axial-plane cleavage in lithologically similar layers in all sets of folds is closely connected with the peripheral distribution of the metamorphic facies around the batholith. In general the intensity of folding and deformation increases toward the batholith. The style of folding for all sets of folds, as well as the type of deformation, is uniform within each metamorphic zone but changes from one metamorphic zone to another. These changes are shown mainly by the orientation of the micaceous minerals.

STYLE OF FOLDING IN VARIOUS METAMORPHIC ZONES

In the area about 40 to 60 km north of the batholith, where most of the rocks of the Belt series were metamorphosed to the greenschist facies or to the lower part of the epidote-amphibolite facies, the folds are large and open. Their amplitudes (300 to 500 m) are much smaller than their wave lengths (1 to 3 km). Gentle dips are common here, and a single formation is generally exposed over wide areas, but asymmetric folds are also common, and overturning occurs in places. Near many fault zones, small overturned folds are common and there is fracture cleavage parallel to the faults. The folds between the faults have straight limbs and round but fairly sharp hinges.

The rocks in the middle zone, 15 to 40 km north of the batholith, where the rocks were metamorphosed to the epidote-amphibolite facies and to the lower subfacies of the amphibolite facies, the folding was more intense (Fig. 6). The amplitudes of the folds are of the same order of magnitude as their wave

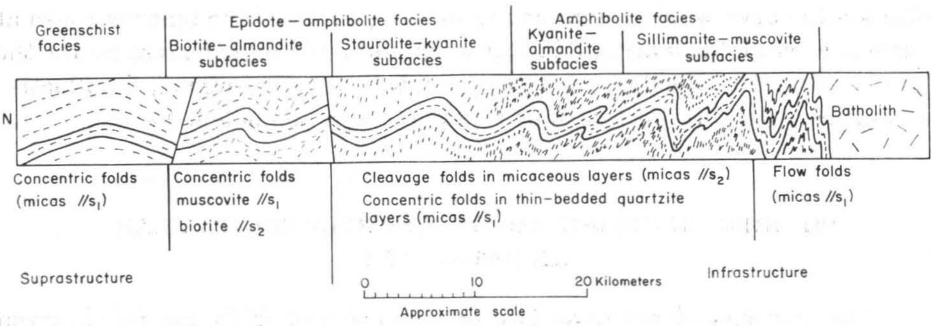

Fig. 6. Schematic section showing changes in style of folding and distribution of metamorphic facies northwest of the Idaho batholith.

lengths, and many small folds occur along the flanks and crests of the large folds. Small folds and wrinkles are commoner in the schist layers than in the interbedded quartzitic and gneissic layers, where the folds have rather round, smooth crests and troughs. Strong overturning occurs in places.

In the zone next to the batholith, and thus in the higher part of the sillimanite-muscovite subfacies, the flanks of the isoclinal folds are parallel and the amplitudes of the folds are much larger than their wave lengths. Small folds are common along the flanks and crests of the large ones. Where migmatitic veins abound, flow folds are typical. The bedding planes have been contorted into intricate folds with sharply turned hinges, and veins, stringers, and irregularly-shaped small masses of pegmatite are common. The veins and stringers are invariably along the bedding, and the larger masses favor the crests of the folds. Many pegmatitic veins are broken into boudins, which indicates that they are older than the last phase of deformation.

These various types of folds were formed around northeastward- as well as northwestward-trending axes, and thus around the trends of the Jurassic folding. In the southern and central parts of the area, folds around eastward-trending axes show similar characteristics.

RELATIONS OF FOLIATION TO BEDDING

In the greenschist facies, where tiny flakes of muscovite and chlorite were crystallized between more or less equant grains of quartz, feldspars, and calcite, the bedding is not clearly discernible; in many places, where shearing was strong and mica flakes lie parallel to the cleavage (s_2), the bedding is obscure. The bedding is distinct where layers of quartzite alternate with pelitic or calcareous layers. Much of the low-grade schist shows only one s plane, the cleavage, and in small outcrops it is impossible to tell whether the bedding is parallel to it or not. Outcrops with distinct bedding are, however, numerous enough to provide adequate material for structural study. Axial-plane cleavage is common in all micaceous beds that are as much as 2 or 3 cm thick. It is missing only in units that consist of thin competent layers separated by paper-thin mica-rich laminae; in these units, the slip occurred parallel to the bedding of the micaceous laminae.

A second cleavage, the fracture cleavage s_3, is common in many localities; it may be parallel to the axial plane of a second, later set of folds or to local faults. In places it was formed after the major period of folding, most likely during the faulting.

In the rocks of the epidote-amphibolite facies and of the lower subfacies of the amphibolite facies the bedding is much more distinct than in the rocks of the greenschist facies. In the pelitic layers, it is generally made conspicuous by distribution of such minerals as kyanite, andalusite, sillimanite, garnet, staurolite, and large flakes of muscovite and biotite, and in the calcareous layers by distribution of hornblende, tremolite-actinolite, and diopside. It is also marked in this zone by alternation of white quartzite layers with dark biotite-bearing layers, in which mica flakes, especially of biotite parallel to the bedding are clearly visible. In many layers the foliation is less distinct than the bedding. It can be detected, however, by a hand lens, because a part or all of the mica flakes are parallel to it. The orientation or the foliation in relation to the bedding varies.

In the thin-bedded rocks, where there is a large difference in the competency of adjoining layers, only one s plane, the bedding, is discernible and the folds were formed by flexural slip. In most of the rocks, however, especially in thick rather homogeneous and moderately micaceous layers, an axial-plane foliation was developed and the folds were formed by slip along these planes.

Studies of the orientation of micas in the Belt series show that where thick homogeneous layers are interbedded with thin layers either of more competent (quartz-rich) or less competent (micaceous) rocks, the mica flakes in the laminae that separate the beds of different competency are alined parallel to the bedding. The surfaces of these laminae resemble slickensides, and clearly acted as slip surfaces during the folding. In the thick homogeneous moderately micaceous beds, micas are subparallel to the axial planes.

Orientation of mica flakes parallel to s_1 and s_2 in the alternating layers that are rich either in micas or in quartz is common in biotite quartzite, biotite gneiss, and quartz-rich schist. In many outcrops competent folded layers, 2 to 10 cm thick, can be peeled off along the micaceous laminae that parallel s_1. In the material between these laminae, the mica flakes are oriented parallel to the axial planes. Thus both planes s_1 and s_2, acted as slip planes during the formation of folds.

In the schist units, where all beds contain 10 to 20 percent of micaceous minerals, all the mica flakes are parallel to the axial planes and the folds were formed by slip along s_2. Many beds contain thin highly micaceous laminae, but in these laminae the mica flakes lie parallel to s_2, because there is no appreciable difference in competency between these and the adjoining slightly less micaceous layers. Where thin-bedded heterogeneous layers with a considerable difference in competency between the individual beds are interbedded with more homogeneous schist, bedding foliation occurs in the former and axial-plane foliation in the latter. Slip was parallel to s_1 in the heterogeneous layers and parallel to s_2 in the more homogeneous schist layers. The distribution of axial-plane foliation in this zone is thus dependent on the character of the material.

In the zone next to the batholith, where tight isoclinal folds are common, the foliation is either parallel to the bedding or deviates from it by a small angle, which rarely exceeds 15°. Where only the flanks of large folds are exposed, the deviating foliation is probably parallel to the axial planes. But folds completely exposed in a single outcrop, say 2 to 10 m wide, reveal an unexpected relation. In the great majority of these the foliation is parallel to the bedding not only along the flanks of the folds but also along their crests; transecting foliation occurs only in some thick rather homogeneous beds of schist. In some other schist layers, however, all the mica flakes are parallel to a wrinkled bedding. These observations raise the following question: was axial-plane foliation ever developed in these outcrops, or were the micas reoriented during a later phase of deformation?

An answer to this question may be found by studying the lithologic character of the rocks concerned in detail. One of the most striking features is the occurrence of two types of pegmatitic veins in the zone next to the batholith. One set of veins consists of quartz and plagioclase with very little biotite; these veins occur always parallel to the bedding. The other set contains also potassium feldspar and muscovite, cuts the metasedimentary rocks and the first set of veins discordantly and are thus younger. It seems that during an early phase of deformation bedding was emphasized by introduction and sweating out of pegmatitic material. The layers next to the pegmatitic veins and stringers were enriched in micas and the strata became distinctly layered, resembling in their heterogeneity, the thin-bedded gneisses. As in the thin-bedded gneiss, the slip in the migmatites thus produced always occurred along the bedding. Axial-plane cleavage was never developed in these parts of the strata.

In many outcrops in the zone adjoining the batholith, the bedding planes are strongly folded and wrinkled and the mica flakes lie parallel to the bedding even though no pegmatitic material occurs along it. The mica flakes are large, many as much as 1 to 5 mm in diameter, and are crystallized in paper-thin laminae that separate layers, 1 to 2 mm thick, of quartz-plagioclase rock. In these rocks the distinct lamination made the bedding an excellent slip surface.

COMPARISON WITH OTHER AREAS

The style of folding in the zone next to the batholith is similar to that found in many areas of migmatized rocks. Here, however, the orientation of the pegmatitic veins and stringers is invariably parallel to the bedding, rather than to both the axial-plane foliation and bedding as is common, for example, in the Precambrian migmatites studied by the author in Finland (Hietanen, 1943, p. 95). It has been pointed out by many investigators (*e. g.* Wegman, 1953, and Kranck, 1957) that the migmatization in Finland was superimposed on rocks that had already been strongly deformed. The axial-plane cleavage was probably formed during this earlier deformation, and the migmatitic veins were emplaced or sweated out parallel to pre-existing *s*-planes.

The occurrence of the pegmatitic veins and stringers parallel to the bedding in the migmatized schist north of the Idaho batholith suggests that the bedding was a prominent *s*-plane, if not the only one, during the migmatization. The boudinaging of many veins shows that these veins were formed at any early stage of deformation. It is therefore concluded that axial-plane foliation was never developed in those migmatized rocks in which it is now missing. The bedding in these rocks was intensified by introduction or sweating out of pegmatitic material along the bedding, and the strata were folded by flexural slip in much the same way as the thin-bedded gneiss with mica-rich laminae. The rigidity of all layers was diminished during the preceding deformation, especially in those parts of the strata where abundant pegmatitic material was introduced, and the flow folds formed are similar to those generally found in highly migmatized rocks.

The migmatization which occurred at any early stage of the major deformation, in Jurassic time, is obviously older than the intrusive rocks of the Early Cretaceous batholith. A later potassium-rich group of pegmatitic veins and dikes, many of which cut the bedding, were introduced in connection with the intrusion of the quartz monzonite batholith.

CONCLUSIONS

In the area studied the grade of metamorphism increases toward the Idaho batholith, passing from a lower subfacies of the epidote-amphibolite facies to the sillimanite-muscovite subfacies of the amphibolite facies. Differences in the course of recrystallization during the metamorphism can be observed in the eastern, central, and western parts of the mapped area. In the easternmost part of the area, where the rocks of the Belt series are gently folded, the trend of recrystallization can be seen, when formations are followed southwestward, to pass from the andalusite-staurolite field through the andalusite-cordierite field to the sillimanite-muscovite field. The occurrence of andalusite instead of kyanite indicates lower pressures. In the north-central part of the area, kyanite, andalusite, and sillimanite are found together and are accompanied by either staurolite or cordierite. Thus the path of recrystallization passes here through the triple point of andalusite, kyanite, and sillimanite, and both staurolite and cordierite are found with all three aluminium silicates. In the western part of the area, where folding was more intense, the mineral assemblages indicate higher pressures during the recrystallization. The path of recrystallization crosses from the staurolite-kyanite field through the kyanite field to the sillimanite-muscovite field.

The inversion curves of the three aluminium silicates and the stability curves of other index mineral divide the PT-field into sub-fields each of which can be considered to represent a subfacies. The staurolite zone, for instance, is subdivided into staurolite-kyanite and andalusite-staurolite subfacies, both of which may cover approximately the same temperature range but differ considerably in the range of pressure. In the amphibolite facies also the conditions under which the three lower-temperature subfacies — the kyanite-almandite, the sillimanite-muscovite, and the andalusite-cordierite subfacies — were formed, are believed to have differed mainly in regard to pressure. The higher-temperature subfacies, in which orthoclase and sillimanite would crystallize instead of muscovite and sillimanite, was not reached in this area, nor was wollastonite found in any of the samples.

Intensity of deformation as well as of metamorphism increases toward the batholith, and the style of folding changes. At distances of 20 to 40 km from the batholith, axial-plane foliation is well developed in the moderately micaceous, fairly homogeneous layers of the greenschist and epidote-amphibolite facies, and of the lower part of the amphibolite facies. It is lacking in the thin-bedded layers where mica-rich laminae alternate with more competent layers that contain only a moderate amount of mica. Slip was parallel to the bedding along the highly micaceous laminae, while the intervening layers yielded by slip along the axial planes. The orientation of micaceous minerals parallel to slippage planes indicates that the rocks recrystallized while they were being deformed.

In the highly metamorphosed zone near the batholith, axial-plane foliation becomes rare. In this zone, pegmatitic veins and stringers and a coarse lamination have intensified the original compositional layering, and folds were formed by bending of the layers and by slip along them. As the rigidity of all layers were reduced near the batholith, these folds are typical flow folds, similar to those generally found in highly migmatized rocks.

The gradual changes in style of folding and deformation in the successive zones of metamorphism north of the Idaho batholith are similar to those found elsewhere in more extensive areas of regional metamorphism, and on a still larger scale at successively deeper levels in the earth's crust.

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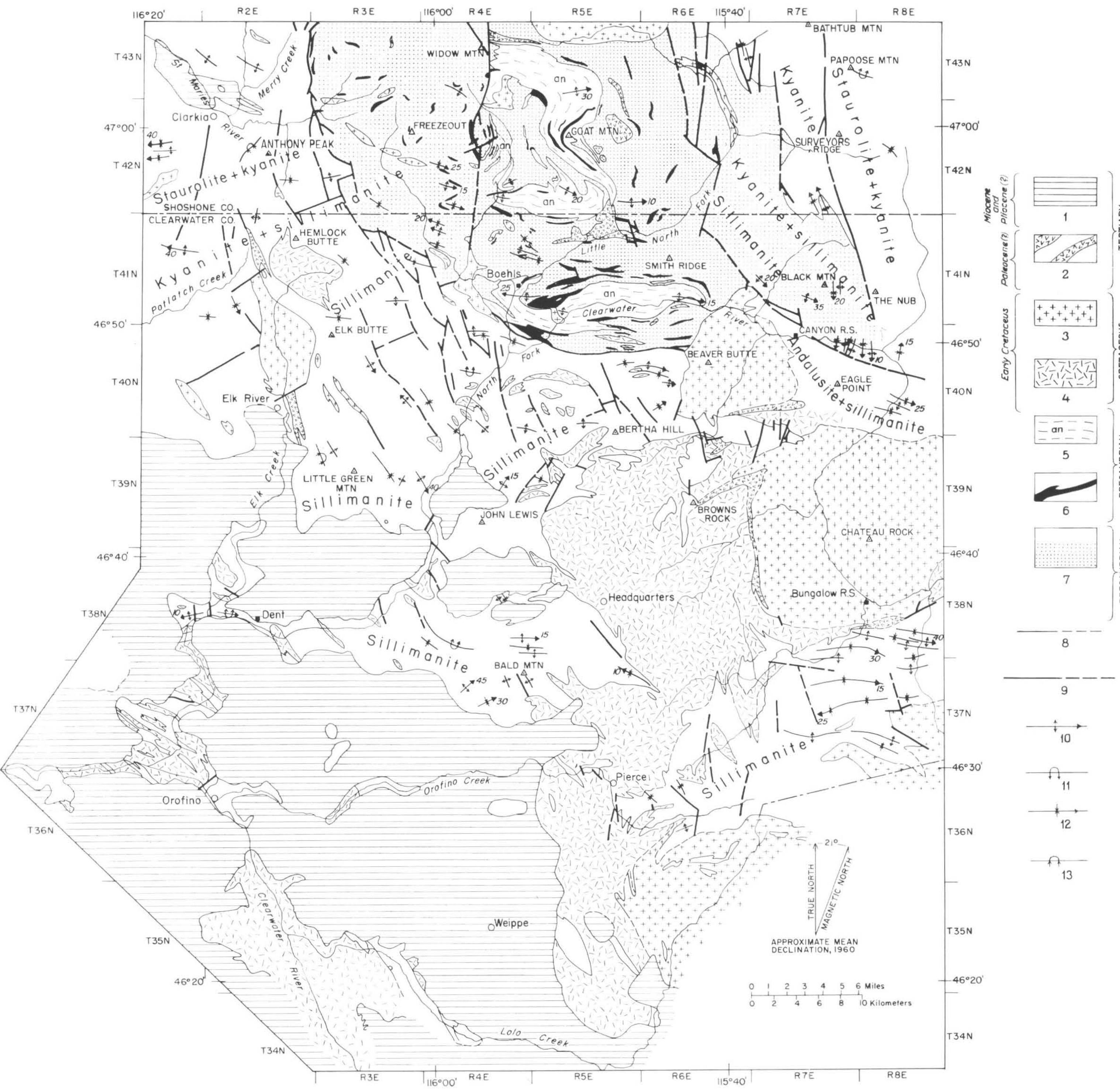

Fig. 2. Sketch map showing the distribution of major rock types and some of the index minerals of the metamorphic facies in western Clearwater County, Idaho. 1 = Columbia River basalt. 2 = Dikes and sills of pyroxene gabbro and granite porphyry. 3 = Granite and quartz monzonite. 4 = Tonalite and quartz diorite. 5 = Anorthosite. 6 = Amphibolite. 7 = Belt series; the largest continuous area of the Prichard formation is stippled; the lowest part is more densely stippled. 8 = Contact; dashed where approximately located. 9 = Fault; dashed where approximately located. 10 = Anticline; showing plunge of axis. 11 = Overturned anticline. 12 = Syncline; showing plunge of fold axis. 13 = Overturned syncline.

4

BEDDING SLIPS, WEDGES, AND FOLDING IN LAYERED SEQUENCES [1]

BY

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ABSTRACT

Relative movements parallel to bedding in layered sequences are difficult to recognize. Relative displacements and sense of motion can be determined where low angle shear planes cut competent beds into wedges which are then telescoped. Where observed these values were found to exceed slippage due to bending in flexure folds. It is therefore assumed that bedding slippage may precede folding because it is uniform over larger areas. A regional study of these phenomena is in progress in the Appalachian area.

CONTENTS

[1] Received February 21, 1961.

INTRODUCTION

Displacements on faults are easily determined if the fault cuts across key beds, dikes, contacts, or any other structures at fairly large angles. Field mapping also proves faults and stratigraphic correlations may indicate »throws.» If the dislocation is parallel to beds, however, it becomes difficult or impossible to detect displacements.

Layers commonly move over each other and such displacements are frequently mentioned but rarely measured. The movement is described in textbooks and is necessary when layered rock sequences are bent into folds. Most descriptions show anticlines in which the relative motion of higher beds is toward the crest on both limbs (Billings, 1954, p. 89).

Measurements of such displacements have been attempted where joints, quartz veins, or similar indicators are available (Hans Cloos, 1948).

In this paper I will try to show that relative movements parallel bedding may be of very much larger orders of magnitude than they should be if they could be attributed to the fold in whose limb they are found. If such movements are properly recognized they can be used in the analyses of folds in a similar manner as drag folds, cleavage, and bedding. They may also lead to some significant conclusions with respect to the folding mechanics of folded belts.

The material for this paper has been collected during many years as a byproduct of other field work.

FIELD OBSERVATIONS

In many well-bedded formations such as the Devonian Catskill, and Jennings, or the Silurian Wills Creek, fractures occur at angles of from 7 to 15 degrees to the bedding planes. They may be closely spaced or many feet apart. They only appear in relatively competent beds between incompetent ones, for instance, in sandstone between shales, harder shale between softer shales, or limestone beds between shales. The competent layer may be a few centimeters or 2 to 3 meters thick or more. Many such fractures have become thrusts with displacements of from 5 centimeters to more than 100 meters.

Figure 1 shows such a small fault in Cretaceous limestone, interbedded with shale at the Val de Fier, S.W. of Geneva, Switzerland. The displacement is about 24 cm. The effect on the limestone bed is a slight thickening at that portion of the section.

Figure 2 also shows a simple wedge in a sandstone bed (»B») in the Silurian of Maryland. The displacement is about 30 cm. The top and bottom

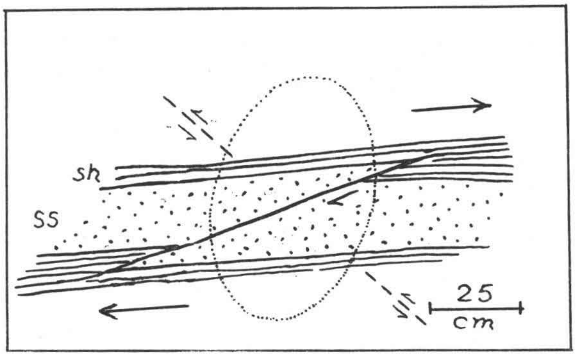

Fig. 1. Small low-angle (16°) reverse fault divides limestone bed into 2 wedges. Displacement 25 cm. Val de Fir SW Geneva (Switzerld.).

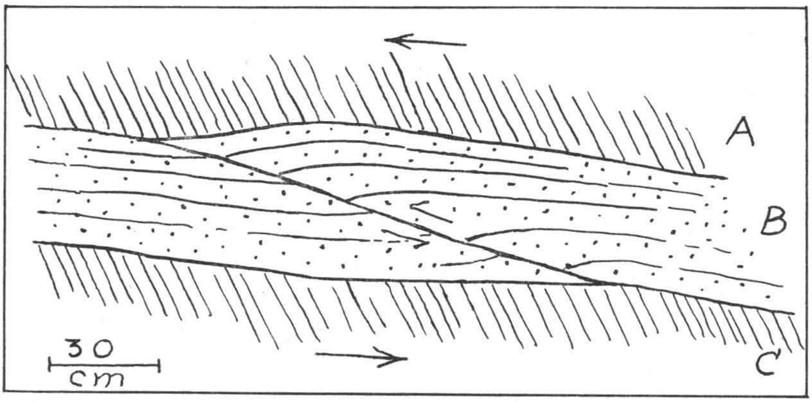

Fig. 2. Sandstone bed cut by low angle fracture with displacement of about 30 cm. Bloomsburg fm. (Silurian) Western, Md. Railroad cut opposite Great Cacapon, West Virginia.

wedges have been bent and the contacts are smooth. The wedging is visible only as displacement within the competent bed, a thickening of that bed and a fracture along which displacement took place. The fracture does not continue into shale beds A and C but continues along the contacts of bed B.

Figure 3 is in the left limb of an anticline. Beds A and C are red shales with strong fracture cleavage normal to bedding. A yellow-weathering coarsely bedded sandstone layer is 150 cm thick and faulted along a low-angle thrust fault with about 90 cm displacement. The left side moved into the bedding plane between B and C, the right side moved into the bedding plane between A and B. The distance x-y has been shortened and the thrust was gently folded. Also — the bed is thickened in the center.

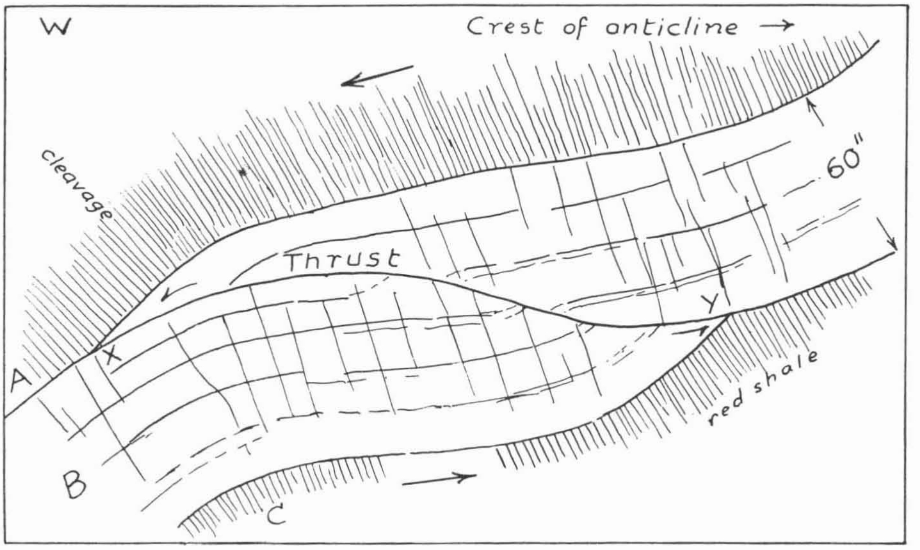

Fig. 3. Sandstone bed 150 cm thick with bent thrust surface and curved edges between red shales. Bloomsburg fm. (Silurian), same loc. as Figure 2.

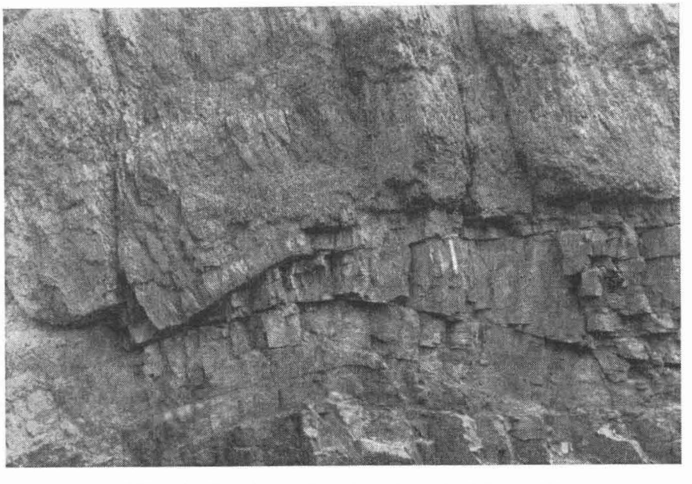

Fig. 4. Photograph of wedges in Figure 3.

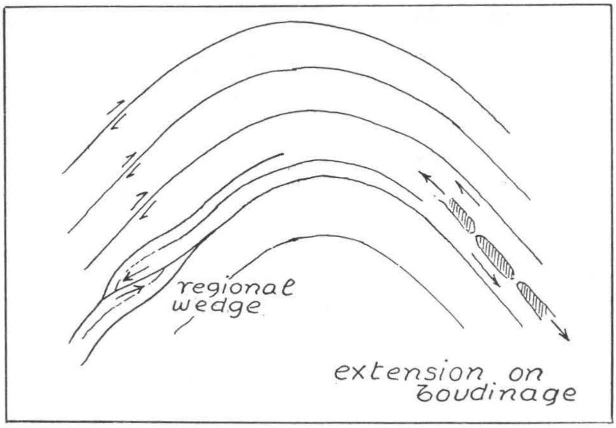

Fig. 5. Schematic fold showing boudinage (right) and wedging (left).

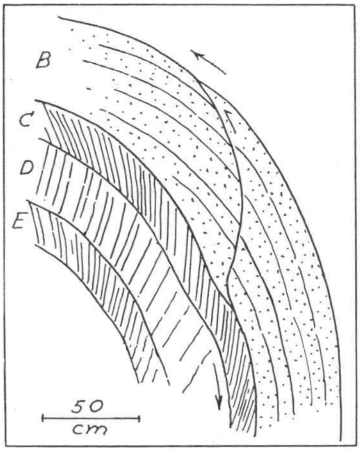

Fig. 6. Wedges in curved sandstone layer. Note also cleavage below sandstone bed B.

The arrows near x and y indicate the obvious relative displacement within the sandstone layer. Since the anticline is to the right the upper beds should move up and to the right relative to the lower ones if the displacement were due to bending of layers in folding. The principle is demonstrated in Figure 5. Where the movement is opposite to that of boudinage which serves to extend a competent bed between incompetent ones (see Figure 5 right limb). The displacement is probably due to a uniform regional bedding plane slip as discussed below.

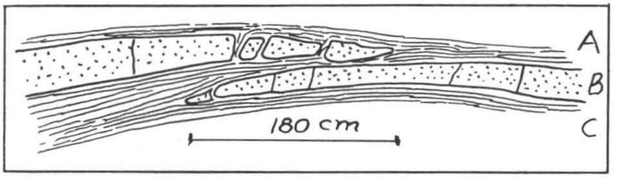

Fig. 7. Sandstone bed in shale. Wedged with boudinage in upper wedge. Devonian near Zell a. d. Mosel, Germany.

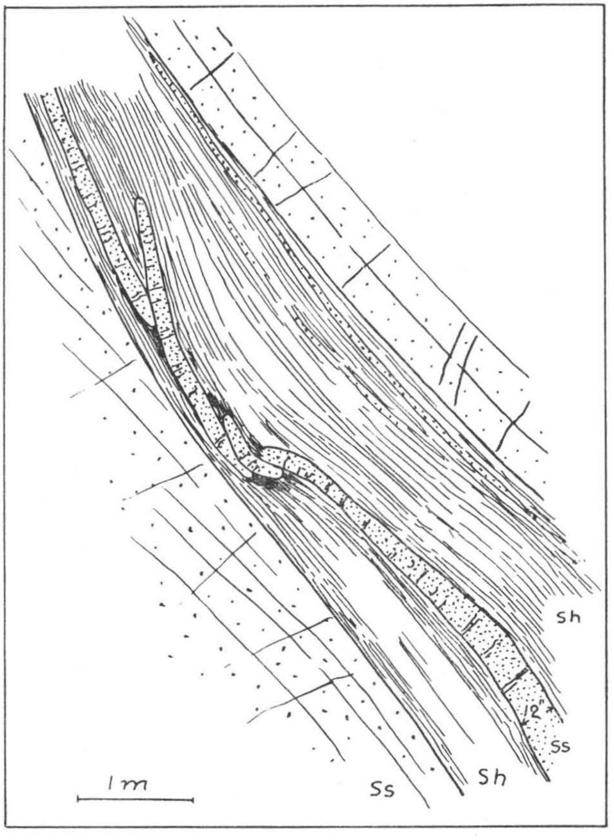

Fig. 8A. Thin sandstone layers in shale telescoped by movement parallel to bedding in shale. Chemung, Md. Total telescoping effect about 150 cm.

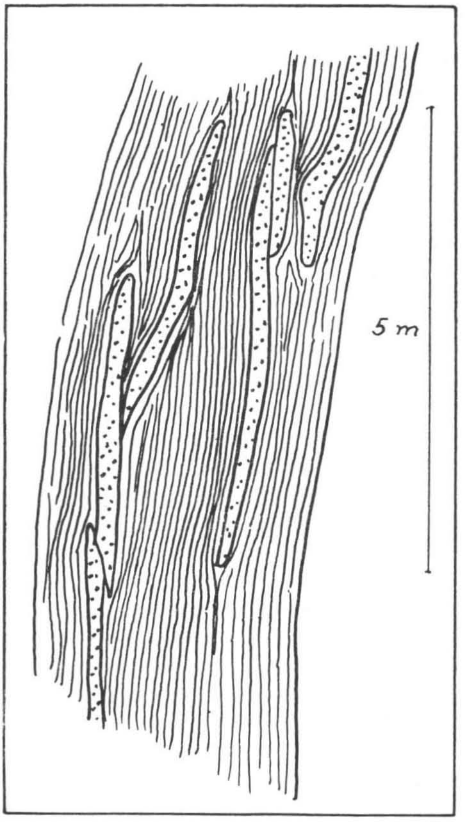

Fig. 8B. Telescoping in 2 beds by about equal amounts.

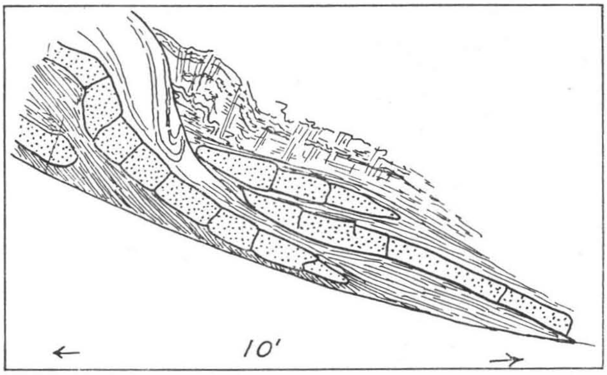

Fig. 9. Telescoped sandstone beds with tension joints which may lead to boudinage. A larger thrust begins to form and shale is crenulated next to it. Total effect about 3 meters.

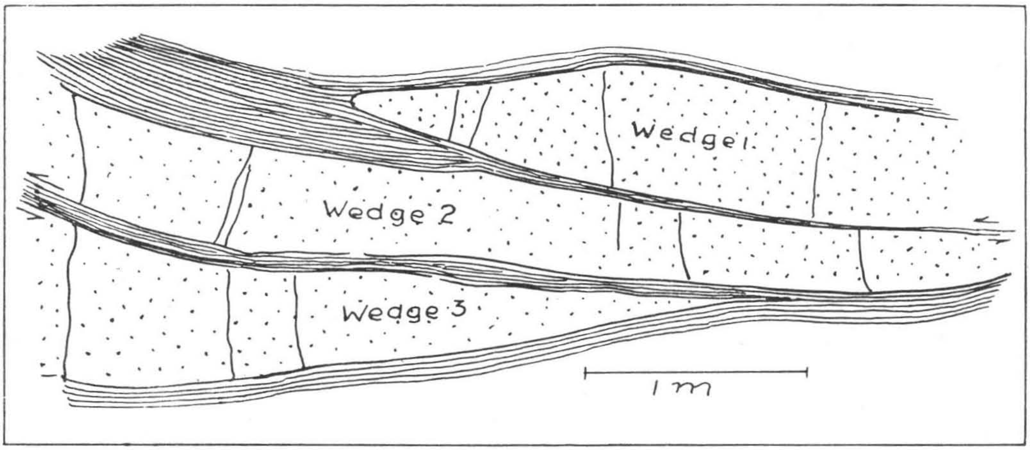

Fig. 10. 3 large wedges of one sandstone layer telescoped in shale. Devonian, Schuld a. d. Ahr, Germany. Total effect at least 8—10 m.

Figure 4 is a photograph of the bed drawn in Figure 3.

Figure 6 shows a similar situation to that of Figure 3. The crest of the anticline is to the left, the upper beds should move upward and toward the crest relative to the lower ones. In bed B a low angle fracture is a movement plane and served the same purpose as the bedding planes; the sense of displacement is the same as in the bedding surfaces.

Thinner sandstone beds show larger displacements and a very common telescoping effect due to movement parallel to the bedding. If the two

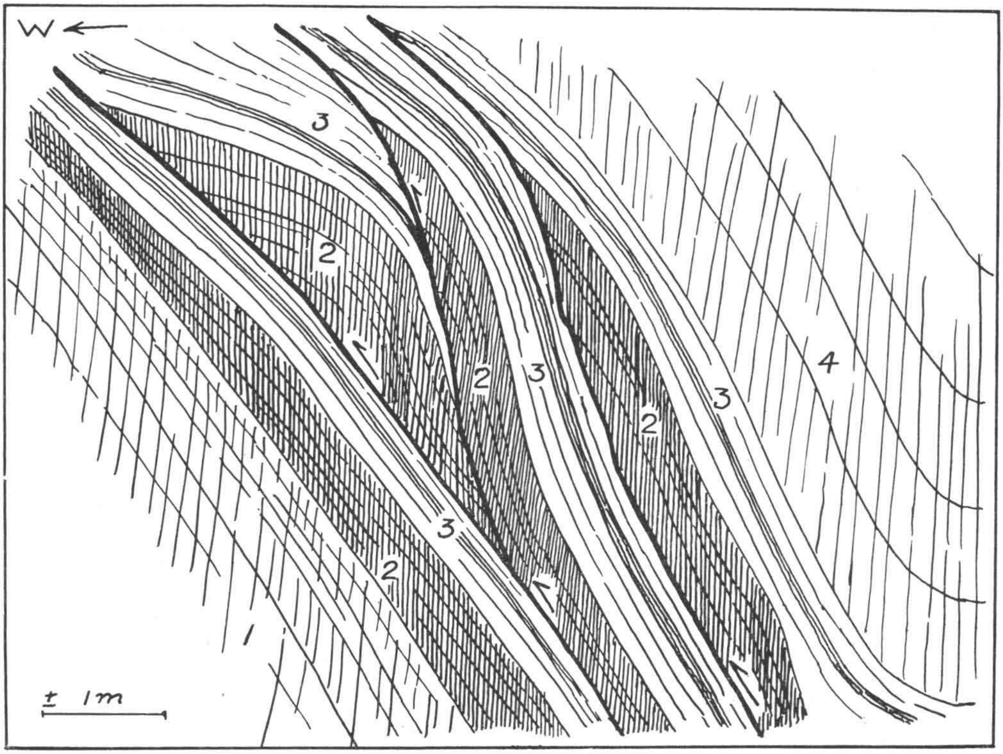

Fig. 11. Wedging in Bloomsburg shale, W. Md. R. R. opposite Great Cacapon. Shale and siltstone layers are repeated by thrusting and wedging. Anticline to left, movement essentially exaggerates bedding slip.

pointed endings of the sandstone bed in Figure 7 are moved into their original position considerable displacement must have moved the sandstone bed into its present position.

Similar telescoping can be seen in sandstone layers of the Chemung in Figure 8 A. The total reduction is at least 120 cm. In Figure 8 B two sandstone beds were telescoped an equal amount which may amount to 5 meters. A larger reduction occurs in Figure 9. The sandstone bed is here resolved into disconnected lenses which are stacked above each other. A small thrust has begun to form and the shale above the top lense is crenulated and a cleavage appears in the most intensely compressed shale area.

Figure 10 illustrates the wedging on a still larger scale. The height of the drawing is about 7 meters. One sandstone bed has been sheared into wedges and then telescoped. During the process the sandstone has thinned and thickened. The cumulative displacement must exceed 10 meters but

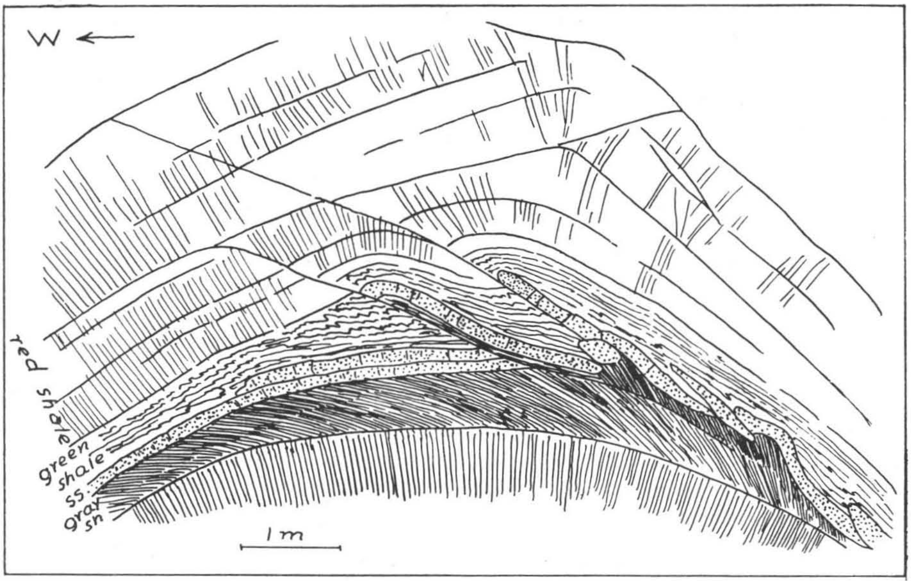

Fig. 12. Sandstone bed telescoped in green shale. Movement from right to left far in excess of bedding slip normal in this anticline. Bloomsburg formation, Round Top Section, Md. (See also E. Cloos, 1951, p. 157. By permission of Dept. Geol. Mines and Water Resources, State of Md.).

the relative displacement within the beds to the right and left must have been much larger.

Figure 11 is a drawing of a portion of the Round Top section in Maryland. A number of beds have been sliced into wedges and telescoped. Between two blocks are two wedges which consist of only portions of the section. Beds 2 and 3 have been affected between beds 1 and 4. Bed 3 can be reconstructed and placed end to end. The displacement is of the order of 150 or 180 meters: a syncline is to the right and an anticline to the left. The thickness of the beds is much too small to cause such large displacements along wedges or bedding planes only due to folding of that anticline. The relative motions indicated by the telescoped beds must be due to more extensive movements.

Wedging and telescoping parallel to bedding has reached still larger proportions in Figures 12 and 13. The two Figures show the north and south sides of a fold in the Round Top railroad cut in Maryland. A sandstone layer ss is over- and underlain by soft greenish shale. In the crest of the fold the layer is doubled and it has been thrust into the crest on two reverse faults. In the east limb the layer is also telescoped and portions of it are much distorted, rolled into pebble-like bodies or small folds. The shale below

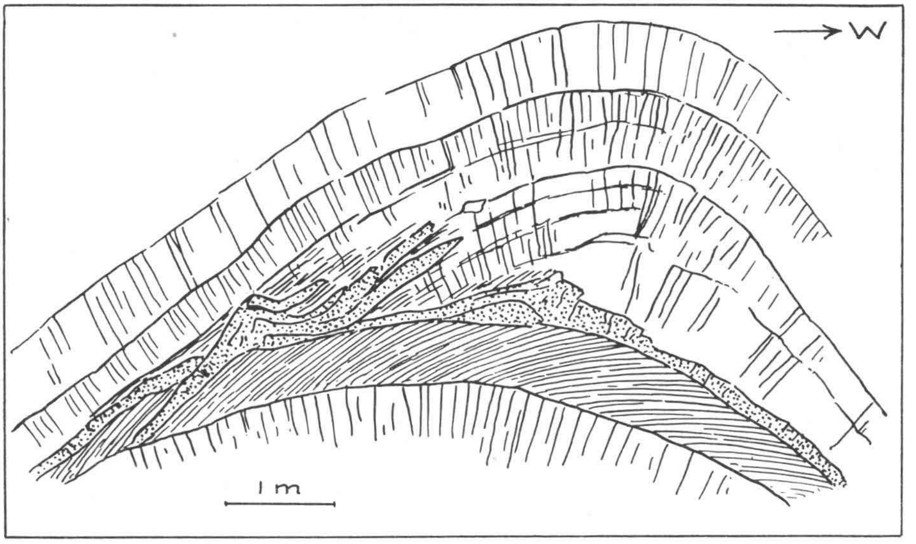

Fig. 13. Same sandstone bed and same structure but opposite side of Railroad cut. Movement from left to right.

the sandstone shows a very distinct cleavage which seems to indicate motion from east to west in the same sense as the sandstone telescoping.

On the other side of the same railroad cut (Figure 13) the details differ somewhat from those in Figure 12 but not fundamentally. The sandstone is also telescoped between shale layers, the motion is from east to west as indicated in the cleavage in the shale below the sandstone bed.

Above and below the key beds are hard red siltstones with a very prominent fracture cleavage. The lower shale is barely bent into a smooth and continuous anticline. The upper shale is also arched up but transected by small faults.

The fracture cleavage is fan cleavage and remains consistently normal to bedding in the two massive layers.

The displacement as shown by the telescoping of the sandstone bed and the cleavage in the shale below exceeds by far the amount which could be due to the anticlinal bend.

The overriding of two anticlines shown in Figure 14 is also due to wedging and slip on bedding planes. The movement plane is slickensided and coated with calcite ¼ inch thick in several layers. The siltstones above possess closely-spaced fracture cleavage almost normal to bedding. The displacements cannot be related to the anticline of the illustration but must be due to displacements of a much larger order of magnitude.

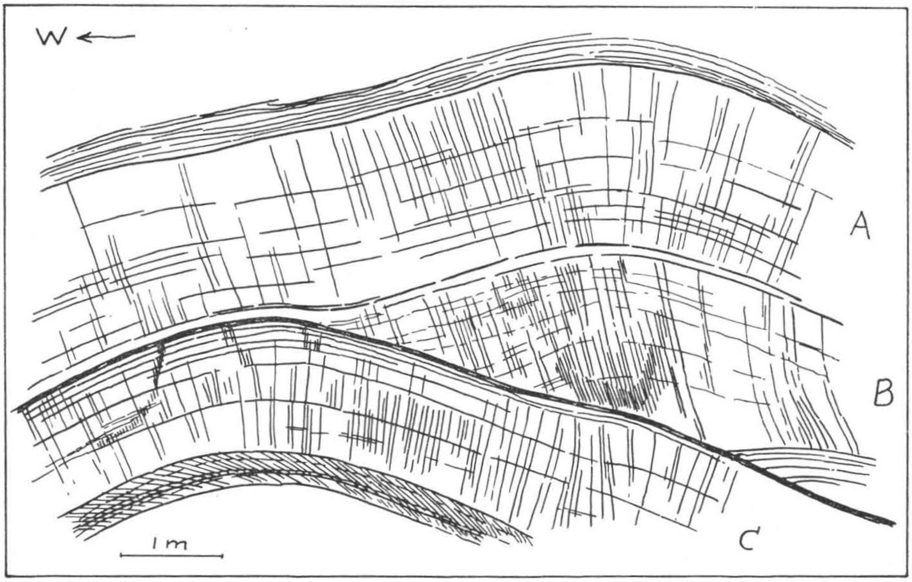

Fig. 14. Wedging on a larger scale. Beds A and B have been thrust over C. In this process a movement surface followed bedding below B, then cut across and continued below A. Bed B is missing on the left side (see also E. Cloos, 1951, p. 154).

Examples of wedges of all magnitudes have been observed at innumerable places in bedded and non-crystalline rocks of the Appalachians, Variscians, the Jura, and the Alps. Wedging is very common.

ORDERS OF MAGNITUDE OF MOVEMENTS

Busk (1929, p. 10) shows that the displacement on bedding planes due to concentric folding depends on the thickness of the affected beds. His figure 10 is here reproduced for convenience as figure 15. If a bed is bent isoclinally through an angle of 180 degrees and the motion has evenly affected both limbs, the displacement y and thickness x are related as follows:

$$y = \tfrac{1}{2} \pi x$$

This means that the slip between beds in a concentric fold is approximately 1.6 times the thickness of the bed. Since many folds are not isoclinal in Appalachian type terraines this value is a maximum. If the measurable displacement is below this value it may be due to the bending of strata. If the value is greater other movements must have occurred, probably prior to the bending in the fold.

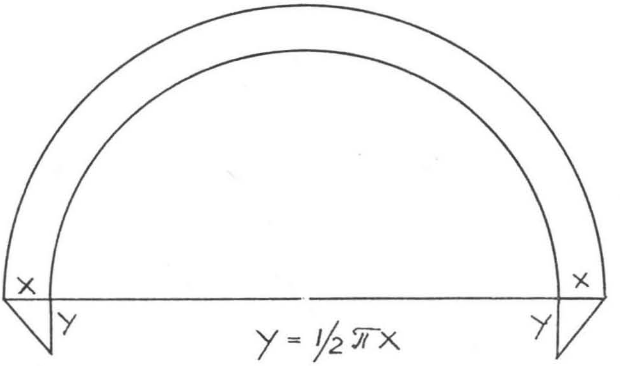

Fig. 15. Amount of displacement in a concentric fold after Busk (1929, p. 10).

The displacement in Figures 1, 2, 3 and 6 could be the function of bending in folds. In Figure 7 the displacement is 7 times the thickness of the sandstone bed; in Figure 8 more than 10 times. In Figure 9 the minimal displacement is 21 times the thickness and in Figure 12 at least 43 times. Additional considerable movement must therefore be responsible for the wedging and telescoping of beds in these examples.

In addition, the displacement shown is only a minimum of the probable motion that has taken place because much flowage may occur before telescoping takes place.

It seems evident that considerable displacements have occurred which cannot be derived from the folding process.

TIME OF DISPLACEMENTS

Wedging and telescoping of beds takes place prior to folding. This cannot be proven if exposures are small and scattered but both limbs of folds must be visible, or a systematic regional survey must show uniform displacements on either limb of folds. If the movement is in the same direction in both limbs, or in several folds, it can be safely assumed that slippage preceded folding.

Figures 12 and 13 show two views of one fold at Round Top, Maryland. The movement is from east to west. In Figure 12 telescoping of the sandstone bed moves up the anticline from the right and across its crest. Figure 11 shows the east limb of an anticline and the movement is up the limb and to the west, and in Figure 14 the motion is also to the west and across the crest of an anticline. Since Figures 12, 13, and 14 are in the Silurian of Maryland

and not more than about 10 kilometers apart it seems that the movement direction is uniform at least for these formations and in a small area.

The displacement shown in Figure 3 may be understood also as a westward component — down the west limb of an anticline. The locality is about 1—½ km from that shown in Figure 14. The displacement should be in the opposite sense if it were due to folding.

The slippage may be dated more accurately by systematic observations over larger areas and in completely exposed folds where both limbs can be seen.

In Figures 12 and 13 more telescoping has occurred on the east limbs of the fold and the accumulation of sandstone layers may be related to the fold — either as an initial thickening which resulted in an anticlinal structure at a place where later folding emphasized the thickening, or as slippage on a gently undulating and slightly folded surface which served as an obstruction where beds telescoped more readily. In either case it would seem that slippage and folding were not separated by a large time interval.

EXPERIMENTAL WEDGES

Along the boundary zone between two blocks that move past each other two shear planes and one set of tension fractures may occur. Both of these are well known and have been produced experimentally many times. They are shown in textbooks such as de Sitter (1956, p. 172), or Billings (1954, p. 104) and I have pictured them in 1955, pl. 3, in a discussion of faulting.

Wedging is due to the intersection of the low angle shear plane (Cloos, 1955, plate 3) with a bedding plane or other plane of separation which need not be primary but can also be an earlier shear plane (see Figure 19 and de Sitter, 1956, Figures 72, 87, 124). Wedges are easily produced experimentally.

Several experiments were performed using two layers of wet clay of the consistency of soft and harder putty. A clay model about 50 cm long, 15 cm wide, and 7—10 cm high was built. The surface was covered by wire netting (window screening or slightly larger) and the wire was used to pull the clay in the direction of the long axis of the model. The motion has to be very slow at a rate of not more than about 1 cm per hour.

Figure 16 is the side view of such an experiment. The top layer was pulled to the left and one wedge is seen in the center of the photograph. Figure 17 shows the side view of an experiment with 3 layers of clay consisting of a soft center layer and two harder ones above and below. The wedges formed very readily in the soft clay. The direction of movement is shown in the displacement of a vertical row of dots above the right edge of the scale.

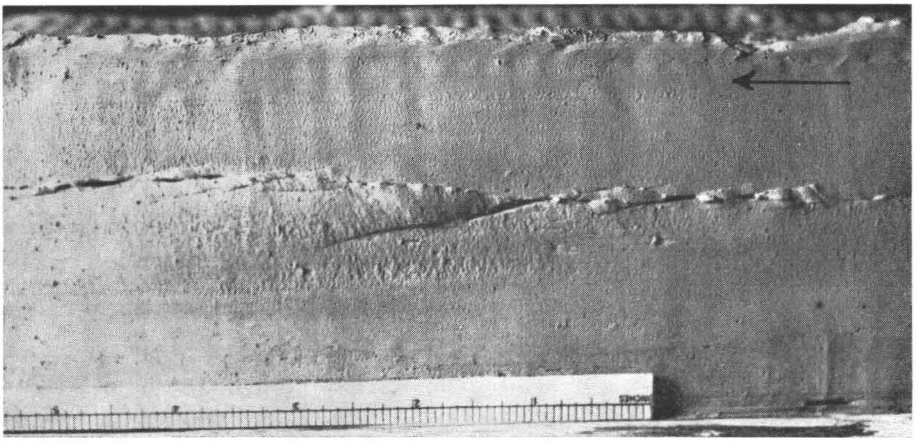

Fig. 16. Wedges in experiment with two clay layers.

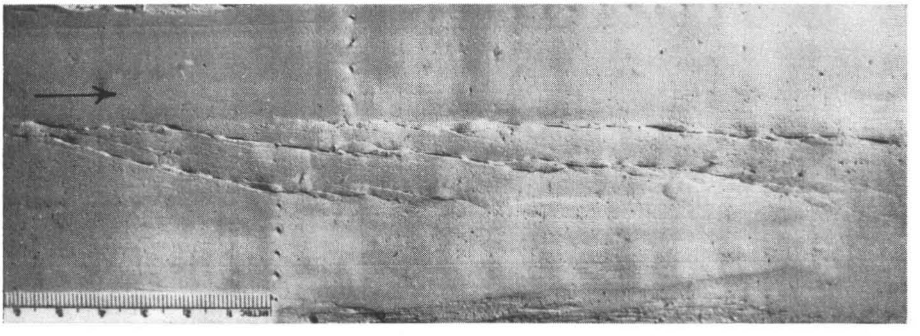

Fig. 17. Wedges in experiment with 3 layers of soft clay. Displacement is indicated by vertical rows of dots above right end of scale.

Wedges in folds are easily reproduced also in wet either layered or unlayered clay. Figure 18 shows wedges in an anticline which consist of 4 layers of slightly differing clay. If compression is applied and the layers are bent slippage on the bedding plane results in wedging.

But the clay need not be layered. If it is not surfaces are formed about where bedding planes would be (see de Sitter, 1956, p. 74 and 100). From then on the model is layered and displacements of the layers result in wedging as shown in Figures 19 and 20.

Figures 18, 19, and 20 demonstrate that wedging which is only due to folding must occur on the limbs of the folds and cannot appear in the crest of anticlines.

Fig. 18. Wedges in left limb of anticline with 3 layers of clay. High angle shear plane is also visible near anticlinal crest.

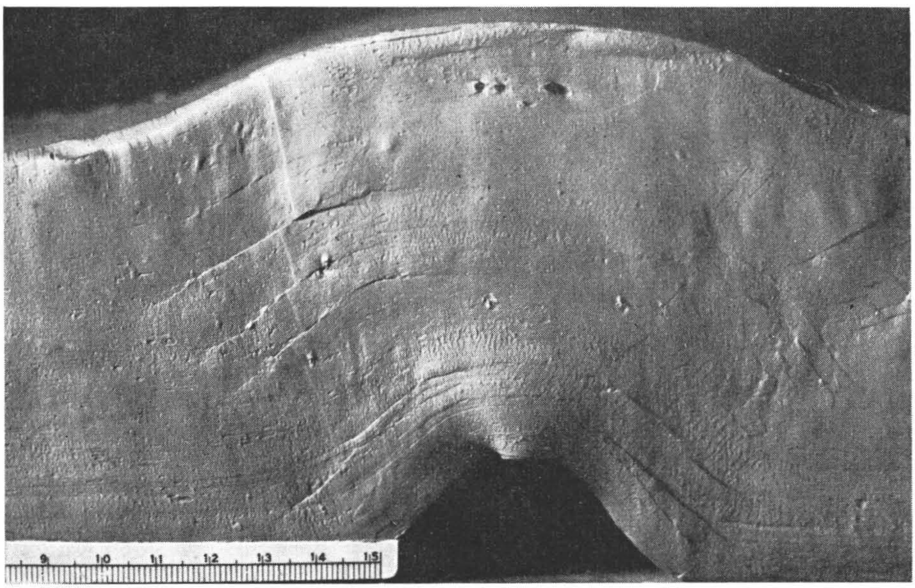

Fig. 19. »Concentric joints» (de Sitter, 1956) in clay experiment. Soon after formation of concentric joints they take the place of bedding planes and wedges form as seen in Figure 20.

Fig. 20. Wedges in experiment of Figure 19 at a later stage; formed only in limbs of folds.

BROADER APPLICATIONS

Bedding plane movements have been mentioned frequently but I do not know that their significance has been realized often or that they were ever measured or mapped systematically. Simple, readily observable criteria are generally lacking.

The observation of wedging in sedimentary sequences is similar to that of boudinage which indicates extension and is explained as a function of folding.

Unfortunately I have as yet not been able to map wedges systematically in the Appalachian area where they are common. I have seen them in the Variscian, the Jura, the Caledonian and elsewhere. In the Alps they occur but may not be as significant.

The few measurements in Maryland suggest a uniform westward slip within the Silurian Bloomsburg formation. The displacement exceeds by far the slippage due to folding and may amount to several hundred feet or more. It is quite well possible that the movements are significant and if plotted on maps would show that during an early phase of deformation slippage preceded folding. The areas affected are many times the size of single folds.

Systematic mapping of such displacements will, we hope, permit the construction of maps which show bedding slips for formations in which wedging is common.

DISCUSSION

Wedging, telescoping of beds and bedding slips are most frequently seen where anticlines are exposed. This may be due to a number of factors: either that anticlines always attract a geologists attention if completely exposed or a connection between anticlines and wedging exists. But since wedging cannot alone be due to folding, anticlines may have formed where wedging has caused thickening of beds. This seems quite possible in Figures 12, 13, and 14. If some slippage has occurred at the beginning of the folding process, wedging may have thickened some areas and anticlines be located at such points.

The fold shown in Figures 17, 18, and 19 was prepared on a thin sheet of aluminium. When compression was applied on both ends the available power was not sufficient to produce an anticline as long as the metal sheet remained flat on the table. Only after a thin trowel was inserted below the aluminium sheet did an anticline rise and hardly any power was necessary to keep the fold moving. It would therefore be reasonable to think that wedging may have a triggering effect in the location of folds. A systematic study of wedges and folds may furnish an answer to the question why anticlines or more generally folds occur just where we find them.

CONCLUSION

Bedding plane slippage is of orders of magnitude which cannot be derived from folding only. Regional components must have been operative at an early stage, either before folding occurred or during its beginning. Systematic regional mapping of slippage through observation of wedges and telescoped beds may well furnish data on the early stages of the folding process and its causes. It would seem highly improbable that wedges could form by horizontal compression without the formation of folds. They may be due to gravity gliding just prior to folding.

A systematic study of wedges and telescoped beds may be a very rewarding project.

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5

ZWEI BAUTYPEN PRAEKAMBRISCHER DEFORMATIONS- UND TRANSFORMATIONSSEGMENTE: KARELIDEN UND SVECOFENNIDEN [1]

VORLÄUFIGE MITTEILUNG

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ABSTRACT

Radiochronological determinations have given about the same age values for the Carelides and the Svecofennides. Some authors have therefore proposed to strike off these names, and to combine the two orogenic zones into one. The present notice shows that there are fundamental structural and evolutionary differences between the two, which give each of them their own individualities: hence separate names are justified. The two fold-belts represent different types in the vast series of basement chains. Deductions from radiochronological determinations need to be confronted with the evidence based on geological fieldwork and tectonic analysis.

INHALT

[1] Eingegangen den 8. April, 1961.

EINLEITUNG

In den letzten Jahren wurden, ausgehend von einigen absoluten Altersbestimmungen, allerlei Theorien über die Svecofenniden und Kareliden geäussert. Es scheint, dass dabei hauptsächlich petrographische Gesichtspunkte und chronologische Angaben ohne vertiefte Berücksichtigung der Grundlagen als Ausgangspunkte dienten. Es ist interessant festzustellen, dass diejenigen Gesichtspunkte, die sich aus der geometrisch-kinematischen Analyse ergeben, kaum erwähnt wurden. Dies ist umso erstaunlicher, als man erwarten könnte, dass bei der Ausarbeitung solcher, auf die Chronologie sich stützender Theorien die Besinnung auf die raumzeitlichen Verhältnisse der Bauformen und Ereignisse in den Vordergrund zu treten hätte. Es wären dabei die Ergebnisse der Forschungen zu berücksichtigen, die sich die Aufgabe gestellt haben, die Entwicklung der Formen verschiedener Grössenordnungen in Zeit und Raum zu bestimmen. Die Gründe dafür mögen dahingestellt bleiben. Sie werden demjenigen, der einmal die Entwicklung des Bewusstwerdens der Probleme untersuchen wird, ein interessantes Kapitel liefern; denn es ist immerhin eigentümlich, dass dort, wo es am ergebnisreichsten wäre, in Zeit und Raum zu denken, diese Betrachtungsweise so spärlich angewendet wird.

Es sei hier nicht auf die kritische Analyse der Altersbestimmungen eingegangen. Wer die Literatur über den Gegenstand kennt, weiss, dass die rohen Ergebnisse oft recht vieldeutig sein können, und ohne stratigraphische und tektonische Analyse nur mit Vorsicht gehandhabt werden sollten. Für das vorliegende Problem hat Edelman (1960) bereits auf wichtige Fragen hingewiesen. Auch die grundsätzlichen Darlegungen über die Altersbestimmungen an Gesteinen jüngerer Kettengebirge, namentlich durch Roques und durch Faul, wären zu berücksichtigen.

In diesen Zeilen soll hauptsächlich auf die F r a g e eingegangen werden, o b e s r a t s a m s e i, d i e S v e c o f e n n i d e n u n d d i e K a r e l i d e n z u s a m m e n z u w e r f e n, o h n e d i e b a u l i c h e n U n t e r s c h i e d e z u b e r ü c k s i c h t i g e n.

Es war uns seinerzeit möglich, eine Reihe geometrisch-kinematischer Analysen in Südfinnland, Ostfinnland und Nordfinnland durchzuführen. Diese Analysen erlaubten für gewisse typische Gebiete, die Erscheinungen in Zeit und Raum zu ordnen. Dabei wurden nicht nur die Erscheinungen einer einzigen Grössenordnung herausgefiltert, sondern es wurde versucht, die verschiedenen Grössenordnungen zu berücksichtigen und zu erfassen. Es wurde also nicht nur ein kleiner Spalt aus dem Spektrum der Grössenordnungen herausgeblendet, wie das z.B. meist bei den petrographischen und petrochemischen Untersuchungen geschieht, bei denen die Verhältnisse einer oder zwei Grössenordnungen, der chemischen Elemente und der Mineralien, berücksichtigt werden, sondern es wurde versucht, von der Gesteinsprobe bis zu den Erscheinungen im Kilometermasstab Beobachtungen zu sammeln und diese zu koordinieren. Diese Verfahrensweise hat gewisse Vorteile, welche die Mühe reichlich lohnen, die durch die Anwendung ganz verschiedener Arbeitsmethoden, entsprechend den verschiedenen Grössenordnungen, entstehen. Es ist natürlich leichter, einen Spalt aus dem Spektrum der Grössenordnungen herauszublenden und die dabei gewonnenen, notwendigerweise beschränkten Ergebnisse nachher durch allerlei frei gestaltete Hypothesen zu vervollständigen. Bei petrochemischen Untersuchungen z.B. hat man die Gesteinstypen nach chemischen Gesichtspunkten geordnet und mit Hilfe der Differentiationshypothese einen Ablauf konstruiert. Dabei wurde oft nicht einmal versucht zu zeigen, ob die Gesteine einer Serie angehören und ob sie wirklich durch eine Differentiationserscheinung mit einander verbunden sein könnten. Diese Extrapolationen zeigen, dass man nicht immer auf einfachste Weise von kleineren auf die grösseren Masstäbe übergehen kann. In vielen Fällen wäre zuerst zu untersuchen, ob die Hypothesen, nach denen solche Extrapolationen durchgeführt werden, wirklich angewendet werden dürfen. So wurden auch, ausgehend von petrochemischen Untersuchungen hie und da Aussagen über die Form der Körper (Intrusionen) gemacht, wobei angenommen wurde, dass ein Gesteinstyp nach dem anderen nach dem Schema abnehmender Basizität, aus unbekannter Tiefe emporgestiegen sei.

Durch die geometrisch-kinematische Analyse erst wird d i e K a r t e z u m R a u m b i l d e. Seine Umrisse werden allerdings mit dem Abstande von der Aufschlussfläche unbestimmter, wobei aber unterschieden werden muss zwischen den numerischen Angaben und der funktionellen Rolle in der Bauformel. Um ein so vollständiges Bild als möglich zu bekommen, ist es daher nötig, namentlich auch die grössten Züge herauszuarbeiten. In diese sollten die Beobachtungen der niederen Grössenordnungen eingefügt werden, um den Bewegungsmechanismus verständlich zu machen.

Die verschiedenen Züge dieses Raumbildes können in einander folgende Ereignisgruppen aufgeteilt werden. Durch eine solche Aufteilung und die Aufreihung der dadurch erhaltenen Bilder kann der Ablauf mit einem Minimum an allgemeinen genetischen Hypothesen festgelegt werden. Diese Bilder können, wie die Bilder eines Films, hinter einander aufgereiht werden. Dadurch entsteht eine C h r o n o l o g i e, die zwar nicht absolut, aber wenigstens r e l a t i v ist. Diese Verfahrensweise der hintereinander gereihten Zustände hat den Vorteil, dass die E r e i g n i s f o l g e sichtbar wird, auch wenn sie nicht mit den geläufigen, aus vereinfachten Mechanismen abgeleiteten Theorien übereinstimmt. Solche Bilderfolgen können im Gegenteil dazu anregen neue Erklärungsmöglichkeiten zu finden, die ausserhalb des Wegnetzes der gewohnten Theorien liegen. Werden in dieses Bild die

Typen und Übergänge der Ablagerungs-, Umwandlungs- und Intrusionserscheinungen eingefügt, so ergibt sich das Panorama eines farbenreichen Geschehens.

In diesem Geschehen kann man zwei verschiedene Arten von Zeugnissen unterscheiden: a) diejenigen, welche an der Oberfläche bei der Aufbereitung, beim Transport und der Ablagerung der Sedimente und der Zwischenlagerung vulkanischer Massen und ihrem Gefolge entstanden, und b) die Ergebnisse der unterirdischen Entwicklung, bei der das Vorhandene umgeformt, umgewandelt und teils in verschiedenen Grössenordnungen entmischt, teils gemischt wurde und die entstandenen Produkte mit anderen, teils endogenen Zuschüssen vermengt wurden.

In den jüngeren Gebirgsketten sind die Zeugnisse der ersten Art besonders augenfällig; in den tieferen Teilen der Gebirgskörper und den tief erodierten alten Kettengebirgen verschleiern die unterirdisch entstandenen Erscheinungen oft die Geschichte der an der Oberfläche aufgeschütteten Baustoffe. Sie treten oft so stark in den Vordergrund, dass es nicht leicht ist, zu den Zeugnissen der ersten Gruppe vorzudringen. Diese Züge müssen daher manchmal durch die Kombination verschiedener Untersuchungen nach und nach sichtbar gemacht werden.

Wenn es gilt, die Frage zu beantworten, ob es nützlich sei, zwei Deformationssegmente zusammenzuschlagen, oder als individuelle Einheiten zu betrachten, sollte nicht ein auf chronologische Bestimmungen zusammengeschrumpftes Wissen, sondern eine möglichst allseitige Betrachtung zu Grunde gelegt werden.

Im Folgenden soll versucht werden, Kareliden und Svecofenniden unter verschiedenen Gesichtspunkten zu vergleichen. Da das Gebiet ausserordentlich weit ist, kann dies nur in kurzen Andeutungen geschehen. Dabei werden wir uns auf die in Finnland gelegenen Abschnitte beschränken.

VERGLEICH ZWISCHEN KARELIDEN UND SVECOFENNIDEN

LAGE UND VERHÄLTNIS ZUR UMWELT

Ein erster wichtiger Unterschied liegt im Verhältnis zur Umgebung.

Die Kareliden in Ostfinnland zeigen deutlich den Rand des Deformationssegmentes gegen den alten Sockel des Vorlandes. Man sieht deutlich, wie der alte vorkarelidische Sockel, schon zu jener Zeit bis zum Grundgebirgsstockwerk abgetragen, unter die Sedimente des karelidischen Zyklus taucht. Dabei ist nicht nur am Kontakte, infolge der Erosionslücke, eine Diskordanz festzustellen, sondern die alten und die neuen Formen haben ganz verschiedene Richtungen. In den Kareliden kann man daher die Bauformen vom alpinotypen Rande her bis zu den Stockwerken mit Grundgebirgscharakter verfolgen. Im Inneren, gegen W, kann man nur eine geographisch-konventionelle, aber keine strukturelle Grenze ziehen.

Ein ganz anderes Bild bieten die Svecofenniden. Sie zeigen nirgends eine strukturelle Grenze, sondern sind nur durch die späteren Bedeckungen begrenzt. Im SE verschwinden sie am Rapakiwimassive von Viborg, weiter westlich tauchen sie unter die Ostsee und die kambrischen Ablagerungen, ohne dass man Anzeichen einer strukturellen Grenze bemerken würde. Im Norden könnte man eine konventionelle Grenze am Rande der Tamperezone ziehen. Sie wäre aber keine wirkliche Begrenzung, da Gebilde, ähnlich den weiter südlich gelegenen, auch weiter nördlich auftreten, allerdings in verschwommenener Form. Der Aufschlussbereich der Svecofenniden in Finnland entspricht daher nicht einem Deformationssegment mit strukturellen Grenzen. Sie zeigen weder die Verbindung mit dem Vor- noch mit dem Rücklande. Sie sind ein durch die Art der Freilegung geographisch bedingtes Bruchstück, ein Torso, der durch seine innere Gestaltung und nicht durch seine äussere Begrenzung charakterisiert wird, etwa in der Art der alten Massive Mitteleuropas.

Im mittleren Schweden liegen die Verhältnisse anders. Die geometrischen Formen der Grenzzonen wurden aber bis jetzt nicht im Hinblick auf diese Frage untersucht.

ZEUGNISSE ALTER OBERFLÄCHENBILDUNGEN

In den Kareliden liegen auf dem vorher tieferodierten Sockel zuerst Quarzite, deren Zusammensetzung zeigt, dass sie umgelagert und ausgereift wurden. Die hauptsächlich dolomitischen Karbonatgesteine sind ebenfalls charakteristisch für die Bedeckung eines alten Tafellandes. Die Bildungen zeigen Ähnlichkeiten mit denjenigen der germanischen Trias, die sich ja weit in die Westalpen hinein erstrecken. Sogar die Lettenkohle findet ihr Analogen in den shungitischen Schichten. Erst in den mächtigen Schieferbildungen beginnt der geosynklinale Einschlag. Wären die Sulfiderze, wie es Borchert für Outokumpu annimmt, aus sedimentären Ausfällungen mobilisiert worden, so hätte man Becken mit Bodenbildungen unter Sauerstoffabschluss anzunehmen. Die alten Sockelbedeckungen setzen sich weit nach dem Inneren bis in die Gegend von Kuopio fort. Erst über den Tafelbildungen setzen nach und nach die orogenen Ablagerungen ein. Über das Liefergebiet ist wenig bekannt. Es erscheint glaubhaft, dass ein grosser Teil des Materials von Westen her geschüttet wurde. Jedenfalls hatte der ältere Sockel schon vor der Ablagerung der jatulischen Quarzite

Grundgebirgscharakter. Die Tiefenerosion des präkarelidischen Sockels fällt also in eine weit frühere Zeit:

Mächtige Sedimentpakete wurden in Ostfinnland von ihrer Unterlage abgeschoren und nach vorne verfrachtet. Es ist daher nicht leicht, die ursprüngliche Stratigraphie herzustellen.

Auch die Bildungen der Svecofenniden wurden sicher auf einem alten Sockel abgelagert. Es war eines der Hauptanliegen J. J. Sederholm's diesen Sockel zu finden. Es ist wahrscheinlich, dass in einer so stark umgeformten und umgewandelten Zone nur wenige Reste von Diskordanzen erhalten geblieben sind. Wo solche sichtbar sind, gehören sie vielleicht jüngeren Ereignissen an. Einen guten Fingerzeig gäben ausgedehnte Vorkommen von ausgereiften Deckgebirgsbildungen. Es gibt zwar mancherlei Quarzite in den Svecofenniden aber nur wenige, die man als alte Tafelbildungen deuten könnte. Man könnte annehmen, dass sie in gewissen Zonen der Granitisation zum Opfer gefallen seien, obwohl solche Quarzite oft recht widerstandsfähig sind. Die Karbonatgesteine der Svecofenniden zeigen eine ganz andere Fazies als diejenigen der Kareliden. Die Mehrzahl von ihnen dürfte der Gruppe angehören, die in Schweden als Leptitformation bezeichnet wird.

In den Svecofenniden nehmen kornsortierte geschichtete Ablagerungen *(graded bedding)*, hie und da mit eigentlichen Warwen, einen breiten Platz ein. Solche kornsortierte Ablagerungen wurden von der Ladogagegend bis nach Schweden angetroffen. Reste ihrer ursprünglichen Struktur sind oft in stark umgewandelten Gneisen noch erkennbar. Wir haben (Wegmann, 1930) ihre Umwandlungen in Cordieritgneise verfolgt und beschrieben. Auch die graphithaltigen Glieder dieser Serie gehen in Gneise über, für die damals der Name Kinzigit eingeführt wurde. Mächtige basische Vulkanite sind weit verbreitet.

Viele dieser Sedimente sind wenig ausgereift, wurden also verhältnismässig rasch in die Ablagerungsbecken transportiert, in denen sie sich absetzten. Ein solcher aktivierter Sedimentumschlag ist nur in einer sich bewegenden, immer wieder Relief bildenden Zone möglich. Es ist daher in dieser Zone mit rascher und tiefer Erosion zu rechnen. Die Lücken, sowohl der alten Sockel, als auch der Liefergebiete der Sedimente, können wahrscheinlich erst nach und nach durch die tektonische Analyse identifiziert werden. Ohne eingehende Studien ist es nicht möglich, die sich folgenden Oberflächenverhältnisse zu rekonstruieren. Die verschiedenen Arten bei der Umformung und Umwandlung zu reagieren können bereits mancherlei nützliche Hinweise geben, wenn die strukturelle Analyse nicht schematisch und die Gesteinsuntersuchung nicht schubladenmässig durchgeführt werden.

Bei der Besprechung der Ablagerungsgesteine ist es vielleicht angebracht, einige Worte über das Jotnium beizufügen. Die jotnischen Sandsteine werden hie und da als »Molasse der Svecofenniden» bezeichnet. Dazu wäre zu bemerken, dass die typischen Molassegesteine am Aussenrande der Alpen die Erosionsprodukte einer sich hebenden mächtigen Gebirgskette darstellen. Die Erosion legte fortwährend frisches Gestein frei, das in einigen Sammelbecken vereint zu mächtigen Fächern geführt und abgelagert wurde. Es liegt in der Regel nicht auf einem alten abgetragenen Sockel, sondern auf dem Deckgebirge des Vorlandes. Es gibt zwar in den Ostalpen kleinere Molassebecken im Inneren des Gebirges. Der Typus ist aber der der perialpinen Trogfüllungen.

Das Bild des Jotniums mit seinen roten Sandsteinen ist ein ganz anderes. Seine Faziesvergesellschaftung gleicht viel mehr derjenigen des Buntsandsteins. Wie er, legt sich das Jotnium weithin auf eine bereits vorher tief erodierte Fläche mit Grundgebirgscharakter, wurde daher nicht während der Hauptphase der Hebung und Erosion abgelagert, sondern erst als der grösste Teil der Gebirgslandschaft bereits abgetragen und eine Fastebene an ihre Stelle getreten war. Diesen Verhältnissen entsprechen auch Fazies und Reifezustand der Sedimente.

Die grosse Masse der vorher abgetragenen Gesteine muss daher an anderen Orten gesucht werden. Svecofenniden und Kareliden bildeten ein Belieferungszentrum für andere Gebiete.

UMFORMUNGEN UND UMWANDLUNGEN

Am augenfälligsten unterscheiden sich die beiden Segmente durch den jetzigen Bau und die damit verbundenen Gesteinsumwandlungen. Diese Unterschiede würden am besten durch eine grössere Reihe von Abbildungen mit den nötigen Kommentaren dargelegt. Ein solches Unternehmen würde aber den Rahmen dieser Mitteilung sprengen. Wir wollen daher versuchen, einige Merkmale auszuwählen und sie kurz darzulegen.

Die Kareliden in Ostfinnland zeigen einen alpinotypen Bau mit liegenden Falten, abgeschorenen und verfrachteten Schichtpaketen, Decken und kristallinen Massiven und Keilen. Da eine Reihe typischer Einwicklungen älterer überschobener Massen durch jünger bewegte Massive vorkommt, ist es möglich, tektonische Phasen zu unterscheiden und ihnen die Gesteinsbewegungen zuzuordnen. Dabei ist die relative Bewegung der höheren Bauteile in der Hauptsache nach Osten gerichtet. In der Gegend von Kuusamo und Paanajärvi drehen zwar die Aufschlusstreifen. Die relative Bewegung der höheren Bauteile ist aber auch hier nach Osten gerichtet. Das Kartenbild kommt durch die kräftige Axialsenkung im Gebiete von Paanajärvi zustande.

Je mehr man aber nach Westen kommt, umso mehr werden alle Bauteile vergneist und hie und da von Graniten durchbrochen. Dadurch nähert sich der Stil mehr und mehr dem der Grundgebirgstypen.

Die Svecofenniden Finnlands zeigen nur Stockwerke vom Grundgebirgstypus, wobei es möglich ist, verschiedene Stilarten zu unterscheiden. Diese Typen sind nicht leicht in wenigen Worten zu beschreiben. Sie sind auch auf vielen geologischen Karten nicht ohne Schwierigkeit zu erkennen, da die Grenzen der petrographischen Typen oft schräg über die tektonischen Einheiten schneiden. In vielen Gegenden wird der Bau durch weitgehende Granitisation verschleiert.

Da viele Gesteine bereits vor der Granitisation stark deformiert waren und die verschiedenen Deformationsstadien des gleichen Gesteins bei der Granitisierung recht verschiedene Gesteinstypen ergeben, so ist eine schubladenpetrographische Einteilung in Beschreibung und Karte wenig nützlich zur Aufklärung des Baues, wenn diese Einteilung nicht gedeutet werden kann. Die Mannigfalt der Gesteinstypen wird dadurch noch erhöht, dass jeder der erwähnten Arten eine Serie weiterer Produkte entspricht, da die Granitisation unter verschiedenen Bewegungsbedingungen geschehen und dadurch recht vielfältige Bilder zustande bringen kann. Erst die kinematische Analyse kann die Verhältnisse klar legen helfen.

Die Generationen basischer Gänge, die schon von J. J. Sederholm als chronologische Stützpunkte benützt wurden, sind auch für die Bewegungsanalyse von grosser Wichtigkeit.

Im Grossen kann man folgende Gruppen von Ereignissen unterscheiden: Wie bereits erwähnt deuten die Merkmale der Ablagerungen auf erhebliche synsedimentäre Bewegungen hin. Da die Stratigraphie einstweilen primitiv ist, können sie kaum unterschieden werden. Weit verbreitet sind die Spuren einer Deformationsgruppe, die in höheren Stockwerken stattfanden. Zu ihnen gehört auch die Platznahme von Graniten, wie sie diesen Stockwerken entsprechen. Darauf folgen die verschiedenen basischen Gänge, in verschiedenen Schüben, die aber wohl nicht über das Ganze gleichzeitig sind. In einer folgenden Phase wurden sie deformiert. Für die Erforschung der Ereignisse in Zeit und Raum sind die basischen Gänge nicht nur chronologische Signale, sondern sie bieten auch Scharen von geometrischen Bezugsflächen.

Die darauf folgenden Granitisationen können in Folge des Bewegungszustandes recht verschieden aussehen und eine grosse Anzahl von Spielarten, von scharfen Kontakten bis zu komplizierten Übergängen, zeigen. An die Granitisationen schliessen sich weitere Intrusionen. Sie zeigen die Form von runden oder ovalen Diapiren, können aber auch langgestreckte mauerartige Gebilde erfüllen.

Viele der aufeinander folgenden Bewegungsgruppen können lokal oder regional in mehrere verschiedene Phasen unterteilt werden. Die jeweils jüngeren Bauformen wickeln die älteren ein, wobei die Achsenrichtungen meist verschieden und nur in Ausnahmefällen parallel sind. Ein scheinbar so einfacher Vorgang wie die Ausdünnung einer Schichtenreihe zeigt eine ausserordentliche Mannigfalt von Erscheinungsformen, die in gewissen Fällen einander im Streichen ablösen können.

Die Formen kleineren und grösseren Masstabes sind daher meist durch mehrere Überprägungen charakterisiert. Wenn diese Komplikationen in einem Aufschlusse nicht immer direkt sichtbar sind, so machen sie sich schon bei einer geringen Erweiterung des Erfahrungskreises bemerkbar.

Da die relativen Verschiebungen nicht nur tangential und vertikal sind, sondern sich auch stark horizontale Komponenten auf steilen Bewegungsbahnen oft bemerkbar machen, entstehen recht verwickelte Strukturen, wie etwa die seinerzeit von uns beschriebenen wirbelartigen Bilder im östlichen Schärenhofe. Komplizierte Strukturen sind sehr häufige Merkmale, man könnte fast sagen, dass sie die Regel sind.

Infolge der Granitisierung und der damit verbundenen Bewegungen stellen sich die Gebiete mit besser erhaltenen Ablagerungen manchmal wie im Migmatitgranit schwimmende »Inseln» mit verdrehter Innenstruktur dar. Die Kontakte sind teils überschneidend teils subparallel, von breiten Übergangszonen bis zu mehr oder weniger scharfen Berührungsflächen variierend. Da derartige »Inseln» verschiedenen Tiefgang haben, ist es möglich, eine Art vergleichender Anatomie aufzustellen. Diese zeigt, dass die Verhältnisse zwischen den »Insel»-gruppen plastischen oder viskosen Bewegungsstilen entsprechen.

Mit der Verfestigung, die wahrscheinlich der Abtragung in der Tiefe voranging, hörten aber die Bewegungen nicht auf, wie die häufigen Übergänge von Migmatitgneisen in Mylonite und Pseudotachylite zeigen, die sowohl in der Stadt Helsinki als auch in der Umgebung häufig angetroffen werden. Solche späte Bewegungserscheinungen wurden auch von Tuominen (1957) für die Gegend von Orijärvi unterstrichen. Sie überprägen die Merkmale mehrerer früherer Bewegungsgruppen.

Die Gruppe der Rapakiwigranite durchschneidet bereits ein verfestigtes Grundgebirge.

Dieses, wenn auch leider recht schematische, Panorama stützt sich auf die Erfassung aller Grössenordnungen. Es ist daher vielleicht nicht mit allen im Schrifttum angetroffenen Vorstellungen vereinbar, da manche derselben auf ganz anderen Wegen gewonnen wurden. Zwei Vorschläge wurden seinerzeit von anderer Seite gemacht, um die geometrisch-kinematische Analyse überflüssig zu machen: a) man sollte nur die im Gelände direkt beobachtbaren und durch das Mikroskop ableitbaren Merkmale berücksichtigen; b) die Kartierung sollte lediglich die Mineralfazies erfassen. Es ist vielleicht nützlich, sich zu diesen interessanten Gesichtspunkten zu äussern.

Nach dem vorangegangenen dürfte es nicht schwer sein, zu verstehen, dass die Ausblendung einiger weniger Grössenordnungen auch eine Schrumpfung des Geltungsbereichs nach sich zieht.

Das Verhältnis der M i n e r a l f a z i e s s t u d i e n zur raumzeitlichen Erforschung liegt wieder ein wenig anders: Der Begriff der Mineralfazies geht vom Gleichgewichte aus. Dies ist eine für die Erforschung der Natur wichtige Abstraktion, wie etwa der Kreis oder das Dreieck. Wäre aber das Gleichgewicht in der Natur überall verwirklicht, so würde dies das Ende aller geologischen Prozesse bedeuten. Es ist immer nur ein Grenzfall. Diese Grenzfälle entsprechen gewissen Verhältnissen von Temperatur, Druck und Feuchtigkeit, d. h. dem, was man als das u n t e r i r d i s c h e K l i m a bezeichnen könnte. Sie entsprechen gewissen Schwellenwerten und ihren Kombinationen. Die unterirdischen Klimate beherrschen gewisse Räume, während gewissen Zeitabschnitten, und können dort ihre Merkmale hinterlassen. Es ist nun eine, oft wiederholte Beobachtungstatsache, dass die Zeugen unterirdischen Klimas, auch wenn ihre Bereiche aneinander grenzen, nicht notwendigerweise demselben Zeitabschnitte angehören müssen, sondern aus ganz verschiedenen Entwicklungsstadien stammen können. Eine Aufschlussfläche schneidet daher sowohl durch synchrone als auch durch metachrone Klimaräume. Um sie zu rationellen Entwicklungsreihen zu verbinden, müssen die raumzeitlichen Verhältnisse bekannt sein. Werden die Angaben, die sich aus den Gleichgewichtstendenzen ergeben, nicht auf diese Weise geordnet, so ist dies, wie wenn man, um eine Klimakarte herzustellen, die Zeugen des karbonischen, jurassischen und tertiären Klimas, samt dem quartären und dem jetzigen nebeneinander notieren würde, nur weil diese Zeugen in dem Gebiete vorkommen. Ohne die Zeitperspektive und die räumliche Begrenzung würde das erhaltene Bild eher verwirrend wirken. Auch die Ergebnisse der Mineralklimatologie können erst rationell verwendet werden, wenn sie in das raumzeitliche Panorama, d.h. in die tektonischen Stockwerke und ihre Entwicklungsphasen eingefügt werden.

EINIGE FOLGERUNGEN

Wir haben versucht, wenn auch recht schematisch, aus den vielen Beobachtungen einige Merkmale der Kareliden und der Svecofenniden herauszugreifen. Bereits diese Auslese zeigt, dass ein Z u s a m m e n w e r f e n b e i d e r G e b i r g s s e g m e n t e e i n e r b e t r ä c h t l i c h e n S c h r u m p f u n g u n s e r e r K e n n t n i s s e d e r i n d i v i d u e l l e n Z ü g e entspräche. Sie würde aber auch ohne vorherige neue Unterteilung kaum die zukünftige Forschung fördern. Im Gegenteil würden — wie es schon hie und da geschah — verfrühte zu weit ausgedehnte Verallgemeinerungen den Fortschritt hemmen. Es würde wahrscheinlich niemandem einfallen, die betische Kordillere, die Alpen und die Karpathen zusammenzuwerfen, weil sie sich ungefähr gleichzeitig entwickelt haben. Man hat sie mit vielen anderen Ketten zur Einheit der mediterranen Ketten zusammengefasst. Die grossen Fortschritte der Tektonik aber wurden gerade dadurch erzielt, dass man sich zum Ziele setzte, — und auch weiterhin versucht — die individuellen Züge nicht nur der einzelnen grossen Segmente, sondern auch der darin unterscheidbaren Zonen herauszuarbeiten und zu vergleichen. Die eingehende geometrisch-kinematische Analyse aller Grössenordnungen dürfte auch in den alten Gebirgssegmenten weiter führer als die Primitivierung der Gesichtspunkte. Allerdings braucht es viel Geduld und eine grössere Anzahl von Methoden und Techniken, um die Knoten der Überprägungen, die sich oft vielfältig durchdringen, nach und nach zu lösen.

Unter den vielen bis jetzt bekannt gewordenen Grundgebirgstypen kann man zwei verschiedene Abteilungen unterscheiden, die, wie das in der Natur meist der Fall ist, durch Übergänge verbunden sind. In den einen kann man eine oft zwar grosse Zahl von Phasen unterscheiden; sie gehören aber mehr oder weniger derselben Folge an. In den anderen Segmenten gruppieren sich die Phasen in mehrere unterscheidbare Ereignisreihen, die nach verschiedenen Plänen abzulaufen scheinen.

Der Typus der Kareliden in Ostfinnland gehört der ersten Abteilung an. Die unterscheidbaren Phasen, von den Ablagerungen über dem Sockel bis zum jetzigen Bau bilden eine grosse Ereignisfolge. In den Svecofenniden zeichnen sich deutlich mehrere solche Ereignisfolgen ab.

Der Entwicklungstypus mit mehreren Phasengruppen rückte in den letzten Jahren deutlich in den Vordergrund des Interesses verschiedener Geologenkreise. Dies hat verschiedene Gründe: Einmal werden auf den Karten immer mehr Einzelheiten notiert, wodurch das Bild immer komplizierter wird; daraus ergibt sich die Notwendigkeit auch solche verwickelte Raumbilder zum Bewusstsein zu bringen und zu deuten. Anderseits bieten diese Forschungen wichtige Unterlagen für die Lösung grundlegender Fragen: Solange man glaubte, dass sich die verschiedenaltrigen Gebirgssegmente nur in ringartigen Bändern konzentrisch um die alten Kerne legten, war diese Frage nicht wichtig. Da aber seit langem Fälle bekannt sind, in denen jüngere Segmente ältere überschneiden, — sie wurden seinerzeit von Eduard Suess notiert — und da sich diese Fälle durch die genaueren Untersuchungen in den letzten Jahrzehnten fortwährend vermehren, werden auch die Methoden zur Auflösung solcher, durch mehrere Phasengruppen überprägter Gebirgskörper immer wichtiger. Die Notwendigkeit sie weiter zu entwickeln wird auch durch die vielen, oft paradoxen Ergebnisse der geochronologischen Bestimmungen unterstrichen.

Bereits früher haben mehrere Verfasser, von anderen Gesichtspunkten ausgehend auf die Probleme der Svecofenniden aufmerksam gemacht. Zwei Lösungen boten sich an: a) eine mit einer Gruppe von Entwicklungsphasen; und b) eine mit Überprägung älterer Strukturtypen durch jüngere (Eskola, 1957; Magnusson, 1960). Die allerdings noch recht begrenzten tektonisch-kinematischen Kenntnisse des Baues scheinen in die gleiche Richtung zu deuten wie die Überlegungen Eskola's und Magnusson's. Sie würden eine vertiefte tektonische Erforschung rechtfertigen. Könnten diese Zeilen dazu ermutigen, so wäre ihr Zweck erfüllt.

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6

AN EXAMPLE OF GRANITE OBVIOUSLY DERIVED FROM RHYOLITIC MATERIAL [1]

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INTRODUCTION

Concerning the theories on the origin of granites, the writer has in earlier papers (Marmo, 1956, 1958, 1958a, 1959) advocated the following way of thinking:

1. The potassium of the granitic rocks may be derived from sediments or may be of juvenile oiigin.

2. In former case, it may have been expelled from clay minerals during regional metamorphism. The clays are generally richer in potassium than plutonic and volcanic rocks and removal of potassium under such circumstances would produce, from the clays, rocks of granodiorite or quartz diorite composition. If potassium is concretionally enriched within the rock, microcline porphyroblasts in a matrix of quartz diorite to granodiorite composition will result (Marmo, 1956a).

3. The potassium thus expelled may reappear in the interstices of other rocks and result in granitization. This especially applies to the so called »synkinematic granites». Eskola has stated (1955, p. 129), »— — — relying upon the mapping and research of the Survey geologists, — — — there may be no synkinematic ideal granites of oiiginal bulk composition in Finland».

4. Under tectonically favourable conditions and sufficiently large free-energy-gradient levels, the expelled material may be concerntrated by hot

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waters and form granitic and aplitic veins and dykes. In extreme cases, even bodies of considerable size and of more or less definite chemical composition typical of the latekinematic microcline granites may result.

5. Pure microcline granites constitute the overwhelming majority of latekinematic granites. On the contrary, orthoclase is the dominant potash feldspar in the postkinematic granites (including the younger Alpine granites and the rapakivi granite).

6. From the above-mentioned and other observations (Marmo, 1958 and 1958a) it has furthermore been deduced that microcline may indicate a slow accumulation of granite-forming material under moderate temperature. If this accumulation is more rapid, or the temperature elevated, orthoclase will be formed instead of or in addition to microcline.

7. From the statements above, the present writer is inclined to conclude that only the orthoclase granites, or those containing microcline of inferior triclinicity (for instance some syenites), may be of truly magmatic origin, but that even for these rocks this origin may not necessarily be the only possible one.

8. Juvenile potassium occurs in potassic extrusives in amounts sufficient to result in granite composition as in recent rhyolitic lavas, in the granite porphyries of the Precambrian, *etc.*

The last mentioned rock types, however, invariably have a texture typical of extrusives with potash feldspar usually forming phenocrysts, but also occurring in the fine-grained groundmass as minute crystals. Concerning such potassic intrusives, the question arises: to what extent may such rocks be transformed by recrystallization under regional metamorphic conditions into rocks of similar chemical composition but having a granoblastic texture especially typical of granites? Furthermore, the plagioclase of the granite porphyries is usually oligoclase, which is also the case in the granites of possible magmatic origin. On the contrary, it is usually albite in true latekinematic microcline granites. Can the granite porphyry be albitized simultaneously with the recrystallization? These two questions are of utmost interest in the problem of the origin of granites.

In the following pages, an example of a granite porphyry grading into an aplitic microcline granite is described and the observations, from the point of view of the thinking epitomized above, are discussed.

PRECAMBRIAN VOLCANICS OF PIHLAJAVESI, CENTRAL FINLAND

On the geological map of Finland, sheet B 2, Tampere (1: 400 000, made by Sederholm) there is a large area of various basic rocks of volcanic origin

Table 1. Chemical composition of some rocks in the area.

1 Feldspar-biotite-porphyrite (277/49) Pihlajavesi, Riihimäki. Analyst: M. Tavela				2 Granite porphyry (51/PP/60) Pihlajavesi. Analyst: A. Heikkinen			3 Aplite (193d/PR/60), between Kotala and Piili. Analyst: A. Heikkinen		
SiO_2	62.31	Quartz	16.62	68.94	Quartz	25.11	77.71	Quartz	38.76
TiO_2	0.77	Albite	31.96	0.35	Albite	31.44	0.07	Albite	26.72
Al_2O_3	19.07	Anorthite	21.41	16.3	Anorthite	11.82	12.4	Anorthite	2.50
Fe_2O_3	1.33	Potash feldspar	17.24	1.00	Potash feldspar	20.02	0.53	Potash feldspar	28.91
FeO	2.02	Biotite	7.42	1.29	Biotite	5.18	0.42	Sphene	0.20
MnO	0.08	Sphene	1.96	0.06	Muscovite	3.80	0.02	Magnetite	0.70
MgO	0.85	Magnetite	1.86	0.93	Sphene	0.87	0.05	$Al_2O_3 . SiO_2$	1.46
CaO	4.85	Al_2O_3 (surplus)	1.02	2.66	Magnetite	1.39	0.56	Rest	0.93
Na_2O	3.80	Rest	0.09	3.70	Rest	0.72	3.15		100.18
K_2O	3.74		99.58	4.40		100.35	4.90		
P_2O_5	0.16			0.05			0.01		
H_2O+	0.50			0.57			0.28		
H_2O-	0.56			0.10			0.06		
Cr_2O_3	0.00			CO_2 0.00	An_{27}		CO_2 0.00	An_9	
S	0.04	An_{40}		100.35			F 0.02		
Total	99.58						100.18		

approx. 20 km WNW of the railway station of Haapamäki. This area was re-investigated by the present writer.

In the whole area, with the exception of its southernmost part, there are only a few outcrops and therefore the boundaries of the varieties of the volcanic rocks composing this area could not be shown on the map. The boundaries between the volcanic rocks and the embracing acid plutonic rocks have in many cases been drawn on the basis of the aeromagnetic maps prepared by the Geological Survey.

The area under consideration is composed of Precambrian extrusives ranging from uralite and plagioclase porphyrites to granite porphyries. In the marginal parts of the area, especially in the NW, they are very strongly metamorphosed and grade into varieties which can be distinguished from the fine-grained gneisses of quartz diorite composition only with difficulty. Amphibole, especially along the margins of this 15 × 7 km area, is often strongly biotitized and varieties containing plagioclase, quartz, biotite and very little or no amphibole, are common. Such varieties usually contain potash feldspar in the ground mass and then approach a granite porphyry composition.

The chemical composition of such a porphyrite is shown by Anal. 1 of Table 1.

The central part of the area containing different porphyrites is shown in Fig. 1.

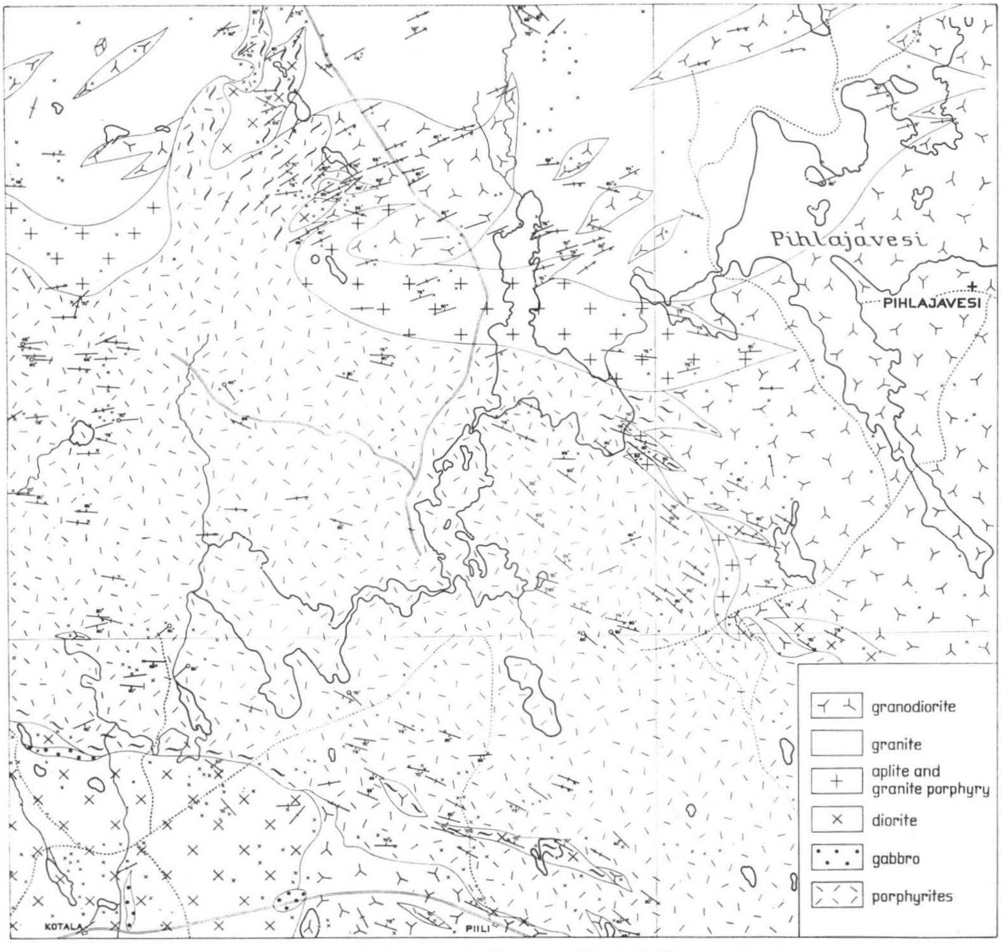

Fig. 1. Geological map of the area N of Piili. 1:120 000.

GRANITE PORPHYRY AND APLITIC GRANITE

The area occupied by the volcanic rocks is embraced by synkinematic granodiorites and quartz-diorites, except in the southwest where it terminates against a large diorite body.

Especially in the north and northeast, however, between the volcanics and porphyroblastic granites, a zone of fine- to medium-grained aplite-like rock up to 1 km broad occurs (see Fig. 1). At several places along the marginal parts of the volcanics, at the contacts with the surrounding rocks, a very similar rock occurs in the form of veins and dykes which brecciate both the feldspar porphyrite and granodiorite and also occasionally the diorite.

At such contacts, diorite veins are often present as well. These also sometimes brecciate the porphyrites as well as the granodiorite, but approx. 300 m N of the railway station of Piili, it may be seen that they are older than the veins of aplite-like rock. The aplitic pink rocks are texturally inhomogeneous, being either true granoblastic aplitic granites or showing distinct porphyritic texture with insets mainly composed of plagioclase, or to a minor extent, of microcline. These insets are embedded in a rather fine-grained groundmass composed of quartz, both feldspars, and of small amounts of micas.

Intermediate textures are often met with and because these differences are distinguishable only under the microscope, the drawing of boundary-lines between different varieties on the map was impossible.

In the Northeast, however, the broad aplitic bodies generally have the texture of a granite porphyry in those parts which are close to the volcanics, but this porphyritic texture fades out northwards where the aplite has a granoblastic texture with only occasional insets which are obviously of a more or less relictic character.

In the narrow dykes along the southern margin of the porphyrites, there brecciating or cutting other rocks, the granite porphyry-like variety mainly occurs at the apexes of the fading out veinlets. A different type of occurrence is found north of the Piili railway station, where a dyke some 15 to 20 m wide has a porphyritic texture at its northern edge, but grades southwards into granoblastic aplite (Fig. 2). At this particular locality, the aplitic veins cut the contact between plagioclase porphyrite and veined diorite which likewise has intensely brecciated the porphyrite. As seen in Fig. 2, the diorite has intruded between the plagioclase porphyry and synkinematic, slightly porphyroblastic granodiorite and close to the former contains abundant angular inclusions of plagioclase porphyrite. This breccia is cut by pink aplitic granite which in places texturally grades northwards into a granite porphyry.

At Kolmisoppi Lake in the north, volcanics of thin alternating beds of feldspar-biotite-porphyrite, plagioclase porphyrite and uralite porphyrite are in contact with granodiorite. This contact is cut at several places by dykes, up to 2 m thick, of pink, aplite-looking granite porphyry which in places grades into a granoblastic normal aplite.

Still further to the north, veinlets of both granite porphyry and aplite occur.

On the basis of the above petrographic observations, it seems that simple textural gradations from a granite porphyry to granoblastic aplite exist almost everywhere in the area. This gradation is seldom discernible with the naked eye, but it is always distinct in thin section.

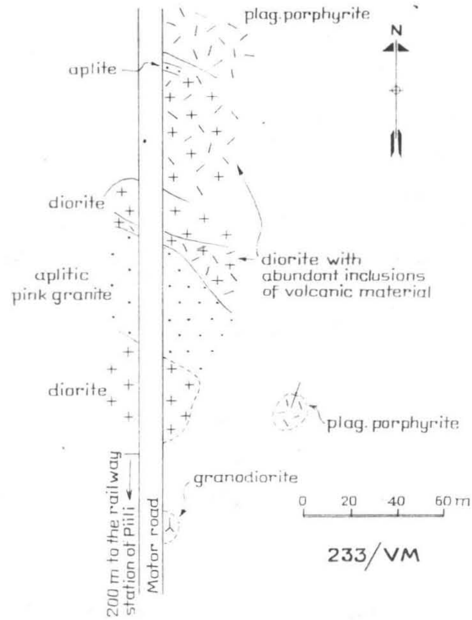

Fig. 2. Road section 300 m N of the railway station of Piili.

As will be shown on the following pages, the actual situation is much more complicated and, in addition to the textural features, there are also other factors which change with the gradation from granite porphyry to aplite.

PETROCHEMICAL ASPECTS

As mentioned on p. 139, there are varieties among the porphyrites of the area which differ from the plagioclase porphyrite by the presence of biotite in addition to or instead of amphibole. This means that biotite-bearinr portions are enriched in potassium. In other varieties, the potash feldspag is also present in the fine-grained matrix. Such varieties approach the composition of a rhyolite, and differ from the above-mentioned aplite-like granite porphyry in their gray colour and abundant biotite, which is sparse in the pink granite porphyry. Anal. 1 of Table 1 illustrates the composition of such a feldspar-biotite porphyrite. Compared with Anal. 2, which illustrates the composition of the pink, aplite-like granite porphyry, it is richer in potassium and the ratio of normative Ab to An is also increased. Consequently, the plagioclase of the granite porphyry contains less anorthite. In Anal. 3,

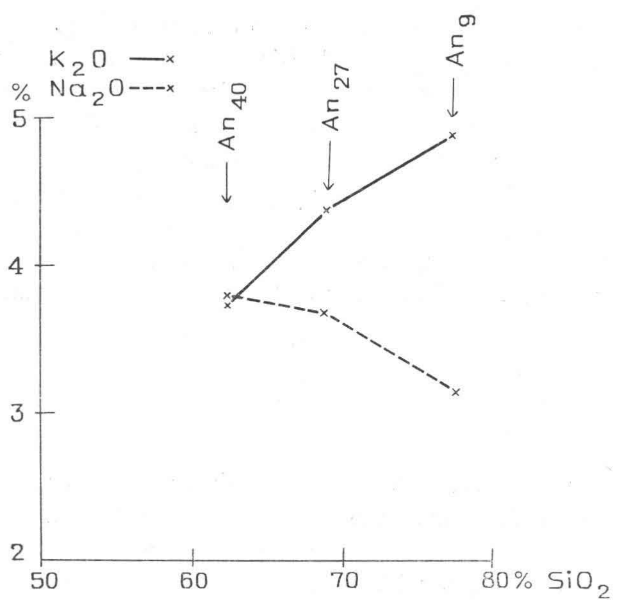

Fig. 3. Variation diagram for alkalies and silica of granite porphyries and granite aplite. For each sample analyzed, the calculated anorthite-content of plagioclase is indicated.

aplitic granite, the plagioclase is already albite with less than 10 % An, and simultaneously the ratio $Na_2O:K_2O$ is considerably higher than in the granite porphyry. From these observations it may be deduced that the gradation from a granite porphyry to an aplite coincides with the change of oligoclase into albite and with a rapid increase of the potassium content.

However, the silica-content as well has been contemporaneously increased.

The alkali-contents of the three rock types considered are shown graphically in Fig. 3. From this graph it may also be seen that the Ca-content decreases much more rapidly, than the sodium content. This decrease is also more rapid than the increase of the potassium-content. Therefore, it may be concluded from the analytical results that the gradation from granite porphyry into aplite implies a considerable increase in the amounts of silica and potassium, and a sharp descrease in the calcium-content. The actual amount of sodium present, however, need not necessarily be changed.

The gradations between the pink granite porphyry and the aplite are often irregular, megascopically hardly discernible, and occur within the boundaries of the same vein or dyke. This phenomenon can be explained only by different conditions of crystallization within the same vein or dyke during the emplacement of the material, with differentiation playing an important role during this emplacement.

The feldspar-biotite porphyrite, on the contrary, differs megascopically from the pink granite porphyry and this difference is much more marked than is the difference between the chemistries of these rocks. The potash metasomatism indicated by the graphs of Fig. 3 could still have taken place, and then the source of potassium could be one of the following alternatives:

1) The surplus potassium of the porphyrites reappeared in the granite porphyry and aplite dykes;

2) The above-mentioned potassic dykes have been derived from the same sources as the potassium which caused the potash metasomatic changes of the porphyrite.

ON THE TRICLINICITY OF MICROCLINE

The ideas proposed and used by the author for his explanations of the granite origins (p. 137) include the statement that if potash feldspar is crystallized under hydrothermal conditions and at a temperature below that at which the disordering of a triclinic lattice mainly takes place, microcline will be formed instead of orthoclase providing the accumulation of material is sufficiently slow. On the other hand, the triclinicity of microcline may not be very high if the temperature is somewhat elevated from that which is optimal for the formation of the highly triclinic microcline, or if it is deposited more rapidly than is necessary for the lattice to obtain as high triclinicity as possible. This implies:

1. If the heterogeneity of the granite porphyry-aplite veins (Figs. 4 and 5) and bodies is due to some kind of differentiation during their emplacement (p. 143), there could also be some difference between the temperatures of the formation of that part of the vein which has a granite porphyry texture and composition, and that which is composed of the later (?) crystallized granoblastic aplite. If on the other hand, the temperature is the same in both cases, the former should possibly have crystallized more rapidly than the latter.

2. Reverse differentiation (metamorphic, accompanied by the albitization of the aplitic portions) is not likely, because it is then difficult to explain the observed differences in the chemistries of granite porphyry and aplite (p. 143).

3. Therefore it seems that some differences in the triclinicities of the microcline extracted from different parts of the same vein may be observable, and in particular that the microcline of granite porphyry (either due to a slightly elevated temperature during its formation, or to a more rapid crystallization than in the aplitic portions), should have an inferior triclinicity than that extracted from the granoblastic aplite of the same vein or body.

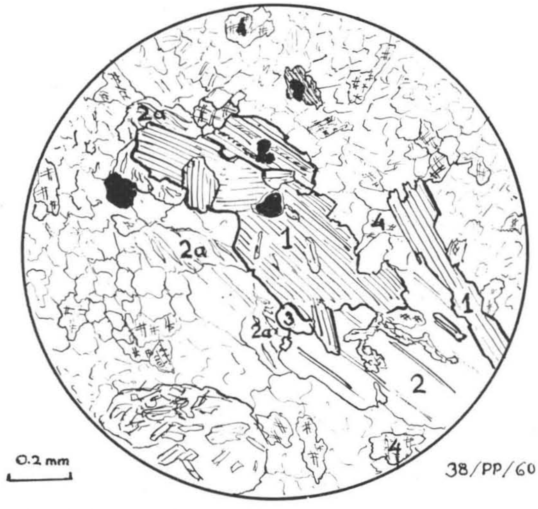

Fig. 4. Fine-grained granite porphyry on the NE. margin of the porphyrite area. A strongly muscovitized grain of plagioclase in the center of figure. 1 = muscotite, 2 = plagioclase, 2a = antiperhite, 3 = fluorite, 4 = microcline; solid black is magnetite. The matrix consists of quartz, plagioclase and microcline.

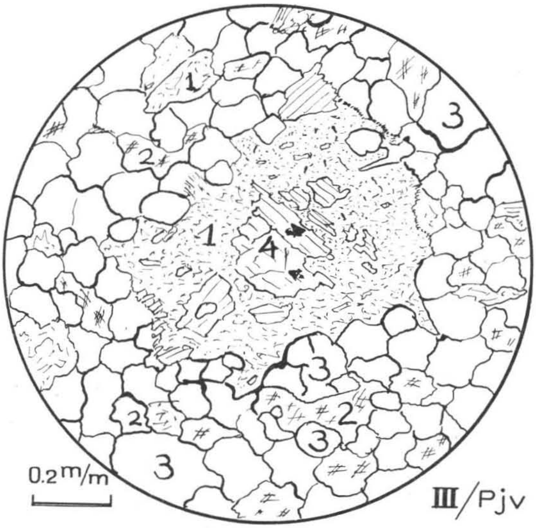

Fig. 5. An occasional inset of highly sericitized plagioclase in a granite aplite. (Kolmisoppi Lake). 1 = plagioclase, 2 = microcline, 3 = quartz, 4 = muscovite.

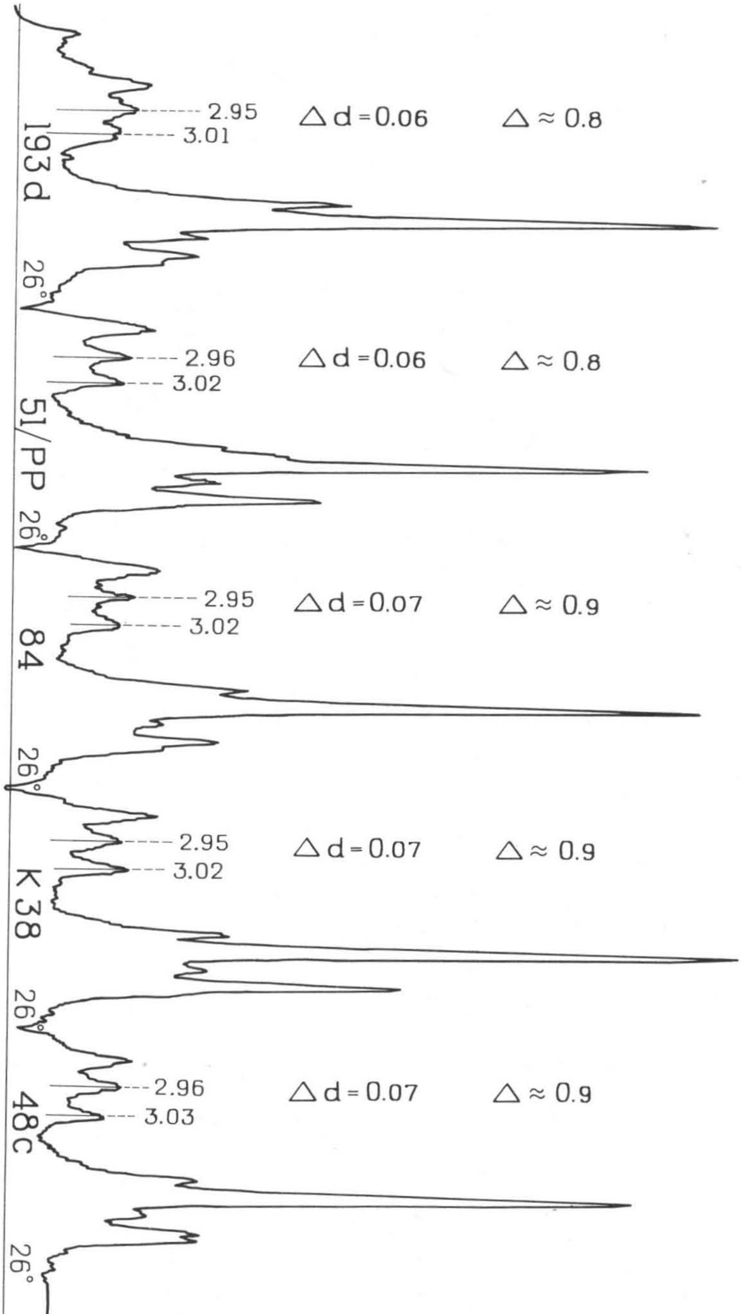

Fig. 6. Determination of the triclinicity of microclines extracted from granite porphyry (193 d, 51PP) and aplite (84, K38, 48c).

To find out whether the above assumptions have any chance of being correct, five samples were choosen for further investigations. Four of them (84, 48c and K38, 193d) are granoblastic aplites, and one (51PP) is a typical granite porphyry containing ragged insets of sericitic oligoclase in a fine-grained matrix of quartz, microcline, and plagioclase. Two of the aplites (K38 and 193d) were even-grained and contained very fresh feldspars (albite and microcline), quartz, and a little fine-scaled muscovite. Aplite 48c contained, in addition, a few highly resorbed remnants of a sericitic plagioclase. Aplite 84 contained, in addition, some biotite and few sericitic plagioclase insets, but the latter are better preserved than those in sample 48c.

In addition to microscopic study, samples 193d and 51PP were analyzed chemically. All five specimens mentioned were powdered and microcline was extracted using heavy liquids. This, as well as x-ray examination of the samples, was performed by Mr. A. Vorma of the Geological Survey of Finland. The results of this investigation are shown in Fig. 6. On the basis of this work it can be concluded that the difference in the triclinicities of granite porphyries (0.8) and granoblastic aplites (0.8 and 0.9), is not clear. This is not in accord with the expectation proposed above. All the examined samples are from different veins. Therefore the small observed scattering (0.8—0.9) is probably only of local importance. From this it may be concluded that the crystallization of both rock types has taken place under more or less similar conditions, and that the variation in chemical composition is due to small-scale differentiation, simultaneous with the earlier formed plagioclase insets accumulated in parts of the veins which thus obtained a typical granite porphyry texture. It may also be mentioned that the values obtained for the microclines of three aplites (0.9) are the same as has been earlier observed by the author as being characteristic of the majority of the microclines of the pre-Cambrian late- and synkinematic granites.

DISCUSSION

The latter of the two above-mentioned alternatives (p. 144), requires large amonts of potassium introduced from the outside and this potassium would have resulted in, in addition to the formation of the granite porphyry and aplite veins and dykes, an invasion of the volcanic rock causing very widespread biotitization and the formation of a rather homogeneous biotite-plagioclase-porphyrite. Such a process, if it took place, was not sufficiently strong to produce potash feldspar other than minute grains occurring in the groundmass of the porphyrites. This seems to be a very unlikely explanation and especially so because it appears to be very difficult to find a sufficiently large and dependable source of potassium. Therefore one would rather imagine that the material extruded was already rich in potassium and had a composition approaching that of rhyolite. Under such circumstances, there is the possibility that after the extrusion of the lava flows, the subsequent activity resulted in the formation of large quantities of water vapour which were enriched in alkalies and silica. These could bring materials rich in potassium into cracks and fissures, and into spaces available between the lava flows and country rock.

In addition to the constituents typical of a granite, these materials probably also contained some lime, magnesia and iron, and therefore would crystallize at the beginning to a granite porphyry containing large crystals of oligoclase embedded in a later crystallized matrix of albite, microcline, and quartz, and also minor biotite.

If this stage of emplacement takes place sluggishly and under strictly hydrothermal conditions, a texture like that observed in the granite porphyries (Plate I, Fig. 1) can be easily understood. The plagioclase insets, already in a solid state, would have formed during the emplacement and to some extent they may have been engulfed by the hydrothermal solutions, thus also explaining the occurrence of the zoned plagioclase insets (see Figure).

The continued accumulation of similar material would then result in an impoverishment of calcium. Therefore, if the crystallization still proceeded, the differentiation would produce solutions from which an aplite containing some scattered plagioclase insets (Plate I, Fig. 2) could crystallize with a matrix rich in silica, microcline and albite. Finally the aplite of well-developed equigranular texture will be formed. This, of course, would contain albite and not any oligoclase because all the lime would have been consumed by the granite porphyry portions of the sequence.

Thus, the apparent gradation from granite porphyry (Fig. 4; Plate I, Fig. 3) into aplite (Fig. 5) can be explained as being due to the differentiation of hydrothermal solutions during crystallization and there is no necessity for regional metamorphic recrystallization.

As matter of fact, the explanation proposed above is here adopted primarily because this gradation is not isochemical. It would be petrochemically difficult to explain how the recrystallization would albitize the oligoclase and not simultaneously produce calcium-minerals such as epidote, which are typical of rocks in which the plagioclase has definitely been albitized (Marmo, 1961). This petrochemical discrepancy, well portrayed by the chemical analyses of Table 1, can be ruled out by the acceptance of the theory of origin outlined above, and which may be epitomized as follows:

The surplus potassium derived from the extruded material of mainly rhyolitic composition was subsequently brought up under hydrothermal conditions and was emplaced in cracks and fissures opened after the consolidation of the lava flows.

The emplacement was sluggish and began with the formation of a granite porphyry with oligoclase insets.

This emplacement coincided with an impoverishment in lime and an increase in potassium and silica, which is in accordance with the chemical analyses. Gradually the emplacing lime-impoverished material could yield equigranular aplite which contains very little calcium but is conspicuously rich in potassium and silica.

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EXPLANATIONS TO THE PLATE I

Fig. 1. Granite aplite (N of Piili). Zoned plagioclase as an occasional inset. N+, magn. 15×. Thin section.

Fig. 2. Granite porphyry (At the N margin of the volcanics). A large inset of plagioclase. N+, magn. 10×. Thin section.

Fig. 3. Granite porphyry (N of Piili). N+, magn. 20×. Thin section.

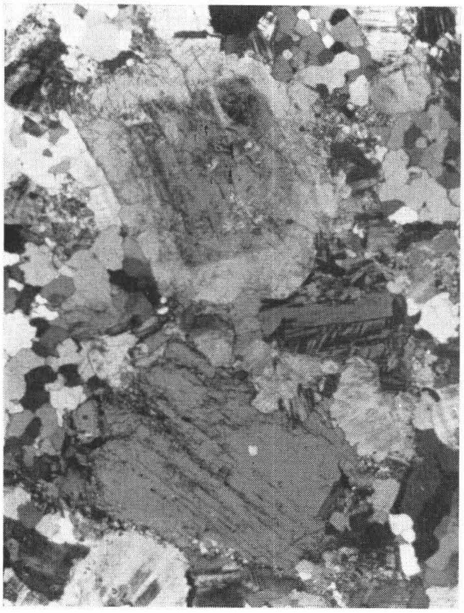

Fig. 1

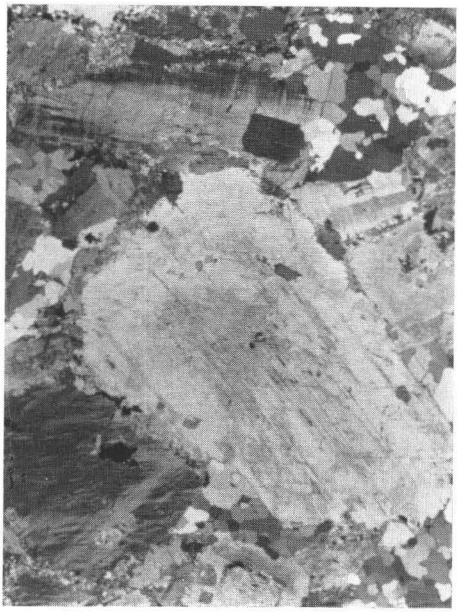

Fig. 2

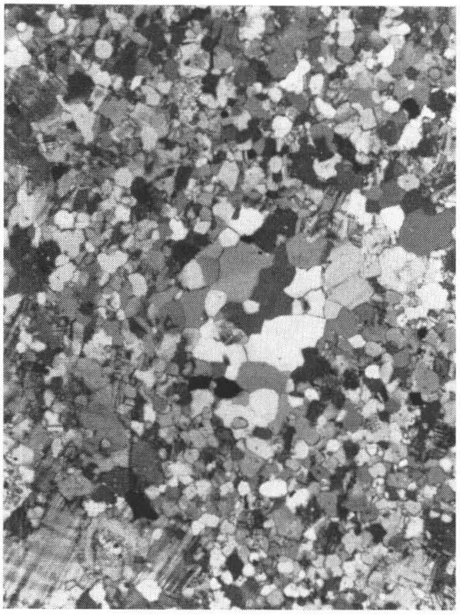

Fig. 3

V. Marmo: An Example of Granite . . .

7

THERMAL METAMORPHISM OF THE VOLCANIC ROCKS OF MT. NYIRAGONGO (EASTERN CONGO)[1]

BY

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ABSTRACT

This paper deals with the phenomena of thermal metamorphism observable in the volcanic rocks of Mt. Nyiragongo, an active volcano in the Eastern Congo, and of its two tributaries Mt. Shaheru and Mt. Baruta.

Based on their mode of occurrence, the rocks of that volcanic complex may be subdivided into a subvolcanic, an intravolcanic and an effusive group (Fig. 1), defined as follows. The subvolcanic rocks have crystallized below the pre-volcanic ground surface. The intravolcanic rocks crystallized or solidified within the cone as radial dykes, tangential cone-sheets or injected masses. The effusive rocks solidified as surface flows in contact with the open air and represent the lavas proper.

In addition to the intravolcanic rocks mentioned, the rocks in which traces of thermal metamorphism in texture and/or mineral assemblage may be detected must be included in this group. For the thermal metamorphism occurring in intravolcanic conditions the designation *ipnism* (Greek ιπνοσ, baking oven) is introduced. The products of *ipnism* or ipnitic metamorphism are designated *ipnates*. Ipnitic metamorphism is caused by a continued heat flow from the molten lava lake and from the feeding channel of the volcano.

Petrographically, the rocks of the Nyiragongo complex represent all-feldspathoidal melilitites and nephelinites, with melilite, nepheline, leucite and kalsilite as the main feldspathoids.

The ipnitic rocks of the Nyiragongo complex represent the only known rock series in which iron-rich olivines of the monticellite-kirschsteinite series are found. These kirschsteinite-rich olivines have not crystallized direct from the melt but are considered products of ipnitic recrystallization. The euhedral olivines in the rocks in which ipnitic features are lacking are calcium-free, usually very rich in the forsterite component.

In rocks in which no traces of ipnitic metamorphism are detectable, the melilite is low-birefringent, rich in magnesium (åkermanitic). In typically ipnitic rocks of the

[1] Received January 17, 1961.

area, the melilite is high-birefringent, often zoned with a low-birefringent core surrounded by a high-birefringent margin. The high birefringence is due to partial substitution of magnesium by iron, representing a ferroan variety of melilite known so far only from this area.

Texturally, the rocks of the Nyiragongo complex vary from porphyritic surface lavas with intravolcanic phenocrysts in a fine-grained effusive groundmass to holocrystalline, coarse-grained, plutonic-looking intravolcanic or even subvolcanic rocks. The textures of the ipnates grade over to those resembling granoblastic textures of metamorphic rocks.

The ipnitic metamorphism of pre-existing volcanic rocks is retrogressive in character, resulting in a mineral assemblage that is stable in a subsolidus temperature range. On increasing ipnism, however, this retrogressive thermal metamorphism changes over into a progressing one and produces a recrystallized mineral assemblage corresponding to that of the original high-temperature crystallization of the rock melt.

CONTENTS

INTRODUCTION

The crystallization of the rock melt that feeds a volcano may be subvolcanic, intravolcanic or effusive, depending on the depth at which the process of crystallization is completed. Since no sharp limits actually exist between these three stages, they may be here defined as follows. The subvolcanic crystallization proceeds at a depth below the level of the prevolcanic ground surface, *i. e.*, beneath the present cone or other products of activity accumulated on the surface. The effusive crystallization or solidification proceeds during or after extrusion in contact with the open air or, in submarine extrusions, with the sea-water. The intravolcanic crystallization or solidification proceeds at a stage intermediate between the subvolcanic and the effusive stages.

The subvolcanic rocks may be entirely plutonic or hypabyssal in character, the only volcanic feature being their genetic connection with a volcano. The effusive rocks are the lavas proper. Texturally, they may vary from volcanic glass without any crystalline particles to relatively coarse-grained holocrystalline rocks, depending on the rate of cooling after the extrusion.

Ample evidence is available indicating that, especially in the inner parts of a flowing lava mass, the rate of cooling may be sufficiently slow to produce a rather coarse-grained, in extreme cases even plutonic-looking texture. The slow rate of cooling may be either an immediate consequence of the thickness of the flow or a result of a continued heat flow from the molten part of the lava mass capable of slowing down the rate of cooling but not sufficient to prevent the crystallization.

The occurrence of rocks here called intravolcanic is not very commonly emphasized among the products of active volcanoes. An example is given, however, by Rittman (1933) from Monte Somma. According to him, the inner wall of the great caldera of this ancient volcano consists to the extent of about one fifth of dykes and sills intruded into the lavas proper. As on Monte Somma, in the main crater of the volcano Mt. Nyiragongo (Eastern Congo) and in the craters of its two great tributaries, Mt. Baruta on the northern and Mt. Shaheru on the southern flank, the share of the rocks that may be termed intravolcanic is by no means negligible. The intravolcanic rocks have not crystallized in contact with the open air. Consequently, on the outer slopes of Nyiragongo and of its tributaries, the intravolcanic rocks may be found only as blocks ejected from the crater or transported by extruding flows. In the inner wall especially of the Nyiragongo main crater, on the other hand, the intravolcanic rocks are well exposed. As has been described by Sahama and Meyer (1958), the lowering of the lava lake level, followed by a gradual collapsing of the inner walls of the upper crater, is responsible for the present-day topography of the top of the mountain. It is this process that evidently made the intravolcanic rocks accessible for study.

The intravolcanic rocks proper in the crater of Nyiragongo represent more or less steep-standing dykes radiating from the center of the crater or tangential cone-sheets that are subvertical or show a very steep dip towards the crater center. In some cases, these dykes are seen to feed surface flows of lava. On the other hand, the tuff layers occupying the lower parts of the Nyiragongo upper crater contain intercalated lenticular rock masses that evidently have been intruded into the crater wall and do not represent any surface flows.

Fig. 1 presents a summary of the kind of volcanic rocks that are found in the Nyiragongo complex. The qualitative scheme of this figure is self-explanatory and needs no further comments.

In addition to the types of intravolcanic rocks mentioned, the main crater of Mt. Nyiragongo and the craters of its two great tributaries still contain rocks of another kind that also must be included in the intravolcanic group of rocks of that volcanic complex. The original solidification and crystallization of these rocks was effusive, intravolcanic or even subvolcanic. Yet, the texture and/or the mineral assemblage shows features indicative of

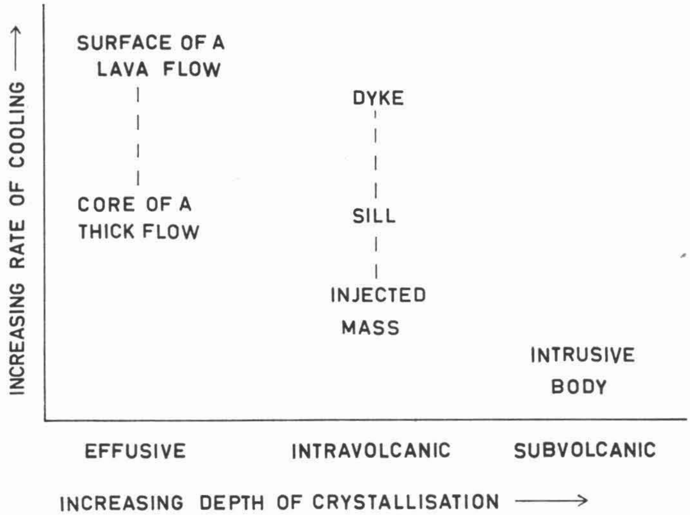

Fig. 1. Mode of occurrence as a basis for textural classification of volcanic rocks of the Nyiragongo complex.

a recrystallization without or with chemical readjustments of the constituting minerals under intravolcanic conditions. Such textural and chemical readjustments represent thermal metamorphism. Being essentially retrogressive in character and even occurring high up in the cone of the volcano, this kind of thermal metamorphism differs entirely from that found in contact aureoles around intrusive rock bodies. For this reason, it is felt that thermal metamorphism under intravolcanic conditions deserves a designation of it sown to differentiate it from ordinary thermal metamorphism. In this paper, thermal metamorphism such as it occurs on Nyiragongo will be termed *ipnism* and the rocks thus produced *ipnates* [1]. Accordingly, the *ipnates* are here defined as resulting from a baking of volcanic rocks in the immediate vicinity of a heat source like that of a permanent lava lake. In the ipnates, the ipnitic features may be merely textural or also pertain to the mineral assemblage with or without a metasomatic addition of volatiles like fluorine, chlorine, *etc.* The ipnitic features may, of course, be more or less pronounced in the rock. According to the definition given, the conception of the ipnates represents no petrographic rock name but a group designation for rocks that have been subjected to thermal metamorphism under intravolcanic conditions. As such it is analogous to the conception of the cumulates recently suggested by Wager, Brown and Wadsworth (1960) for igneous rocks that have been formed by crystal accumulation.

[1] Greek *ιπνοσ*, baking oven. This term was orally suggested by Mr. Aulis Häkli, now of the Outokumpu Co, Outokumpu, Finland.

SUMMARY OF THE PETROGRAPHY OF THE ROCKS

As has been tentatively suggested by Sahama and Meyer (1958), the lavas exposed in the inner walls of the Nyiragongo crater, above the upper platform, show the following general succession.

The lower beds of lava represent rocks very rich in melilite. In the progress report mentioned, these rocks were provisionally termed melilite-nephelinites with or without small amounts of olivine and leucite. According to the most distinctive mineral of these rocks, the phases of activity that produced the lavas were called the melilite phase and the transitional phase. Later microscopic studies, embracing a more extensive collection of specimens, have revealed the fact that, in the most typical representatives of these lava types, melilite exceeds nepheline in quantity and, accordingly, the rock should rather be termed nepheline-melilitites. A detailed description of these rock types, known so far only from the Nyiragongo complex, will be given on a later occasion.

On top of the melilite-rich lavas, there is a series of beds characterized by an abundance of leucite (leucite phase) and, finally, on top of the mountain by nepheline (nepheline phase).

Of the lava beds of the Nyiragongo inner pit, between the upper and the lower platform, the specimens available to the author for this study were but few. A microscopic examination of these specimens makes it most probable that, among the lavas exposed in the walls of the inner pit, the types ranging from melilite-bearing to melilite-rich play a very important role.

The quantitative importance of the rocks belonging to the melilite phase makes it understandable that rocks with ipnitic features in the main crater of Mt. Nyiragongo are mostly restricted to the lavas of this phase. On Mt. Shaheru and on Mt. Baruta, ipnitic rocks have so far been found only among loose blocks. It is a remarkable fact, however, that the ipnitic rocks found on these two tributary volcanoes are all more or less rich in melilite.

Most of the effusive lavas of the craters of Mt. Nyiragongo and of its two great tributaries are porphyritic or microporphyritic. In the lavas of the melilite phase, the phenocrysts of melilite usually exceed those of other constituents in quantity as well as in size. An exception to this rule is offered by the glomeroporphyritic rocks containing aggregates of phenocrysts of the light constituents nepheline, nepheline-kalsilite and, in a few instances, even leucite. As was emphasized by the author on a previous occasion (Sahama, 1960), the aggregates of phenocrysts of such glomeroporphyritic rocks have been formed through a rise of the light constituents under turbulent currents of the molten lava. Accoıdingly, those aggregates represent exceptional enrichments of the phenocrysts of these minerals. The crystallization of the phenocrysts or aggregates of phenocrysts is intravolcanic or even subvolcanic.

On the other hand, the crystallization of the groundmass evidently was effusive. The groundmass is mostly very fine-grained, in extreme cases almost cryptocrystalline. In some rocks, the size of the phenocrysts is quite variable, showing a virtually continuous gradation down to the grain size of the groundmass. In other rocks, a sharp difference exists between the size of the phenocrysts and that of the grains of the groundmass.

OCCURRENCE OF OLIVINES OF THE MONTICELLITE-KIRSCHSTEINITE SERIES

The lavas and the intravolcanic rocks belonging to the melilite-rich series quite often contain small amounts of olivine. Olivine-rich rock types are scarce. In rocks of the entire Nyiragongo complex in which no ipnitic features may be traced under the microscope, the olivine belongs to the forsterite-fayalite series and is relatively rich in the forsterite component. On the other hand, the Nyiragongo complex represents the only area in which iron-rich olivines of the monticellite-kirschsteinite series have been found (Sahama and Hytönen, 1957a and 1958). The occurrence of these calcium olivines is of especial interest in studying the ipnates of that volcanic complex.

An excellent example of the mode of occurrence of kirschsteinite in the ipnitic rocks of the Nyiragongo complex is offered by specimen S. 80 (= VM. 395), a loose boulder from the Shaheru crater.

The same rock has been found in several blocks on the NE. inner wall of the crater, close to the elephant path leading down to the crater bottom. In that spot, the wall is almost entirely covered by loose material in which the blocks occur embedded. The blocks were found in the middle and upper parts of the wall. Because the rock has so far not been found *in situ*, its mode of occurrence is not known. The bottom of the old Shaheru crater, now covered by vegetation, forms a horizontal lava plain. No sink-hole is observable. In the NNE. part of the crater, a young flow of nepheline aggregate lava came down from the Nyiragongo main cone and filled a part of the Shaheru crater bottom. It is possible that an originally existing sink-hole was filled by this flow. The relatively thick vegetation on the spot makes it difficult to prove or to disprove the existence of a sink-hole under the nepheline aggregate lava. It is possible, however, that the blocks represent material ejected from such a sink-hole. On the other hand, the possibility cannot be excluded, that the blocks are derived from a dyke cutting through the layers of the wall or even from an injected lava horizon in the wall. The rock of specimen S. 80, to be called a complex kirschsteinite-melilite-sodalite-nephelinite, is holocrystalline. The bulk chemical composition and a short summary of the constituents were given by Sahama and Meyer (1958). For that reason, only a brief account of the petrography of the rock will be given here.

The following constituents were identified: nepheline ($Ne_{68}Ks_{32}$), clinopyroxene (Sahama and Meyer, 1958), melilite, kirschsteinite (Sahama and Hytönen, 1957a), sodalite, götzenite and combeite (Sahama and Hytönen, 1957b), kalsilite (Ne_0Ks_{100}), delhayelite (Sahama and Hytönen, 1959), magnetite, perovskite, apatite, brown hornblende and pale brownish mica. Later on, the following data for magnetite have been obtained. Chemical composition calculated free from admixed insoluble clinopyroxene (analysis by Pentti Ojanperä, 1959):

SiO_2	1.27 %	MnO	1.52 %
TiO_2	19.86	MgO	1.01
Al_2O_3	3.02	CaO	0.74
Fe_2O_3	26.62	P_2O_5	0.23
FeO	45.73		

Accordingly, the molecular ratio of the magnetite component to the ulvite component $Fe_3O_4 : Fe_2TiO_4 = 2 : 3$. $a_0 = 8.485 \pm 0.002$ Å (powder pattern). Space group Fd3m (Weissenberg photographs). The single crystal photographs supported by microscopic study in reflected light on polished specimen did not reveal any exsolution texture. Sp. gr. = 4.73. In addition, small amounts of phillipsite not mentioned in the previous reports were detected.

The rock contains aggregates, up to 4—5 mm in diameter, of subhedral to euhedral nepheline phenocrysts (Fig. 1, Plate I). In the margin of the aggregates, the nepheline crystals are larger than in the core in which the interstices are filled with constituents of the mesostasis, often with phillipsite and delhayelite. The mesostasis between the aggregates shows an average grain size of 0.1 mm. The aggregates evidently are direct crystallizations from the molten lava. In the ipnates of the area, as in specimen S. 80, the aggregates of nepheline or nepheline-kalsilite are considered relictic, not affected by the more or less incomplete ipnitic recrystallization of the mesostasis. In this mesostasis, no distinct order of crystallization can be established. Especially clinopyroxene, götzenite, magnetite and nepheline show a tendency to subhedral or even euhedral development. Combeite mostly occurs as relatively large grains, up to 3—4 mm in size, detectable in the specimen by the unaided eye as cream-colored spots.

In specimen S. 80, kirschsteinite never shows the crystal forms of an olivine. The grains occurring in the groundmass are irregularly shaped, sprinkled with nepheline inclusions and bearing the appearance of small poikiloblasts. Some spots of kirschsteinite are not uniformly oriented but consist of an extremely fine-grained, highly birefringent mass in which single crystals can hardly be distinguished. Except as single grains or spots in the mesostasis, kirschsteinite quite characteristically forms coronas around the melilite crystals. In addition, the mode of occurrence of kirschsteinite as an alteration product of melilite is evidenced by the larger lath-shaped melilite crystals the grain size of which (up to 1—2 mm) clearly exceeds that of the average mesostasis. The larger melilite crystals evidently represent original phenocrysts of the lava. The relationship between the melilite and the kirschsteinite phases in these phenocrysts is illustrated in Fig. 2. Despite the fact that these phenocrysts do reveal different stages of replacement of melilite by kirschsteinite, they have usually preserved their original shape well. Optically, the still unaltered part of the melilite is of the ordinary

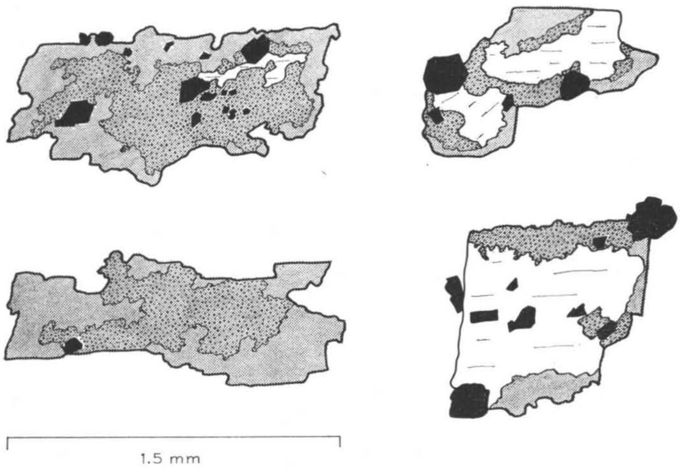

Fig. 2. Progressive replacement of melilite by kirschsteinite. Specimen S. 80 (= VM. 395). Loose block from the NE. inner wall of the Shaheru crater.

White: unaltered melilite.
Shadowed: kirschsteinite of uniform orientation.
Shadowed with dots: extremely fine-grained kirschsteinite of not uniform orientation.
Black: magnetite.

low-birefringent magnesium-rich type characteristic of the melilitites of the Nyiragongo area. Chemically, the alteration of melilite into kirschsteinite means a very strong shift in the ratio of Mg to Fe^{2+} in favor of iron. In the mesostasis of the rock, magnetite is very abundant, occurring as small well-developed crystals. In and immediately around the spots where kirschsteinite occurs, the magnetite crystals are, however, less numerous or virtually absent, indicating that they have partly been used up for forming kirschsteinite.

The microscopic observations briefly explained in the foregoing evidently indicate that, in specimen S. 80, the mineral kirschsteinite represents no direct crystallization from the melt but is formed through alteration of melilite and recrystallization of the lava already solidified. This interpretation is supported by observations made of other rocks in the area that contain olivines of the monticellite-kirschsteinite series. In thin section, an unequivocal distinction between olivines of the monticellite-kirschsteinite series and those of the forsterite-fayalite series is sometimes not easy. In most instances, however, the distinction may be made with sufficient certainty by just using approximate measurements or estimations of birefringence, of the optic axial angle and of refringence according to methods applicable to thin section work. For reference, those Nyiragongo olivines were used for which accurate X-ray, optical and chemical data were available. The results of this study may be summarized as follows.

1) No olivines of the monticellite-kirschsteinite series were found as euhedral phenocrysts clearly exceeding in size that of the average groundmass. The euhedral phenocrysts of olivine found in a number of the Nyiragongo melilitites could in all cases be proved to belong to the forsterite-fayalite series.

2) Magnesium-rich members of the monticellite-kirschsteinite series, *i. e.*, monticellites proper, have so far not been found on Nyiragongo. The olivines of this series in the specimens collected extend in composition from about the middle of the series towards its iron end.

3) The relatively iron-rich olivines of the monticellite-kirschsteinite series found in the groundmass of the Nyiragongo area rocks never show euhedral development.

4) Together with well-developed crystals of leucite, nepheline, clinopyroxene, iron-rich melilite, *etc.*, kirschsteinite-rich olivines have been found in vesicles of ipnitic blocks transported by extruding lava flows. Such blocks occur in abundance on top and on the uppermost slopes of the southern part of the Nyiragongo main cone. In these vesicles, the kirschsteinite-rich olivine often shows a good crystal form known for olivines of the forsterite-fayalite series with (110), (120), (010) and (021) as dominating faces.

OCCURRENCE OF MELILITE

In the melilitites of the Nyiragongo area in which no ipnitic features are detectable, the melilites occurring as large phenocrysts are optically negative, low-birefringent ($\omega - \varepsilon$ varying between 0.003 and 0.005). Chemically, these melilites are rich in magnesium and relatively poor in iron. A few analyses of melilite from these rocks have been published by Sahama and Meyer (1958). For comparison, an additional analysis of such a melilite is given in No. 1, Table 1.

On the other hand, it is quite a remarkable fact that the melilites found in the ipnitic rocks of the area are of a clearly different kind. These melilite crystals, especially when occurring as phenocrysts, are heavily zoned. The core of the crystals is relatively low-birefringent. Towards the margin, the interference color rises rapidly and the boundary between the low-birefringent core and the high-birefringent margin is often quite sharp. The zoning cannot be made readily visible in black-and-white photomicrographs. Therefore, a highly typical melilite crystal is illustrated in the drawing of Fig. 3. In the core of this crystal, the birefringence amounts to ca. 0.005 and in the outermost margin to ca. 0.013. The low-birefringent core is strictly euhedral and, apparently, represents an original phenocryst crystallized from the melt. The high-birefringent margin, on the other hand, is typically an-

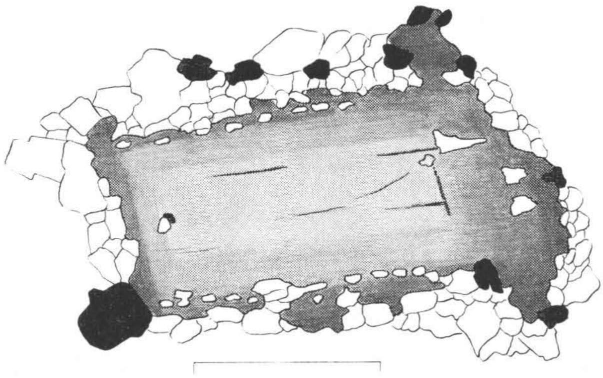

Fig. 3. Strongly zoned melilite phenocryst in an ipnitic melilite-leucite-nephelinite. Specimen R. G. 4924. Nyiragongo. Drawn by Toini Mikkola.

hedral extending into the interstices between the grains of the groundmass. It seems evident that the growth of the melilite crystal continued after the crystallization of the groundmass, enclosing some of the neighboring nepheline grains in its marginal parts. These nepheline grains occur now as inclusions in the melilite crystal. The high-birefringent melilite is also optically negative and single crystal X-ray photographs reveal the same symmetry as for the low-birefringent melilite.

The mode of occurrence of zoned melilite that is reproduced in Fig. 3 is very common in the ipnitic rocks of the Nyiragongo area. Similar zoning of melilite is found not only in large phenocrysts but also in crystals of the groundmass. As indicated by the interference color, the zoning is quite sharp in some rocks. In other specimens it is but slight, in extreme cases even lacking. In such cases, the melilite crystals look virtually homogeneous, with a high birefringence from the core to the margin. Wherever the melilite is zoned, the interference color rises from the core to the margin. A reverse zoning, a high-birefringent core surrounded by a low-birefringent margin, has not been observed. In some rocks, as in the ipnitic blocks carried by flows of nepheline aggregate lava on the southern rim of Mt. Nyiragongo, homogeneous high-birefringent melilite occurs in the vesicles together with kirschensteinite-rich crystals of olivine. These melilite crystals are elongated prisms in which the faces (100), (110), (120), (101), (111), (121) and (001) have been detected.

For chemical analysis, it is not possible to separate the high-birefringent margin from the low-birefringent core. Therefore, the cause of the high birefringence in the margin was tested by an X-ray microanalyzer at the Department of Mineralogy and Petrology of the University of Cambridge,

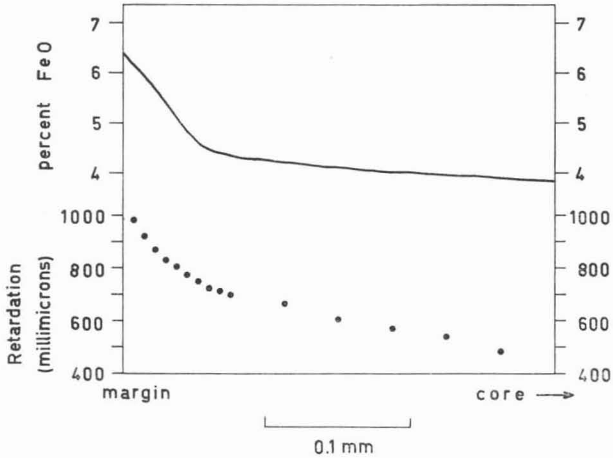

Fig. 4. Variation of optical retardation (bottom) and FeO content (top) in the zoned margin of a melilite phenocryst. Specimen VM. 362. Ipnitic block from the notch of the southern rim of the Baruta crater.

England. As a standard, a piece of fused melilite was used that, after fusion, had been chemically analyzed. The standard was mounted on the same glass as the thin section and was ground simultaneously with it. A run was made starting from the homogeneous core through the zoned margin of the crystal recording the intensity of the emitted radiation of iron. On a number of points along the same path, the optical retardation was measured with a Berek compensator. The results of these measurements are summarized in Fig. 4. As is shown by the figure, the curves indicating the variation of the optical retardation and the iron content calculated as FeO, respectively, run almost parallel to each other. In the large core, the slopes of both curves are very gentle and become steep in the strongly zoned margin. Accordingly, it seems that the increase of the birefringence in the margin of the crystal is mainly caused by an increased substitution of magnesium by iron. This conclusion is further supported by analytical data obtained on some high-birefringent Nyiragongo melilites in which the zoning is extremely weak or almost absent and, therefore, in which a chemical analysis may be expected to correspond to the high-birefringent variety of the mineral. The results of two analyses of such high-birefringent melilites are summarized in Nos. 2 and 3, Table 1. No. 2 represents a virtually homogeneous melilite in an even-grained ipnitic rock and No. 3 a homogeneous melilite occurring in vesicles of an ipnitic block carried by the nepheline aggregate lava. A glance at the table reveals the fact that the main difference between the low-birefringent and the high-birefringent melilites is in the Mg:Fe ratio. This

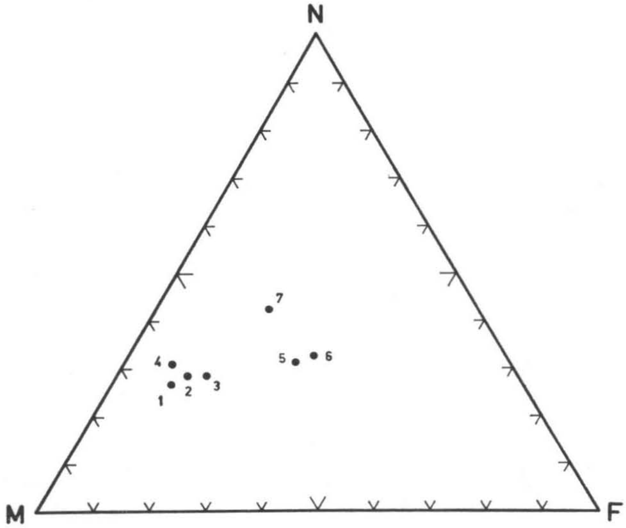

Fig. 5. Graphical presentation of the chemical composition of some analyzed Nyiragongo melilites. Calculated in molecular percentages.

Coordinates: N .. $NaCaAlSi_2O_7$ (hypothetical alkali melilite).
M .. $Ca_2MgSi_2O_7$ (åkermanite) + $Ca_2Al_2SiO_7$ (gehlenite).
F = $Ca_2FeSi_2O_7$ (iron åkermanite) + $Ca_2Fe_2SiO_7$ (iron gehlenite).

Specimens: 1 .. VM. 569. Non-ipnitic. Phenocrysts in lava from the Nyiragongo crater.
2 .. FEAE No. 83. Non-ipnitic. Phenocrysts in Kabfumu lava.
3 .. FEAE No. 93. Non-ipnitic. Phenocrysts in nepheline aggregate lava.
4 .. S. 96 = VM. 355. Non-ipnitic. Ferrian melilite. Phenocrysts in lava from Baruta.
5 .. R. G. 22778. Prismatic crystals in vesicles of an ipnitic block in nepheline aggregate lava. High-birefringent.
6 .. VM. 391. High-birefringent crystals in an ipnitic block from Shaheru.
7 .. VS. 217. Anhedral ferrian melilite in slowly cooled effusive leucite-nephelinite. Block at foot of the big talus, Nyiragongo upper platform.

difference is further illustrated in Fig. 5 in which the compositions of the melilites of Table 1 and those of the melilites previously published by Sahama and Meyer (1958) are plotted in terms of the melilite components. The high-birefringent ipnitic melilites represented by Nos. 5 and 6 in the figure clearly differ from the group of the low-birefringent melilites (Nos. 1—4).

TEXTURAL CHARACTERISTICS

As was already mentioned in the introduction of this paper, the ipnitic recrystallization found in the rocks of the Nyiragongo area may be more or less advanced, varying from rocks with completely recrystallized textures to lavas in which only slight traces of post-extrusional thermal effects are observable. The texture resulting from an ipnitic metamorphism depends, of course, on the entire thermal history of the rock, *i. e.* on the history of the original crystallization of the rock and of its ipnitic baking. If two rocks of different modes of occurrence of Fig. 1 and, consequently, of textures differing from each other are subjected to a complete recrystallization under identical conditions, they will yield similar textures. Thus, an original fine-grained porphyritic surface lava will recrystallize with a texture identical to that of an intravolcanic or even subvolcanic rock. If, as is mostly the case with the ipnates of the Nyiragongo complex, the ipnitic metamorphism is incomplete, the resulting texture of the recrystallized rock will largely depend not only on the conditions of ipnism but also on the texture of the original rock.

Fig. 6 summarizes schematically the supposed cooling histories of the volcanic rocks of different modes of origin found in the Nyiragongo complex.

The coordinates in the figure, temperature and time, are merely qualitative and need not necessarily be the same for the curves Nos. 1—4. *A* indicates the temperature range in which the crystallized phases are in equilibrium with the still molten part of the lava. In the range *B* the temperature is below the solidus of the system. Accordingly, the rock is entirely crystalline; yet, the temperature is high enough to allow recrystallization. In the range *C* the rate of crystallization is too slow to have any notable effect on the texture or mineral assemblage of the rock. Absolute values for the limits between these temperature ranges depend on the bulk composition of the rock

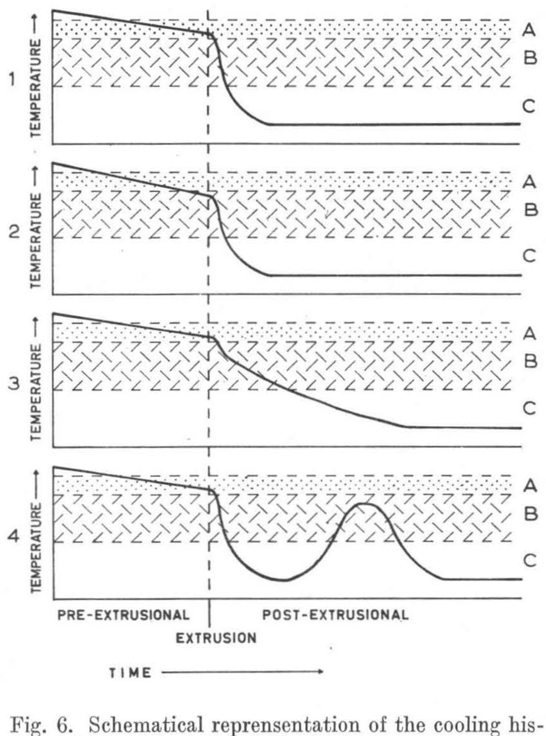

Fig. 6. Schematical reprensentation of the cooling history of volcanic rocks of the Nyiragongo complex.
1. Cooling history of the surface of a lava flow.
2. Cooling history of a lower intravolcanic or subvolcanic block.
3. Cooling history of the core of a thick flow.
4. Cooling history of an ipnitic block or rock mass.
A. Temperature range in which some crystalline phases are in equilibrium with the melt.
B. Higher subsolidus range in which recrystallization may occur.
C. Lower subsolidus range in which virtually no recrystallization can occur.

and on local contents of volatiles. The bulk chemical composition of rocks of the Nyiragongo complex mostly varies within relatively narrow limits. It may be assumed, therefore, that the effect of this variation upon the limit between *A* and *B* is relatively slight. On the other hand, the content of volatiles is known to vary quite considerably and may affect the solidus. Irrespective of the absolute temperatures involved, *A* represents the temperature range in which the phenocrysts are formed. When, on cooling, the upper limit of the range *B* (higher subsolidus range) is reached, the rock has become entirely crystalline. There may still occur crystallization caused by gases or solutions circulating in the rock. In the range *C* (lower subsolidus range), only a secondary crystallization of zeolites or calcite in the vesicles may take place. All intermediate cooling histories between those presented in curves Nos. 1—4 may actually be expected to occur. In the following, the dependence of the rock texture on the cooling history will be briefly discussed.

As is illustrated by curve No. 1 of Fig. 6, the temperature of an extruding lava, with its phenocrysts floating in the molten mass, must lie within the range *A*. The rapidity of the temperature drop after the extrusion results in a porphyritic texture with a fine-grained groundmass. Such a texture, illustrated in Fig. 1 of Plate II, is common in rocks collected on the surfaces of flows. On the other hand, if the extruding flow is relatively thick, its core will cool down considerably more slowly than its surface (curve No. 3, Fig. 6).

On the outer slopes of the mountain, only the surfaces of flows are accessible for collecting specimens. The inner walls of the Nyiragongo main crater, however, offer a good cross-section of the entire upper cone. In the following, a place on the crater wall will be described at which the core of a thick surface flow seems to be exposed for study.

In the SE part of the upper platform there is a big talus of blocks, the largest ca. 3 m in diameter. The entire wall is some 170 m high and the top of the talus more than a third of it. The wall itself consists of alternating tuff and lava layers and, on top of the wall, there is a continuous layer of nepheline aggregate lava, some 15—20 m thick. A schematical drawing of that part of the crater wall is given in Fig. 7. In that place, the crater wall is subvertical and extremely difficult of access. For that reason, specimens from the upper wall could not be collected in situ. From below it may be seen, however, that exactly above the talus there is a notch cutting through the upper tuff and lava layers. This notch is filled with material marked with dots in Fig. 7. It seems evident that the material filling the notch represents surface lava the extrusion of which is anterior to that of the nepheline aggregate flows. The blocks of the talus give an idea of the kind of rocks occurring in the notch. As has been mentioned by the author on a previous occasion (Sahama, 1960), the talus, especially on its two sides close to the crater wall, is the best locality for collecting lava with the giant leucite aggregates. The origin of this giant leucite lava has not been accurately established, but it seems most probable that the blocks come from a lava layer high up in the wall.

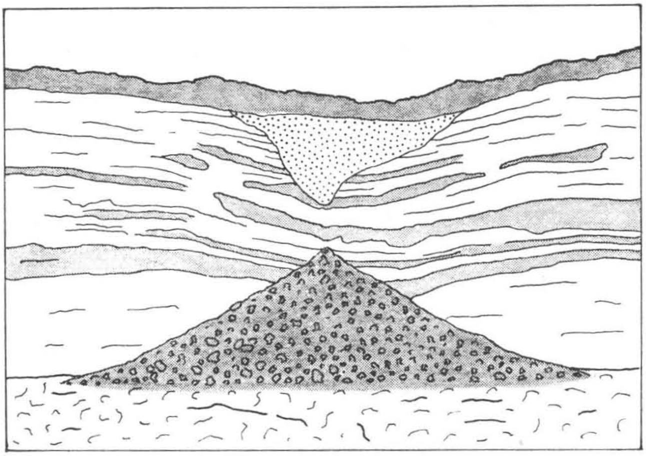

Fig. 7. Schematical drawing of a portion of the SE. inner wall of the Nyiragongo crater, above the upper platform.

The front of the talus consists mostly of very coarse-grained rocks that texturally look more like miarolitic plutonic rocks than like volcanic lavas. Fig. 3 of Plate I illustrates the texture on a polished section of a hand specimen of such a rock and Fig. 3 of Plate II gives a microscopic view of the same specimen. The coarse-grained rock is a holocrystalline leucite-nephelinite and may be termed miarolitic rather than vesicular. The numerous cavities are coated by leucite trapezohedrons, now multiply twinned, by stout nepheline prisms and by platy crystals of titanian clinopyroxene, sometimes with greenish margins. Microscopically, the texture is hypidiomorphic-granular. No trace of a fine-grained groundmass can be detected. Accordingly, the rock represents no accumulation of intravolcanic phenocrysts but a very slowly cooled lava in which no definite order of crystallization can be traced.

The only exception to this observation is offered by the ferrian melilite, which is present in small amounts (not visible in Fig. 3, Plate II) and an analysis of which is given in No. 4, Table I. The mode of occurrence of this melilite differs entirely from that of the ordinary low-birefringent melilite in the lavas of the Nyiragongo complex. This melilite never shows any traces of crystal forms but just fills the interstices left between the nepheline, leucite and the clinopyroxene crystals. It evidently represents a very late crystallization of the rock, formed at a stage when the main rock mass had crystallized and the temperature had dropped. This interpretation would explain the fact that this melilite is richer in iron and alkalies than are the low-birefringent melilites summarized in Fig. 5. The mode of crystallization also explains the observation that the iron is mainly in the ferric state.

The rocks illustrated in Figs. 1 and 3, Plate II, represent the two textural extremes among the effusive rocks of Nyiragongo between which all gradations may be found. Among the intravolcanic rocks of the main Nyiragongo crater, the textures also vary. Especially the rocks occurring as relatively narrow dykes are mostly porphyritic, with a more or less fine-grained groundmass and cannot be distinguished texturally from rapidly cooled surface lavas. Fig. 2, Plate II, gives the texture of an injected intravolcanic rock mass. The texture of this rock is entirely different from that of a typical surface lava. Some constituents may show a tendency to euhedral or subhedral development (nepheline; bottom left in the figure), whereas others appear anhedral (clinopyroxene; top right corner).

The supposed cooling history of typical ipnitic rocks is illustrated in curve No. 4, Fig. 6. The cooling history immediately after the extrusion is identical to that of a surface lava. In the ipnitic reheating of the rock, the temperature is raised from the range *C* to the range *B*. If the temperature rise extends close to the upper limit of the range *B*, the ipnitic recrystallization will mainly affect the texture of the rock and not the mineral assemblage. If, on the other hand, during the ipnitic reheating the temperature reaches only the lower part of range *B*, the mineral assemblage produced by the recrystallization can not remain unchanged but will correspond to an assemblage stable in a lower temperature range. Fig. 4, Plate II, illustrates the texture of a rock with partial ipnitic recrystallization. The main constituents, clinopyroxene, zoned melilite and nepheline, are anhedral and the entire texture resembles that of a granoblastic metamorphic rock. Another example of an ipnitic texture is illustrated in Fig. 2, Plate I. In this rock, large anhedral crystals of clinopyroxene enclose numerous euhedral nepheline prisms in random orientation. The patchy appearance of the texture caused by the poikilitic clinopyroxene crystals is plainly visible in the photomicrograph.

In principle, the cooling histories presented in curves Nos. 3 and 4, respectively, differ from each other merely in the fact that curve No. 4 shows a post-extrusional temperature minimum after which the temperature starts rising again. This temperature minimum may result in a very gently sloping curve and the cooling history passes gradually over to that of curve No. 3. For the cooling histories represented by both these curves, it is essential that the rock has been baked long enough in a temperature range sufficiently high to cause recrystallization. Accordingly, it is not always possible on the basis of a microscopic study, to tell whether the rock has been subjected to real reheating or whether the cooling of the rock has just been sufficiently slow to produce the texture and mineral assemblage found. Since no sharp boundary exists between the slowly cooled and the reheated volcanic rocks of the Nyiragongo complex, the term ipnate may be assigned to all the volcanic rocks of the area in which readjustments in texture and in mineral assemblage are evident.

Curve No. 2 of Fig. 6 gives the supposed cooling history of a lower intravolcanic or subvolcanic block that has been rapidly transported to the surface by an extruding lava mass. The shape of this curve is virtually identical to that of No. 1 with the only difference that, already prior to extrusion the temperature of the rock must have passed the range A and must lie within the range *B*. In other words, the temperature must have reached the subsolidus range before extrusion. The texture of the rock will be plutonic in character and may resemble that of Fig. 3, Plate II, the origin of which, however, was different.

MINERAL ASSEMBLAGE

It was known already to Lacroix (1933) that some of the ejected blocks from the top of Mt. Nyiragongo have a bulk chemical composition intermediate between the families of missourite and ijolite. These rocks, classified by him as the family of niligongite, contain in their typical development titanian clinopyroxene, nepheline, leucite, magnesian melilite and, as accessories, titanian magnetite, apatite and often perovskite. Texturally, the rocks are relatively coarse-grained, hypidiomorphic-granular or doleritic. Later on, the conception of the family of niligongite was discussed and slightly redefined by Denayer (1956).

Both chemically and mineralogically, the niligongites proper of Mt. Nyiragongo correspond fairly well to the nepheline aggregate lavas of the mountain. The leucite content is sometimes slightly higher in the niligongites, but this difference seems hardly to be more pronounced than are the variations in the leucite content found among the lavas of the nepheline aggregate flows. The main difference between the niligongites and the nepheline aggregate lavas are textural. The textures of typical niligongites indicate a slow crystallization at relatively high temperatures (No. 2, Fig. 6). Occurring as angular blocks transported by extruding lavas, the niligongites evidently crystallized prior to extrusion and represent a lower intravolcanic or even subvolcanic analogue to the effusive nepheline aggregate lavas. Accordingly, the niligongite blocks may be interpreted as originating from a crystallized coating grown on the walls of the feeding channel of that volcano. As such they may be taken as direct crystallizations from the rock melt.

The genesis of rock types corresponding to the niligongite blocks may, however, be interpreted in an entirely different way, as illustrated in the scheme of Fig. 8. Originally the rocks may have been holocrystalline intra-

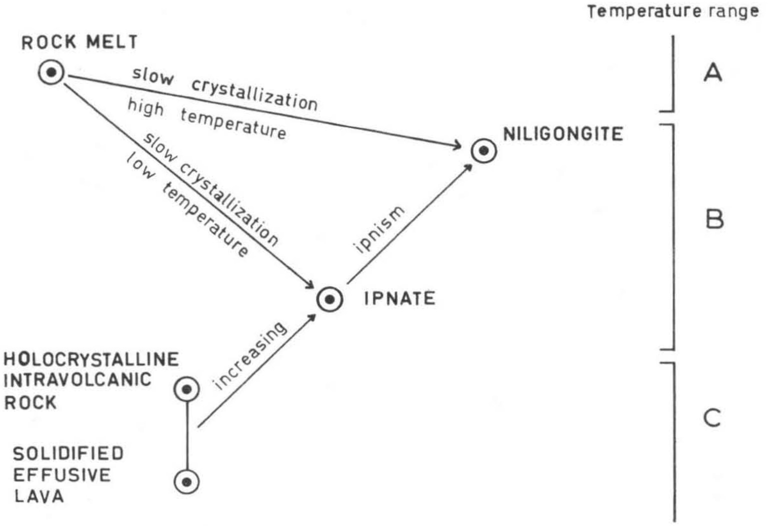

Fig. 8. Scheme of processes yielding ipnitic mineral assemblage.

volcanic rocks injected into the strata of the mountain or solidified effusive lavas buried by later flows and occurring now in the lower horizons of the cone. If such rocks are subjected to ipnitic reheating and recrystallization at a temperature not far from the upper limit of range *B* in Fig. 6 and if that ipnitic metamorphism is carried on till no relictic features in textures and mineral assemblage are left, the rocks will not be distinguishable from products of direct slow crystallization of the melt.

Accordingly, a niligongitic rock may be taken as an end product of one of two different processes, *viz.* a slow crystallization of a rock melt or a strong ipnitic recrystallization of a pre-existing volcanic rock. Both processes are possible and may well have been effective on Nyiragongo.

The rocks called ipnates in this paper have been marked by a single point in the scheme of Fig. 8. This is done simply for clarity. In reality, the ipnates, such as they are found in the Nyiragongo complex, represent a series of rocks of progressing ipnitic metamorphism and, accordingly, should have been marked by a series of points along the line of the increasing ipnism. As is further indicated in the figure, the slow crystallization of a rock melt occurring at low temperatures will result in a mineral assemblage similar to that of reheated and recrystallized rocks. Such a slow crystallization or recrystallization at low temperatures may be caused by the presence of volatiles and may occur in the core of a thick flow or in an intravolcanic mass.

As was emphasized in the case of the niligongites proper, it seems that also the ipnates may have been formed in two different ways. Which of these two ways better fits the observed relictic features and mode of occurrence must be studied separately for each particular rock.

Starting from a pre-existing volcanic rock with a mineral assemblage originally formed through crystallization of the melt at high temperatures, the ipnitic metamorphism means passing over into a new mineral assemblage that is stable in a lower temperature range. The metamorphism is, accordingly, retrogressive in character. If, however, on increasing ipnism, the rock is subjected to baking at a higher temperature, the mineral assemblage will be adjusted to correspond to that new, higher temperature range. The direction of metamorphism will change from a previously retrogressive into a new, progressing one. In intravolcanic conditions, as on Nyiragongo, in which temperature gradients in the rock masses may be very steep and irregular locally and, in addition, in which the temperature and time of baking may often be insufficient, the equilibrium will apparently mostly not be attained. For that reason, the mineral assemblages actually found in the ipnitic rocks of the Nyiragongo complex represent a continuous series that, both in a retrogressive and in a progressing direction, may be expected to grade over to the mineral assemblages of the rocks with no ipnism. Some of the constituents typical of the Nyiragongo volcanic suite do not seem to be notably affected chemically by the ipnitic metamorphism whereas others evidently are highly sensitive to the conditions of crystallization or recrystallization.

For the three iron- and magnesium-bearing constituents, clinopyroxene, melilite and olivine, the difference in composition between the ipnitic and the non-ipnitic rocks may be summarized as follows.

Non-ipnitic	Ipnitic
Titanian augite	Salite
Åkermanitic melilite	Ferroan melilite
Forsteritic olivine	Kirschsteinitic olivine

Clinopyroxene is an exceedingly abundant constituent in the rocks of the Nyiragongo area, occurring both as phenocrysts and in the goundmass. From the fine-grained groundmass, the mineral is usually quite difficult to isolate for chemical analysis. Therefore, most of the analyses published by Sahama and Meyer (1958) have been made of the mineral that occurs as large phenocrysts in the rock. According to the analytical data available of such clinopyroxenes of Nyiragongo lavas with no traces of ipnism, the mineral represents a titanian variety, rich in aluminium, and may be called titanian augite. This mineral is mostly zoned and often shows a beautifully developed hour-glass texture. In the margin of the phenocrysts and in the groundmass, the brownish violet color of the titanian augite often passes over into greenish. It is not the purpose of the present paper to enter any further into a discussion of the variation of the composition of clinopyroxene in Nyiragongo lavas. It may only be stated that, apparently, the titanian augite represents the kind of clinopyroxene that crystallizes from the melt of these rocks at highest temperatures. This conclusion is further supported by the fact that the phenocrysts of titanian augite show traces of incipient resorption and alteration into another clinopyroxene optically different from the titanian variety of the mineral. Therefore, the titanian augite may be taken as the representative of the pyroxene family that is most typical of the mineral assemblage of high-temperature volcanic rocks of the Nyiragongo complex.

On the other hand, only one chemical analysis of ipnitic clinopyroxene is available at present, *viz.* that of the clinopyroxene of specimen S. 80 published by Sahama and Meyer (1958). This clinopyroxene represents a salite, free from aluminium and poor in alkalies. The mineral is green in color. Similar green clinopyroxenes seem to be common in the ipnitic rocks of that area. Therefore, it may be justified to consider salite a representative of the clinopyroxene family in the mineral assemblage of the Nyiragongo ipnitic rocks.

With respect to melilite and olivine, the difference between the ipnates and the rocks of the Nyiragongo complex without traces of ipnism are quite striking. In contrast to the low-birefringent åkermanitic melilite and the calcium-free forsteritic olivine of the rocks of non-ipnitic character, the high-birefringent ferroan melilite and the kirschsteinitic calcium olivine are characteristic of the mineral assemblage of the ipnates.

The exsolution of a highly potassian solid solution of nepheline and kalsilite is so rapid that it occurs even in surfaces of lava flows and is post-extrusional. Accordingly, the assemblage nepheline+kalsilite in the groundmass of the rocks is not indicative of ipnitic recrystallization. In large phenocrysts, however, the perthite-like exsolution texture with narrow nepheline lamellae in a kalsilite base may be considered indicating an absence of ipnitic baking. On the other hand, a homogeneous kalsilite core surrounded by a co-axial homogeneous nepheline margin, interpreted as the final stage of the exsolution process (Sahama, 1960), evidently requires a certain ipnitic baking and may be considered distinctive for the ipnates. It seems, however, that the minimum ipnitic baking necessary for the completion of the exsolution is too weak to produce a mineral assemblage deviating from that of ordinary lavas.

Of the silicates containing the volatiles fluorine, chlorine and the sulfate anion, the minerals sodalite, götzenite, combeite and delhayelite have been found on Nyiragongo. Of these minerals, sodalite and götzenite are relatively common, combeite has been found in a couple of specimens and delhayelite only once so far. Götzenite is a typical constituent of an ipnitic mineral assemblage. Sodalite occurs in rocks of doubtless ipnitic character. Whether it can also belong to the mineral assemblage of the lavas or intravolcanic rocks that have not been subjected to ipnitic baking, is not definitely clear.

Leucite, apatite, perovskite and fluorite do not seem to be chemically affected by ipnitic metamorphism.

RETROSPECT

Despite the doubts raised against the existence of $NaCaAlSi_2O_7$ as a stable compound in pure synthetic form, this formula was adopted for the hypothetical alkali melilite end member of the Nyiragongo melilites. Following Neuvonen (1955), that hypothetical end member was found useful in presenting diagrammatically the compositions of the Nyiragongo melilites. The average molecular ratio of this alkali melilite component to the åkermanite component in the low-birefringent melilites of the area is roughly 1: 2. The content of the iron åkermanite and the iron gehlenite component is relatively small. In contrast to pure synthetic åkermanite, these Nyiragongo melilites are optically negative with a slightly lower birefringence.

The high-birefringent melilites of the ipnitic Nyiragongo rocks show a content of the alkali melilite component roughly similar to that of the low-birefringent melilites of the area. Like pure synthetic $Ca_2FeSi_2O_7$, the Nyiragongo ferroan melilites are optically negative. Compared with the optical properties of the synthetic $Ca_2MgSi_2O_7$—$Ca_2FeSi_2O_7$ series summarized by, *e. g.*, Schairer and Osborn (1950), the refractive indices of the Nyiragongo ferroan melilites are considerably lower and the birefringence is higher.

The iron end members of the two most typical ipnitic constituents of the Nyiragongo volcanic rocks, kirschsteinitic olivine and ferroan melilite, are both known from artificial slags. On Nyiragongo, these minerals evidently are formed at the expense of the åkermanitic melilite. Stoichiometric reaction equations illustrating the breakdown of the åkermanitic melilite and the formation of the kirschsteinitic olivine and the ferroan melilite could be written in several ways. The correctness of no such reaction equation can, however, be proved in the complex system offered by these rocks. The reaction indicates a strong shift in the ratio of Mg to Fe in favor of iron, the required addition of which may occur at the expense of iron oxides (titanian magnetite, iron oxide dust present in many lavas) or of the dark green volcanic glass (sideromelane) of the groundmass. In the formation of the kirschsteinitic olivine, the original alkali melilite component will yield nepheline, as is illustrated by the intimate intergrowth of nepheline and kirschsteinite in the coronas around the partly altered melilite phenocrysts of specimen S. 80 from Shaheru.

In this connection, it is of interest to note that, according to Bowen, Schairer and Posnjak (1933), pure åkermanite is stable only above 1 325°C and pure iron åkermanite only below 775°C. According to the preliminary information published by Schairer and Osborn (1950), on the other hand, the system åkermanite-iron åkermanite is not binary at its iron end but yields on crystallization 'olivine' in addition to iron-rich melilite. Before more data on this system, especially in the subsolidus range, become available, the chemical reactions involved in the ipnitic metamorphism of the Nyiragongo rocks cannot be discussed in any more detail.

The occurrence of sodalite, götzenite, combeite and delhayelite is evidently autopneumatolytic. Autopneumatolysis, the importance of which in the volcano-petrological phenomena has been emphasized especially by Rittmann (1960), may occur in connection with ipnitic baking. Whether or not an autopneumatolytic addition of iron occurred in connection with the ipnitic metamorphism of the Nyiragongo rocks, cannot be judged on the basis of the material available at present.

Because of the very difficult accessibility of the inner pit of the Nyiragongo crater, the main part of the specimens available so far for study has been collected in the upper crater, above the upper platform. For that reason, it is quite understandable that the rocks with more or less pronounced ipnitic features have been found mostly among blocks ejected from the crater or transported by extruding lava masses. Ipnitic rocks occurring *in situ* would presumably be exposed mainly at the bottom of the inner pit, close to the present-day lava lake. The lava lake has been visited by man only twice, *viz.* in 1958 and in 1959. Especially on the later occasion, undertaken by the expedition of the Belgian National Center for Volcanology (Centre National de Volcanologie), an extensive geophysical and topographical survey of the entire crater was carried out (Evrard, 1960). The collections made by that expedition and by future expeditions may further illuminate the phenomena described and discussed in this paper.

Among the ejected blocks found in many volcanoes, Lacroix (1930) distinguished a certain type which he called *blocs réchauffés*. According to his definition, this designation applies to blocks of irregular form that have been transported and ejected in an entirely solid state as well as heated up during the transport. In the nomenclature used by Lacroix, the *blocs réchauffés* are set up in contrast to the bombs proper, particularly the angular *bombes en groûte de pain*, the form of which indicates their having been ajected in a highly viscous (not solid) state, preventing the attainment of a rounded form during the flight through the air. Consequently, the designation *blocs réchauffés* does not apply to the mineral assemblage of the rock and, as such, represents no analogue to the petrographie conception of an ipnitic block.

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The generous assistance of all these institutions and individuals is gratefully acknowledged.

Table 1. Chemical composition of some melilites from the Nyiragongo area. Analyst: Pentti Ojanperä.

	1	2	3	4
SiO_2	42.60 %	41.42 %	42.04 %	42.21 %
TiO_2	0.15	0.11	0.06	0.07
Al_2O_3	6.02	6.79	7.03	8.37
Fe_2O_3	1.17	0.41	0.00	8.38
FeO	2.18	8.45	8.03	0.92
MnO	0.11	0.22	0.22	0.20
MgO	8.95	4.79	5.33	4.69
CaO	35.04	33.59	32.92	29.98
Na_2O	2.85	3.45	3.29	3.75
K_2O	0.31	0.37	0.50	1.08
P_2O_5	0.24	n. d.	n. d.	n. d.
CO_2	0.00	0.00	0.00	n. d.
H_2O+	0.13	0.31	0.30	0.30
H_2O-	0.04	0.09	0.02	0.02
Total	99.79	100.00	99.74	99.99[1]
ω	1.637	1.646	1.647	1.672
ε	1.634	1.636	1.636	1.661

[1] Including 0.02 % Li_2O.

1. Melilite occurring as large phenocrysts in a porphyritic nepheline-kalsilite-melilitite (melilitophyre). Specimen VM. 569. Southern inner wall of the Nyiragongo crater, ca. 85 m above the upper platform.
2. Melilite from ipnitic melilite-leucite-nephelinite. Specimen VM. 391. Loose block from the eastern inner wall of the Shaheru crater.
3. Melilite occurring in vesicles of melilite-leucite-nephelinite. Specimen R. G. 22778. Southern rim of the Nyiragongo crater.
4. Anhedral ferrian melilite in slowly cooled effusive leucite-nephelinite. Specimen VS. 217. Block on foot of the big talus, Nyiragongo upper platform.

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EXPLANATIONS TO THE PLATE I

Fig. 1. Complex kirschsteinite-melilite-sodalite-nephelinite with aggregates of nepheline phenocrysts. Specimen S. 80 = VM. 395. Loose block from the NE. inner wall of the Shaheru crater. No analyzer. Magnification 7 ×.

Fig. 2. Complex melilite-sodalite-nephelinite. Large poikilitic crystals of anhedral clinopyroxene causing a patchy appearance of the rock. Specimen VM. 394. Loose block on the E. inner wall of the Shaheru crater. Nicols crossed. Magnification 3 ×.

Fig. 3. Polished surface of a holocrystalline leucite-nephelinite. The same specimen as in Fig. 3, Plate II. Magnification 2 ×.

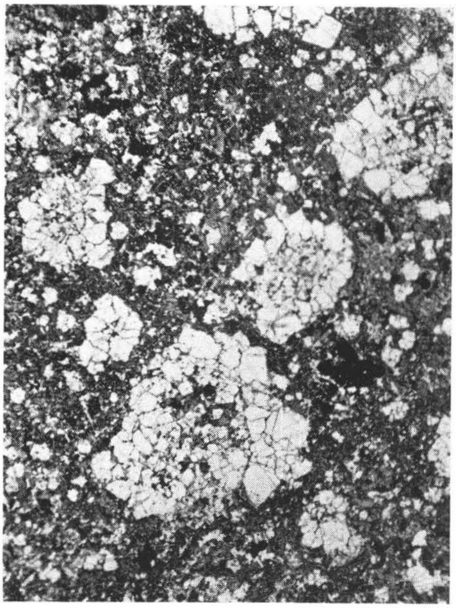

Fig. 1

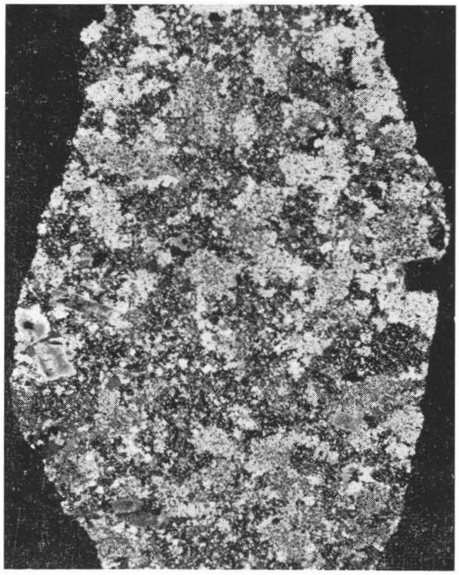

Fig. 2

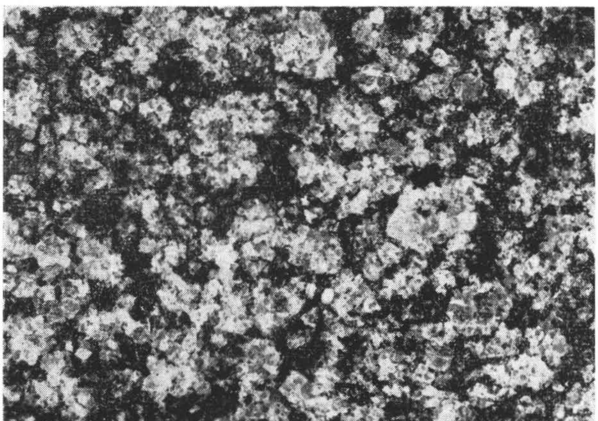

Fig. 3

Th. G. Sahama: Thermal Metamorphism of the Volcanic Rocks . . .

EXPLANATIONS TO THE PLATE II

Fig. 1. Surface lava. Porphyritic nepheline-kalsilite-melilitite (melilitophyre) with large melilite phenocrysts and small euhedral crystals of melilite and nepheline (kalsilite) in a very fine-grained, dark mesostatis. Specimen VM. 569. Southern inner wall of the Nyiragongo crater, ca. 85 m above the upper platform.

Fig. 2. Injected intravolcanic rock mass. Holocrystalline, even-grained nepheline-melilitite. Specimen VM. 208. Southern inner wall of the Nyiragongo crater, just above the upper platform.

Fig. 3. Core of a thick flow. Holocrystalline, coarse-grained leucite-nephelinite with ferrian melilite (not shown in the figure). Specimen VS. 217. Fallen block from the foot of the big talus, SE upper platform of the Nyiragongo crater.

Fig. 4. Partly recrystallized ipnitic block. Holocrystalline, even-grained melilite-nephelinite. Specimen VM. 362. Southern rim of the Baruta crater.

All photomicrographs taken without analyzer. Magnification 30 ×.

Fig. 1

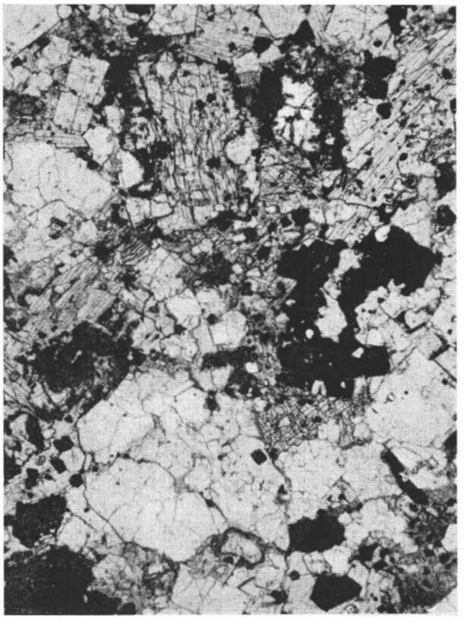

Fig. 2

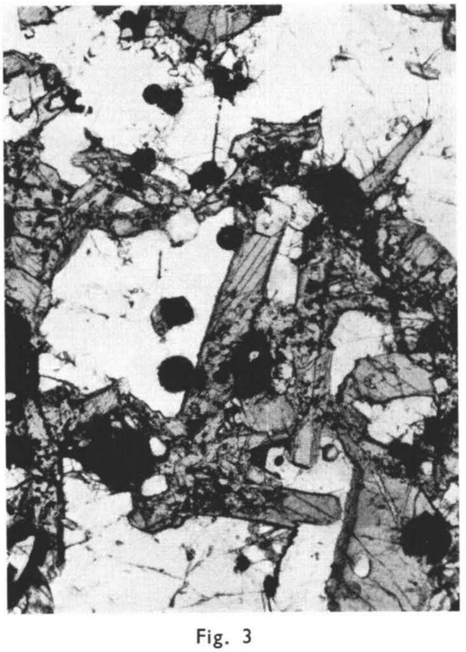

Fig. 3

Fig. 4

Th. G. Sahama: Thermal Metamorphism of the Volcanic Rocks . . .

8

CONTACT MICRORADIOGRAPHIC DISTINCTION BETWEEN NEPHELINE AND KALSILITE IN THIN SECTION [1]

BY

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ABSTRACT

A contact microradiographic method using vanadium radiation to distinguish between nepheline and kalsilite in thin section is described. The method was applied to the volcanic rocks of Mt. Nyiragongo, Eastern Congo. For phenocrysts containing the nepheline and kalsilite phases in co-axial orientation, the method is illustrated by the figures of Plate I. In the fine-grained groundmass of the rocks, the distinction between the two minerals must be made by comparing the thin section with the microradiographic film under the microscope.

As has become evident through the general reconnaissance of the phase relations in the $NaAlSiO_4$—$KAlSiO_4$ system by Tuttle and Smith (1958), the nepheline-kalsilite solid solution series shows a large gap in the subsolidus region. The very existence of a two-phase area in the phase diagram explains the joint occurrence of nepheline and kalsilite as separate phases in a rock. So far, the only rocks in which these two minerals actually have been found to occur together as common rock-making constituents are the volcanic rocks of Mt. Nyiragongo in Eastern Congo. As has been described by the author on a previous occasion (Sahama, 1960), the rocks of that volcano contain kalsilite partly as complex phenocrysts in co-axial orientation with the nepheline phase and, on the other hand, as grains separate from nepheline in the fine-grained groundmass.

The optical properties of nepheline and kalsilite determinable in thin section are quite similar to each other. Both minerals are optically uniaxial

[1] Received December 19, 1960.

negative. The refractive indices and the birefringence of kalsilite are but very slightly different from those of nepheline. These differences are, however, still sufficient for distinction between the two phases, provided that both phases occur in a complex phenocryst in co-axial orientation. In polished sections in reflected light, the distinction between the nepheline and the kalsilite phases of a complex crystal may be made by etching with alcoholic sulfuric acid (Trojer, 1956). If, on the other hand, nepheline and kalsilite occur as separate grains in the groundmass of the rock, a distinction between them is not possible by optical means. If the rock contains other easily etched constituents, the etching may not be suitable for identification. The mere presence or absence of kalsilite in addition to nepheline may, of course, be tested by powder pattern, preferably to be prepared of the nepheline fraction extracted by heavy liquids from the rock. If the mode of occurrence of kalsilite in the groundmass of the rock is to be studied, methods like X-ray microscopy must be used.

For the purpose, contact microradiography was used. A thin section of the rock to be studied was ground to a thickness of 0.01—0.02 mm. After the Canada balsam had been dissolved in xylol, the loose rock plate was placed under a binocular microscope. A suitable part of the plate was cut off with a razor blade and placed in a microradiographic camera. A photographic plate ('Maximum Resolution' plates by Messrs. Kodak Ltd, London) was inserted immediately in contact with the rock plate and the lid of the camera replaced. Preliminary tests made with these photographic plates showed that the microradiograms obtained were mostly not sharp. It was found that, in the process of dissolving the Canada balsam and loosening and drying the thin rock plate, this rock plate usually becomes slightly curved. When placed in the camera, the curved rock plate does not make good contact with the photographic emulsion and renders a more or less diffuse microradiographic image. For that reason, the plates in which the photographic emulsion is mounted on a rigid glass plate were abandoned and, instead, films were used. These films, a small batch of which was made by Messrs. Kodak Ltd, contained the same extremely fine-grained emulsion as the 'Maximum Resolution' plates, but mounted on a thin film basis. The front face of the camera lid was machined very slightly curved. When replaced, it presses the flexible film and the thin rock plate into a slightly curved form in which the contact between the rock plate and the film is good. On using this arrangement, the microradiographic images obtained were sharp. On using low power, the microradiographic film could be studied under the microscope like an ordinary thin section.

The exposures were made using vanadium radiation. Because X-ray tubes with a vanadium target were not aavilable to the author, a tube with a copper target was used. A sheet of metallic vanadium was placed in the path of the copper radiation. The secondary vanadium radiation emitted entered the microradiographic camera. The soft vanadium radiation is selectively absorbed by calcium and potassium. Because nepheline and kalsilite do not contain any large amounts of calcium, the difference in blackening on the film is caused only by a difference in the potassium content. Being considerably richer in potassium, a kalsilite grain appears white on the microradiogram whereas a nepheline grain appears gray.

Figs. 1 and 2 of Plate I illustrate the method applied to crystals with nepheline and kalsilite phases in co-axial orientation. Figs. 1A and 2A represent ordinary optical photomicrographs taken in polarized light between crossed nicols and Figs. 1B and 2B photomicrographs taken of the microradiographic films. The difference in absorption between the two phases is evident.

The mode of occurrence of kalsilite in the fine-grained nepheline-bearing groundmass of a volcanic rock was studied in the following way. The thin rock plate of which the contact microradiogram had been prepared was mounted on glass in Canada balsam, covered with cover glass like an ordinary thin section and placed under the microscope. The microradiographic film was placed on top of the thin section in such a position that it exactly matched the rock plate. The microscope was first focussed on the rock plate and a grain was sought that, in respect to its optical properties, represented either nepheline or kalsilite. The microscope was now focussed on the microradiographic film and the blackening of the grain observed. If the grain appeared white on the film, it evidently represented kalsilite. If, on the other hand, it appeared gray, the grain represented nepheline. Because the rocks of Mt. Nyiragongo also contain other potassium or calcium minerals, like leucite, melilite, clinopyroxene, *etc.*, that cannot be easily distinguished from kalsilite on microradiographic film, it is necessary to compare the thin section with the film.

Using the contact microradiographic method explained in the foregoing, a number of rocks from the Nyiragongo complex have been studied with respect to the mode of occurrence of kalsilite in the groundmass. The method is relatively time-consuming and tedious, but it has proved to be successful for the purpose. Kalsilite seems to prefer crystallizing entirely anhedrally in the interstices between other constituents of the groundmass.

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EXPLANATIONS TO THE PLATE I

Fig. 1. Complex phenocryst with kalsilite kernel surrounded by nepheline margin. Specimen S. 88 = VM. 358. Baruta. Magnification 60 ×.

Fig. 2. Complex nepheline-kalsilite phenocrysts with perthite-like eksolution texture showing nepheline lamellae in a kalsilite base. Specimen FEAE No. 83. Kabfumu lava. Magnification 50 ×.

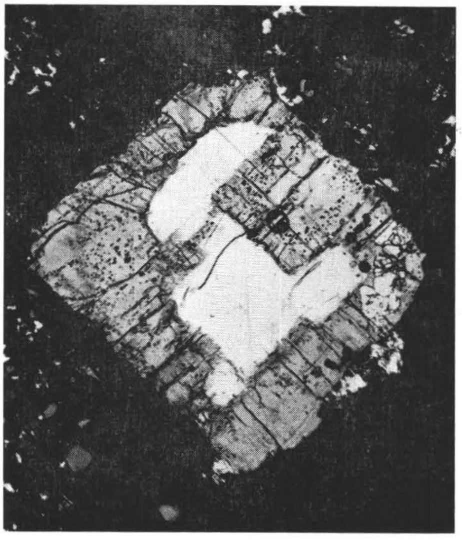

Fig. 1 A. Optical photomicrograph.

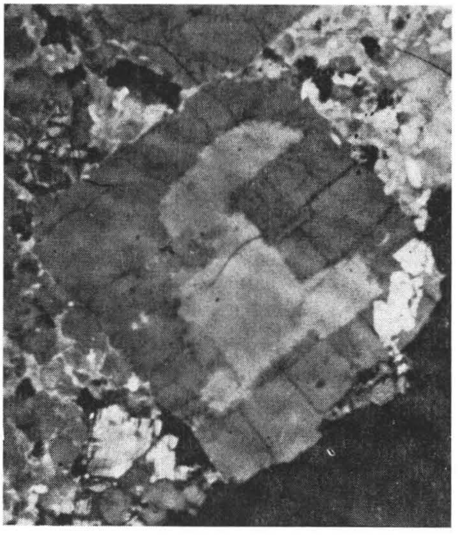

Fig. 1 B. Contact microradiogram.

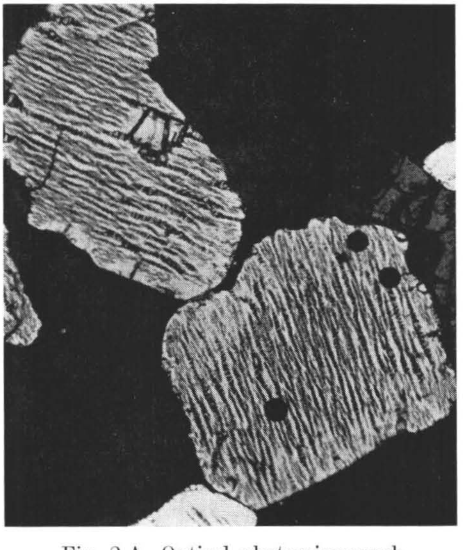

Fig. 2 A. Optical photomicrograph.

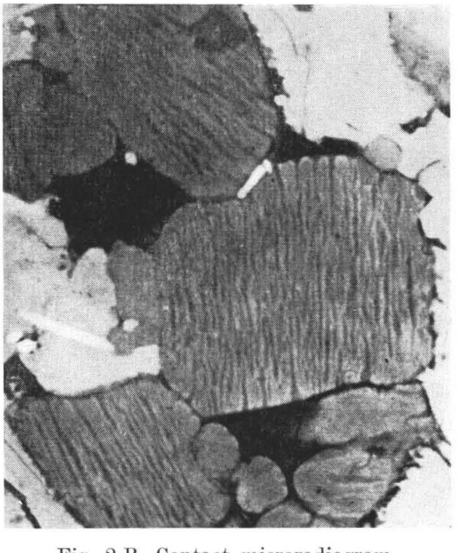

Fig. 2 B. Contact microradiogram.

Th. G. Sahama: Contact Microradiographic Distinction . . .

9

CANARIAN CALDERAS [1]

A short review based on personal impressions, 1947—1957

BY

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Professor Emeritus, Academy of Åbo, Finland

CONTENTS

PREFACE

In the autumn of 1949, the author read a paper at a meeting of the Finnish Society of Science (Helsinki) entitled: »Om calderabildningar med särskild hänsyn till Kanarieöarna.» It contained an account of the author's

[1] Received March 2, 1961.

observations during a visit to the Canary Islands the previous year, supplemented, of course, by data from the literature and from topographic maps of the islands. Now I am in a somewhat improved situation as far as my Canarian field observations are concerned, for after the first voyage I had the good fortune to spend parts of the years 1948, 1950, 1953, 1954 and 1957 in the archipelago. On these journeys I saw many things not known to me before in the Canarian calderas, these mysterious depressions in the otherwise no less mysterious volcanic landscape of the »Happy Islands» *(Insulae Fortunatae).*

Notwithstanding his possession of a wealth of new impressions, the author does not feel qualified to give a full account of the caldera phenomena in the Canaries and to explain their origin definitely in all cases. Much investigation is still needed to gain a clear insight into the volcanologic history of the islands, so this short *exposé* has no far-reaching pretentions. It is merely a modest contribution to a fascinating problem that has preoccupied investigators since the days of Leopold von Buch at the beginning of the 19th century.

THE CANARY ISLANDS AS A VOLCANOLOGIC PROVINCE AND ITS SPECIAL FEATURES

The Canary Islands may be considered one of the classical regions in the world for the study of volcanic phenomena. A great number of geological investigators have visited these islands, which are easy to reach by following the transatlantic traffic lanes to the southern continents. Like the Azores and Madeira, the Canaries are volcanic in nature, and the same can be said of the Cape Verde Islands farther south.

There are 12 islands in the Canary group, of which only 7 are of considerable size: La Palma, Hierro, Gomera, Tenerife, Grand Canary, Fuerteventura and Lanzarote (Fig. 1). The remaining, insignificant ones, are called *islotes.* The archipelago lies within the western African marginal zone, above the 4 000 m bathymetric line. They are not shelf islands in the proper sense, but they are not true oceanic islands either. The composition of their volcanic products is rather different from the typical, chiefly basaltic oceanic islands. The Canaries consist to a great extent, at least in the larger islands, of salic material. This circumstance is not without importance as to the mode of volcanic manifestations. In short, many of the lavas here are of high viscosity and were rich in volatile matter. There is, however, a wide range in the composition of the volcanic products not only in a single island, but also within the insular province as a whole. That makes the islands an interesting field of study for petrographers as well as for volcanologists.

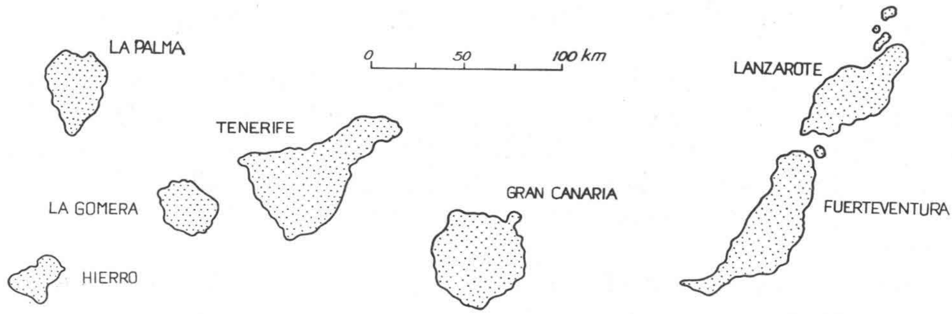

Fig. 1. The Canarian archipelago, key map.

Except for the easternmost islands, the Canaries are rather independent volcanic edifices rising from a deep sea. They have all had, save for the easternmost ones, an independent volcanic life, and their evolution has been rather complicated each in a special way, as indicated in the works devoted to several of the islands (see the bibliographic list).

In the present author's view, the insular region was originally occupied by a broad headland attached to the adjacent continent of Africa — an area now confined by the 4 000 m bathymetric line. In early Tertiary time, this headland was broken into pieces by sets of fault lines; blocks were left standing, separated by deep hollows. Only the easternmost of the islands still adhered to the continent in later times. The »chunks» thus formed consisted of the same Sial material that composes the continent. The old basement is still visible in some of the islands.

Upon these basement »chunks», very much dissected by fissures and resting on a Sima zone at a considerable depth, great masses of volcanic material were piled up in the course of time. The building-up process was by no means continuous, but frequently interrupted by periods of weathering and erosion and also by ruptures in the crust.

In this way a number of fairly independent volcanic edifices were created, the five outer ones rising from a deep sea — quite remarkable formations on the earth's surface, if we measure their height from the sea floor. Only the easternmost islands are of a more flattened shape, and they are surrounded by a relatively shallow sea, separated, however, from the African mainland by a sound with a maximum depth of about 1 000 m.

The Canarian geologic province is consequently volcanic in composition, at least when the submarine parts of the islands are excluded. These parts are, indeed, by far the greater ones, so we cannot make the generalization that the islands are entirely volcanic. In fact, an old basement found in some deeper parts of the islands reveals the presence of granular plutonic rocks (Fuerteventura, Grand Canary, Gomera, La Palma).

Since we know very little about the composition of the basement and never will be able to reach down there for closer investigation, it is better to speak of the Canaries as a volcanic province. The volcanic rocks — lavas, tuffs, agglomerates, *etc.* have formed the mountains, and still products were brought to the surface by eruptions in some of the islands during later centuries.

However, the islands offer not only »positive» geologic records — the volcanic rocks. There are also plenty of »negative» volcanic forms in the landscape; depressions of different kinds, among them the so-called calderas. On the other hand, exogenic forces have left their scars on all the islands — the *barrancos* furrow the slopes in an infinite number. Moreover, marine abrasion has worked on the coasts, especially on the side of the trade winds, creating high sea cliffs. One class of geologic record, however, is extremely sparse: the sedimentary material transported down the slopes by running water and also accumulations made by the wind. The only island in which sediments have been deposited on a somewhat larger scale is Fuerteventura, the one with a deeply denuded surface. We cannot classify the pyroclastic deposits of superficial character as sediments; in fact, they attain considerable thicknesses in some parts, as on the southern and southeastern sides of Tenerife, the largest of the islands.

The frequent destructive intervention of the exogenic agencies in the long volcanologic history of the Canaries has made interpretation of the different volcanic phases in many parts extremely difficult.

THE SO-CALLED CALDERAS IN THE ISLANDS AND THEIR VARIOUS TYPES

The Canary islands are really a wonderland of nature; in few places in the world are there such majestic manifestations of the endogenic forces we call volcanism. Not only is there a multitude of stately volcanic cones and their lava streams frozen into a fantastic jumble of glassy boulders and crests covering vast areas of the island slopes, but negative volcanic forms are also abundant. These include more or less rounded depressions — craters and giant-craters or calderas — not to speak of other kinds of irregular depressions confined by rugged precipices. In this paper we will treat the more rounded depressions, called calderas by the Spaniards (after the Portugese word *caldeira).* A multitude of forms of different sizes and origin has been grouped under this name, as we will see in the following descriptions.

The Canarian calderas, in the broadest sense, have long since been the object of study by many investigators, Spaniards as well as foreigners. The first scientifically trained observer to visit these islands was Leopold von Buch; that was in the year 1815. He was followed by a great number of geologists, most of whom have described the volcanic phenomena (see bibliography) and also expressed their opinions about the formation of the calderas. It would be desirable to recapitulate here all these opinions, but lack of space obliges us to confine ourselves to short comments.

Leopold von Buch (1825), the pioneer, found in Caldera de La Palma and in Caldera de Tirajana and in Caldera Bandama, Gran Canaria, and also in Caldera de las Cañadas, Tenerife, good examples to illustrate his »elevation hypothesis» according to which a crater is formed in the summit of a geotumor pushed up by subterranean forces. He interpreted the barrancos radiating from such an »elevation crater» as radial tension cracks in the rising tumor! These ideas re-appeared in Webb and Berthelot's great work on the Canaries (1839); and a later disciple was Déville (1846). But Charles Lyell, who also visited the islands in the first half of the last century, rejected the elevation hypothesis, running water being in his opinion the chief agent of excavation of these remarkable depressions. His ideas are expressed in the famous work »Principles of Geology» (Lyell, 1868) — especially regarding the caldera of La Palma, the remarkable hollow that has since become world famous.

More recently Gagel (1908) has thoroughly examined this Palma caldera and his conclusion agrees with Lyell's: running water has, in this case, done the chief work. As for the still larger Caldera de las Cañadas in Tenerife, Gagel is of the opinion that it must be the result of gigantic explosions and hence a true volcanic (negative) form. A similar view is also expressed by Navarro (1917), who spent a long time in the study of all these phenomena. During the last decade the present author also undertook a renewed study of the Tenerifan »circus», partly in company with the Canarian geologist T. Bravo. This colleague will in the near future give a detailed description for his Spanish public not only of the Tenerifan caldera but also of those in the other islands, in connection with the appearance of his »Geografía general de Canarias,» of which the first volume came out in 1954. Of other geologists lately interested in the problem of the calderas, Macau Vilar (1960) should be mentioned. We will refer to the opinions of these writers in the following chapters.

Before entering into a more detailed treatment of the thesis in question, we should bear in mind that a variety of depressions differing both in size and in shape and genesis have been brought together under the term *caldera*. This inexactness in nomenclature (by no means confined to the Canaries) has held sway for many decades, as pointed out in the great compilatory work on volcanism by von Wolff (in 2 volumes, 1914 and 1931).

It was a good thing that the volcanologic literature was enriched with the special treatise on calderas by Howel Williams (1941), who brought order to the conceptions and definitions. Also B. G. Escher has treated the subject in his many publications, chiefly referring to Indonesian volcanoes. Finally, A. Rittmann (1960) has given a comprehensive although brief report on calderas; but there would also be other authors to mention.

This literature on volcanism and calderas has, of course, been of great use to the author in his efforts to undeistand the real nature of all the depressions met with in the Canaries. Much, however, has to be left for future work. — Theoretical knowledge is one thing and interpretations in the field another: in bewilderment the newcomer stares at first into these tremendous abysses, unable, of course, to grasp any idea about their formation. It is only after many tiresome excursions from bottom to top and down again along criss-cross routes, that one may begin to gain some insight into these secrets of nature.

SPECIAL CHARACTERISTICS OF CALDERAS IN THE DIFFERENT ISLANDS

Now we may proceed with a brief description of the more remarkable types of so-called calderas in the Canaries and try to explain their mode of formation in every case, starting with the more important ones belonging to the larger islands. The reader is referred to topographic maps of at least 5 of the islands, however, on a very reduced scale. Some pictures will complete the verbal description.

TENERIFE

This, the largest and best known of the islands (Fig. 2) occupies a central position in the archipelago. It covers an area of 2 058 sq. km and includes the highest elevations, with Pico de Teide (3 711 m) the loftiest peak in the whole Spanish empire. This often described island (Hausen, 1956; bibl. list) is known not only for its great central volcanic top, but also for its centrally situated caldera, measuring at least 200 sq. km in area. It is called Caldera de las Cañadas, and its bottom lies about 2 000 m above sea level. We shall here study in some detail this remarkable surface formation, nowadays easily reached by motor roads.

Caldera de las Cañadas has been formed in the summit region of a great volcano that occupies the chief part of the island (Fig. 3). This volcano seems to be composed chiefly of salic (phonolitic, trachytic) lavas

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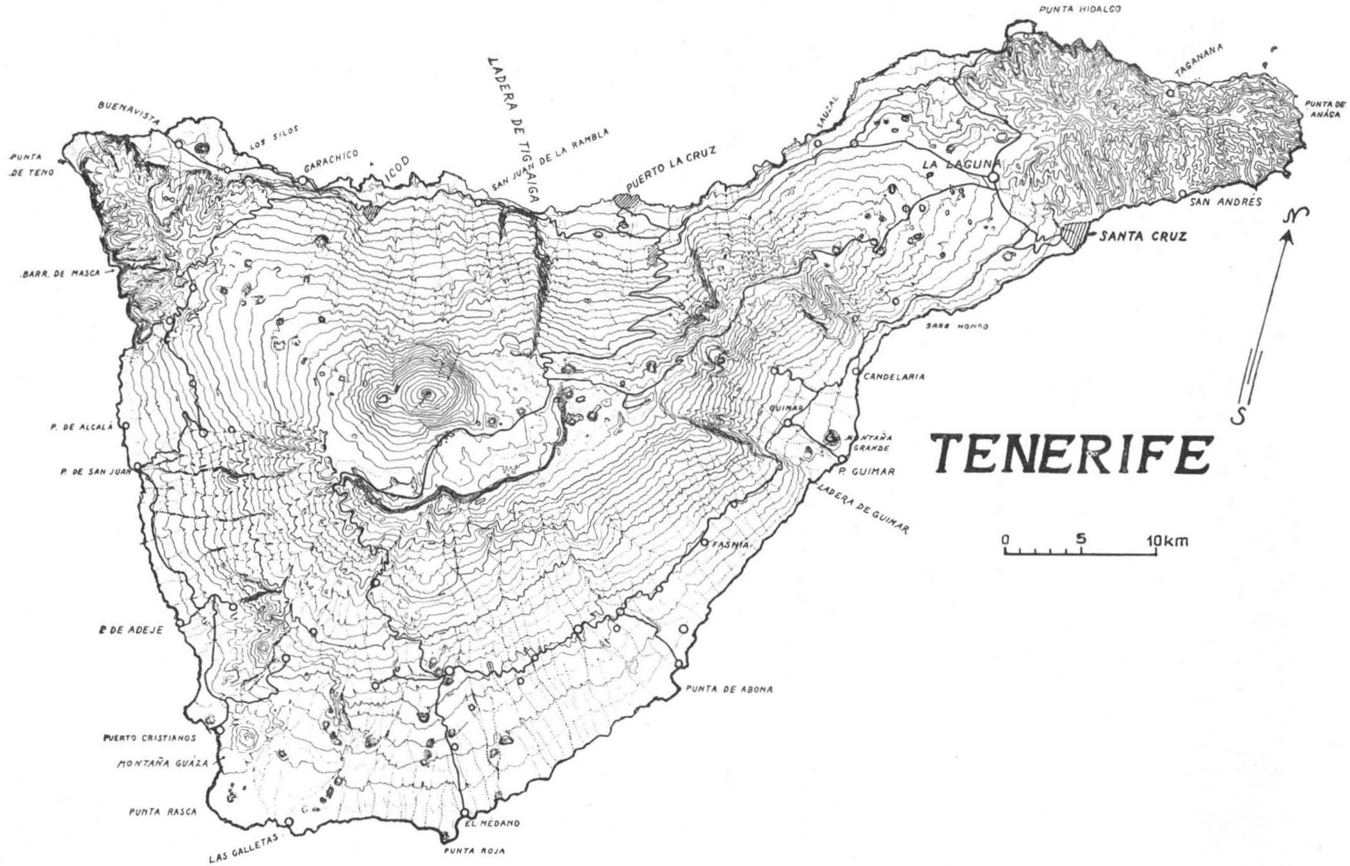

Fig. 2. Copy of a topographic map of the island. Scale 1: 50 000, with contour interval of 100 m.

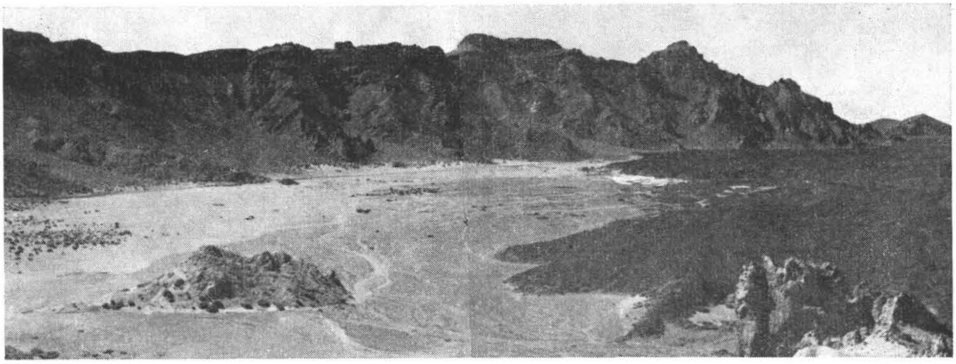

Fig. 3. Part of the southern wall of Caldera de las Cañadas, Tenerife, dissected by weathering and erosion. Sands of the caldera fill and (to the right) tongues of lavas from Pico Viejo. Looking southwest. Photo T. Bravo.

and tuffs. Older than this gigantic volcanic edifice is a lava and tuff formation of basaltic composition, with a number of later plugs and dikes of trachytes. — The twin volcano Pico Viejo—Pico de Teide, firmly united to a single body with two summits, rises in the central part of Caldera de las Cañadas. This complex volcanic apparatus stands as a new generation inside the caldera of the old volcano, like the cone of Vesuvius inside the fragmental ringwall of Monte Somma.

The formation of Caldera de las Cañadas has a rather complex history that may be difficult to disentangle. Some of the geologic records have been destroyed while others have been covered forevei by younger volcanic materials. We shall, however, try to follow the probable events that led to this spectacular negative volcanic surface form.

Omitting the oldest phase in the development of the island structure, we must imagine the existence of a gigantic phonolite volcano resting on an older basalt formation. The salic volcano was of the central strato-volcano type with countless sheets of lavas, tuffs and agglomerates. The cone reached a height that certainly surpassed that of the present Pico de Teide — perhaps 5 000 m above sea level. It did have a top crater, and the lavas ran down in all directions. As time passed, the activity ceased and the sides were attacked by erosion. Many barrancos were formed and some of them can still be seen (decapitated) in the better preserved southern part.

At the close of the Tertiary period (?) there was a sudden revival of activity, explosion catastrophes followed one after another, and enormous masses of bombs, blocks, lapilli and glass-ashes were thrown into the air. The airborne material drifted southwards owing to the prevailing trade winds and formed a thick mantle on the southern and southeastern slopes.

The pale-coloured pumiceous mass is mostly well stratified, but numerous unconformities indicate the presence of several breaks in the eruptions. A great part of this material drifted out over the ocean and formed a floating pumice »carpet» on the surface. This material was then caught by the southward-moving Canarian current and floated far and wide, and a large amount may have been carried by the mid-Atlantic current to the West Indies. Some of the pumice material probably reached the islands there and accumulated on the shores (?) The great explosions of Krakatoa in the Sunda archipelago in 1883 also produced enormous masses of pumice that drifted great distances over the Indian ocean.

Owing to the expulsion of these masses of salic material from the central Tenerifan volcano, a lack of support in the substructure of the volcano took place and the summit collapsed, forming a great caldera, or more exactly, t w o calderas. One developed somewhat later than the other, and between the two a ribbon was left standing (the present jagged row of »Los Azulejos»).

As time passed, an infilling process began in the two hollows, with material sliding down from the circumwallation. Stagnant wateı may have accumulated in the bottoms. The lake suıfaces rose to the height of thresholds and an outlet was formed to the west. In this way at least the western caldera was drained, and perhaps the eastern one was emptied to the north.

The sides of the twin caldera were heavily attacked by weathering and erosion during a long period of time, and certainly great landslides occurred here and there along the circumference. Embayments took shape and the caldera was greatly enlarged.

But later on, perhaps at the beginning of Quaternary time (?), a new volcanic phase started: a vent was opened inside the western caldera and a cone began to rise. As time passed, the cone grew to a considerable height, emitting from its crater salic (phonolitic) lavas, which rapidly filled the western caldera. In this way Pico Viejo came into existence.

The activity of the cone ended, so it seems, with an explosion and a small top-caldera was formed in this stratovolcano. Except for a blow hole left open at the western rim, its bottom was soon filled with lavas.

But another nearby vent developed soon after the completion of Pico Viejo. The Teide cone began to grow and finally reached a great height matching Pico Viejo, with its top crater, La Rambleta. Salic (trachytic) lavas were also emitted from this crater, inundating the entire surrounding eastern caldera. The cone grew firmly with the older one into a single great form with two summits separated by a saddle pass.

The latest phase of the Teide activity consisted of the building of a small cone inside La Rambleta—El Pitón. This top attained the maximum height of 3 711 m.

The flood of lavas not only inundated the caldera, but a great amount spilled down the northern slopes of the island, filling the volcano-tectonic valley of Orotava down to the seashore. Earlier lavas from Pico Viejo had found their way down to Icod, farther west, and still other lavas ran in broad streams to the west coast.

This central activity has now ceased, at least from the two central vents. A small eruption took place in 1799 from a secondary vent at Pico Viejo (Chahorra), and a final small eruption occurred at a place more to the west in 1909 (Chinyero).

Those portions of the twin calderas not filled by lava have been covered by gravels and sands deposited by running water (Fig. 3).

The evolution sketched in the foregoing closely resembles the history of Crater Lake in Oregon, U. S. A. (Williams, 1941), except that the old caldera Crater Lake is occupied by water and not by lavas and detritus.

Montaña Blanca, a cupola-shaped trachyte mass covered with whitish lapilli, has also contributed to the filling process in the eastern caldera. There are also some later cinder cones that have emitted their share of lava streams.

Not many othei depressions of a caldera nature are to be found in Tenerife. The author has covered the entire area but can report only the following calderas: On the south side of the lofty ridge of Cumbre de Pedro Gil, somewhat above the small town of A r a f o (Valle de Guimar), there is a kind of rounded depression of more insignificant size. It is closed, but in the south there is a narrow passage down to the Guimar valley. In the middle of this passage a small cinder cone has grown, having originated in the year 1705. A lava stream ran down the valley and menaced the town of Guimar. The caldera may be considered an explosion hollow dating back to fairly ancient times; a curious revival of the activity is represented by this small volcano.

There is also a rather important caldera situated in the southwestern corner of the island, to the north of the small harbour of Puerto Cristianos. It is called C a l d e r a d e l R e y and lies at a low elevation at the southern foot of the lofty Montaña del Conde (1 000 m). Its dimensions are 1 500 × 1 200 m and in depth it is only about 100 m. The bottom is a cultivated plain. The sides are steep 30°—40° slopes. The material in the walls consists of a mixture of fine pumice material (brownish-yellow), mingled with stones and boulders of lava. A rough stratification with a distal dip is perceptible. It is evident that the caldera is of the explosion type. The material in the walls was deposited at intervals and finally a collapse resulted in the enlargment of the orifice. The conduit seems to have reached down across the puzzolane formation and also across various lava sheets hidden beneath the same region.

Fig. 4. Topographic map of the island of Grand Canary. 100 m contour interval.

Hans Hausen: Canarian calderas.

In Tenerife there are several depressions that are not really calderas, but volcano-tectonic grabens such as Valle de Orotava, Valle de Guimar, Valle de Guerra and Valle de Segovia (Hausen, 1956). In the literature are some explanations of the causes of these negative land-forms (Rothpletz, 1889; Bravo, 1954; Hausen, 1956). These depressions will not, however, be considered in this paper.

GRAND CANARY

This shield-shaped island of 1 532 sq. km has a physiognomy that differs radically from that of Tenerife, even if we disregard the coastal contours. Grand Canary is far more dissected than its neighbour. Erosion has had time to work more continuously and there have also been other forces in operation dissecting the island into the present insular volcanic »ruin» (Fig. 4, Plate I).

At first — there is no dominating peak overlooking the island, no large top caldera in the central highland. Instead, the centre consists of a rather smooth highland lying above the 1 500 m contour line. From this elevated ground, a system of barrancos radiates in nearly all directions to the coasts, so that the general impression is that of an erosion landscape.

Closer scrutiny will, however, reveal several irregularities and also many forms not brought about by erosion alone. In two of the island sectors, in the west and in the southeast, there are very deep embayments into the very core of the central highland, both with relatively narrow outlets. They are called Caldera de Tejeda and Caldera de Tirajana, respectively. In the eastern sector there is also a kind of semi-caldera or a natural »amphitheater»- the head region of Barranco de Telde.

Moreover, plenty of volcanic constructive forms are also to be found in the island — mostly, however, of rather subordinate size. The northern and the eastern declivities are especially dotted with adventive cinder cones with a rather fresh aspect. There is also, in the extreme northeast, a small peninsula, La Isleta, that has been formed by a group of relatively young volcanoes and later united with the mainland by alluvial sedimentation.

Our chief interest in this connection lies in the two aforementioned »calderas», which will be briefly described below. We also have in some areas small calderas evidently of volcanic origin, chiefly, it would seem, of the maar type. They will also be briefly characterized.

C a l d e r a d e T e j e d a (Fig:s 5 and 6) is a strange depression, with its long axis extending east-west. It has not the slightest similarity to a true volcanic caldera. At first sight it gives the impression of an erosion

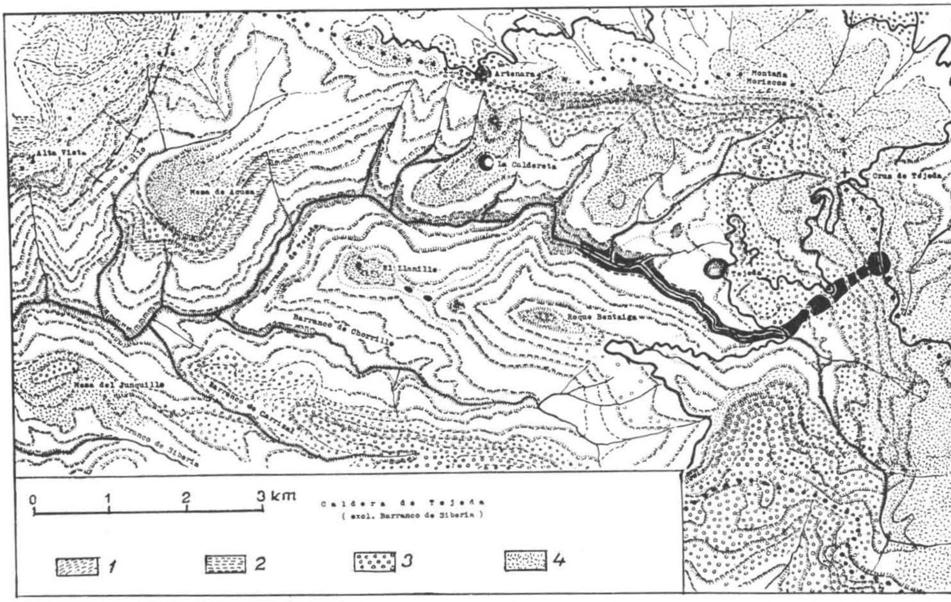

Fig. 5. Topographic-geologic map of Caldera de Tejeda (excl. Barranco de Siberia). White, unmarked area inside the caldera: old trachytes, dislocated. 1 = trachytes in Alta Vista. 2 = trachyphonolite in Mesa de Acusa. 3 = Roque Nublo agglomerate. 4 = Post-Miocene olivine basalts. Dark stream to the right: young basalt lava following Barranco de Tejeda. Dotted line: watershed. Copied from the topogr. map of the island on the scale of 1: 25 000. Long direction: E—W.

landscape with a system of deep gorges that are united farther west to a narrow canyon that crosses a mountain barrier on the way to the coast. The basin measures 14 × 10 km. Its depth is at Villa de Tejeda, 1 000 m below the northern edge (Los Moriscos) and that point lies 1 750 m above sea level. The area is crossed by a master gorge, Barranco de Tejeda, which has three tributaries joining from the left, all deep gorges. Between them there are sharp divides, partly with standing erosional pinnacles. The area is enclosed by a watershed, except in the west, where the cross-cutting gorge forms a break.

One may ask how this tremendous abyss is to be explained. Is it really the sole record of running water in combination with weathering, or have other geological forces, perhaps of an endogenic nature, also been at work here?

Looking at the topographic map of the island and especially at the radiating pattern of barrancos, one must be impressed by the great achievements of erosion in t h i s w e s t e r n s e c t o r (and also in the southeast: Caldera de Tirajana). There must have been some circumstances facilitating the deep erosion.

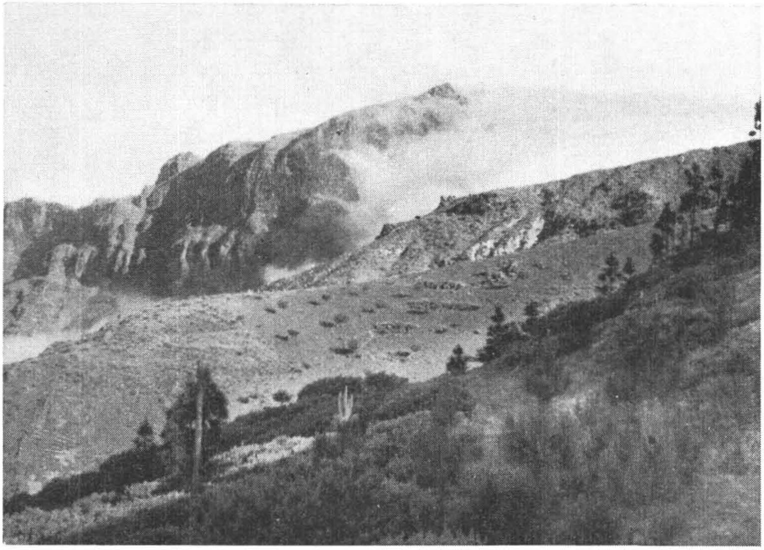

Fig. 6. Northern wall of Caldera de Tejeda as seen from the eastern rim, Cruz de Tejeda (1 500 m above sea level). The thick banks in the precipices consist of a coarse agglomerate with intercalated lavas (tephrites). Photo H. Hausen.

Judging from the geologic conditions inside the area, it is evident that the evolution culminating in the present state has been a long process. At first there was a tectonic subsidence along several fissures, in the north, in the east, in the south and in the west. A kind of tectonic basin came into existence. This was to a considerable degree filled with agglomeratic masses, in parts reassorted by water. The materials issued from vents in the highland area. Later this formation was repeatedly covered with basalt lavas and the bottom of the basin rose. But soon new displacements occurred and the basin again deepened. A considerable amount of detrital material then accumulated; water became stagnant here, and an outflow was opened over the mountain barrier to the west. This outlet was continually deepened, and drainage inside the basin was established. The gorges were successively deepened, probably in connection with a rise of the island in pre-Quaternary time. Remnants of the old bottoms are still to be seen here and there, as at Mesa de Acusa, Mesa de Junquillos and El Llanillo. Old terraces on the cross-cutting gorge in the west are also visible (Presa de las Niñas).

In still later times (Quaternary) new fractures occurred and cinder cones grew along the borders and inside the caldera.

Caldera de Tirajana, the other spectacular depression in this island, is of a somewhat different nature (Fig. 7), although the erosion forms

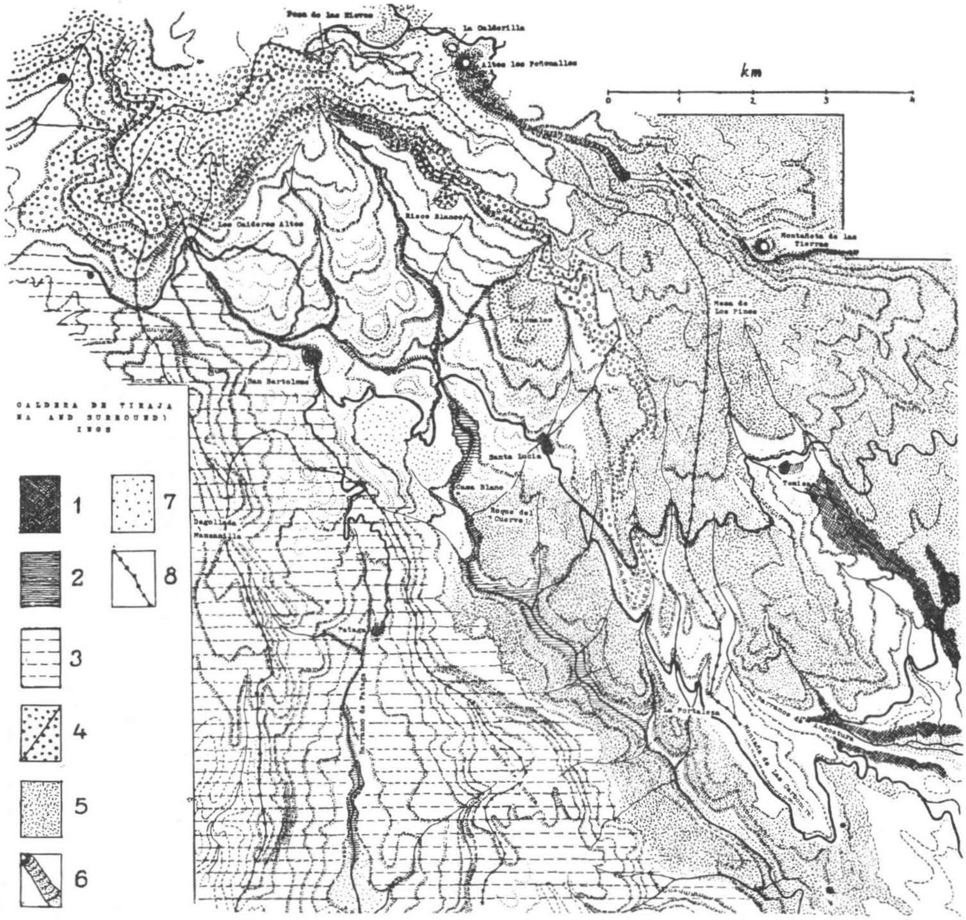

Fig. 7. Topographic-geologic map of Caldera de Tirajana, Grand Canary. 1 = old basalts, 2 = rhyolites (older series), 3 = nepheline phonolites, 4 = Roque Nublo agglomerates, 5 = Quaternary basalt lavas filling Valle de Tirajana, 6 = young olivine basalt lava stream in the highland, 7 = Quaternary sands and gravels, 8 = watershed surrounding the caldera area. Copied from the topogr. map of the island on the scale 1: 25 000. The map is oriented N—S.

also in this area seem to dominate the physiognomy. A somewhat caldera-like limitation is to be found at least on the north side, where the border of the central highland forms precipices several hundred meters high. Close to this border rises the highest point of the island, Pozo de las Nieves (1950 m), a hilltop belonging to the highland ground and cut off by the precipices in the south side.

The semi-circular shape of this northern part of the depression has led many investigators to assume the caldera to be of volcanic origin — an explosion and collapse caldera. Recently Macau Vilar (1960) advanced the idea that the caldera is the result of a gigantic collapse as a consequence of a withdrawal of magma from below the area. This magma was used up by volcanic eruptions in the surroundings. He considers the depression as a collapse caldera of a type described by Williams (1941) in his systematic study of all the calderas in the world. Later on the caldera was enlarged by landslides.

The present author has devoted considerable time to the study of this caldera and to all its geologic particulars. He must confess he cannot accept Macau Vilar's hypothesis *(op. cit)*, for several reasons. First, in the surroundings there is no such quantity of lavas which could have caused the collapse of that tremendous hollow. Secondly, there is no closed basin typical of that kind of collapse. And finally the volcanic beds exposed in the outlet gorge — Barranco de Tirajana — continue undisturbed northwards as far as La Rosiana, if not farther. Barranco de Tirajana strikes me as the master canyon of tectonic origin, and along it erosion has worked back into the highland block assisted by great landslides from all sides. The slides may have played a far more important role than Macau Vilar seems to have suggested.

The erosion-landslide hypothesis does not, however, exclude the appearance at some intervals of tectonic displacements in the area. The outlet barranco Valle de Tirajana certainly of tectonic origin, marks the dividing line between two different formations, phonolites in the west and agglomerates and basalts in the east. There were other displacements as well along the large cuesta in the west, which continues southwards down the length of Barranco de Fataga.

But fractures inside the caldera were obviously few, as seen from the absence of cinder cones.

If we now try to follow the development of Caldera de Tirajana from its beginning as a depression in this sector of the island, we must first consider the set of fractures that crossed the area and along which displacements occurred. There is a long line or zone running from Degollada de Tirajana in the northwest in a southeasterly direction with a course coinciding with that of the present Barranco de Tirajana. It is a very important line, since the land to the northeast of it has subsided in relation to the land west of it. [1] Along the same line (of post-Miocene age?) erosion began to work. The young, narrow gorge was continually widened and its head migrated backwards, and finally a relatively open valley with steep sides was formed. An early stage of the valley-making process is seen in a very coarse delta fan in the left side not less than 900—1 000 m above sea level. Lower down,

[1] It represents a part of the Bourcart dislocation line that crosses the entire island diagonally in the dir. NW — SE. See the map by J, Bourcart (1937).

opposite Santa Lucía-La Rosiana, there is a broad rock terrace at a level of about 625 m.

Another fault line, in the west side of the caldera and continuing down Barranco de Fataga, has opened the way for the erosion of a valley similar to Barranco de Tirajana. Between these two tectonical valleys stands the horst ridge Cumbre de Amurga.

Inside this, rectilinear fracture system, which was formed later than the great basalt lava floods in the east (end of the Tertiary epoch?), the regimen of exogenic forces continued with erosion, weathering and great intermittent landslides, all of which are still active processes. At one time, perhaps in early Quaternary epoch, the Tirajana valley was flooded by olivine basalt lavas, which came from the left side (the highland border) and filled the old valley bottom to a considerable height. Later, owing to the rise of the island in Quaternary time, a young canyon was cut into the fill and lava terraces were formed. In the middle course of this canyon, strange erosional pinnacles known as Las Fortalezas have been formed.

The present semi-caldera-like form of the northern part of Caldera de Tirajana owes its existence primarily to such factors as mechanical disintegration of the outcropping volcanic beds, to springs appearing between the heads of the sheets in the precipices, and to rock slides. Such a circus-like morphologic end-result is also a common feature in the Hawaii islands, where the geologic and the climatic conditions are rather similar.

A good illustration of »circus-making» caused by exogenic forces may be found in the semi-caldera of Tenteniguada in the eastern sector of the island. Here is another edge of the central highland, in which weathering and erosion have created a broad embayment — the head of the long valley Barranco de Telde. Acute form — elements, a number of volcanic necks of hauyne phonolite lavas that reached the surface of the highland prior to the opening of the semicaldera, give a special accent to the mountain scenery.

Besides the afore-described so-called calderas, which can scarcely be considered of volcanic origin, we have a number of calderas of true volcanic nature but of quite insignificant dimensions. In the islands such depressions have received the very adequate name *calderetas*, the diminutive form of the word calderas.

The best known of such calderetas in Grand Canary is Caldera de Bandama (should be Caldereta de B.) 10 km from the capital, Las Palmas, on the east side of the island (Fig. 8). It is only 870 m (max.) across and 180 m in depth. The bottom is dry and sandy. In the immediate vicinity rises a coke-black cinder cone, Mont. de Bandama, and its lapillis have slid to the bottom of the caldereta and covered its western wall.

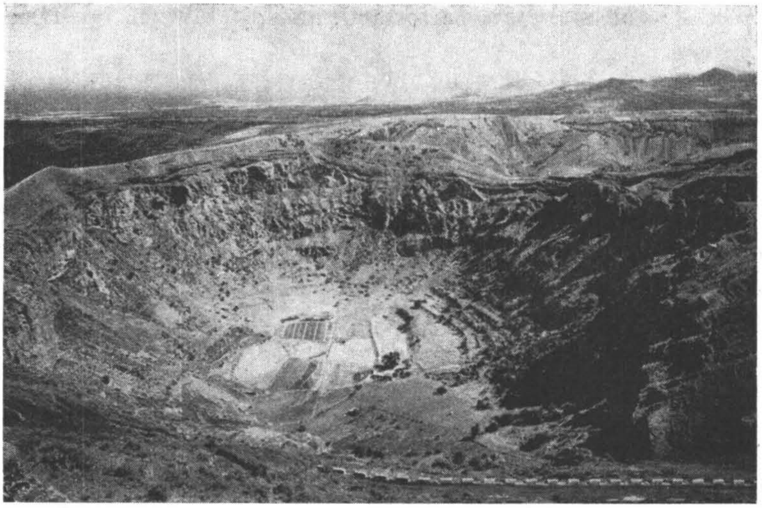

Fig. 8. Caldera de Bandama, an explosion maar that has perforated the rock ground consisting of phonolite lavas and agglomerates in banks; the rims covered by ejecta of stones of salic material. Looking east from the slope of the recent cone, Montaña de Bandama. Eastern side of Grand Canary, above Jinamar. From a postcard.

Macau Vilar (1960) has tried to explain the origin of the small caldera as the result of a collapse owing to withdrawal of lava from below during the eruption of the nearby cone. The present author cannot agree with that conception on the following grounds:

As already shown by von Buch (1825), the hollow in question is fringed by a rather thick well-stratified mantle of gravels, *i. e.*, lapilli and stone-ejecta with a distal dip. These layers cover a basement of »trachyte» and some agglomerate or conglomerate. All these layers are of salic composition, except for a basalt lava sheet in one side. The covering lapilli, *etc.*, demonstrate that the hollow owes its existence to explosions that carried the salic material from some depth. The hollow is old and weathered, whereas the nearby cone is of recent in age. The latter consists of basaltic material, and it belongs to the generation of adventive cones so common on the eastern and northern slopes of the island.

To use a collapse theory in this case is not necessary. There are many, similar cones in the island and not one has such a depression in the vicinity! Caldera de Bandama may be attributed to the group of explosion maars, a very fine example, indeed.

The explosions may have been provoked by ascending juvenile gases. When reaching the ground water table these vaporized the water and now the joint gases found their way through a diatreme to the surface. A complex of salic volcanics was perforated, and ejecta from it settled around the maar.

Of the other similar calderetas La Caldereta in the central highland, a short distance to the north of the cinder cone Los Peñonallos, may be mentioned. It is a rounded, not very deep depression in a lava sheet of hauyne phonolite, the sides showing columnar jointings. The bottom is covered with gravels (lapilli) from the cone nearby and the rims also have a covering sheet of lapilli of the same origin. In this case, the hauynophyre lava has been perforated, either by explosion or by collapse, although the latter seems less probable. The circumstances should, however, be more closely investigated.

Not far from this point lies another caldereta, Caldera de los Marteles, close to the eastern rim of the central highland, 1 500 m above sea level. It has somewhat greater dimensions, the sides varying in height. The caldereta lies depressed in the left head - arm of Barranco de Guayadeque, a large canyon with a rather steep gradient that leads down to the east coast.

The situation of Caldera de los Marteles is somewhat complicated. It is not a depression inside an old rock ground, except in the north and east. In the south it is barred with a ridge of lapilli in loose condition; in the west rises a steep wall, which is simply the cross section of a volcanic cone that stands close to that side.

The author is of the opinion that this caldereta is the result of a collapse in the cinder cone of which remains still stand at the sides. The bar of lapilli is the lower part of the original south flank. A collapse has occurred in the summit of the cone after explosions that ejected the loose material. No lava stream has been emitted here. The cause of the collapse seems to have been exhaustion: withdrawal of the clasmatics. The age of this volcano cannot be great since it is located in the head barranco of Guayadeque that was formed at the close of the Tertiary period or at the beginning of Quaternary time. Barranco de Guayadeque is incised into the thick series of post-Miocene lavas. Hence the caldereta in question differs considerably from that of Bandama and cannot be classified as a maar.

Pinos de Galdar is a small explosion crater-caldera on the road between Valleseco and Artenara at an altitude of about 1 450 m. It lies in the vicinity of a sub-Recent cinder cone, Montañon Negro.

LA PALMA

This northwestern island in the archipelago has an area of 730 sq. km and is pear-shaped (Fig. 9, Pl. II). Like Hierro, it rises from a rather deep sea and lies close to the great submarine threshold that separates the 4 000 m

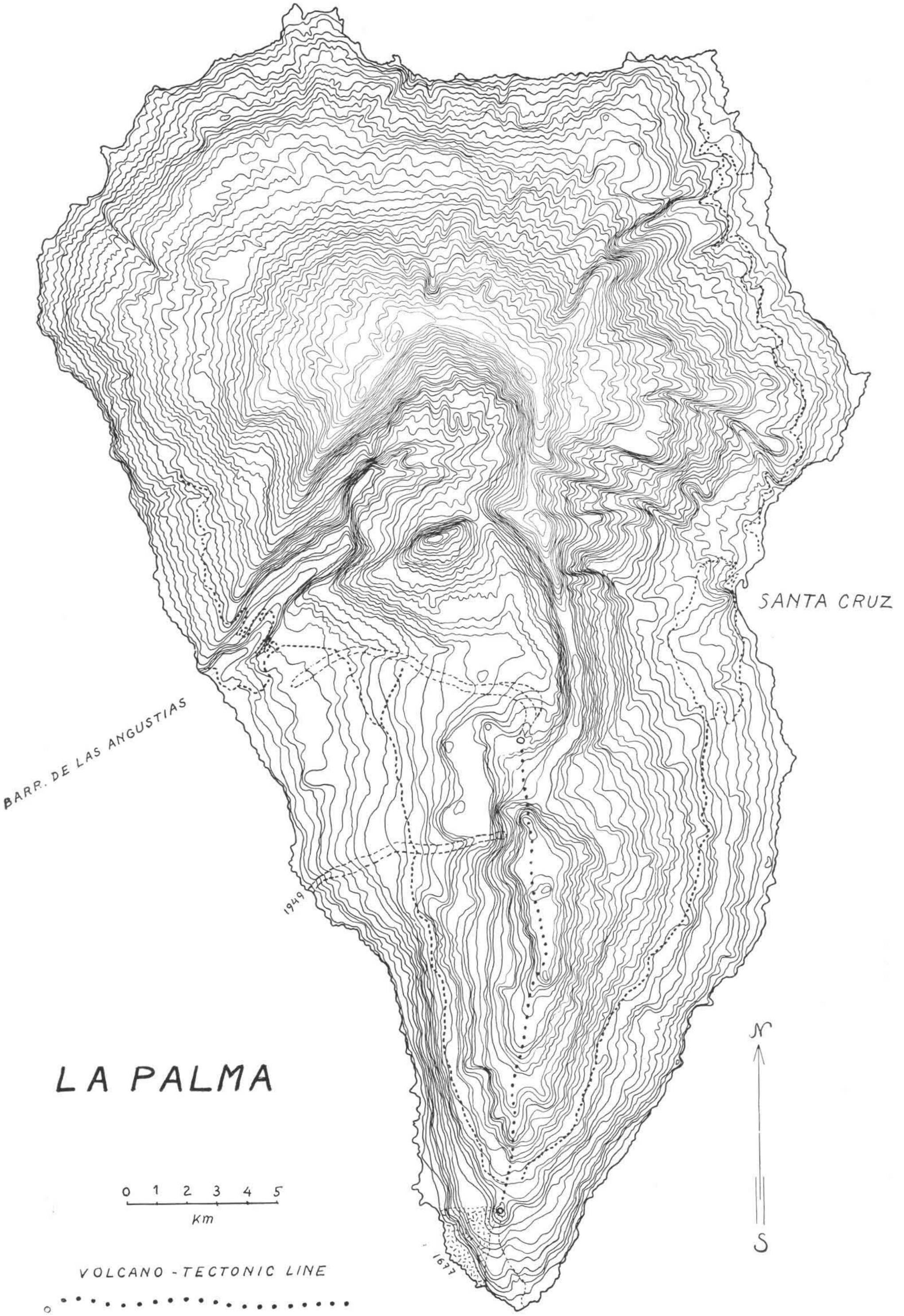

Fig. 9. Topographic map of the island of La Palma with 50 m contour interval. The dotted line represents the main volcano-tectonic fissure of the island.

Hans Hausen: Canarian calderas.

bathymetric line from the abyss of the so-called Canarian basin. In spite of the deep surroundings, the island rises to heights of about 2 500 m (max. 2 423 m) in the northern enlarged part. The coasts are mostly steep.

La Palma is well known the world over for its caldera, the so-called Caldera de Taburiente, a rather narrow depression in the midst of the northern part of the island. This curious rounded hollow, 8 km across N—S, has been compared with the cleaned kernel of a pear. The caldera is not closed, but has a deep outlet-gorge, Barranco de las Angustias, leading down to the sea in the southwest. The caldera is located on the highest part of the island, and at first glance on the map one is inclined to think it repiesents a kind of top caldera to a large volcano composing the whole northern part of La Palma. The differences in height between the bottom of the caldera and its borders are considerable (max. 1 600 m). On nearer scrutiny, the caldera offers many irregularities difficult to explain, and it is no wonder that many geologists since the first visit by L. von Buch in 1815 have expressed divergent opinions.

The northern and eastern wall of the caldera together form a semi-circle somewhat damaged by erosion. The eastern wall continues with a curvature into the southern wall but with a break represented by the broad saddle-pass La Cumbrecita. The southern wall, of short extension, is only the north side of Montaña Bejonado (2 000 m) that looks like a tectonic ribbon. The west side of the caldera is somewhat broken into steps *(escalonado)*, and in the southwest there is the outlet opening, the narrow beginning of Barranco de las Angustias. This barranco is walled in on the right side by the very high escarpment El Time that looks like the immediate continuation of the western wall. Also the left side of the barranco is an escarpment. It seems that this outlet has been opened by great fractures extending from the coast into the caldera.

From the borders of the caldera, the surface of the island slopes to the sea rather continuously, at least to the west, north and east. To the south is a long ridge, Los Rancones, which can be followed as a kind of backbone along the whole southern part of the island and which varies somewhat in height (many parts reaching the 2 000 m contour line). In the southern-most part, the ridge steadily decreases in elevation down to the sea.

To the west of the ridge is the El Paso woodland, which may be a tectonic depression. It extends northward to the pass La Cumbrecita as a kind of highland valley.

Los Rancones may be considered not only as a topographic backbone, but also as a volcano-tectonic line, for several outbursts have taken place, the latest one in 1949.

If we now look somewhat closer at Caldera de Taburiente (visited by the author on three separate journeys), we can first state that this remarkable hollow in the highest part of the island reaches down to its old fundament, which consists of decomposed spilite lavas and various kinds of intermediate and basic granular intrusive rock masses. They have been described in detail by Gagel (1908), and we do not need to recapitulate his results here. This formation is capped with a very thick mantle of basaltic lavas and tuffs, the stratigraphic limit being easily perceptible in the steep caldera walls. A multitude of dikes crosses these rock complexes.

Now we should consider the probable formation of the caldera, one of the most remarkable in the whole world.

After the interpretation by von Buch (»elevation crater»), Lyell (1868) rejected the idea and recognized the dominating role of running water. Also Gagel (1908) seems to believe that these exogenic forces may be responsible for the excavations, since the old basement could offer but little resistance. Later, the German volcanologist Reck (1928) presented a new explanation without having seen the caldera himself. He found a more complicated history of volcanic actions alternating with erosion. I have previously (Hausen, 1949) considered all his views, and since the paper by Reck is easily available, we do not need to recapitulate the events he postulated.

Finally, Bravo (1954) has given a short explanation of the caldera based on field examinations. He is of the opinion that the chief work has been done by erosion, and the old basement was an easy victim to this process. A more detailed account will be given in a later publication (according to a communication in Tome I of his »Geografía general de Canarias»).

On the basis of what he has seen during his visits in the caldera, the present author cannot deny the fact that tectonic forces have been at work shattering the kernel of the island. There are at least two important fault lines that converge into the caldera; one from the south and the other from the southwest. There are also innumerable fissures filled with magma that cross the formations in all directions. Great displacements have taken place along the right side of Barranco de Angustias, where the wall of El Time now extends. This dislocation opened the way for running water to transport loose material to the sea. Hence, the excavation work was initiated under control of fractures. The formation of the amphitheatre head, the caldera itself, was facilitated by the »rotten» condition of the basement formation (Gagel, 1908; Bravo, 1954). Explosions at any time seem less probable, as there are no masses of ejecta in the surroundings that may have belonged to the deeper part of the island now exposed in the caldera.

In the bottom of the caldera are some higher rocky remnants corresponding to an earlier stage of the downcutting process. In Miocene time the island lay more than 300 m deeper than now, according to finds of marine fossils on the side of Barranco de Angustias. The circumference of the cal-

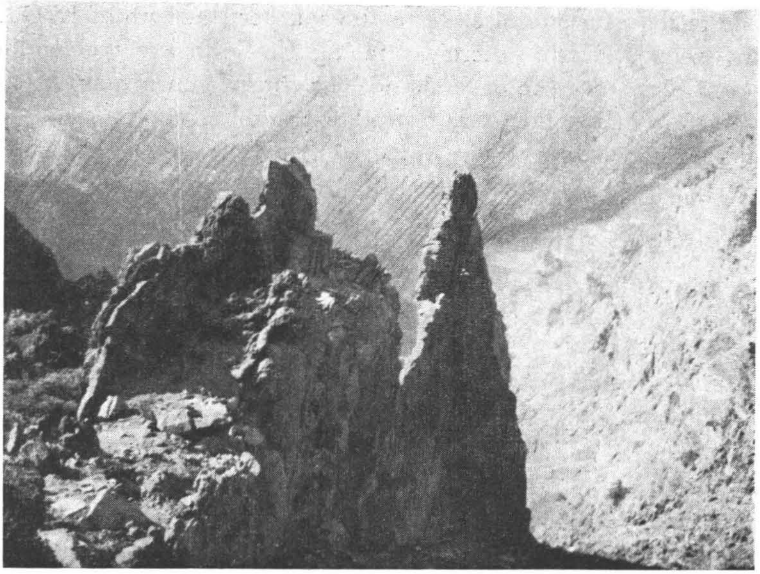

Fig. 10. Vertical dike plates standing at the northern edge of Caldera de Taburiente, La Palma. Looking south into the caldera. Photo H. Hausen.

dera has in later times been greatly enlarged, as may be seen from the walls with their pinnacles of standing hard dike plates (Fig. 10).

In the island there are certain kinds of small explosion hollows or maars. One of these, called Caldera de San Antonio, not far from Fuencalientes in the southern part of the island, was visited by the author. It is well walled in, lies at an altitude of about 650 m, is 50 m deep and has a diameter of 350 m. No lava has been emitted from this hollow. It is a typical maar, formed by explosions; masses of loose ejecta lie around on the slopes. Among these erratics are both lavas and deep-seated rocks, such as olivinite. The caldereta was active in the year 1677, and lower down on the slope a lava stream issued, reaching the sea shore.

Caldera de San Antonio lies on the same volcano-tectonic line that begins in Caldera de Taburiente and along which the ultimate eruptions took place (1949).

HIERRO

The small island of Hierro (formerly called Ferro), 278 sq. km in area, lies in the extreme southwest of the archipelago in a rather isolated position surrounded by deep waters. Its peculiar shape (Fig. 11, Pl. III) is chiefly due to a broad embayment in the northwest — El Golfo. The island is relatively

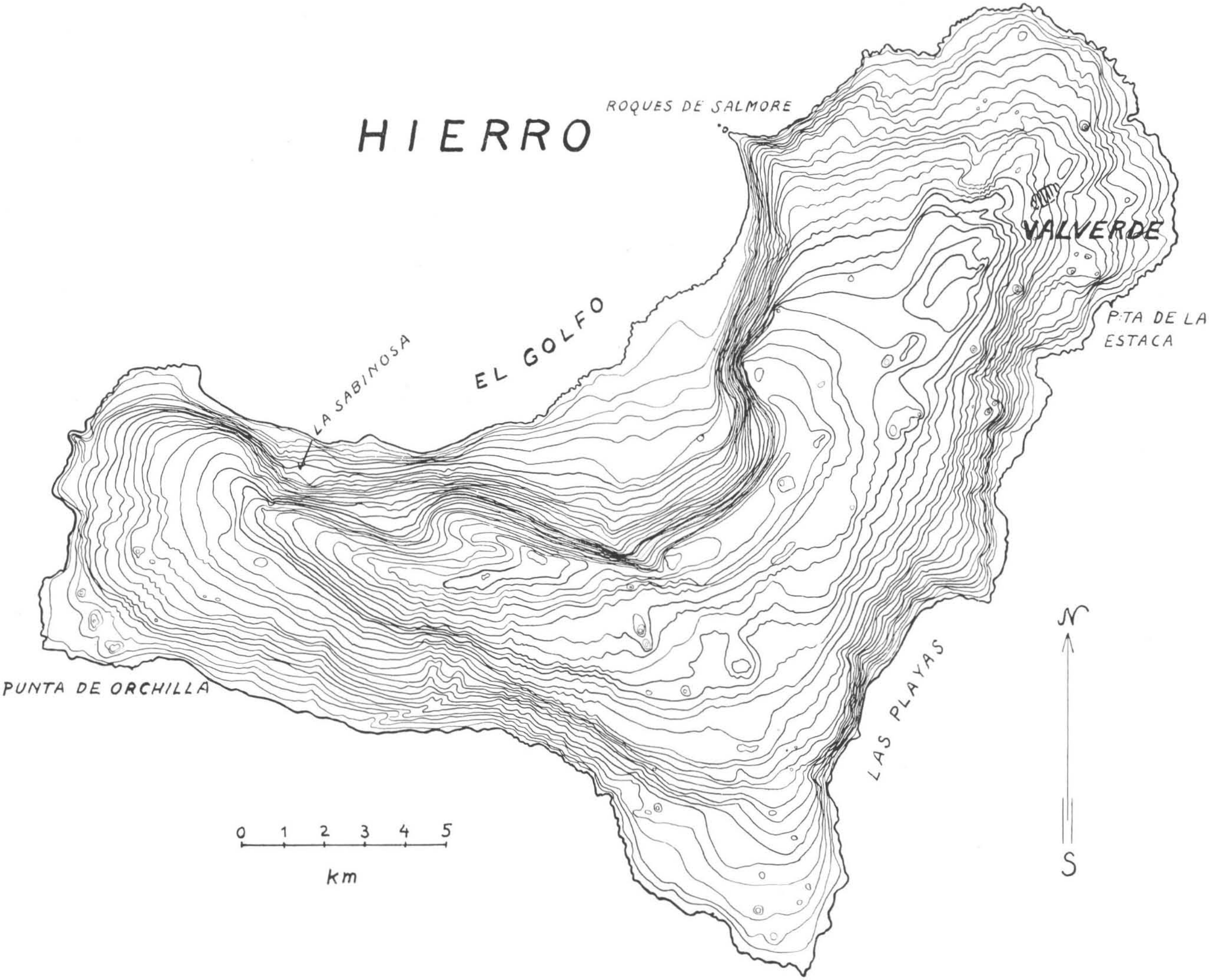

Fig. 11. Topographic map with 50 m contour intervals showing the strange outlines of the island, produced partly by fractures, partly by marine abrasion.

high, culminating at 1 501 m close to the southern margin of El Golfo. The shores are steep and the precipices facing El Golfo are tremendous walls about 900 m high. On the other hand, the upper (inner) part of the island is relatively flat, here and there dotted with cinder cones.

Among earlier visitors to this remote island was von Knebel (1906), who expressed the opinion that the broad embayment in the northwest, with a corda of about 15 km, is the remaining part of a former giant explosion caldera of a large basaltic volcano, which has since been reduced in circumference by faulting and abrasion.

I devoted only a week to excursions on Hierro and have not been able to arrive at any clear idea about the formation of El Golfo. Since there are no loose ejecta in the surrounding highland that should have been strewn around the site of explosions, it seems to me the assumption of an explosion-caldera is not well founded. It is most likely that a special type of faulting controlled by a semi-circular set of fractures occurred in the northwest after the island had been piled up from the deep lying basement. That there are such faults in the bay at the edge and foot of the precipices is indicated by the presence of several cinder cones, one of which was visited at the village La Frontera.

The former shield volcano of Hierro has suffered not only from the fractures referred to but also from continuous attack by the surf, so that high sea cliffs have been formed in many sectors.

There are also some small crater-calderas worth mentioning on the island; one is in the vicinity of Valverde and another on the northern coast. They seem to be small collapse-calderas produced by draining off of the lavas into the deeper part of the volcanic throat.

The island is apparently quite old, as evidenced not only by the presence of the Golfo embayment, but also by the advanced stage of the marine abrasion. Curiously enough, barranco erosion has only slightly affected the surface.

LA GOMERA

This relatively small island, 380 sq. km in area, has a rounded shape (Fig. 12, Pl. IV) and is a smaller replica of Grand Canary in the west. It is relatively high (max. 1 484 m) and is surrounded by deep waters. The distance to the nearest coast of Tenerife is only 26 km, and it seems the two islands had been united in the past.

La Gomera is, in comparison with the other islands, nearly devoid of volcanic cones or explosion craters, not to speak of volcanic calderas. Its surface is furrowed by a radial pattern of barrancos that have cut deeply into the island block, especially on the north side. Much of the higher ground is covered with decomposition products. In short, it has all the characteristics of an old island, as is also manifested by the steep abrasion coasts.

To speak of La Gomera in this connection seems superfluous, owing to the features referred to. Nevertheless, on the windward side there are some wide valleys, Valle Hermigua and Valle Hermoso, that seem to have a certain similarity to Caldera de Taburiente in La Palma. Gagel (1925) has pointed out such an affinity, and he does not attribute a volcanic origin, as in the case of the caldera of La Palma to these valleys. He considers them the only result of erosion.

We may briefly describe the main features of these two valleys and their probable formation. The author visited them during a rather short trip and many details have not been registered.

Valle Hermigua represents a deep embayment in the northeastern side of the island. Its head is broad but not semi-circular. Erosion has been vigorous, here and there only a narrow crest separates it from the head of Barranco de San Sebastián, which opens the way to the east coast. In fact the ridge is so sharp that a tunnel has been opened across it, for otherwise traffic by motor-car would be impracticable. Valle Hermigua has a master barranco that leads to the coast in the north, but it has not excavated the entire breadth. It is of interest to note that landslides have played an important part here. This phenomenon has been facilitated by the presence of an old, highly decomposed basement formation soaked with water on this windward, humid side of the island. We owe to Gagel (1925) recognition of this old formation, which also appears along the north coast.

Valle Hermoso farther west is a rather similar topographic feature. It likewise has a master canyon incised into the bottom and the head is again not of the semi-circular shape. Instead, a great rectangular corner suggests the presence of two sets of fractures along which subsidences may have taken place. The valley seems to have been cut into soft ground of pale-coloured phonolites (?), extremely decomposed and disintegrated. This ground also occurs in the valley walls. Only some necks of trachytes stand as hard rocks overlooking the scene (Roque del Valle). From the crest line to the west of the valley I could see that the bottom is divided by a threshold into two steps, which have later been cut by a young canyon.

The weathered older formation appearing in the barrancos of the north side of the island has seemingly facilitated the excavation of the two valleys. In other sectors of the island it appears as if a harder basalt formation of later age has restricted the erosion to a number of narrow, canyon-like barrancos, the longest of which is Barranco de San Sebastián. They all have their heads in a smooth upland, the remnant of a formerly wider shield-shaped island surface covered with a thick mantle of reddish brown weathering soil.

Both Valle Hermiqua and Valle Hermoso formerly had a greater extension to the north. They are mere rudiments of their earlier shape, for the incessant marine abrasion has pushed the coastline back and high sea cliffs have been formed in some places. The shortening of the valleys has also caused a steepening of the longitudinal profile with the consequence of rejuvenation of the river erosion. In other directions, the barrancos do not show any thresholds or young canyons as far as has been proved; it seems the erosion of the island has reached a relative standstill. Barranco erosion here belongs chiefly to an earlier humid period with running water (the Pliocene uplift ?).

As pointed out, no volcanic forms have been observed on the island, a rare case in the Canaries. La Gomera is the most conspicuously eroded land in the archipelago and the oldest of the western islands.

FUERTEVENTURA

This large island, the one of the *Purpuraries* (or *»Islas orientales»)* with an area of 1 725 sq. km (Fig. 13), differs considerably from the other islands by its dominating erosion surface (Hausen, 1956). The volcanic formations that compose its ground are mostly of old age and they have undergone both mechanical disintegration and chemical decomposition and also denudation by running water and by the stormy trade winds. The landscape has an open desert-like aspect not very different from that of the adjacent coast land of Africa.

There are both constructive and destructive volcanic forms to be observed in some parts of the island. These represent superimposed elements in the topography and indicate a revival of volcanic activity after an extremely long period of repose, when chiefly exogenic forces had been at work.

There are two main regions with young volcanic forms in Fuerteventura: one in the extreme north and the other in the middle part of the island. There are also scattered smaller cinder cones. Many of the younger volcanoes have sent lava streams into the surroundings, covering considerable areas with rugged, blocky and glassy basalts. Several of the larger cones are of interest, because they enclose top crater-calderas, sometimes of larger size (depth and diam.), and they will be briefly described here. (For further details see Hausen 1956.)

There is a great number of volcanoes in the northern region, all cinder cones of Quaternary and sub-Recent age. They lie in a NE—SW row, and it seems that Montaña del Faro in the small Isla de Lobos also belongs to this group. Masses of basalt lavas have flooded a great deal of this part of the island. Many of the cones have a large top crater-caldera open on one

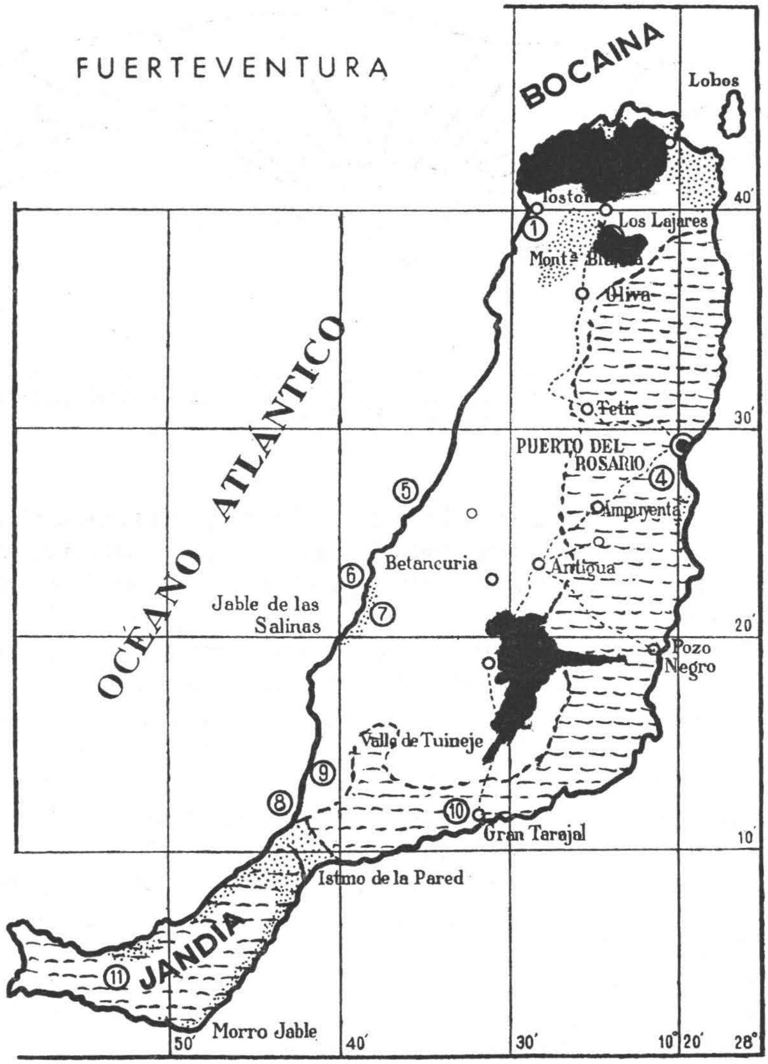

Fig. 13. Black areas: Young volcanoes and their lavas. Dashed areas: basaltic table land. White: old spilites (dislocated). Number refer to occurrences of sedimentary rocks.

side, size exaggerated in relation to the cone itself. The bottom of such a caldera is in at least two cases filled with chilled lava bubbles (Fig. 14). The material in the walls of the cones may be scoriae, slag and lapilli. Most of these crater-calderas seem to have been formed by explosions, but a final phase has comprised a new rise of lavas from the conduit, accompanied by emissions from the flanks of more lava flows into the surroundings.

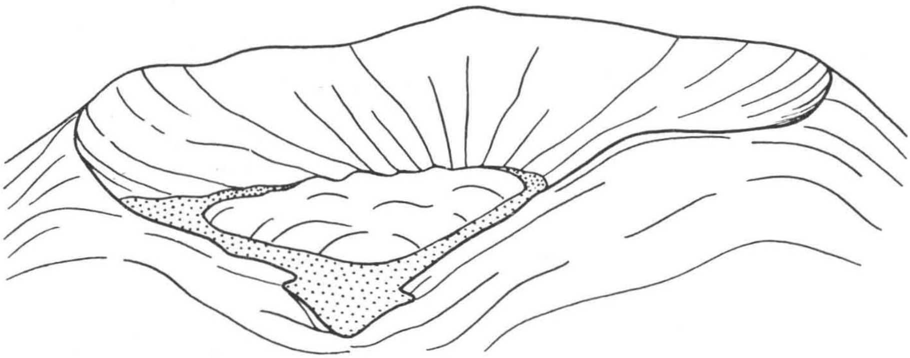

Fig. 14. Caldera de San Rafaél, northern part of Fuerteventura. A cinder cone with a wide crater-caldera, the bottom of which is filled with chilled lava bubbles. Looking south. Field sketch by H. Hausen.

All the lavas are olivine-rich basalts, the leading types in the latest stages of Canarian volcanism. Since these lavas are of low viscosity, the great explosions may at least in some cases be attributed to contact with basal ground water and sea water soaking the basement rocks.

The other group of younger volcanoes lies farther south within the broadest part of the island. They are of a somewhat different age, however; the oldest is Caldera de Gairía (461 m), some distance south of Antigua. This is a stately cone of slag and scoriae with a deep crater-caldera open to the northeast. From here lavas of olivine basalts have moved to the east, but huge masses of pyroclastics have also been ejected and settled down on the south flank. It was an activity of Strombolian nature. The steepness of the crater walls and the deep bottom speak in favour of a final drain into the throat. The volcano seems to be of Quaternary age. Its sides are covered with *tosca* (lime-crust).

The more recent volcanoes in the vicinity are of a rather ordinary kind. They have sent large masses of lava into the surrounding lowland.

Finally, we should look at the southern appendage of Fuerteventura — the peninsula of Jandía *(»Dehesa de Jandía»)*. It is separated from the main bulk of the island by a low, sandy isthmus at Matas Blancas. Jandía is a high and sharp-crested mountain ridge with a maximum height of 807 m and a total length of 25 km. Its axis runs NE—SW to W in a slight bow and the cross-profile is asymmetric; the windward side *(barlovento)* is very steep down to a coastal flat — the leeward side *(sotavento)* is more gentle ending in a sandy coast beach. A number of barrancos dissect the long slope, slightly radiating from the crest line.

This peninsula is mentioned here because it has been considered the remaining half of a giant caldera volcano (stratovolcano), the missing part of which has disappeared into the sea in the northwest. The chief advocate of this idea has been S. Benitez Padilla (1945), who stated that the cause of destruction has been marine abrasion over a considerable time.

The present author must confess he cannot share the opinion about the former existence of such a gigantic explosion caldera, for he has not been able to find any traces of this caldera. The great escarpment seems to be a topographic feature created by faults and modified by subsequent marine attack and weathering and erosion of the slopes.

As far as supposed explosions are concerned, no traces of the ejecta are to be found in the broad southern slopes. As I have tried to show in an earlier paper (Hausen, 1956) the basalt formation in Jandía is contemporaneous with the so-called »cordillera oriental» accompanying the whole east coast of the island. The basalt lavas composing the same have issued from a train of fissures along the west coast of the island, now hidden by the ocean. This eastern cordillera and Jandía are the remaining parts of a basaltic tableland formation that once covered the entire island (and more) but has in later times been largely destroyed by erosion, so that the subjacent old formation of spilite lavas has in parts been dismantled.

LANZAROTE AND ALEGRANZA

In the extreme northeast of the archipelago there are some islands of a certain interest in this connection.

Lanzarote, 796 sq. km in area, is a »purely volcanic island» *(La Isla de los Volcanes*, as it is called in the tourist literature) crowded with several hundred cinder cones mostly of moderate size, with none dominating the landscape as a central cone (Fig. 15). True volcanic calderas, if we mean the type represented by Caldera de las Cañadas in Tenerife, are not found here. Instead, we have plenty of »horseshoe-calderas» in the summits of many of the larger cones, always with the low rim facing north. Some of these attain considerable circumferences (Fig. 16). They are all the result of explosions with the ejection of slag, lapilli and ashes. All the air-borne material has come under the influence of the strong north to northeast trade wind, hence the asymmetric form of the calderas. [1]

But there is also another type of crater-caldera in the summit of the volcanoes. The best example is to be found in the north of the island in Montaña de la Corona (610 m). Here is a deep and rounded crater-caldera open to the north in its upper part, and this one has been formed

[1] These wind directions seem to have persisted since Quaternary time.

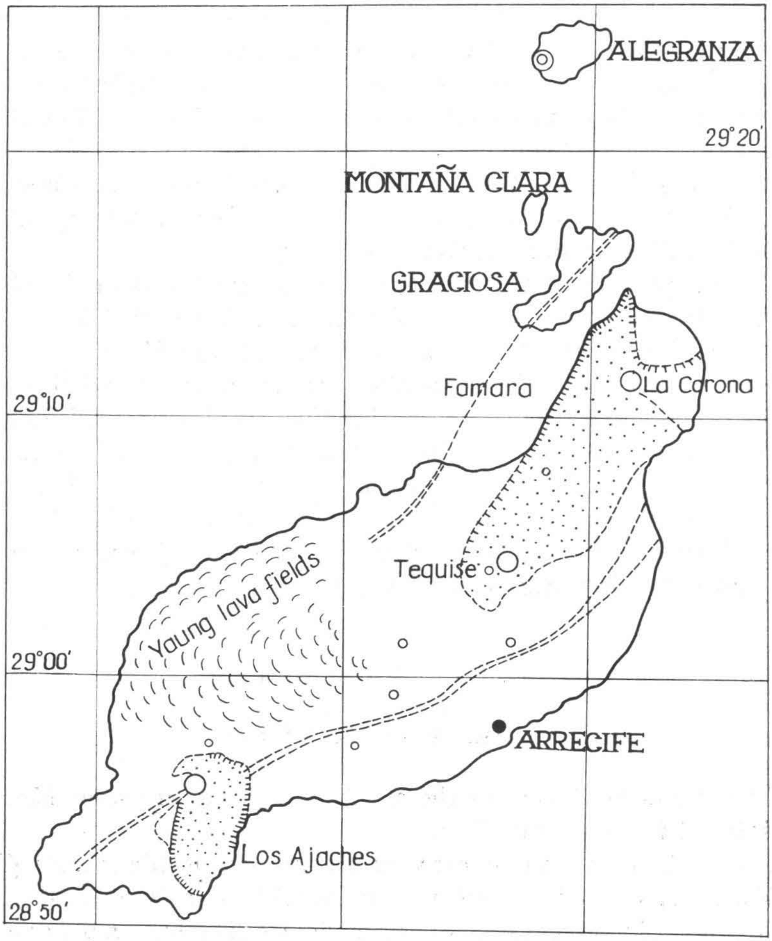

Fig. 15. Contour map of the island of Lanzarote with Graciosa, Montaña Clara and Alegranza. Some of the crater-calderas are indicated, also some of the (older) volcano-tectonic lines. Dotted areas in the north and in the south: old basaltic table-land formation.

not by explosion, but by the draining off of lavas into the throat-interior after the eruptions had ceased. In the threshold of the northern side one can still observe the consolidated lava stream that escaped the crater when it was filled with lava up to this threshold.

A similar emptied crater-caldera is M o r r o Q u e m a d o in the vicinity of Magues, not far to the south of the former volcano. This caldera is a closed form, but lava has escaped through a fissure in the eastern flank and flooded the valley of Magues. The draining of lava was consequently not

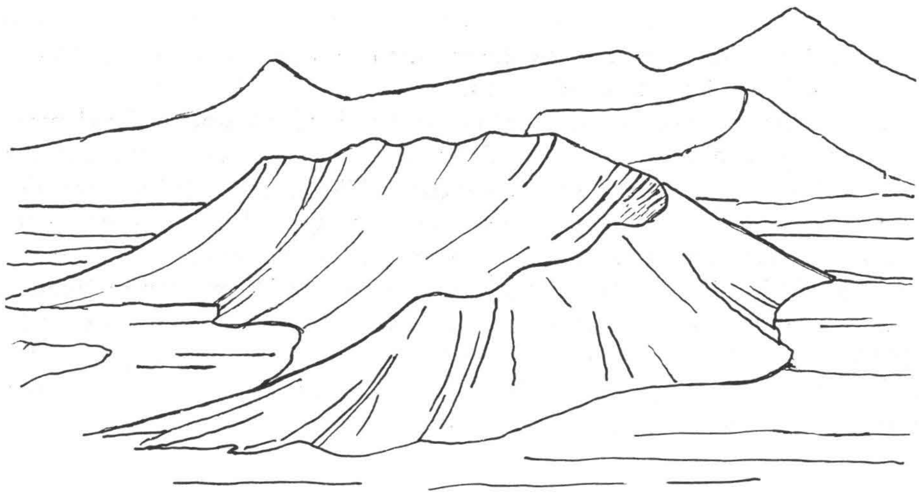

Fig. 16. A typical »horse-shoe» crater-caldera of clasmatic material surrounded by lava fields in the region northeast of Yaiza, Lanzarote. Asymmetry due to the trade wind (from the left). Looking southeast. Field sketch by H. Hausen.

conduced into the interior, but outwards. Somewhat similar is also the small Caldera de Guatifay at Teguise, farther south. It is likewise closed on all sides, but lava has escaped from the south flank.

All these volcanos consist of olivine basaltic material. They differ in age, ranging from the Quaternary to the Recent period.

One of the latest groups of volcanoes in the island, Tinguatón, (eruptions 1824) also has a kind of crater-caldera, the bottom of which shows empty vertical shafts, from which hot water issued after cessation of lava emissions through an opening in the north wall (Hausen, 1959).

At last we must consider the small island of Alegranza (12 sq. km) lying to the north of Lanzarote and resting on the same submarine shoal that surrounds both Graciosa and Lanzarote. Alegranza, briefly visited by the author, is of interest because of its large caldera on the west coast, called simply La Caldera (289 m). It has a diameter of about 1 200 m and is closed. During rains there is a shallow lake in the bottom. The walls consist of basalts, whether lava or clasmatica I do not know, since I only passed by on the east side. It looks, however, as if the encircling wall is composed of slag and lapilli because there is an abundance of such materials in the southern surroundings. Most likely the caldera is of the explosion type and is of rather old age, as can be seen from the achievements of marine abrasion on the west side, where precipices 200 m high have been formed.

The very insignificant islote Roque del Este lying between Lanzarote and Alegranza seems to be the remnant of a tuff caldera, almost completely destroyed by marine abrasion.

Now we have ended our wanderings in search of calderas in the Canaries, and have found a great number of depressions partly of volcanic nature, partly not, or perhaps only to some degree. The chief work has been done by exogenic forces. There is in fact a bewildering multitude of types both in size and shape and also in age and state of preservation. It is a difficult task to find plausible explanations as to the mode of formation of them all, as already pointed out. A scrutiny of the existing geologic literature suffices in itself to give the »calders question» an air of mystery. In the next chapter we shall devote some pages to classifying all the types met with in the islands according to shape and mode of formation.

AN ATTEMPT AT A CLASSIFICATION OF THE CALDERAS ACCORDING TO FORM AND GENESIS

If we first consider the group of true volcanic calderas with better preserved original forms, at least with regard to circumference, we obtain two subcategories: large-scale explosion-collapse-calderas and small sink hole-explosion-calderas. There seems to be only one representative of the former category in the Canaries- that of Caldera de las Cañadas in Tenerife. Only its eastern and southern walls have been preserved, whereas in the north some short segments are to be found. In the other directions younger lavas cover the rim completely. Nothing is to be seen of the bottom, the caldera has been filled up to the thresholds and a new volcanic form has grown inside it.

We have several types of the second sub-category. First there is a kind of sink-hole, smaller collapse-caldera without any visible traces of explosions (although there are some dubious cases). Arafo in the east side of Cumbre de Pedro Gil, Tenerife, is such a caldera. There is also Caldera de los Marteles in Grand Canary, although it seems the depression originated in a cone, the summit of which seems to have collapsed after explosions exhausted the throat. La Caldereta on the same central highland also seems to belong to this group of explosion and collapse.

Then we have types of phreatic explosion calderas (calderetas) without any cones. They can be designated as maars. A fine example is Caldera de Bandama in Grand Canary.

Another type includes the summit-crater-calderas in cones of slag and lapilli. Lava seems to have filled the throat but later on it was drained back into the depth. La Corona in Lanzarote is an illustrative example.

Further there is the type of »horse-shoe» calderas (craters) enclosed in larger cinder cones. They are either empty or the bottom is filled with lava bubbles (Fuerteventura). The asymmetric form of the cone is due to the prevailing wind (F. and Lanzarote).

There are also non-volcanic calderas, large depressions with outlet-barrancos and semi-circular (amphitheatre) heads. They seem to have originated in a sector of an island previously shattered by fractures. In these cases, weathering has deeply softened the rock formations. Great land-slides along the flanks have worked hand in hand with the erosion by running water and spring action in the precipices. These calderas are of great dimensions, and inside they have the aspect of rough erosion landscapes.

A special tectonic type seems to have originated by semicircular faults and segmental displacements, opening wide embayments in the island block. A prominent type is El Golfo in Hierro, but there are also some other islands with embayments of a tectonic nature such as the NW: coast of Grand Canary. The fractures are accompanied in parts by young cinder cones.

Pure erosion valleys with broad heads are found in La Gomera on the windward side, where the rock ground has suffered from intense weathering, and therefore offers little resistance to erosion. Here land slides have also taken place, widening the valleys. These forms are not really »calderas», although there is a certain transition to the type represented, for instance, by Caldera de Taburiente in La Palma.

If we now take all the cases under a common glance, we may state that the more apparent negative forms in the configuration of the islands are the erosion valleys and — depressions. With the younger cones, when present, they dominate the landscape panorama. Only Caldera de las Cañadas in Tenerife is an exception: its magnitude is sufficient to give a special stamp to the island in question. All the crater-calderas that are very numerous and widely distributed (especially in the eastern islands) are relatively small and mostly hidden inside the cones.

The famous Caldera de Taburiente in La Palma is, in fact, also of the category of the narrower calderas, at least in relation to the circumference of the island itself. It is the depth of this caldera that makes it distinctive; an infilling process, as in Caldera de las Cañadas in Tenerife, has not taken place.

Classification of the Canarian calderas

	Examples
1. Volcanic caldera, in the proper sense, formed by explosions and subsequent collapse and embayments.	Caldera de las Cañadas, Tenerife. (Really a twin caldera)

2. Smaller collapse calderas.	Caldera de los Marteles, La Caldereta, Gran Canaria; Caldera de Arafo, Tenerife.
3. Smaller calderas formed by phreatic explosion and subsequent sink.	Caldera de Bandama, Grand Canary; Caldera de S. Antonio, La Palma; Caldera del Rey, Tenerife.
4. Crater-calderas in the summit of cones formed by withdrawal of lavas down the throat.	Caldera de la Corona, Lanzarote; Caldera de Gairía, Fuerteventura; Caldera del Pico Viejo, Tenerife.
5. Crater-calderas of horseshoe form in cinder cones of larger dimensions.	Several cones in Lanzaronte and Fuerteventura.
6. Erosion and landslide calderas formed in fault-dissected parts of the island.	Caldera de Tejeda and Caldera de Tirajana, Gr. Canary; Caldera de Taburiente, La Palma.
7. Amphitheater calderas (semi-calderas) produced by semi-circular faults on an island slope.	El Golfo, Hierro.
8. Erosion and land-slide calderas.	Valle Hermigua and Valle Hermoso, Gomera.

CALDERAS AS A GENERAL PHENOMENON IN THE MACARONESIAN INSULAR AREA

By the expression Macaronesia is meant the assemblage of archipelagoes strewn over the vast surface of the Mid-Atlantic ocean in the middle and low northern latitudes, or down to the 15°N parallel. These archipelagoes are: the Azores, Madeira with Porto Santo and Desertas, Selvagem (Grande y Chico), Canary Islands and the Cape Verde Islands farthest to the south. They are not oceanic islands in a proper sense since they all lie on submarine socles. The Azores are on the Mid-Atlantic ridge, Madeira and Selvagem lie on ridges connected with Africa, and the Canary and the Cape Verde islands are on the continental slope.

In all these archipelagoes, calderas are a fairly common phenomenon; they are a part, so to speak, of the volcanic nature of the islands. But there is, of course, a wide range of types and sizes that cannot be fully understood until a general survey of them has been made. The best formed calderas (caldeiras) appear to be in the Azores, and these were admirably described a hundred years ago by Hartung (1860) in a great work provided with excellent panorama pictures. Also the great caldera in Madeira, Gran Curral, has been described and illustrated by the same author (Hartung, 1864). Later Gagel (1913) filled out our knowledge of Madeira in connection with his study of the La Palma caldera. As far as the Cape Verde islands are concerned, Friedlaender (1912) has given a fairly full account of the calderas in this region. Of more recent literature on Macaronesian volcanoes and calderas there is the well known compilation by von Wolff (II 1931). Lately an interesting study in the Azores has been published by Krejci-Graf (1956). A general discussion of the calderas in Macaronesia will lie outside the present scope. It should only be stated that the Tenerifan type of Las Cañadas with a central peak reappears in several of the archipelagoes, and calderas without a central cone also occur. Besides, there are many »calderas» of non-volcanic or of volcano-tectonic origin.

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10

ON THE GLACIAL GEOLOGY OF THE HÄMEENLINNA REGION, SOUTHERN FINLAND [1]

BY

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ABSTRACT

The movements of the continental ice sheet are described according to the evidence provided by striations and till fabric. Dealt with in addition are glacial forms met with in the area explored, including drumlins, moraine ridges (end moraines), eskers and marginal stratified drift. The last-mentioned formations belong, as parts, to the Second and the Third Salpausselkä. During its retreat, the ice sheet behaved in different ways in different zones and sectors.

CONTENTS

PREFACE

The present study is based on the mappings of surficial deposits carried out by the author since 1952 in the surroundings of Hämeenlinna, principally

[1] Received March 3, 1961.

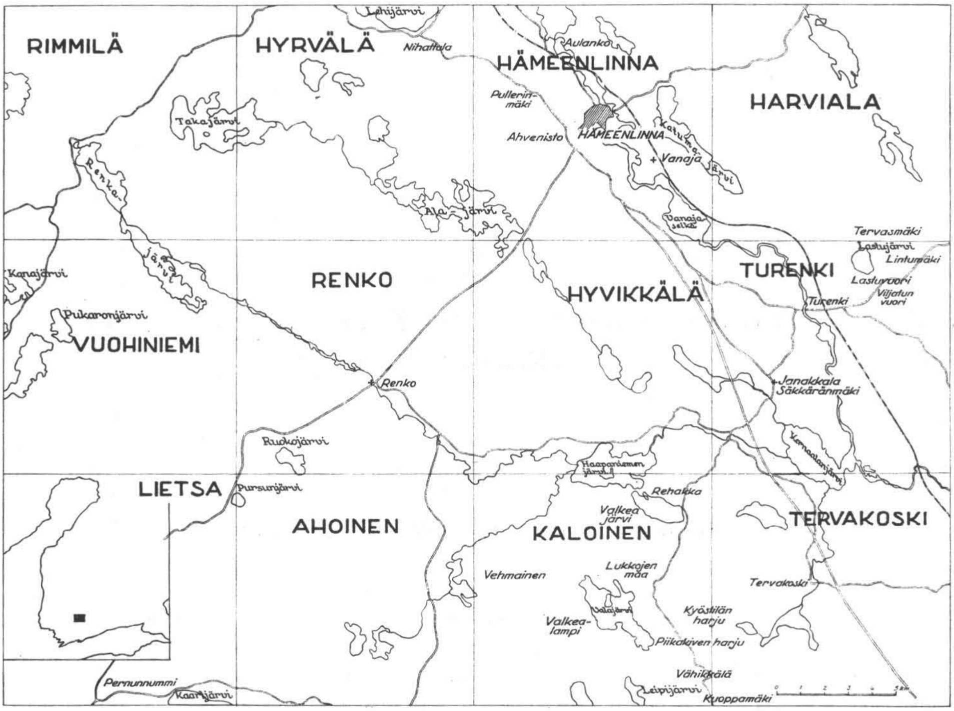

Fig. 1. The Hämeenlinna region investigated, its map sheet division, location in southern Finland and the place names mentioned in the present study.

in the region south of the town. The mapping operations proper have been performed in the area included in the topographic map sheet 1 : 100 000 of Hämeenlinna (No. 2131) and which in the following will be referred to as the Hämeenlinna Region.

The most interesting formations in the area under study are the Second and the Third Salpausselkä, fragmentary parts of which are met with in various localities. Previously, the Salpausselkäs in the region had been described by Leiviskä (1920). Since the appearance of Leiviskä's extensive work, new Finnish topographic maps have been published on a scale of 1 : 20 000, which has considerably facilitated following the Salpausselkä formations in the terrain. More recently, Okko (1957) has published a study dealing with the rather small area of Jylisjärvi, which borders directly on the northeastern corner of the Hämeenlinna region.

Figure 1 shows the area investigated, its map sheet division and the place names mentioned in the present study.

In addition to the author, the following persons have carried out mappings of surficial deposits in the area: Mr. J. Ahtiainen, M. A., Dr. R. Kanerva,

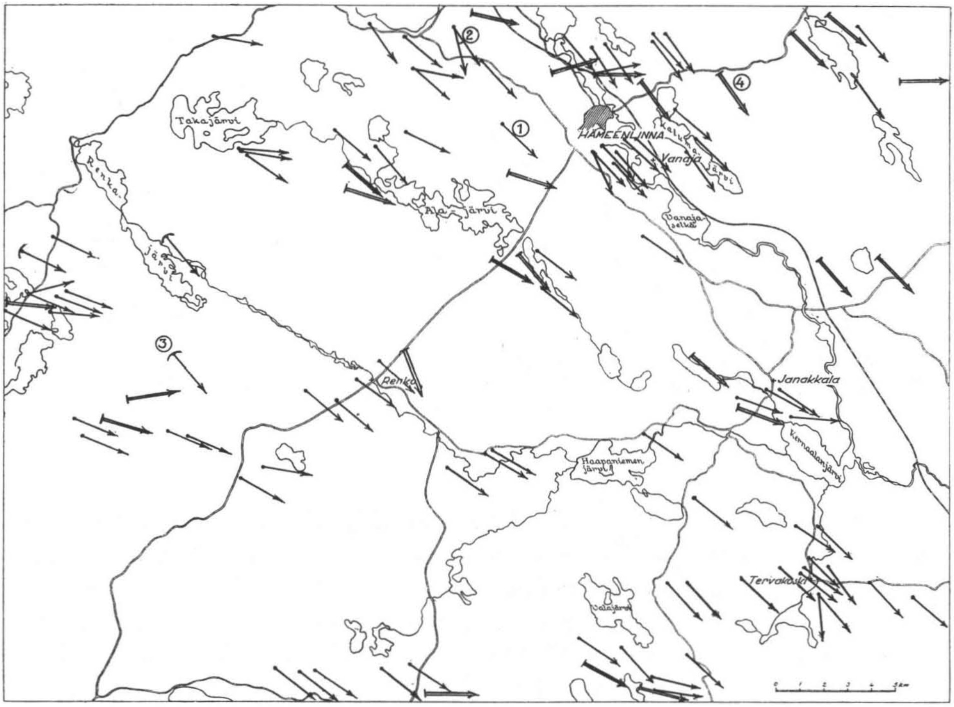

Fig. 2. The striations and till fabric in the region surveyed. 1 = trend of striations, 2 = trend of striations together with intermediate values, 3 = trend of erosion facets, 4 = preferred orientation of stones in till.

Mr. K. Korpela, M. A., Mr. P. Oivanen, M. A., and Mr. R. Tynni, M. A. I wish in this connection to express my appreciation of their conscientiously performed work.

GLACIER MOVEMENTS ACCORDING TO THE STRIATIONS AND TILL FABRIC

According to the striations, the strongest erosion direction in the Hämeenlinna region was, on the whole, from northwest to southeast (Fig. 2). In the western part of the area, the direction tended to turn slightly toward the west-northwest. Some three-quarters of the striae observations fall between azimuths 295° and 325°. Certain deviations from the prevailing erosion trend are, however, to be noted.

In the locality covered by the Hyrvälä map sheet, striae have been met with to the south of both Takajärvi and Lehijärvi which exhibit approxi-

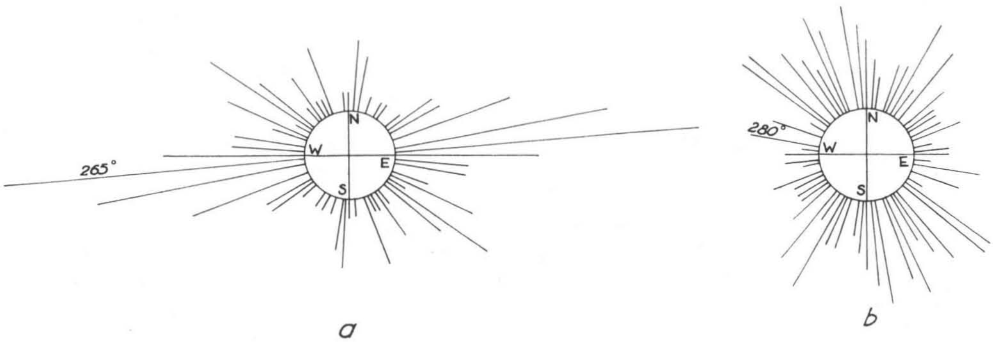

Fig. 3. Two fabric analyses from the till section at Aulanko. Analysis *a* is from a depth of about 1 ½ m, *b* from a depth of about 6 m. At bottom of section is bedrock, on which the trend of the striations is 260°.

mately a westerly trend. In the latter locality the westerly striae occur together with northwesterly ones as crossing striae, but it has not been possible to determine their relative ages.

Certain other observations regarding the westerly erosion direction have also been made. On the shores of Kanajärvi and Pukaronjärvi, in the Vuohiniemi map sheet, westerly striations have been observed in a couple of places. Similar observations have been made, furthermore, on the northern side of the town of Hämeenlinna, along the road leading to Aulanko, as well as at Janakkala, at the northern end of Kernaalanjärvi. In the Kaloinen sheet, westerly striae occur on the eastern shore of Leipijärvi as well as on the southwestern side of Ruokojärvi, at Renko.

Judging by the striations alone, it is difficult, however, to tell whether this westerly erosion trend should be interpreted as an independent glacier movement or whether it signifies only a temporary and local change in direction from the northwest-southeast flow of the ice sheet.

The same trends appear in the till fabric as in the striations. A northwesterly trend is commonest in the till fabric, too. Of the maxima of Fig. 2, seventeen observations reveal a northwesterly trend. But, in addition, a westerly orientation has also been met with in the till fabric — and even more distinctly and abundantly than among the striations. As many as seven fabric analyses indicate deposition of till that had taken place from the west; particularly clear are the fabric analyses made north of Hämeenlinna and along the shore of Leipijärvi. The former observation was made from the same pit at the bottom of which westerly striations were also met with. The upper part of the till bed, with a depth of about 1 ½ meters, has an admirably clear till fabric (Fig. 3 a). On the other hand, the fabric in the lower part of the till is indistinct, or no trend whatsoever can be perceived

Fig. 4. Till overlying on surface of glaciofluvial drift on the Second Salpausselkä, on the eastern shore of Leipijärvi. The till exhibits a weak laminated structure.

(Fig. 3 b). The deposition of the till and the final erosion of the bedrock thus plainly took place here from the west.

The analysis indicating the westerly trend of the shore of Leipijärvi was made from the summit of the high ridge belonging to the Second Salpausselkä. The ridge consists for the most part of sorted glaciofluvial gravel and sand. Even at the top of the ridge, the surface layer, measuring nearly two meters in thickness, consists of predominantly fine-sandy till, which, partly laminated in structure, had obviously been thrust on top of the sorted material from the west (Fig. 4).

Judging by the strations and the till fabric, the last local oscillation of the ice sheet took place, on the whole, from the west. The same kind of conclusion was previously reached by Sauramo (1929), too, in writing: »In Central Finland, Häme, the margin readvanced on a wide front for more than 10 kilometers, as appears from the great disturbances in the varved sediments.»

Sauramo ran across disturbances in the varved clay in three places in the commune of Hattula, north of Hämeenlinna. In connection with mapping operations performed in the area, similar disturbances have been met with, in addition, south of Lehijärvi, in the village of Nihattula, as well as in the railroad yard of Hämeenlinna Station. Disturbances have also been met with in certain places in the glaciofluvial material, possibly caused by a local advance of the glacier. One observation was made on the southwestern shore of Katumajärvi. Violent disturbances had completely demilished the original stratification in this locality, but no till has been encountered overlying the glaciofluvial material. Similar disturbances have been observed in the area of the Tervakoski mills as well, providing further evidence of a glacial advance. Sandy till overlies the disturbed sorted material in this area to a depth of some 1 1/2 meters. A comparable observation has been made, in addition, a couple of kilometers to the northwest from Tervakoski.

GLACIAL DRIFT

The glacial drift in the area may be classified on the basis of whether the material is sorted or unsorted. The former includes all the glaciofluvial formations in the area, while the latter consists of till. The glacial drift may also be classified according to whether its forms have any particular trend or not. In the former case the morphological trend conforms to the prevailing movement of the ice sheet or then runs perpendicular to it. In the latter case, the glacial drift is either lacking in forms of its own, like the ground moraines, or its forms lack any set trend, as evidenced by the ablation moraines.

TILL

Till comprises almost the exclusive component of ground moraines, which predominates among the till formations in the area. Moreover, till is the prevailing material in drumlins and in moraine ridges running parallel to the glacial margin, besides occurring to some extent in a washed state in ablation moraines. All told, till deposits cover forty percent of the land area under investigation. Actually, they are present on a considerably larger scale, for they are also met with in most cases underlying younger surficial deposits, such as littoral accumulations, clay and peat. In different parts of the region, the amount of till varies, however, to a great extent. Whereas in the areas covered by the Hämeenlinna, Turenki and Tervakoski map sheets, for instance, it accounts for thirty-five, thirteen and thirty-one percent, respectively, of the land area, the corresponding percentages for the Vuohiniemi, Hyrvälä and Harviala map sheets are fifty-six, fifty-two and fifty-one percent.

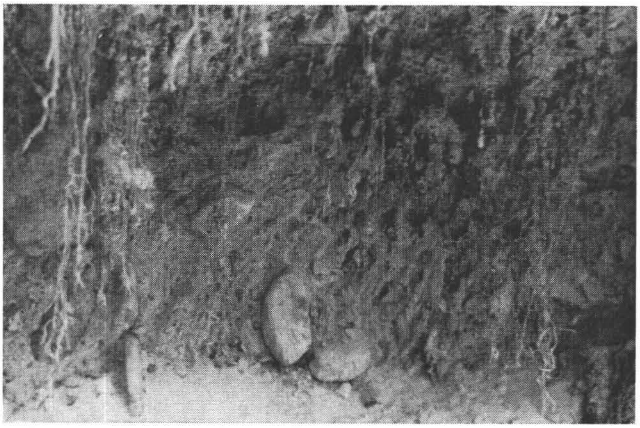

Fig. 5. Till with vertical schistosity in the Ahoinen map sheet. The majority of the elongated stones lie in a vertical position.

The ground moraines in the region lack surface forms peculiarly their own. Their topography is wholly determined by the morphology of the local bedrock. This is due mainly to the fact that the thickness of the till in the ground moraine localities is very slight. It appears to average only between three and five meters.

The till of the Hämeenlinna region exhibits relatively few structural features. In places there occurs till revealing a weak laminated structure (Fig. 4), as described by the present author previously from eastern Finland (Virkkala, 1949). In the area of the Ahoinen map sheet, vertically schistose till has been observed in a certain section, with even the stones lying in quite a vertical position (Fig. 5). Such a structure evidently was created during the melting stage of the ice, when watery drift oozed up from underneath the ice into a crevasse (Hoppe, 1952). A flamy structure (Virkkala, *op. cit.)* has further been noted to occur to a slight extent in the local till (Fig. 6).

The most characteristic feature of the mechanical composition of the till is its wealth of boulders, cobbles and pebbles. Quite prevalent are the pebbles, especially large ones, which are consequently all the more abundant; and generally they are angular and uneroded. Especially in the surficial parts of the till deposits and on the very surface of the ground, the content of boulders and stones is highest. To some extent this is due to secondary factors, the washing effect of waves and the tendency of ground frost to lift stones toward the surface. But primary factors can be cited, too. Surface stoniness is often most marked in places where the till cover is thin or where there are rock outcrops in abundance. The surface stoniness of the ablation moraines is further due to the fact that the finest material was

Fig. 6. Indistinct and folded sorted lenses in till exhibiting a flamy structure.

the first to become released from the melting ice, whereas the boulders and pebbles broke loose last and accordingly were deposited on top of the till. Some of the surface boulders occurring in till may, in addition, be material borne by ice floes.

The commonest component of the finest material in the till, *i.e.*, that measuring under 20 mm, is sand. Of some fifty samples studied to determine their mechanical composition, sandy till accounted for 67 percent, gravelly till for 12 percent and fine sandy till for 21 percent. Silty and clayey till has not been met with in the region. Figure 7 presents certain cumulative curves illustrating the till as well as the average grain size composition of fifty till samples. According to the curve depicting the last-mentioned, the mean grain size of the till in the class of material measuring under 20 mm is 0.4 mm.

An extraordinary type of mechanical till composition has been observed in the zone of the Second Salpausselkä and its vicinity. The spaces between the characteristic, angular till stones are filled with sorted fine sand and sand, with the finest grain sizes almost totally lacking (Fig. 7, Curve 1). It is obvious that the sorting of the material between the stones took place in part already during the sedimentation stage of the till, when the melt-water trickling out of the frozen drift washed the finest fractions out of the till. In part, again, it is possible to conceive of this type of till having originated in such a way that during the period of subglacial sedimentation stones had dropped out of the ice into the sorted material.

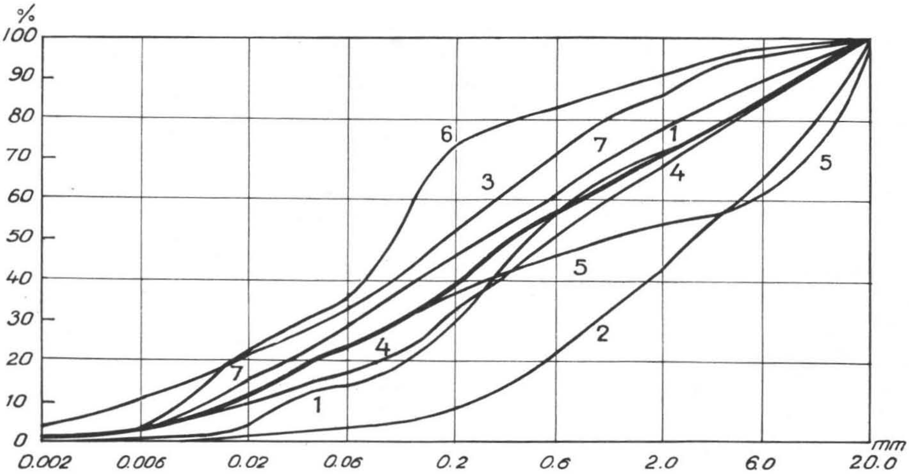

Fig. 7. Cumulative curves for the till in the Hämeenlinna region. The thick line represents the average mechanical composition of approximately fifty till samples. 1 = Salpausselkä till, Turenki, 2 = material of a small end moraine, Renko, 3 = fine sandy till, Tervakoski, 4 = material of a small end moraine, Lastujärvi, 5 = gravelly till, Hämeenlinna, 6 = fine sandy till, Rehakka, 7 = sandy till, Vähikkälä.

ABLATION MORAINE

The till lacking a morphological trend not only appears as a thin ground moraine cover but also exhibits surficial forms of its own. Especially in the western part of the Hämeenlinna region, moraine hummocks are characteristic of extensive areas (Fig. 8). They are usually between five and fifteen meters high (Fig. 9), with kettle-like hollows situated in between them, though in most cases the depressions are only more or less open and indefinite in form. In places among the hummocks there are small ridges whose trend varies or corresponds to the prevailing direction of the ice movement.

Ablation moraines are typical representatives of dead ice, *i.e.*, they are forms created by ice slowly melting in place (Hoppe, 1952; Lundqvist, 1937; Tarr, 1909). The ice had contained an abundance of unevenly distributed drift, and it was the melting of the fragments of ice covered with this drift that brought about the turbulent surface topography (Fig. 10).

Although individual moraine hummocks usually lack a morphological trend, extensive areas featured by ablation moraines often tend to stretch out as elongated zones running parallel to the most conspicuous direction of the movement of the ice sheet (Fig. 8). This feature is most plainly evident in the southern part of the Vuohiniemi sheet.

Among ablation moraines one meets in places also with hummocks and short ridges composed of sorted drift. They are mostly situated along

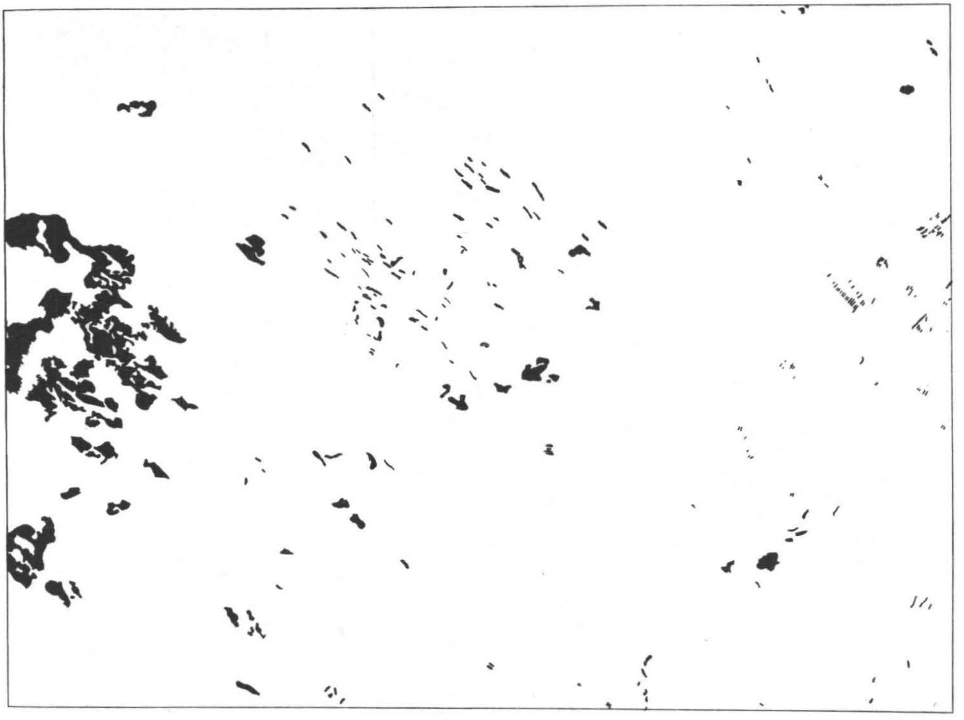

Fig. 8. Moraines in the Hämeenlinna region. The indefinite black figures indicate ablation moraines, met with especially at the western edge of the region. In the middle is situated the drumlin field of Renko-Alajärvi and in the eastern part are numerous end moraines, both large and small. Scale ca. 1: 320 000.

Fig. 9. A large moraine hummock situated in the ablation moraine area in the northern part of the Vuohiniemi map sheet.

Fig. 10. A detail from the ablation moraine area located in the Vuohiniemi map sheet. The ruled lines represent bogs, the space between the elevation curves is 5 m, scale 1: 20 000.

extensions of long, continuous esker chains. In the area of ablation moraines, the eskers regularly end or at least break off and lose their form.

For the most part, the material of the ablation moraine hummocks consists of till. It is not, however, the same kind as the till met with in areas of ground moraines but is usually more or less washed. It contains fine grain sizes only to a very slight extent; on the other hand, its material is characterized by an abundance of pebbles and boulders. The degree of washing is highest particularly in those parts of ablation moraines that are situated in the extensions or vicinity of esker chains. There it may be observed how gradually the washed till changes into distinctly sorted and stratified glaciofluvial drift.

Ablation moraine areas of slight extent are to be found outside the western part of the Hämeenlinna region. They are situated mainly in an exceedingly fragmentary belt a few kilometers wide, which extends from the southwestern part of the region to its northeastern corner (Fig. 8). The largest continuous moraine hummock area in the belt lies a few miles east of the church of Renko. A couple of small ablation moraine areas are also known in connection with the Second Salpausselkä, being located west of Tervakoski. Here and there in other places, too, ablation moraine hummocks or small groups of hummocks occur.

Ablation moraines are limited in the main to the supra-aquatic level or, at any rate, to areas that at the end of the Ice Age have been covered only with shallow water.

GLACIAL DRIFT PARALLEL TO THE ICE MOVEMENT

The glacial drift conforming to the movement of the ice sheet consists either of till or of sorted material. The former includes the drumlins, the latter the numerous eskers occurring in the region.

THE DRUMLINS

Drumlins occur in relative abundance in the area lying between Alajärvi and Renko, where dozens of them are to be run across (Fig. 8). Many of the drumlins have the character of rock drumlins, whose form and trend have been more or less clearly determined by the local bedrock and its fracture lines (Simonen, 1949). Included, however, are true drumlins, which consist wholly of till. On the basis of outward appearance it is often impossible to decide in which group a drumlin belongs. Solitary drumlins are met with here and there in other parts of the region as well.

The majority of the drumlins, however, lack any distinctive forms, as compared, for instance, with many other drumlin areas in this country. In some cases, the terrain reveals rather a more or less clear trend than being characterized by true drumlins. Among the formations, however, are numerous quite distinct drumlins. Most commonly, the drumlins have a fairly gentle slope, amounting to ridges five to ten meters high and from 100 meters to a kilometer long.

The drumlin area is fairly clearly demarcated. Its southeastern margin is bordered by an esker extending from the church of Janakkala in the southeast. Having reached the drumlin area, the esker abruptly terminates and appears on the southwestern shore of Alajärvi in the guise of just a few slight glaciofluvial occurrences. On the northwestern side of the drumlin area, again, the esker reappears as quite an imposing formation.

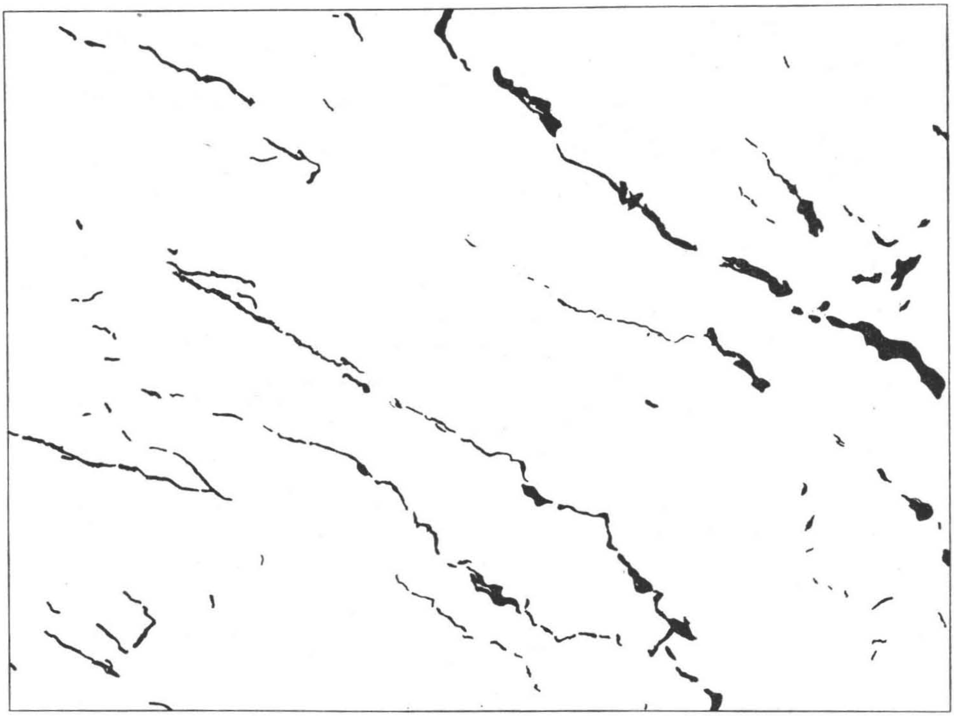

Fig. 11. The eskers of the Hämeenlinna region. Scale ca. 1: 320 000.

The material composing the drumlins varies considerably in the region. To some extent it consists of washed till or, at least, there occur sorted lenses within the till, which brings to mind till described as exhibiting a flamy structure (Virkkala, 1949). On the other hand, the drumlins of the region commonly consist of the same sort of sandy till marked by a heavy content of cobbles and pebbles; this is typical of the ground moraines of the entire area investigated. The occurrence of boulders on the surface is a general characteristic of the drumlins too. In some cases, the drumlins lack cobbles and boulders, small, angular pebbles being so much the more abundant. The stones of the drumlins in these cases have no preferred orientation. Where there is a smaller content of stones, again, their orientation conforms to that of the ridge.

THE ESKERS

Some ten eskers, esker chains or parts of eskers are to be met with in the area investigated (Fig. 11). The largest eskers, namely, the Hämeenlinna

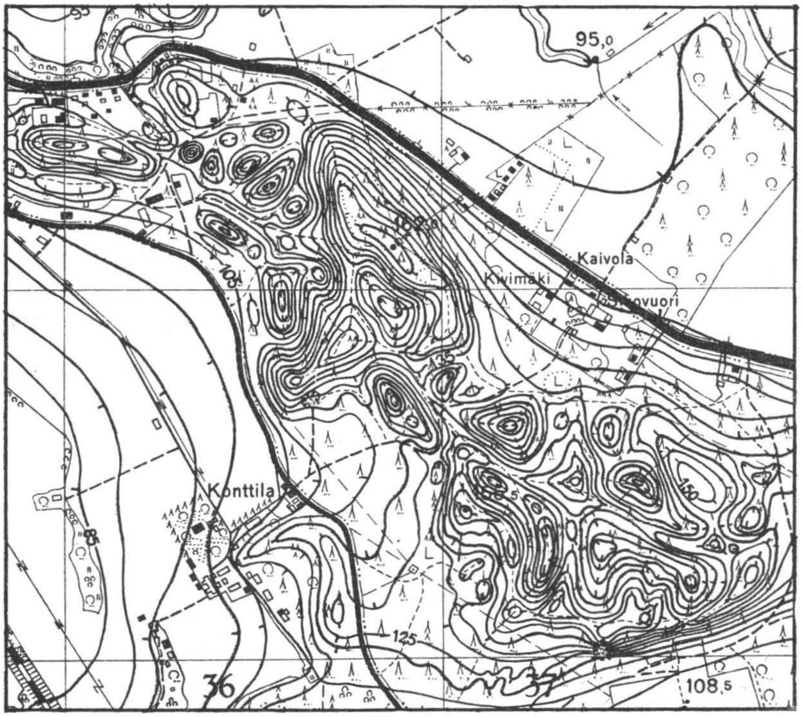

Fig. 12. Pitted esker topography on southeastern side of Turenki. Scale 1: 20 000.

—Turenki, Janakkala—Takajärvi, Renko and Vehmainen—Ruokojärvi chains, by and large run parallel to the watercourses. They are not situated, however, at the deepest points of the preglacial valleys, now submerged, but rather on the slopes of the valleys. In some cases, the eskers run at an angle across a valley, as, *e.g.*, at Turenki. The smallest eskers tend also to run across the local watersheds.

The ridge-like esker is by no means the sole esker form in the region. In many an instance, the esker has been broken up into a »kame-and-kettle field,» where large esker kettles alternate with hummocks and chains of hummocks. The largest esker kettles are as much as between sixty and seventy meters deep, at least on some sides, and, considered in general, the latter figure may represent the maximum depth for an esker kettle (Flint, 1957). The usual depth of these esker kettles, however, is less than thirty meters.

This kind of esker topography created by dead ice is particularly conspicuous in the surroundings of Turenki (Fig. 12), on the western and northwestern sides of Hämeenlinna, to the northwest of the church of Renko, on

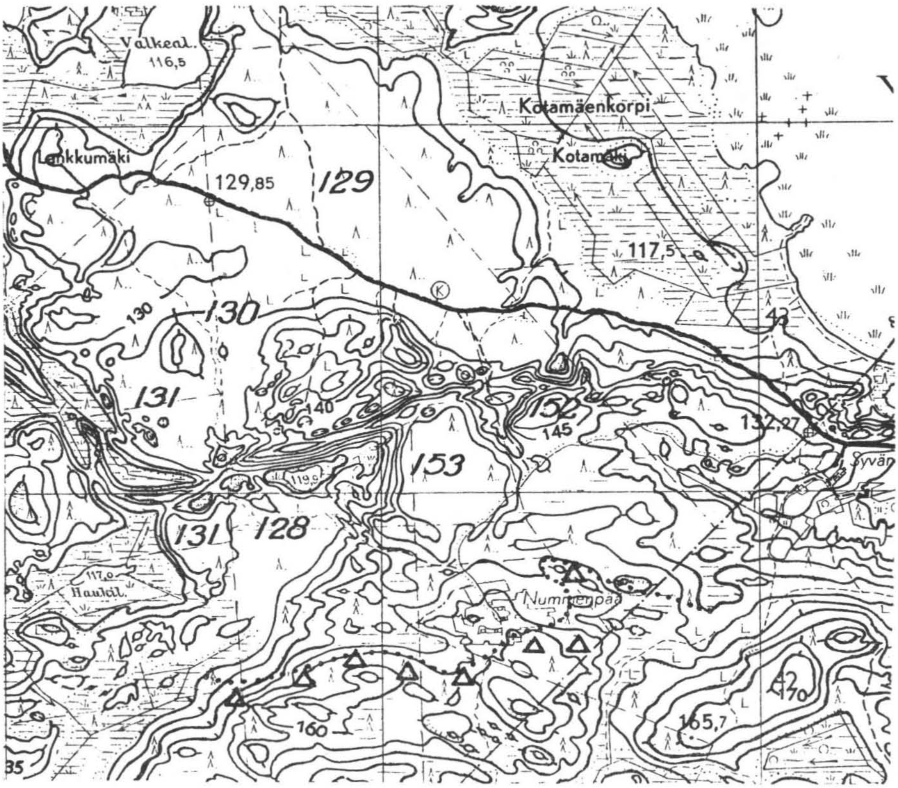

Fig. 13. Elevations of esker plateaus in the southern part of the Kaloinen map sheet. The elevation of two of the esker plateaus corresponds to the maximum height of the Second Salpausselkä, 152—153 meters. On their western and northern sides spreads broad Valkea-lammenkangas, which has levelled out at an elevation of approximately 130 meters. The numbers are levelled plateau elevations, the dotted line represents the border between the till and the sorted drift and the triangles indicate till. Space between elevation curves is 5 m. Scale 1: 20 000.

the southwestern side of Tervakoski and in the southwestern corner of the area included in the present survey. All these localities, together with numerous smaller occurrences, lie within the zones of the Second and Third Salpausselkä. In the intermediate and outlying areas, the form of the eskers is more definitely ridge-like and fewer esker kettles occur.

In many places the highest points of an esker spread out to form plateaus of lesser or greater extent. In the zone of the Second Salpausselkä, the summits reach an elevation of 150—155 meters, but some distance within they fall to a niveau some 25—30 meters lower. According to Sauramo (1958), this signifies a sinking of the level of the Baltic Ice Lake situated in the Baltic basin to the level of the ocean. On the northwestern side of

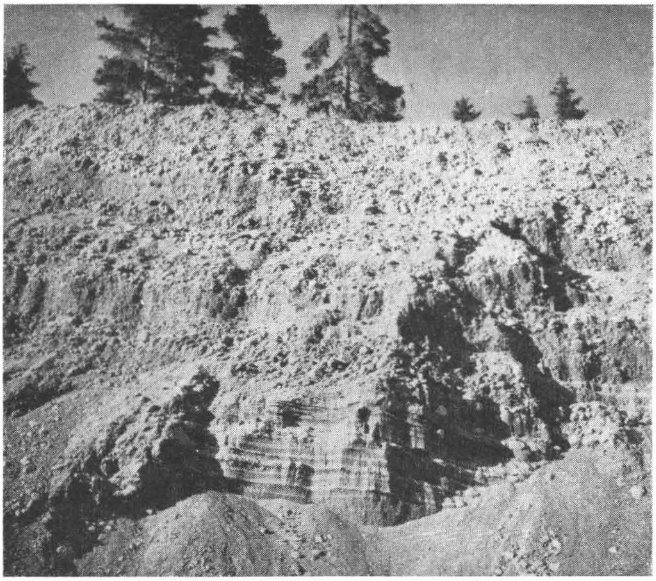

Fig. 14. Coarse glacioflucial drift from Pullerinmäki, Hämeenlinna esker. Layers are gently inclined in the distal direction. Downward, in the middle, is finer material, which resembles the structure of varved sediment.

Turenki, the elevation of the esker plateaus attains to 130 meters and in Hämeenlinna to approximately 135 meters. In Renko, on the western side of Valajärvi, the elevation of Valkealammenkangas is about 130 m, whereas that of the flat summit situated immediately to the southeast 152—153 m (Fig. 13). North of Valajärvi, the plateau of Lukkojenmaa is 157 m, above sea level but that of Valkeajärvi, situated to the northwest of this locality, has an elevation of 127 m. The lower elevation rises steadily and gently toward the northwest. On the northwestern side of Renko it is 135 m, on the northern side of Takajärvi 140 m and in the northwestern corner of the area 145—150 m. South of Renkajärvi a short, solitary esker ridge reaches an elevation of 165 m. Everywhere else on the proximal side of the Second Salpausselkä the eskers fail to rise even to the elevation of the lower plateau level.

Besides the four large eskers mentioned in the foregoing, numerous shorter and smaller chains occur in the region (Fig. 11). They are less continuous and more often take the form of a typical ridge. Especially in the aforementioned ablation moraine area, short, fragmentary eskers gradually change in material composition to slightly washed till, which is typical of ablation moraines. It is not even nearly possible always to draw the line between an esker and an ablation moraine.

The internal structure of the eskers is everywhere in the region quite true to type and exhibits clear ice-contact features. The variations in grain size, which are often very marked, the collapse of glaciofluvial drift after the melting away of ice fragments buried underneath, slight dislocations and other minor disturbances characterize the internal structure of the eskers. In many places, however, the structure remains undisturbed. The bedding is generally very distinct in the eskers. The beds have a distal tilt, whether slight or steep. In cross-section the bedding mostly appears horizontal, conforming to the surface or lying at an angle. Cross-bedding is likewise commonly observed in the eskers. Figure 14 gives one example of the structure of the esker of Hämeenlinna.

MARGINAL GLACIAL DRIFT

Marginal glacial drift embraces formations consisting of both till and of glaciofluvial drift. The former include moraine ridges of various sizes, the latter transverse eskers running parallel to the edge of the ice sheet as well as proglacial, delta-like plains. The marginal glacial drift of the area under investigation is met with in the Second and the Third Salpausselkä, which run discontinuously and in a variety of shapes at an angle approximating a line from the southwest to the northeast across the region.

MORAINE RIDGES AND ABLATION MORAINES

For the greatest part, the moraine ridges in the region — annual moraines (De Geer, 1889), De Geer moraines (Hoppe, 1959), washboard moraines (Mawdsley, 1939) — are smallish, five to ten meters high at most and less than a kilometer long, but usually only between one and five meters high and about a hundred long. These ridges consist chiefly of till and consistently run from southwest to northeast. In form they are usually somewhat asymmetrical, with the distal flank generally steeper, though also ridges whose proximal side is steeper have been observed. In the area investigated, they lie almost exclusively in the zone of the Second Salpausselkä and its distal margin. They are exceedingly scarce beyond it. All told, nearly 300 of them have been met with in the region in connection with mapping operations.

Moraine ridges occasionally occur in isolation, primarily outside the Salpausselkä zone. In most instances, they are to be found there, too, in small groups, either overlapping or lined up in a row. In Turenki there

Fig. 15. Small end moraines from Turenki. Dots indicate fine sand, triangles till, circles sand and gravel, lines clay and silt. Scale 1: 20 000.

occur some twenty moraine ridges in a row over a distance of about 1 1/2 km (Fig. 15). The most extensive group of them, however, is situated in the southeastern corner of the Harviala map sheet, where about forty have been mapped within an area of slightly more than one square kilometer. From this locality they continue in nearly equal abundance to the side of the Jylisjärvi sheet (Okko, 1957).

The material composition of the moraine ridges varies considerably. The largest stones and boulders are situated on the surface of the formation, the summit of which is in spots completely covered with boulders (Fig. 16). In many cases, the material consists of washed till, the sorting of which generally becomes clearer upon proceeding downward. Curve 2 of Figure 7 represents the grain size composition of a certain moraine ridge consisting of washed till.

Neither are moraine ridges composed of the typical dense basal till rare. On the eastern shore of Lastujärvi there rises a hill some 20—30 m high, the northern part of which consists on the surface of till and the southern part of sorted drift. On the hill are situated some twenty small moraine ridges, the material between which at the surface consists to a depth of about 1 1/2 m of sand. The material of the moraine ridges, on the other hand, is dense and hard basal till, as in a ground moraine, and identical

Fig. 16. Stony summit of small end moraine, Turenki.

to the rest of the till on the hill (Fig. 7, Curve 4). In excavating work the observation has been made that the material of the moraine ridges continues underneath the sand to merge into the till composing the hill. The ice sheet had thus in this area built up the moraine ridges during the deposition stage of the basal till, and the surficial sand in between should be regarded as some kind of littoral accumulation.

The moraine ridges described in the foregoing have been described as having resulted from the annual oscillations that occurred during the retreat of the ice (De Geer, 1889; Sauramo, 1929) or from calving of the ice (Hoppe, 1948; Mölder, 1955). In the event that the moraine ridges are understood to indicate the annual retreat of the ice sheet, this would mean, in the Turenki area, for instance, that the 4 km broad belt formed by the Second Salpausselkä, where these ridges occur, had evolved during a period of only fifty or sixty years. Sauramo's (1923) clay chronology, however, calls for a period three or four times longer. It therefore appears evident that each moraine ridge had evolved during the course of several years (Okko, 1957). Accordingly, they could not be annual moraines, though in respect to form, composition and position the resemblance is complete.

Besides the aforedescribed minor moraine ridges, till occurs in ice-marginal deposits also as solitary ridges on a larger scale running parallel to the glacier margin. In the region under investigation such formations are met with only in association with the Second Salpausselkä, though not many in number.

Approximately four kilometers north of Tervakoski, there is a moraine ridge about 25 m high and 1/2 km long with a southwest-northeast trend.

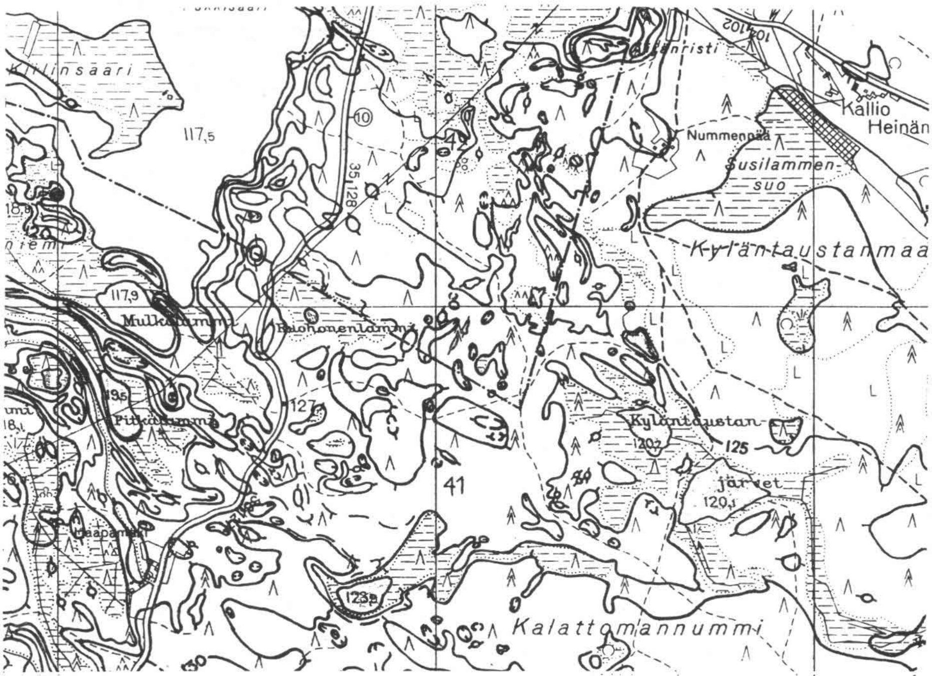

Fig. 17. Undulating proximal part of Pernunnummi, from which several eskers start at left margin of picture. Ruled lines are bogs, space between elevation curves 5 m. Scale 1: 20 000.

At its southwestern end, the ridge borders on a high rock outcrop, which clearly provided support while it was forming.

At the western end of the village of Vähikkälä, there are situated a few narrow-crested and fairly high moraine ridges with approximately a north-south trend. Through one ridge a large cut has been made alongside the Vähikkälä—Vehmainen highway, revealing that the material consists of the sandy till typical of the region. The distal slope of the moraine ridge, on the other hand, is covered with sorted littoral sand. Farther north the same ridge consists wholly of sorted drift. The material of the Second Salpausselkä in this region thus appears to be composed of till in many cases where it forms narrow, esker-like ridges.

On the northwestern side of Tervakoski, within the zone of the Second Salpausselkä and forming an organic part of it, there is a small ablation moraine area. It includes a few ridges running parallel to the margin of the ice sheet and to the movement of the ice as well as some roundish and steep-sloped moraine hummocks 20—25 m high. Except for this minor occurrence, nowhere else in the region have moraine hummocks belonging to the Salpausselkä marginal formation been met with.

Till occurs in places in the marginal formations of the region not only as independent forms but also as an even covering on the surface of the sorted glaciofluvial drift. Mention has been made already in the foregoing (p. 219) of the till-covered transverse esker situated on the southern shore of Leipijärvi. In the southwestern corner of the region investigated is situated part of the extensive ice-marginal formation known as Pernunnummi, which belongs to the Third Salpausselkä and the proximal edge of which forms typical dead ice terrain, featured by hummocks and kettles. In numerous places the sorted drift there is covered by a more or less distinct layer of till usually measuring less than a meter in thickness. This layer did not form over the glaciofluvial drift as a result of any thrust forward by the glacier but evolved from drift situated at a higher level in the ice — as the ice disappeared this material was the last to be deposited over the sorted drift (Fig. 17).

STRATIFIED DRIFT

Besides the moraine ridges and ablation moraine, the Salpausselkäs in the Hämeenlinna region also include numerous ice-marginal formations composed of glaciofluvial drift. In part they are ridges, transverse eskers, paralleling the glacier margin, in part, again, more extensive, mainly proglacial, delta-like plains.

The transverse eskers in the region do not compose long, continuous ridges, but short, fragmentary formations, met with more often in the Second than in the Third Salpauselkä. The largest ridges in the northeastern part of the Turenki map sheet, Lastuvuori, Tervasmäki, Lintumäki and Viljatunvuori attain to an elevation of approximately 160 meters, which, according to Sauramo (1958), corresponds to the maximum elevation of Salpausselkä in the region. In form the transverse eskers are in many cases asymmetrical, their proximal side being steeper and higher than the distal side. Along the proximal flank are commonly situated esker kettles and other ice-contact features, too, indicating the ice margin to have been closed there during the genesis of the formation. In places ice-contact features can be noticed also in the distal part of the transverse esker.

From Turenki toward the southwest the marginal formation disappears, but in the vicinity of Janakkala church it forms a few large, solitary hummocks, including Säkkäränmäki, Kierteenmäki and Lystinmäki. Farther south, on the shore of Kernaalanjärvi there are numerous transverse eskers of varying sizes. The most distinct transverse esker, however, is Piikakivenharju, situated on the northern side of the village of Vähikkälä. Its southern part is composed of till, as has been pointed out in the foregoing (p. 234).

Finally, on the southern side of Vähikkälä there is an arched glaciofluvial accumulation, Kuoppamäki, the Finnish name of which in itself reveals the occurrence there of numerous esker kettles.

On the Third Salpausselkä a transverse esker has been met with only at the proximal edge of the aforementioned Pernunnummi, except for a short stretch of about 1 1/2 km on the eastern side of Takajärvi.

Typical of all the transverse eskers is a more or less asymmetrical form. The bedding throughout these formations generally slopes in a distal direction. Dislocations have often taken place in the bedding of the proximal part, having been caused by the withdrawal or melting away in place of the ice supporting the proximal edge. The disappearance of the ice wall caused the material to collapse in the direction of the removed support, disturbing the original bedding.

In addition to the transverse eskers and associated kames, glaciofluvial drift occurs in the marginal formations of the Hämeenlinna region as delta-like plains, some small, some more extensive. They are not deltas proper in the sense that it is possible to distinguish in them parts originating above and below the waterline. Nor do they bear vestiges of channels along which the last meltwaters flowed across the delta surface. Moreover, they lack the typical delta structure, including bottom-set and top-set beds and the fore-set beds in between. The plains are generally covered by littoral accumulations and it appears in many cases that the surface had subsequently undergone secondary levelling. Possibly, at least some of these plains represent incomplete deltas, which failed to grow up to the water level but over the surfaces of which littoral accumulations later spread in thick layers, levelling out the original surface forms.

The deltas of the Second Salpausselkä in the Hämeenlinna region are very small in area. The large transverse eskers of Turenki mentioned in the foregoing (p. 235) have barely reached the highest level of the Second Salpausselkä — an elevation of about 160 meters — but have not spread sidewards to any noteworthy extent. The highest point of Turenki esker is 162 meters.

On the northern side of the village of Vähikkälä there extends the first minor delta, Kyöstilänharju, the surface of which is 156 m above sea level. This elevation of the delta continues on a level for a distance of several kilometers in a proximal direction from the Second Salpausselkä (*cf.* p. 230).

The most extensive plateaus occur in the zone of the Third Salpausselkä. The Ahvenisto district of Hämeenlinna has several plateaus of small extent at an elevation of 135 meters (Pullerinmäki, and the plateaus of the »dog school», of the local cemetery and of the sanatorium). At the southeastern end of the transverse esker of Takajärvi there is likewise a small plateau at the same elevation. The Renko plateau, on the other hand, is considerably larger, comprising a total of some thirty square kilometers. There the maximum elevation of 136—137 m has been reached only in its proximal part, which in places contains numerous kettle holes created by dead ice. For the most part, the plateau is situated, however, below the 120 m level of the elevation curve. The distal part of the delta is by no means level but broken up by ice kettles, postglacial river erosion and islets of till and bedrock. It has not nearly attained to the level of the highest waterline but represents an incomplete stage of the process involved. The esker of Renko runs across the middle of the plateau. This indicates that the plateau as a whole had not evolved simultaneously, although the main parts had formed while the margin of the ice sheet remained in the proximal part of the plateau. In its distal part the material of the plateau consists of very fine sand, in the middle of fine sand and in its proximal part of gravel and sand as well as coarser fractions, too.

Only half the size of the Renko plateau is the one situated in the vicinity of Pursunjärvi, to the southwest. This plateau is likewise only indistinctly delta-like, being uneven on the surface in places. Its proximal parts achieve an elevation of nearly 140 meters, while its distal parts are about 115 m above sea level. The main part of the plateau lies between the elevations of 120 and 130 meters.

The most extensive of the deltas of the Third Salpausselkä is Pernunnummi (Fig. 17), which is situated in the area under investigation only in part. Including the lakes, bogs and islets of other surficial deposits, it comprises a total area of some fifty square kilometers. In the proximal part, to which belong the aforementioned transverse esker and an area of pitted landscape, the plateau reaches an elevation of 130—135 m. That of the distal part is about 120 m, while the main part of the plateau is situated approximately 125 meters above sea level. Broad stretches of the delta are quite as flat as a table, but numerous kettles, of which the largest are covered by lakes, are characteristic of its surface. From the proximal margin of the plateau there extend five eskers, of which, however, only one is of considerable size; it covers quite a distance lengthwise, reaching as far even as the distal part of the delta, on the shore of Kaartjärvi. Along the southern margin of Pernunnummi runs a minor esker, which beyond the delta, both to the northwest and to the southeast, immediately begins to expand considerably.

DEGLACIATION

Among the glacial formations of the Hämeenlinna region, several different zones may be observed in the direction of the retreat of the continental ice sheet (Fig. 18). The zone situated farthest to the southeast and, at the

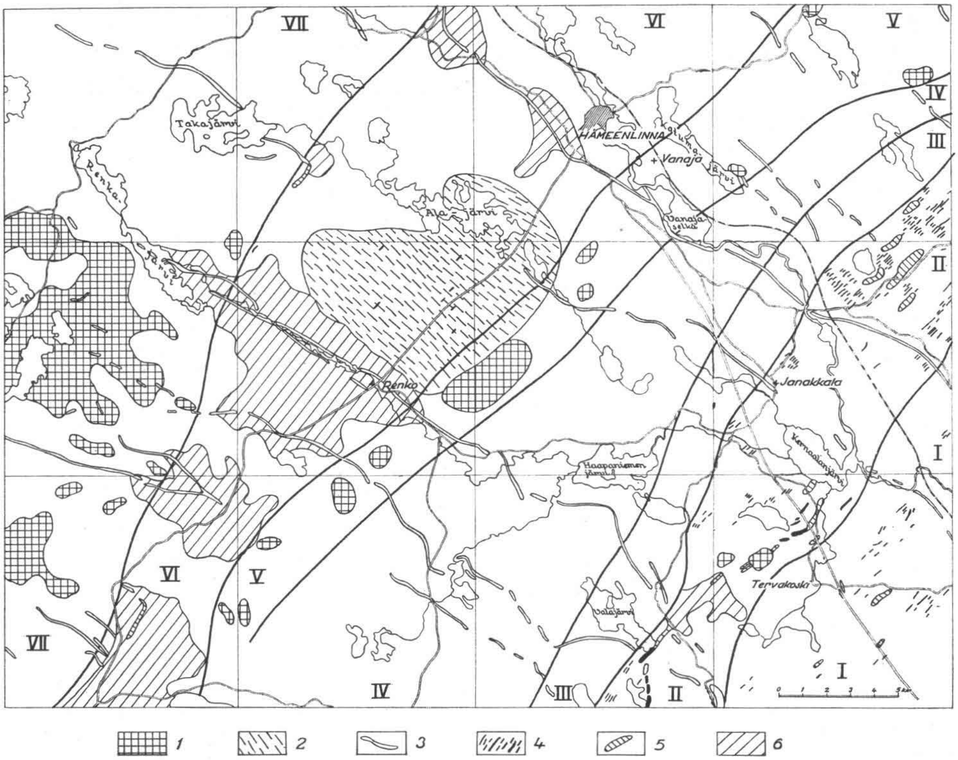

Fig. 18. Retreat of continental ice sheet and certain glacial forms in the Hämeenlinna region. 1 = ablation moraine, 2 = drumlins, 3 = eskers, 4 = end moraines, 5 = transverse eskers, 6 = delta-like marginal formations, I—VII = functional zones of retreat of the ice sheet in the region investigated.

same time, the one first to be released from the grip of the ice lies between the First and the Second Salpausselkä (Fig. 18, I). It is characterized in the region surveyed by numerous small moraine ridges and transverse eskers, which occur in isolation or in small clusters. The retreat of the ice sheet did not, therefore, take place in the area between the First and the Second Salpausselkä uniformly or continuously, but there occurred halts of greater or lesser duration, and even temporary advances of the ice again. The retreating margin became active from time to time, which means that the ice grew thicker, probably as the result of increasing precipitation.

In the belt comprising the Second Salpausselkä, Okko (1957) noted three stages in the Jylisjärvi area: series of outer marginal moraine ridges, during the genesis of which the ice had been rather thick and active, a zone of transverse eskers and deltas, which represents a standstill stage of the ice sheet, and an inner marginal moraine arch, formed by the re-activization of the ice. No such tripartition has been noted in the Hämeenlinna region, where the large moraine ridges parallelling the glacier margin, characteristic of the Jylisjärvi area, largely are lacking.

The belt formed by the Second Salpausselkä in the region under consideration is characterized above all by the large transverse eskers (Fig. 18, II). They indicate a standstill of the ice margin for a considerable period of time at the point marked by each esker. The transverse eskers must be understood in part to be responsible for the formation of the deltas in front of the ice margin. To some extent, again, they exhibit ice-contact features on both sides. This would signify their having evolved through deposition within marginal crevasses running parallel to the edge of the ice sheet (Hyyppä, 1951).

Moraine ridges of various sizes are also typical of the Second Salpausselkä within the region investigated. At Turenki and in the area to the northeast of it, they are of slight proportions, resembling annual moraines, but farther south there are larger moraine ridges, too. Judging by certain of the circumstances considered in the foregoing, the margin of the ice sheet had advanced in this locality along a rather broad front, pushing forward material to form end moraines, disturbing the underlying beds and accumulating till on the surface of the glaciofluvial drift. The later direction of movement of the ice sheet was slightly more to the west than its previous movement.

In the valley of Vanajavesi, in the vicinity of Turenki, the ice sheet in the belt of the Second Salpausselkä behaved in a slightly different way. Its advance cannot be observed so clearly here. In the valley the glacier has also been thicker than in the higher surroundings. It had nearly stagnated here and peacefully melted in place. Evidence of this is the mighty esker of Turenki, characterized in this locality by a pitted landscape, which presupposes a stagnant but relatively thick covering of ice. The Turenki esker runs without interruption across the Second Salpausselkä, which at this point breaks off for a distance of many kilometers.

Within the belt comprising the Second Salpausselkä, it is thus possible to distinguish different kinds of sectors, where the ice had acted in different ways. It has already been pointed out in the foregoing that one functionally deviating ice sector is the one described by Okko (1957) from the Jylisjärvi area.

From the Second Salpausselkä there extends in a proximal direction a zone about 2—3 km broad wherein no morphological ice -marginal formations occur (Fig. 18, III). The plateaus in this zone remain at the same elevation as the Second Salpausselkä. At the proximal edge of the zone, however, their elevation sinks to a level 25—30 m lower down, as revealed by the observations made in the areas of the Kaloinen and Turenki map sheets.

This change took place quite suddenly: the surfaces of the plateaus descend to the lower level within a distance of a few hundred meters (Fig. 13).

In the following zone (Fig. 18, IV), which is some 5—10 km broad in the Hämeenlinna region, the ice sheet retreated along an even front. The elevations are pretty much the same in the various parts of the area. The ice melted at a rapid rate, as the large eskers indicate. In places they have spread out into extensive plains, of which the largest are the ones at Valkealampi and Valkeajärvi, in the Kaloinen map sheet. The plains evolved, according to Sauramo (1958), so as to conform with the highest Yoldia shore of the Baltic water level, *i.e.*, the YI-level.

The next zone, proceeding in the proximal direction, is featured by an extremely fragmentary ablation moraine (Fig. 18, V). In the Hämeenlinna region this zone is quite narrow, measuring only 2—4 km in breadth. It consists of several ablation moraine areas of slight extent, which topographically fall into the transitional area between the higher and the lower ground. On the southern side of the region surveyed the zone broadens out some 10—15 km, although ablation moraines do not form any extensive continuous areas here either. The ice sheet had largely disappeared from this zone by melting in place.

Toward the northwest the next formation is the marginal formation of the Third Salpausselkä (Fig. 18, VI). As in the case of the Second Salpausselkä, distinct marginal formations are lacking in the Vanajavesi valley of the Hämeenlinna region. The extensions of the Hämeenlinna esker at the western end of the city and on the southeastern side of Lehijärvi belong, however, to this zone.

In the area between Alajärvi and Renko, the ice sheet behaved differently from what it had in the rest of the Hämeenlinna region. A conspicuous feature is the occurrence here of a rather extensive drumlin area, signifying the action of a thick glacier in motion. The esker running toward the drumlin area from the southeast breaks off and terminates, only to continue again at the northwestern end of Alajärvi, where formations belonging to the Third Salpausselkä, small plateaus and a short transverse esker, are situated. The absence of ablation moraines and eskers from the area lying between Alajärvi and Renko indicates, together with the occurrence of drumlins, active ice. This signifies the fact that the drumlins had evidently evolved during a fairly late stage of glaciation.

At the southwestern end of the drumlin area is situated the extensive plateau of Renko. To the southwest, again, are the plateau of Pursunjärvi and, after a break of a few kilometers, the broad expanse of Pernunnummi.

The last zone running parallel to the ice margin in the region under investigation is the inner zone of the Third Salpausselkä (Fig. 18, VII). Here, too, different sectors can be distinguished. On the northern side of

Renkajärvi, the ice retreated along an even front, creating imposing eskers, which in places spread out into extensive plateaus. At the southern end of Renkajärvi, on the other hand, the ice disappeared by melting in place, as the broad, though not quite continuous ablation moraine areas indicate. The ice had here for the most part retreated across dry ground. An ablation moraine follows for a long distance on the proximal side of the Third Salpausselkä, and considerable ablation moraine areas are met with also in the southern part of the region investigated.

The proximal zone of the Third Salpausselkä is the last zone giving evidence of the mode of retreat of the continental ice sheet met with in the region.

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11

THE DECOMPOSITION OF ALNÖ ALKALINE DIKES BY PERCOLATING WATER [1]

BY

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The exceedingly rapid weathering of the sövite dikes at Alnö Island has previously been described in my Alnö Memoir (von Eckermann, 1948). Of the generally equally rapid decomposition of some of the conesheets and radial dikes emanating from deepseated levels of the volcanic feeding channel an account was given in my paper on the dikes of the mainland northwest of the island (von Eckermann, 1958). The essential data concerning this decomposition, gathered at that time, may be summarized as follows:

1. Crumbling of the dikes in the outcrops, due to normal weathering, is rapid, but ground-water percolating through the country rocks and the dikes produces an even more rapid dissolution of some of the dikes.
2. An increased flow and pressure of the ground water, caused for instance by the dammed up lake upstream from the Bergeforsen hydraulic powerstation, further increases the rapidity of dissolution.
3. In carbonatitic dikes with chilled margins containing carbonic acid, the decomposition starts along the contacts towards the wall rocks and proceeds very rapidly. Within the dikes of both carbonatitic and femic composition, the alteration process starts along fissures and slipping planes, and may be either fast or slow depending upon the mineral composition of the rock.
4. At a pH of 5.6 and a water pressure of 2.4 kg/cm^2, an average depth of 9.5 mm of dike rock is removed yearly at each »dissolution surface».

[1]) Received February 22, 1961.

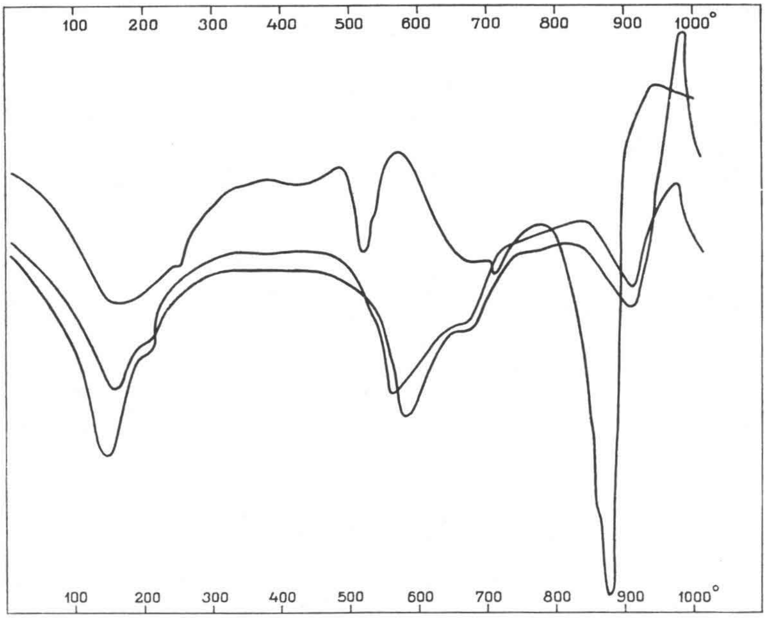

Fig. 1. Differential thermal diagrams of clays from altered kimberlitic dikes.

5. The operation process is twofold: a) the removal of calcite by the action of carbonic acid as soluble bicarbonate, and b) the hydration of minerals, mainly montmorillonite, causing disruption of fabric by swelling.

6. The ultimate product of decomposition, not carried away in solution, is a clayey, plastic mixture of clay minerals, such as halloysite, montmorillonite and illite, garnet, apatite, perowskite, quartz, natrolite and calcite. Three differential thermal curves of clays from various dikes are shown in Fig. 1.

During the last ten years the investigation of the decomposition process of the dikes has continued and better instruments and methods have been developed to measure the rate of dissolution. My earlier experiments were first carried out on small cubes of the size of 1 cm³, inserted in glass tubes, their diagonals between two opposite corners parallel to the length of the tubes. By forcing water through the tubes and reversing the direction of flow every 10 minutes all six sides of the cubes were equally exposed to the action of the water.

By this method, however, no information was obtained as to the permeability of the rock. Neither did it give satisfactory answers to the questions of the changes of permeability in relation to the alteration and the

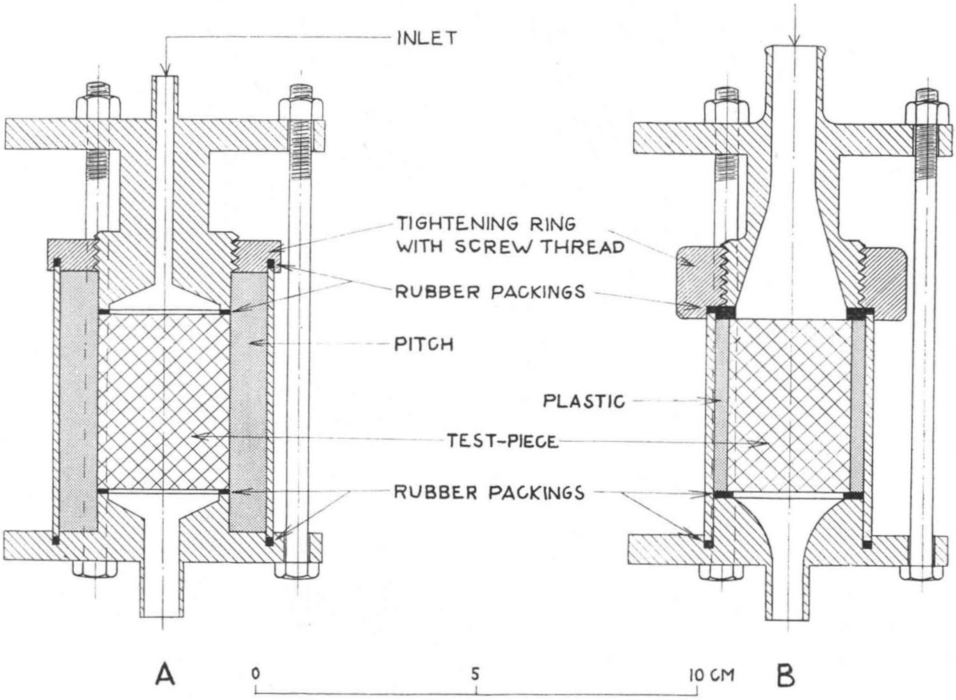

Fig. 2. Apparatutes used for the determination of the rate of dissolution of Alnö dikes.

appertaining time factor, nor did it reproduce the actual conditions within the dikes, where the action of the water is a three-dimension and not a two-dimension one. At the proposal of Mr. Nils Berg, civil engineer and director of the firm of consulting engineers »Vattenbyggnadsbyrån» in Stockholm, a new series of experiments was carried out with 6 apparatuses constructed by the said firm. Cylindrical test-pieces 30 mm in diameter and 5, 10, 20, 40 and 80 mm long, respectively, were drilled out of the center of a thick kimberlitic Alnö dike at Bergeforsen, and mounted in the apparatuses, of which a schematical drawing is shown in Fig. 2A and a photograph in Fig. 3. Subsequently the upper surfaces of the test-pieces were put under a water pressure of 0.5 kg/cm^2, which was maintained until water started leaking through the 5 mm samples. It was then raised to 1 kg/cm^2 in the cases of the thicker samples and maintained until the 10 mm one showed leakage, and so until it reached the ultimate pressure of 8 kg/cm^2 in the last sample. The pH of the water was kept at 5.55—5.65.

The amount of leakage per unit time was measured and from it the permeability »k» was calculated according to Darcy's formula

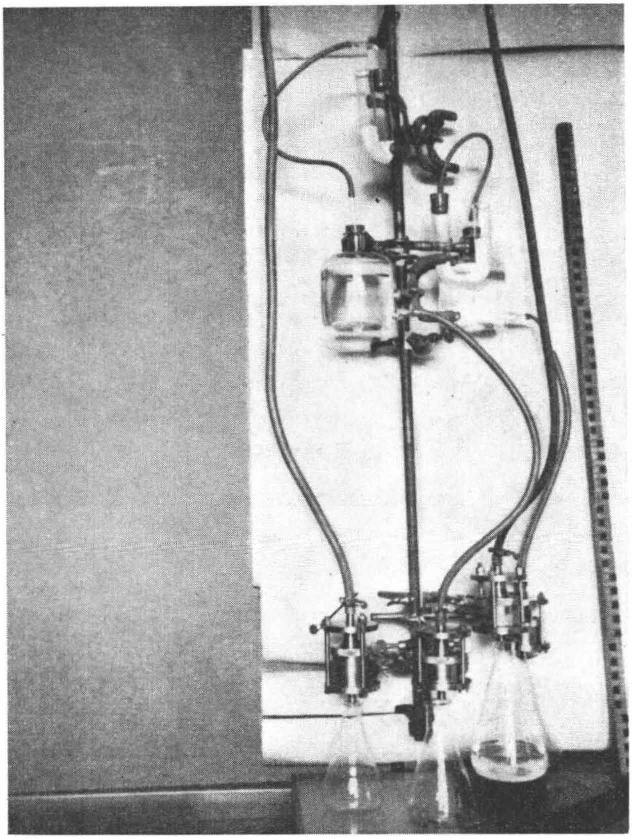

Fig. 3. Assemblage of apparatuses used for the determination of the dissolution ratio of the kimberlitic dikes. Two apparatuses are put under 0.5 and two under 1.0 metres water pressure. Two of the U-shaped vessels absorbing the CO_2 of the air are seen. During the experimental run the beakers which collect the percolating water were placed in a freezing-mixture.

$$\frac{Q}{FT} = \frac{k \cdot H}{h}$$

where Q = volume of water in cm² during the time T seconds,
F = cross-section of test-piece in cm²,
H = head of water in cm,
h = thickness of test-piece in cm.

In case of a perfectly homogeneous unalterable dike rock, the »k» calculated from the water measurements should be the same in all cases as the ratio H/h is kept constant — but the actual results showed a very wide

Table 1.

Water pressure in metres	Loss in mg/cm²/24 hours		Average	Number of tests
	Minimum	Maximum		
0.5	1.96	2.18	2.06	5
1.0	1.88	2.36	2.15	3
2.0	2.29	3.72	3.00	2
4.0	3.91	4.25	4.08	2
8.0	4.63	5.42	5.02	2

margin of variation. For instance, the time required before the first wetness was observed on the bottom surfaces of the samples varied from 2 hours and 13 minutes to 642 hours and 12 minutes. In the case of three samples no percolation had taken place even after 720 hours. Altogether 34 »runs» were made on different samples. This series was followed by a shorter one of 5 »runs» on 10 mm test-pieces at a lower pressuregradient (20 percent of the main series). The latter gave slightly lower »leakage-time», but confirmed the previously established reasons for the variation of the experimental results, viz.:

1) Unsuitable construction of the apparatuses used, the inlet for the water being too small and gas bubbles, arising from the decomposed rock, obstructing the flow;

2) Inhomogeneous rock samples;

3) Dissolution of calcite which successively altered the permeability;

4) Alteration of the pH-value of the water during its passage through the test-pieces, especially in the case of the long ones,

5) The micas acting like flaps in a valve when exposed by the removal of surrounding calcite through leaching. Up to a point increased water pressure was found to lead to increased dissolution of calcite, as illustrated by Table 1. Later, however, in the case of the thicker test-pieces, when a sufficient amount of mica flakes had been freed, not only did the dissolution but also the water percolation decrease. A shaking of the apparatus proved generally sufficient in such cases to »close the valves» completely and stop the flow of water.

Even if the earlier experiments on the aquatic alteration of the Alnö dikes into clay supplied many interesting data of great technical value in connection with the construction work at the Bergeforsen hydraulic power station, I found the scientific problems underlying the great variation of dissolution data insufficiently solved, and I undertook the running of a new series of tests with better laboratory equipment and carefully chosen samples of dike rocks of different composition. In consequence, I redesigned the

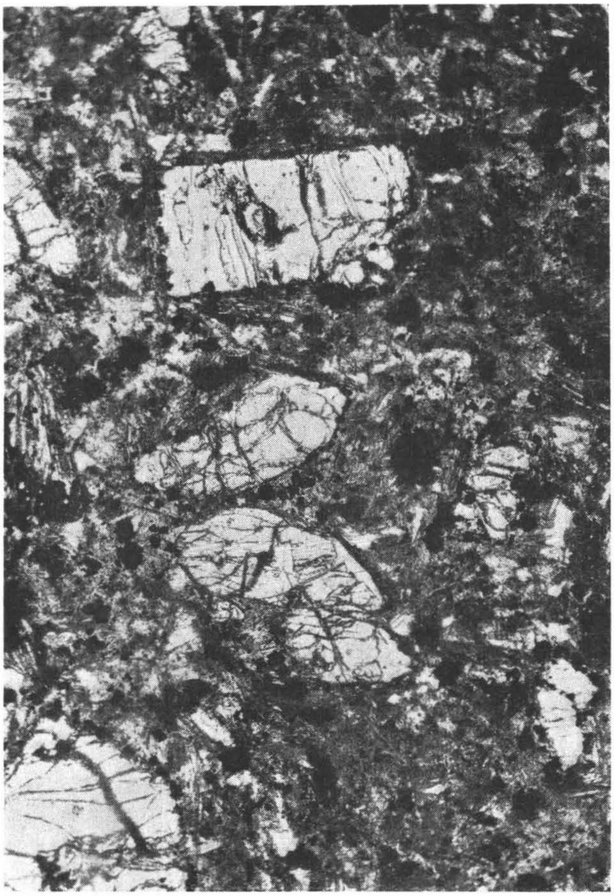

Fig. 4. Autometamorphically altered kimberlite. The upper rectangular olivine crystal is altered to serpentine with spots of calcite, the two below retain some of the original mineral intermixed with calcite and the one in the low left corner consists of serpentine and kaolinite. The melilite laths are all altered to clay minerals, garnet and calcite, although their contours are still locally discernible. Ordinary light. Magnification × 45.

test-apparatus as shown by Fig. 2B, giving it a very wide sloping water-inlet, carefully polished on the inside in order to prevent any adherence of gasbubbles. The cylindrical sides of the testpieces were encased in plastic, thereby avoiding the previous not easily checked sealing with pitch which had to take place after their »mise en place». All samples were of 5 mm thickness in order to avoid the mica flap-valve action. The pH-value of the water used was 5.6. The experiments were run at a room temperature of +20 C° and at a water pressure of 0.5 metres.

Table 2.

	Loss in mg/cm²/24h	Average loss
Sample no. 1	0.77	0.71
» » 2	0.53	
» » 3	0.67	
» » 4	0.86	
» » 5	1.99	1.97
» » 6	1.67	
» » 7	2.16	
» » 8	2.13	
» » 9	2.68	2.76
» » 10	2.89	
» » 11	2.51	
» » 12	2.97	
» » 13	2.01	1.48
» » 14	1.67	
» » 15	1.21	
» » 16	1.01	
» » 17	0.43	0.50
» » 18	0.53	
» » 19	0.62	
» » 20	0.43	

Thinsections from the five groups of samples were examined and the following was noted:

No. 1—4: The dissolution is mainly one of interstitial, presumably calcitic, carbonate. In a slide of the rock executed before the testing, the calcite of the interstices contains minute inclusions, which may be carbonic acid. The melilite is slightly affected, the typical characteristic striation accompanying metamorphism being fairly strongly developed. The olivine is unaffected.

No. 5—8: The dissolution mainly affects the metamorphosed melilite. The fine grained carbonate of the pseudomorphs is mostly gone, as are part of the clay minerals. The fresh melilite shows signs of incipient destruction. Some of the serpentine of the olivines has disappeared. The larger calcite crystals remain unaffected.

No. 9—12: The dissolution of the melilite is the same as in the previous group. The minute garnet grains are still more or less held in place by the remaining swollen clay minerals and the original positions of the melilite crystals may still be seen. Many phlogopitic mica flakes are sticking out in the interstitial cavities formed by dissolution, but are still too strongly encased to act as flaps closing the leakage-pores. Most of the alteration products of the olivine have been removed, probably mechanically.

No. 13—16: The dissolution seems to be confined to the unrecrystallized melilites and to interstitial calcitic carbonate. In this case ,too, the thinsections of the original rock showed minute vesicles, presumably containing carbonic acid. Some serpentine was removed mechanically.

No. 17—20: The comparatively small dissolution in this case is found to be concentrated in the joints and interstices between the fairly large carbonate crystals. The water carried away occasional flakes of serpentine and minute phlogopite flakes.

The following rock-specimens were chosen:

No. 1 to 4: Kimberlite with almost intact melilite and fresh olivine (Plate I, Figs. 1 and 2).

» 5 » 8: Kimberlite with about 50 percent of the melilite altered to a microscopic aggregate of carbonate, garnet, clay minerals and quartz, but with fresh olivines.

» 9 » 12: Kimberlite with all melilite altered to pseudomorphs and the originally monticellitic olivine altered to calcite, serpentine and some unidentified micaceous product (Fig. 4).

Table 3.

Samples no:	Loss in weight in mg/cm²/24h.							
	21	22	23	24	25	26	27	28
pH-value: 3	3.99	4.05						
5	2.94	2.97	3.03					
6			2.61	2.65				
7				2.40	2.38			
8					2.22	2.24		
9						2.07	2.08	
10							1.95	1.97
11								1.92

» 13 » 16: Kimberlite with serpentine *(etc.)*-pseudomorphs of olivine and the melilite-pseudomorphs autometamorphically more or less recrystallized into larger garnet- and calcite crystals, most of the clay minerals replaced by greenish chlorite and yellowish phlogopite, still indicating roughly the original contours of the melilite crystals.

» 17 » 20: Ouachititic kimberlite, where the recrystallization has gone another step further, the original melilite crystals being only occasionally discernible, the amount of calcite increased and the garnet developed as fairly big crystals with melanitic kernels. All clay minerals were lacking.

This suite of specimens indicates the autometamorphic alteration of the original melilititic olivine-basalt (= kimberlite rich in melilite), caused by the action of increased concentration of carbonic acid during its rise in the volcanic funnel.

The test-pieces were weighed before and after a run of 24 hours and the resulting losses in weight given in mg/cm² are compiled in Table 2.

Consequently, there is no doubt about the rate of decomposition of the kimberlitic dikes being intimately connected with their degree of autometamorphic alteration. But it is also dependent on the pH-value of the percolating water. In order to determine the variation a series of 8 samples no. 21—28, taken from the kimberlite of the previous samples 9—12, was tested under the same conditions as the above mentioned ones, but with water of varying pH-values. The results are compiled in Table 3. They indicate a very strong increase of decomposition at lower pH and a slow »asymptotic» decrease at higher pH towards a fixed minimum lying above zero.

The water of high pH was obtained by diluting a saturated solution of lime-hydrate. The dissolution of the kimberlite occurring at high concentra-

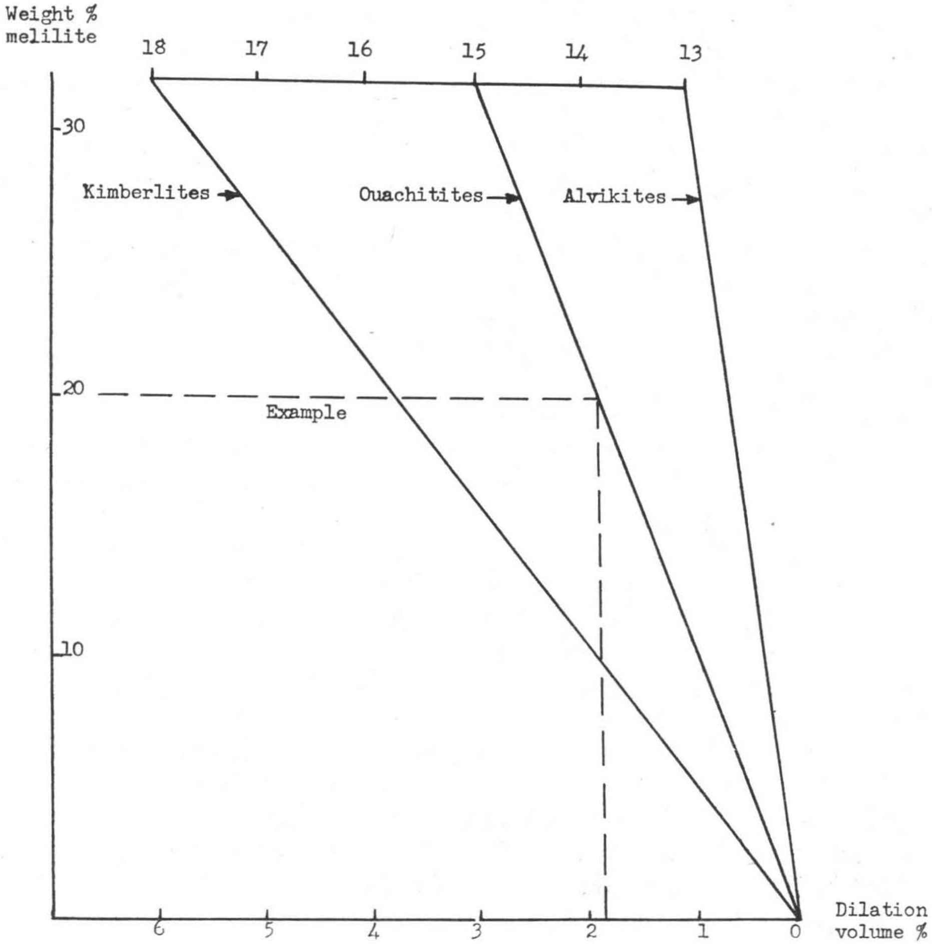

Fig. 5. Dilation of dike rocks on hydration of clay minerals. Example: An Alnö dike containing 20 percent primary melilite and 14 percent clay minerals in the melilite pseudomorphs increases 1.9 vol. percent.

tions of Ca-ions in the water, can only be explained by disintegration of the crystal aggregate on swelling of the clay minerals and by the presence of free CO_2 as vugs in the calcite.

One would think that the swelling of the clay minerals must lead to a closing up of all pores in the altered rocks and a prevention of any percolation of water. From available analyses and geometrical determinations of the mineral compositions of the calcitebearing Alnö dike rocks, I have compiled the appended monogram, Fig. 5. It shows that the increase in volume of the rocks itself is comparatively small, and that it will in general be totally offset by the dissolution of the much larger volume of calcite. I have calculated the shrinkage of kimberlites, totally altered into clay, to be from 5 to 12 percent. Actual compression tests undertaken by the laboratory of the Swedish State Water Power Board gave values of about 10 percent. While the majority of the altered kimberlite showed a gradually increased compression at increased pressure some indicated a limited intermediate sudden shrinkage preceded and followed by a gradual one. Further experiments carried out at my laboratory have shown that this is invariably the case for dikes rich in natrolites and lime-zeolites. A possible explanation for this irregularity is apparently the breakdown of the zeolite-structure.

During my study of the dikes within the Alnö region I have also found that tinguaitic and nephelinitic dikes are occasionally subject to very rapid weathering. This was also observed in a few cases underground at Bergeforsen. Laboratory tests, however, made with the assistance of the apparatus described above and using soft water of pH = 6.5, did not show any very pronounced rate of dissolution. The answer to this problem was given when samples of the ground water percolating through these dikes were later analyzed. While the water from Indalsälven river contains in mg/l 26 Ca, 15 Na and 49 bicarbonate (HCO_3), the same water after having passed through the calcite-rich montmorillonitic zone upstream from the power station was found to have an increased content of 150—200 Ca and 250 HCO_3, the Na oscillating between 8 and 25. If this type of water then encounters dikes rich in natrolite (pseudomorphs of nepheline) the Na-content increases considerably, occasionally reaching 110 mg/l, while the Ca-content may drop to as low as 46 and the bicarbonate remains at a high level. The explanation seems to be ion-exchange of Na and Ca and the formation of calcium-zeolites in the dikes and bicarbonate of soda in the water.

The ground water circulating through the country rocks around Alnö must necessarily encounter many Alnö dikes containing an excess of CO_2, and be correspondingly enriched in HCO_3. This, in connection with the ion-exchange may be consequently taken as the reason for the very strong and rapid weathering of most nephelinitic and tinguaitic dikes and the scarcity of observable ones in the outcrops. Diabase (dolerite) is a rock, generally fairly resistent to weathering but within a certain radius of Alnö, say about 10 km, even this is apt to be deeply covered by its own detritus, the reason being the attack of water on its carbonatized plagioclase feldspars. This type of diabase is mentioned in one of my previous papers on Jotnian rocks (von Eckermann, 1945, p. 62, 65—66, Pl. VIII), where the description and analysis of a dike west of Östrand Pulp Mill is given. The very same dike was found again at 4 200 m from the inlet in the Östrand soft water tunnel (von Eckermann, 1945, Pl. XX, dike no. 59), where it is in part even more altered. This alteration, which turns the labradorite of the rock into a mixture of andesine and calcite or even into calcite, kaolin, sericite and some quartz is very old, being connected with the Alnö intrusion and the accompanying soaking of the country rocks with carbonic acid gases escaping under high pressure from the volcanic center. The same phenomenon was observed in the case of practically every diabase dike encountered in the tunnels at Fagervik and Östrand as well as at the hydraulic power station at Bergeforsen. In the latter case, however, the dikes are few and very small and the carbonatization is mainly concentrated at intersections with carbonatitic Alnö dikes.

So far, the Alnö intrusion seems to be the only one of its kind where a complete set of contemporaneous conesheets, ring dikes and radial dikes have been found, or at least described in any published paper. I wonder if underground investigations of the same magnitude as those carried out in connection the Bergeforsen hydraulic powerstation would disclose at several of the manifold carbonatite occurrences, dikes which have escaped notice due to deep weathering under tropical conditions. In conclusion, I call attention to the Plate II, Fig. 1 and 2 showing what weathering can do to carbonatites during the course of a few years even in a temperate climate like Sweden.

Acknowledgement — I am indebted to the Bergeforsen Kraft AB for providing the rock specimens investigated and for permitting the publication of data obtained during my earlier investigations of the decomposition of the dikes at the Bergeforsen hydraulic power station.

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EXPLANATIONS TO THE PLATES

PLATE I

Fig. 1. Kimberlite. The monticellitic olivine crystals are only very slightly altered to serpentine and the melilite is fresh. Ordinary light. Magnification 45 ×.

Fig. 2. Melilite crystals in kimberlite. Ordinary light. Magnification 125 ×.

PLATE II

Fig. 1. Sövitic carbonatite, photographed in a road cut south-west of Ås, Alnö Island, in 1948.

Fig. 2. The same road cut as above, photographed from about the same direction in 1958. The dark xenolith below and to the right of the hammer in Fig. 1 is still recognizable at the upper right (arrows).

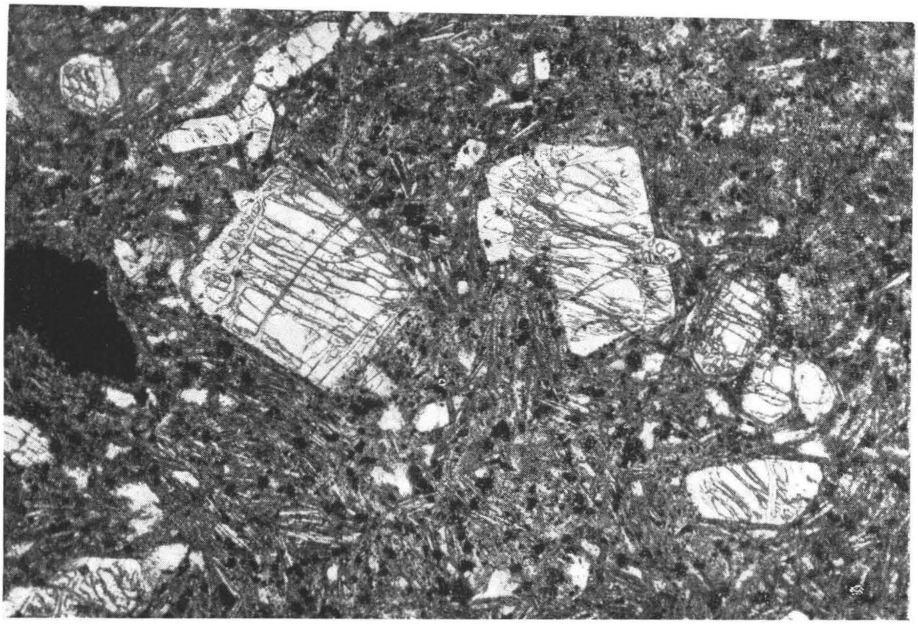

Fig. 1

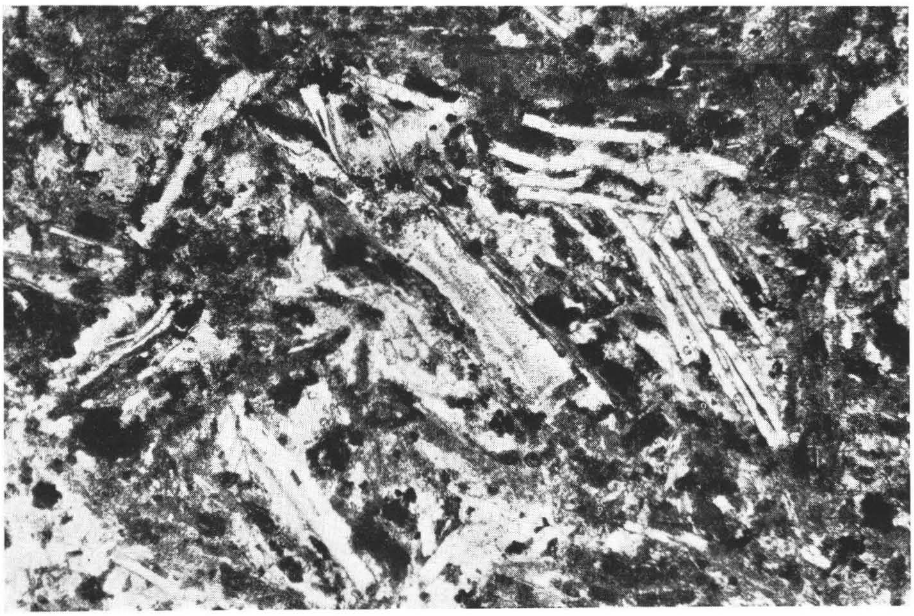

Fig. 2

Harry von Eckermann: The Decomposition of Alnö Alkaline Dikes . . .

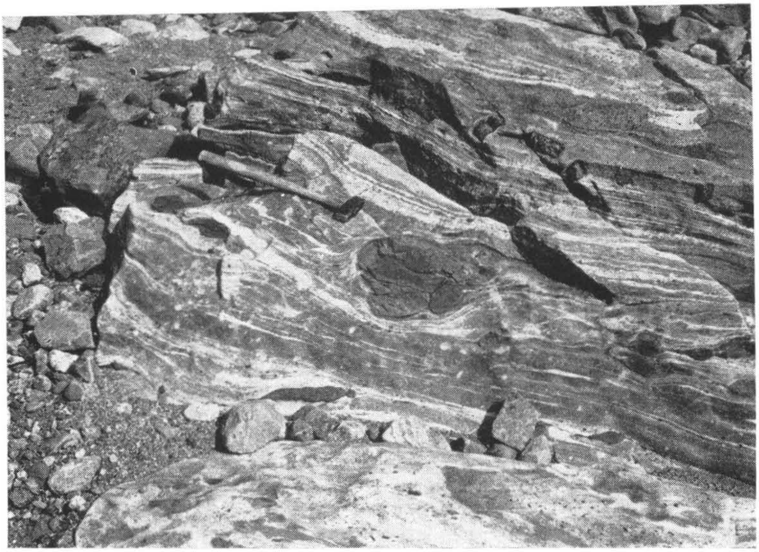

Fig. 1

Fig. 2

Harry von Eckermann: The Decomposition of Alnö Alkaline Dikes . . .

12

ASPEKTE DER FINNISCHEN GEOLOGISCHEN FORSCHUNG [1]

VON

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AUSZUG

Der Verfasser beschreibt zuerst die verschiedenen Methoden und Stile, deren man sich bei der Entwerfung der Felsgrundkarten bedient hat. Danach werden die von verschiedenen Forschern vorgelegten Einteilungen der Granite und anderen Tiefengesteine behandelt. Nach der Meinung des Verfassers sind die von J. J. Sederholm, W. Wahl und A. Simonen angewendeten Einteilungsprinzipe nicht zweckentsprechend.

Zum Schluss wird vom Verfasser gezeigt, dass die jetzige Alterseinteilung des Felsgrundes von Finnland nicht mehr als solche aufrechtzuerhalten ist. Diejenigen Bildungen, die stratigraphisch unterhalb der subbothnischen Diskordanz liegen, sind zu einer gesonderten, älteren svionischen Formationsserie zu rechnen. Die oberhalb der Diskordanz liegenden bothnischen Bildungen schliessen sich lückenlos an die Kareliden an und bilden mit diesen die karelobothnische Formationsserie.

INHALT

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[1] Eingegangen den 31. März 1961.

VORWORT

Die geologische Forschungsarbeit ist in den letzten Jahrzehnten in Finnland sehr rege gewesen, und sie hat auch viele wichtige und wertvolle Resultate ergeben. Man könnte jedoch fragen, ob man sich immer bemüht hat, die geologische Forschung in jeder Beziehung mit der Vorurteilsfreiheit und Überlegung zu entwickeln und weiterzuführen, wie sie die wissenschaftliche Arbeit im allgemeinen voraussetzt. So scheint es mir, als ob man in einigen Fällen, z. B. im Zusammenhang mit der geologischen Kartierung, von theoretischen Ideen durchdrungen gewesen wäre, die keinen Grund der Realität besitzen und die demnach die Entwicklung der Forschungsarbeit nicht günstig beeinflusst haben können. Es kommt mir auch vor, als ob einige wichtige frühere geologische Errungenschaften allzu wenig Aufmerksamkeit auf sich gezogen hätten. Jetzt, im Jubiläumsjahr der finnischen geologischen Forschung, dürfte es am Platze sein, im konstruktiven Sinne einige gewonnene Resultate und Auffassungen zu überprüfen.

ÜBER DIE GEOLOGISCHE KARTIERUNG

Zu Beginn der geologischen Aufnahmearbeit in Finnland, bei der Übersichtskartierung von Südfinnland, wurden auf den Karten (neben den Bodenarten) nur die Felsaufschlüsse mit den ihnen zukommenden Bezeichnungen oder Farben zur Darstellung der petrographischen Zusammensetzung aufgetragen. Dies war natürlich eine sehr objektive Kartierungsmethode, da auf den Karten nur zur Aufzeichnung kam, was man wirklich sehen und beobachten konnte. Auch heutzutage ist man gezwungen, nach dieser Methode in solchen Gebieten zu verfahren, wo die Aufschlüsse sehr spärlich sind. Im allgemeinen, und speziell für die Felsgrundübersichtskarten, war diese Methode jedoch nicht zu empfehlen. Erstens sollte man doch aus den Felsgrundkarten nicht ausschliesslich die petrographische Zusammensetzung der Felsen ersehen können. Sie sollten auch ein Bild von der wirklichen Verbreitung der verschiedenen Gesteine, unabhängig von ihrer Erdbedekkung, der Erscheinungsweise derselben und zudem auch vom tektonischen Aufbau des kartierten Gebiets usw. geben. Sie mussten, mit anderen Worten, wirklich lebendige g e o l o g i s c h e Karten sein. Dies setzte selbstverständlich voraus, dass man auch die Zwischenräume zwischen den Felsaufschlüssen mit demjenigen Gestein bezeichnet, das man auf Grund gewisser geologischer Tatsachen als dort vorkommend anzunehmen hat. Bei der Entwerfung von derartigen Übersichtskarten stand man anfangs — und tut dies teilweise auch jetzt noch — im grossen ganzen auf dem Standpunkt mechanischer Interpolation. Diese Interpolationsmethode könnte man in

Kürze folgendermassen charakterisieren. Wenn zwei Felsaufschlüsse aus demselben Gestein bestehen, bezeichnet man auf der Karte das zwischen ihnen befindliche, von losen Bodenarten bedeckte Gebiet als aus demselben Gestein bestehend. Wenn sie dagegen aus verschiedenartigen Gesteinen bestehen, verläuft auf der Karte die Grenze zwischen diesen Gesteinen ungefähr mitten zwischen den beiden Aufschlüssen. Eine solche Methode führte natürlich zu vielfachen Fehlern im Kartenbilde. So z. B. erhielten harte, der Denudation gut widerstehende und daher besser aufgeschlossene Gesteine, wie Quartzite und Granite, auf der Karte eine unverhältnismässig grosse Ausdehnung. Immer mehr fing man jedoch an, Aufmerksamkeit auf die Beschaffenheit des unter der Erdbedeckung verborgenen Gesteinsgrundes zu richten. Es lassen sich auch darüber mehrere ziemlich zuverlässige Schlüsse ziehen. Schon die Topographie des Gebietes gibt sehr oft Winke in dieser Hinsicht. Die Moräne, sowie die Vegetation, verrät stellenweise recht deutlich den darunter befindlichen Gesteinsgrund. Eine weitere Erklärung, speziell in tektonischer Hinsicht, geben die aeromagnetischen Karten. — Ich habe schon früher mit Hilfe einiger Beispiele gezeigt (Saksela, 1936), wie das geologische Kartenbild einiger Gebiete sich gründlich änderte, wenn man sich von der mechanischen Interpolation freimachte.

Als ich in den Jahren 1930—1934 mit der geologischen Übersichtskartierung Ostbothniens beschäftigt war, erwies es sich als notwendig, der tektonischen Erscheinungsweise der Tiefengesteine besondere Aufmerksamkeit zuzuwenden, um eine möglichst zuverlässige und lebendige Karte von dem Gesteinsgrund zustandezubringen. Es zeigte sich herbei, dass sich die genannten Gesteine ungezwungen in zwei Gruppen einteilen lassen, deren tektonischen Erscheinungsweisen in dem Masse von einander abweichen, dass sie dem ganzen Kartenbild ihr eigenes, spezielles Gepräge aufdrücken. Die eine von den Gruppen bezeichnete ich als synorogen, die andere wieder als spätorogen. Es erwies sich dann, dass diese Gruppen auch zwei verschiedenaltrige, nichtkomagmatische Eruptionsserien repräsentieren, und dass sich an beide verschiedenartige Tiefengesteine von Peridotiten bis Granite und Granitpegmatite anschliessen. Später habe ich gezeigt, dass dieselben tektonischen Tiefengesteinsgruppen deutlich auch anderswo in Fennoskandia, wo genauere Felsgrundkartierung ausgeführt worden ist, vorkommen (Saksela, 1953). Wir können ohne übertreibung veıallgemeinernd sagen, dass für jeden orogenen Zyklus gerade die erwähnten Tiefengesteinsgruppen charakteristisch sind. — Eigentümlicherweise hat man jedoch in den letzten Felsgrundkartierungen in Finnland magmatektonischen Fragen nur in geringem Masse Beachtung geschenkt. Sehr befremdend ist aber, dass man bei der Einteilung der Tiefengesteine den verschiedenen Gruppen Na-

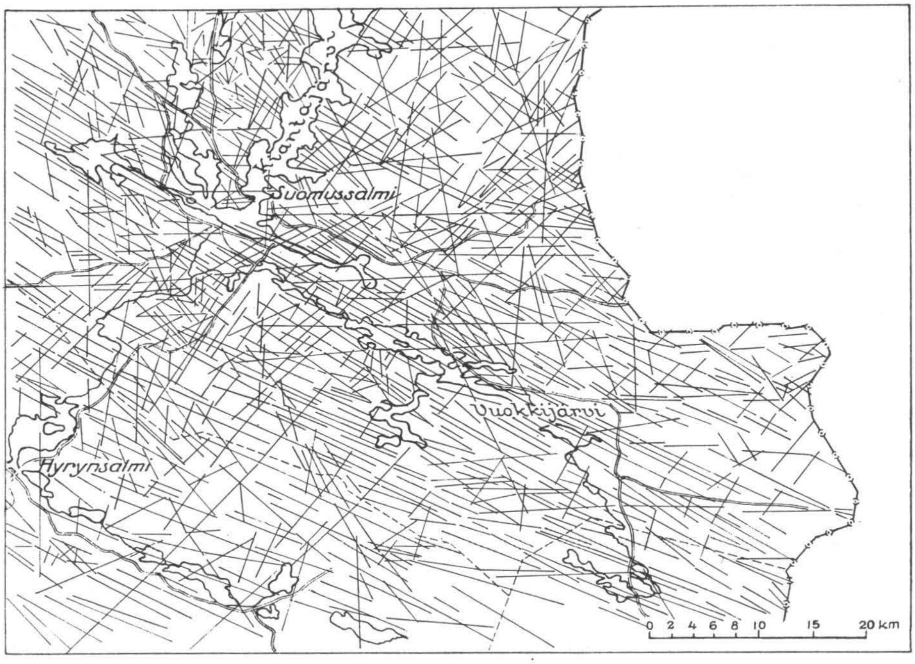

Abb 1. Die morphologisch beobachteten Bruchlinien im Suomussalmi—Hyrynsalmigebiet. Nach S. Penttilä (Virkkala, 1960).

men beigelegt hat, die auf die tektonische Erscheinungsweise derselben hindeuten, obschon die Einteilung in der Tat ausschliesslich auf Grund petrographischer Merkmale ausgeführt worden ist. Es ist natürlich, dass ein solches Verfahren geeignet gewesen ist, mancherlei Verwirrungen hervorzurufen. Ich werde später auf die Sache zurückkommen.

In den letzten Jahrzehnten hat man bei der Entwerfung von Felsgrundkarten einen »geschmeidigen» Stil angenommen. Man ist bestrebt, wenn irgend möglich, die von verschiedenen Gesteinen gebildeten Gebiete auf den Karten als langgestreckte Streifen, die dem allgemeinen Streichen folgen, zu bezeichnen. Eine Ausnahme scheinen in dieser Hinsicht nur die Rapakiwigranite, sowie einige Granite der 3. Gruppe Sederholm's zu machen. Diese Gesteine kommen auf den Karten als mehr oder minder rundliche Flecke vor, deren Grenzlinien schonungslos die Parallelstrukturen der umgebenden Gesteine durchschneiden. Als Beispiel für derartige »geschmeidige» Felsgrundkarten führe ich die von Erkki Mikkola (1941) entworfenen Kartenblätter Muonio, Sodankylä und Tuntsajoki an. Offensichtlich ist man auf diese Weise in vieler Hinsicht zu einem lebendigeren und wirklichkeits-

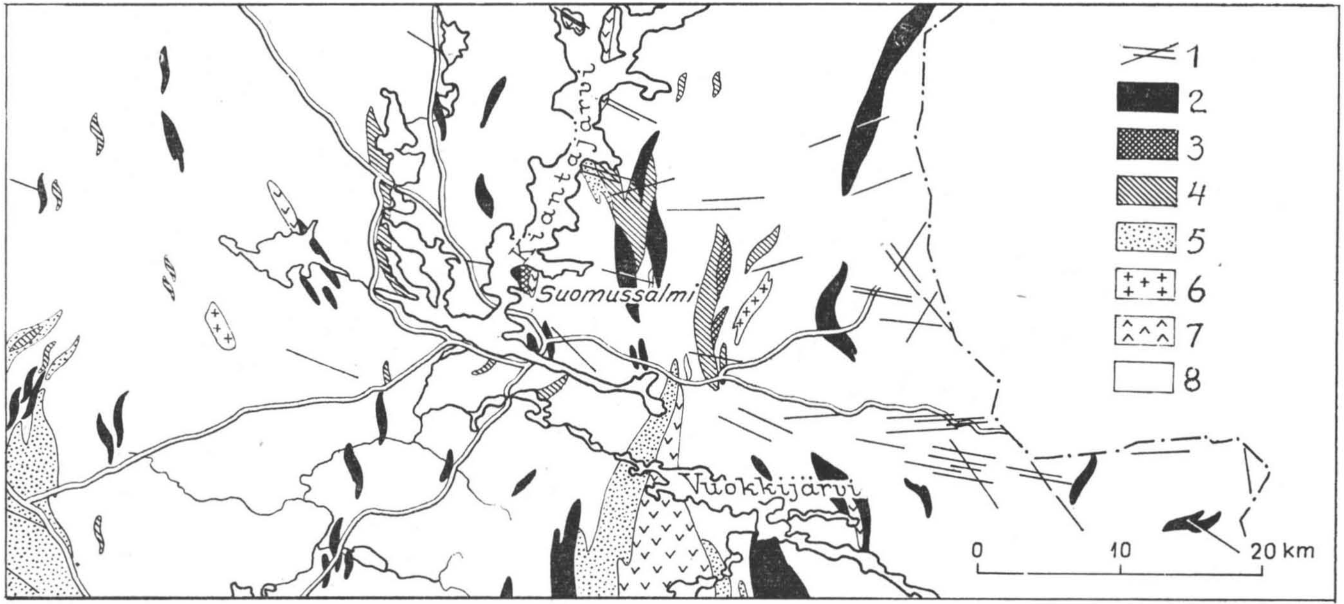

Abb. 2. Geologische Übersichtskarte des Suomussalmi—Hyrynsalmigebiets. Etwas verallgemeinernd nach Matisto (1958) umgezeichnet. 1. Diabasgänge; 2. Granite; 3. Ultrabasische Gesteine, hauptsächlich Peridotite und Serpentingesteine; 4. Amphibolite; 5. Quarzite und Quarz-Feldspatschiefer; 6. Gabbros, Diorite und Granodiorite; 7. Basische Vulkanite; 8. Gneisgranite.

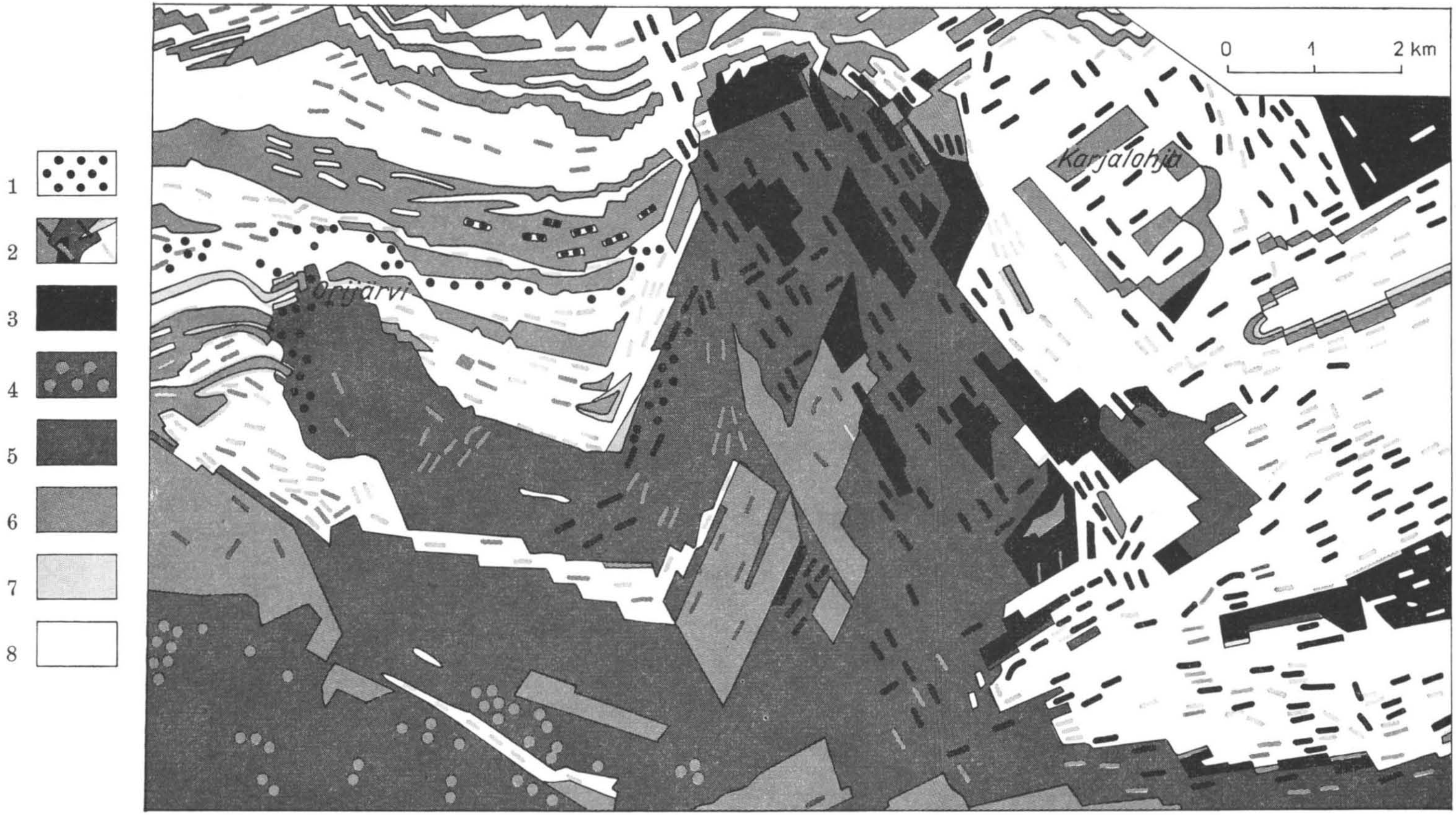

Abb. 3. Geologische Karte des Gebiets zwischen Orijärvi und Karjalohja. Nach der von Tuominen (1958) im Jahre 1957 entworfenen Karte umgezeichnet. 1. Quarzeinsprenglinge; 2. Mischgesteine (vgl. die folgenden Zeichen); 3. Granitische Gesteine; 4. Breccierter Amphibolit, mit den Gesteinen der folgenden Gruppe vermengt; 5. Gneisige quarzdioritische, granodioritische und trondhjemitische Gesteine; 6. Amphibolit, Diopsidamphibolit und Amphibolitgneis; 7, Kalkstein und Skarn; 8. Leptit und Leptitgneis, die oft Porphyroblasten von Cordierit, Almandit, Sillimanit oder Andalusit enthalten.

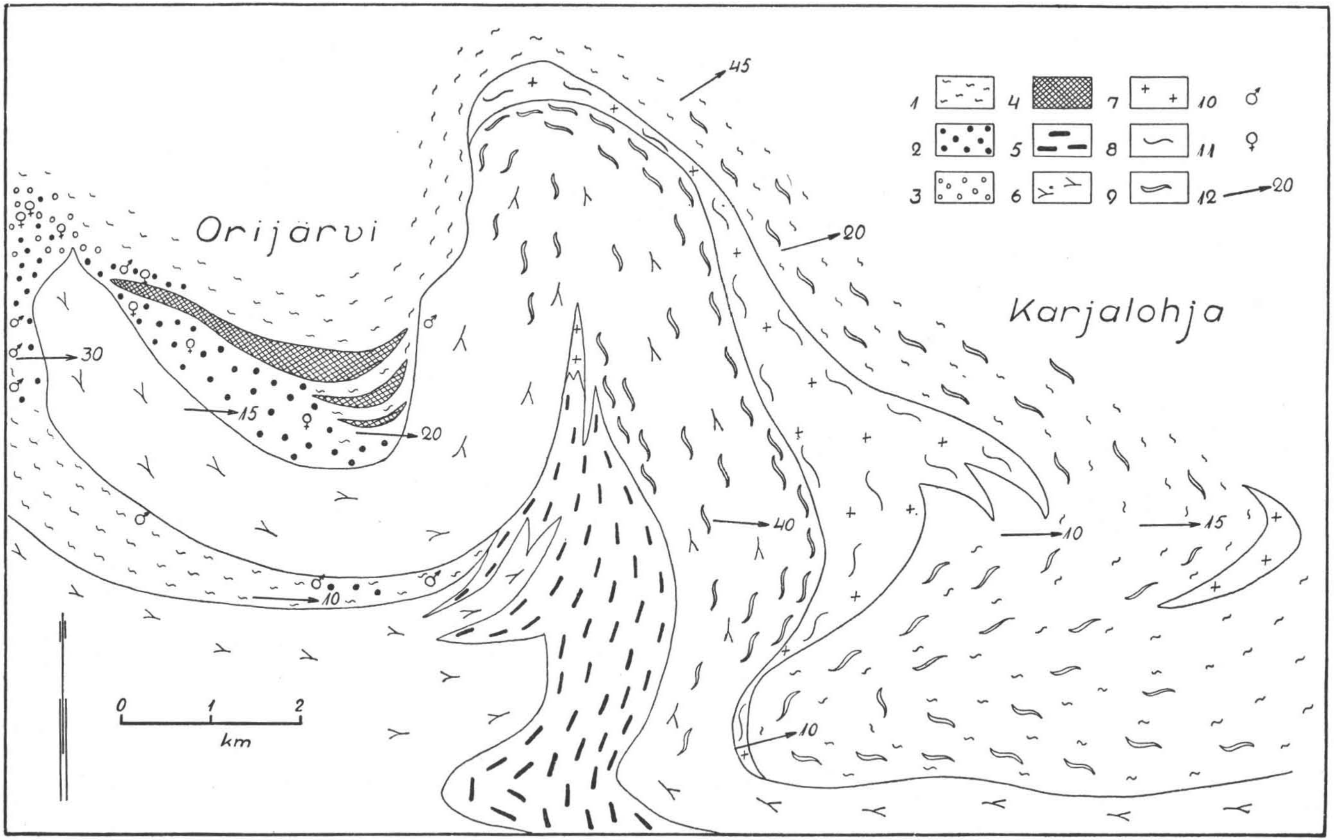

Abb. 4. Geologische Karte des Gebiets zwischen Orijärvi und Karjalohja. 1. Die Leptitformation; 2. Cordierit- und antophyllithaltige Gesteine; 3. Andalusithaltige Gesteine; 4. Amphibolit; 5. Diorit; 6. Oligoklasgranit oder »Gneisgranit»; 7. Mikroklingranit; 8. Fragmente von älteren Gesteinen im Mikroklingranit; 9. Schlieren von Mikroklingranit oder migmatitisierte Gesteine; 10. Magnetitvorkommen; 11. Sulfidvorkommen; 12. Die Achsialrichtung. Tuominen und Mikkola (1950).

getreueren Kartenbild gelangt. Es ist aber ebenso wahrscheinlich, dass der angenommene geschmeidige Stil oft zu einem entgegengesetzten Resultat geführt hat. So kommen auch andere Granite als die oben erwähnten, und namentlich alle spätorogenen Tiefengesteine, als »steife» klumpenförmige Massive vor. — Ich möchte auch in diesem Zusammenhang betonen, dass es bei den Kartierungsarbeiten wichtig ist, seine Aufmerksamkeit auch auf die tektonische Erscheinungsweise der Tiefengesteine zu richten.

Mit dem Obigen ist keineswegs gemeint, dass die tektonische Forschung in Finnland im allgemeinen der Vergessenheit anheimgefallen wäre. Nachdem C. E. Wegmann gegen Ende der zwanziger Jahre d. Js. mit seinen Untersuchungen in Ostfinnland gezeigt hatte, dass die Überlegungen und Arbeitsmethoden, die für junge Gebirgsketten gelten, sich auch auf die präkambrischen Gebirge anwenden lassen (Wegmann, 1928, 1929a, 1929b, 1929c), fingen die finnischen Geologen an, sich nun ernsthafter mit tektonischen Fragen zu befassen. Neben der Faltungstektonik ist dann auch die Spaltentektonik Gegenstand der Untersuchungen gewesen (vgl. z. B. Härme, 1960). Ein klares Bild vom Reichtum des finnischen Felsgrundes an Spalten und Bruchzonen gibt die beigefügte (Abb. 1), von S. Penttilä entworfene Bruchlinienkarte des Suomussalmi—Hyrynsalmigebietes (Virkkala, 1960; vgl. auch Penttilä, 1961). Die Bruchlinien sind, ursprünglich im Massstab 1: 100 000, auf Grund von gewissen auf den ökonomischen Karten sichtbaren morphologischen Zügen konstruiert worden, aber offensichtlich repräsentieren diese, wenigstens zum grössten Teil, auch die im Felsgrunde vorkommenden Bruchlinien. Die kartierenden Geologen sind natürlich speziell für diejenigen Bruchlinien interessiert, längs denen grössere Verwerfungen geschehen sind und die demnach Einfluss auf das Kartenbild haben können. Selbstverständlich ist es von der Grösse des angewendeten Massstabes abhängig, in welchem Masse die Bruchlinien des Felsgrundes im Kartenbilde sichtbare Deformationen der von verschiedenen Gesteinen gebildeten Gebiete verursachen. Wie aus der Felsgrundkarte des Suomussalmigebiet im Massstab 1: 400 000 hervorgeht (Abb. 2), scheint keines von den zahlreichen Spaltensystemen des Gebiets einen deformierenden Einfluss auf den allgemeinen Stil der kleinmassstäbigen Übersichtskarte gehabt zu haben (Matisto, 1958).

In diesem Zusammenhang sei auf die von Tuominen (1957) im Massstab 1: 100 000 entworfene Karte über das Orijärvigebiet aufmerksam gemacht, die meiner Meinung nach ein ziemlich irreführendes Bild von dem Einfluss der Bruchlinien auf das Kartenbild gibt (Abb. 3). Keinesfalls kann die Spaltentektonik, auch wenn der Massstab viel grösser wäre, zu solch einem Kartenbild führen, wo die von allen Gesteinen gebildeten Teilgebiete ausnahmslos als geradlinig begrenzte kantige Figuren vorkommen. Es wäre zu wünschen, dass ein derartiger »kubistischer» Stil in den kommenden

Übersichtskartierungen nicht ohne weiteres angenommen würde, denn m. E. kann sie wohl hauptsächlich im künstlerischen Sinne von Bedeutung sein. Vergleichshalber mag eine von Tuominen und Toivo Mikkola (1950) nur einige Jahre früher entworfene Karte über das Orijärvigebiet vorgelegt werden (Abb. 4).

ÜBER DIE EINTEILUNG VON GRANITEN UND ANDEREN TIEFENGESTEINEN

DIE GRANITEINTEILUNG VON J. J. SEDERHOLM

Wie Sederholm (1928) berichtet hat, wurden bei den hauptsächlich in SW-Finnland, aber auch in den Gegenden nördlich des Sees Ladoga in den achtziger und neunziger Jahren des 19. Jahrhunderts ausgeführten Kartierungen, ausser den Rapakiwigraniten, nur die sog. älteren und jüngeren Granite ausgeschieden. Die ältere Gruppe umfasste hauptsächlich gneisartige, vorzugsweise graue, plagioklasreiche Granite, die jüngere wieder (speziell in SW-Finnland) mikroklinreiche, meistens rötliche Granite. Diese hinsichtlich der petrographischen Charakteristik nach und nach in gewissem Masse modifizierte Zweiteilung stützte sich zum grossen Teil auf die angenommenen Altersverhältnisse der Granite zu den bothnischen Schiefern in der Gegend von Tampere, die nach Sederholm altersmässig zwischen den beiden Granitformationen gelegen waren. In den Jahren 1914—1915 gelang es Mäkinen (1914, 1915) festzustellen, dass einige Gneisgranite die bothnischen Schiefer durchsetzten, sodass sie wenigstens teilweise ihrem Alter nach postbothnisch waren. Als man aber dann in Schiefergebiet von Tampere Konglomerate fand, deren Gerölle zum Teil aus gleichartigen gneisigen Tiefengesteinen bestanden, stellte es sich heraus, dass irgendwo auch präbothnische Gneisgranite vorkommen müssen.

Weitere Erklärung in obenerwähnter Hinsicht erhielt man, als Sederholm auf der Insel Bockholm in Enklinge (Kirchspiel Kumlinge) ein Bodenkonglomerat entdeckte, das unmittelbar auf Gneisgranit lag (Abb. 5; Sederholm, 1930). Das Konglomerat, den Sederholm (1934) später sorgfältig beschrieben hat, enthält in reichem Masse Gneisgranitgerölle, die ihrer chemischen Zusammensetzung nach genau gleichartig sind wie der in der Nähe vorkommende Gneisgranit. In gewissen Masse kommen auch Gerölle von Glimmerschiefer, sowie von basischen Gesteinen vor. Das Konglomerat geht allmählich in ein Gestein, das man eine metamorphosierte Arkose nennen könnte, und weiter in einen Glimmerschiefer über. Am nächsten an dem Kontakt mit dem Gneisgranit kommt ein Gestein vor, das nach Sederholm aus verwittertem granitischem Material gebildet worden ist. Der Gneisgranit durchschneidet seinerseits eine ältere Sedimentformation. Es ist klar,

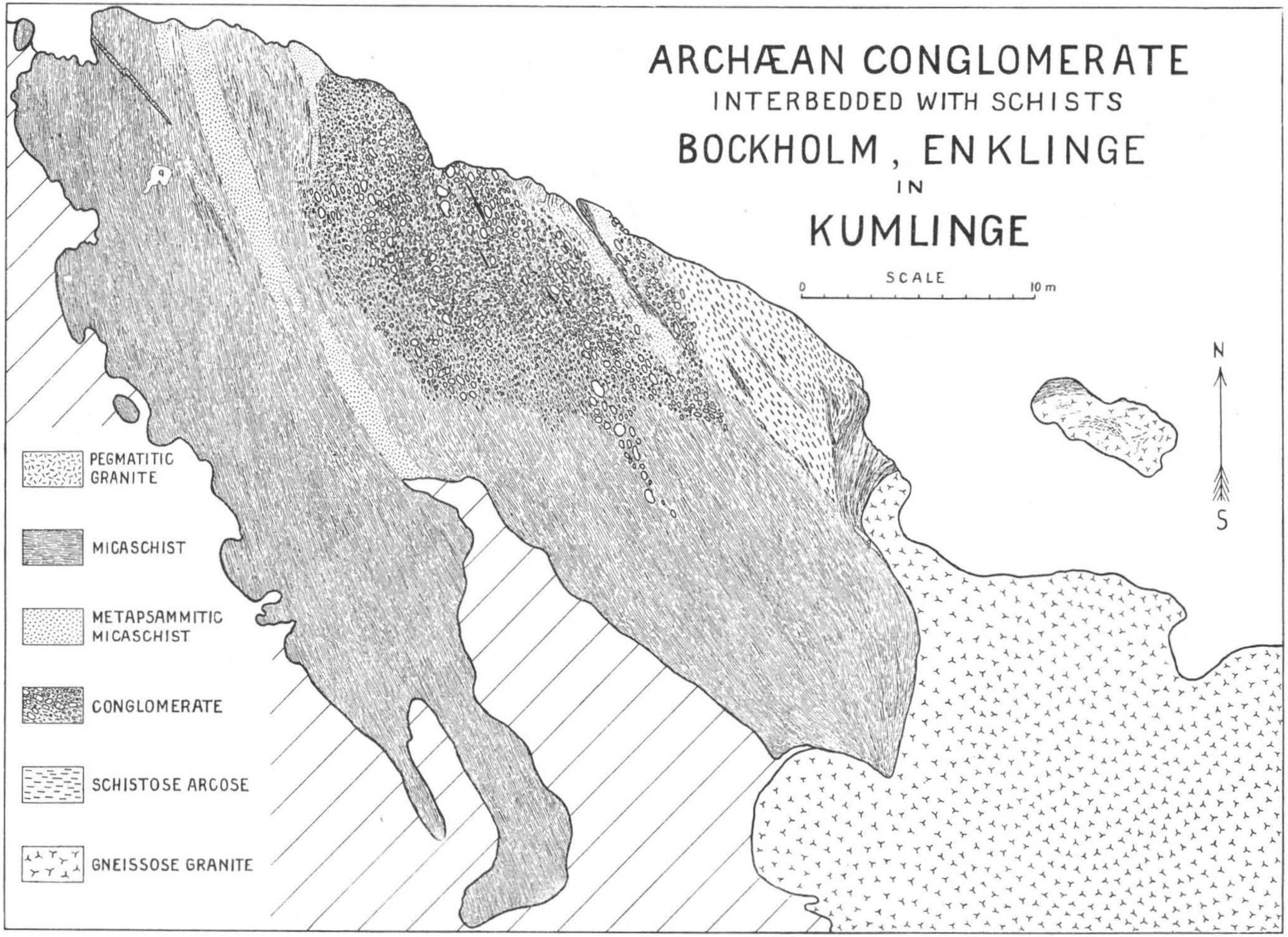

Abb. 5. Das Bodenkonglomerat auf der Insel Bockholm in Enklinge, Kirchspiel Kumlinge, Nach Sederholm (1930).

dass wir hier mit einer Diskordanz erster Ordnung zu tun haben. Im Enklingegebiet kommen also Gneisgranite vor, die viel älter sind als diejenigen Gneisgranite, die die bothnischen Schiefer durchschneiden. Ebenso treten dort suprakrustale Gesteine auf, die viel älter sind als die bothnischen Schiefer. Diese ältere Formationsserie können wir im Einverständnis mit Sederholm Svionium nennen.

Was die jüngeren Granite Südfinnlands (die Granite von Hangötypus) betrifft, erstrecken sie sich, wie Sederholm (1928) berichtet hat, in östlicher Richtung gegen den Ladogasee und von dort über die Gegenden von Saimaa gegen NW, wo sie sich mit den Graniten des grossen zentralfinnischen Gebiets vereinigen. Nirgends ist es gelungen, diese Granite geologisch von einander zu trennen. Von diesen »jüngeren Graniten» kann man in den letztgenannten Gebieten jedoch gewisse saure Tiefengesteine ausscheiden, die, wie Sederholm bemerkt, schon vor dem Empordringen der jüngeren Granite schiefrig geworden waren und demnach älter als diese sind. Die postbothnischen Granite teilen sich also nach Sederholm in zwei Untergruppen, obschon dieser später mitteilt (Sederholm, 1932), dass es nicht ganz klar ist, ob die Intrusionen der beiden Granitgruppen Phasen von derselben Periode von plutonischer Aktivität repräsentieren, oder ob sie sich an verschiedene Perioden anschliessen.

Es war das Studium der gegenseitigen Verhältnisse der Schiefer- und Granitformationen in Ost- und Nordfinnland, das zur Ausscheidung einer dritten Hauptgranitgruppe führte. Wie Sederholm berichtet, wurden im Laufe dieser Untersuchungen mächtige Konglomeratformationen angetroffen, die nicht nur Gneisgranitgerölle, sondern auch Gerölle von massigen Mikroklingraniten, turmalinführenden Pegmatiten und anderen derartigen Gesteinen enthielten, die an diejenigen der postbothnischen Tiefengesteinsgruppe erinnerten. Auf Grund dieser Beobachtungen, wie auch auf die stratigraphischen und Kontaktverhältnisse der erwähnten Suprakrustalformationen gestützt, wurden diese von denjenigen Schiefern, die von den postbothnischen Graniten durchdrungen waren, getrennt und zu einem speziellen, kalevischen System gerechnet. Man nahm anfangs an, dass dieses jünger war als alle Granite vom Grundgebirgstyp, aber im Laufe dei Untersuchungen wurden schon am Anfang dieses Jahrhunderts Granite angetroffen, die die Gesteine des Systems durchsetzten. Zu diesen »postkalevischen» Graniten wurde später einige Granitmassive in SW-Finnland gerechnet. — Sederholm (1932) berichtet, dass es möglich ist, dass in westlichen Fennoskandia zwei Untergruppen von postkalevischen Graniten vorkommen.

Sederholm teilte also die Granite nach ihrem Alter in vier Hauptgruppen (die postsvionischen, postbothnischen und postkalevischen Granite, und die Rapakiwigranite), von denen die zweite und dritte Gruppe sich in zwei Untergruppen teilten. Jede von den drei ersten Gruppen ist während einer bestimmten orogenen Periode emporgedrungen. Die Rapakiwigranite haben nach Sederholm nicht an den orogenen Bewegungen teilgenommen.

Sederholms Graniteinteilung hat nicht in jeder Beziehung allgemeine Anerkennung gefunden. Dies beruht m. E. in erster Linie darauf, dass man allgemein der Meinung war, dass die von Sederholm vorgelegte subbothnische Diskordanz nicht so bedeutend ist, dass sie Bildungen aus zwei verschiedenen orogenen Zyklen von einander trennte. So erwähnt z. B. Eskola (1941), dass die von Erkki Mikkola 1928 und 1931 in Lavia und Suodenniemi angetroffenen und von Sederholm (1931) eingehend beschriebenen zahlreichen Bodenbreccien und Konglomerate mit Granit- und Dioritgeröllen nicht auf eine grosse Diskordanz hinweisen. Die Seltenheit der Gerölle von wahren Tiefengesteinen und das Fehlen von Migmatiten in ihnen sollen von einer derartigen Sachlage zeugen (vgl. auch Eskola, 1953). Über die Konglomerate des Enklingegebiets erwähnt Eskola im selben Zusammenhang, dass die Beweise nicht eindeutig genug seien, die Geologen davon zu überzeugen, dass hier zwei von einem präbothnischen Granit getrennte Serien von Suprakrustalformationen vorkämen. Zum Schluss erwähnt Eskola, »dass nunmehr ziemlich sichergestellt ist, dass das Svionium und das Bothnium zum selben orogenen Zyklus der Svekofenniden gehören.» Dies scheint immer noch eine ziemlich allgemeine Auffassung der finnischen Geologen zu sein. So z. B. berichtet Simonen (1953), der sich der Auffassung Eskolas anschliesst, dass die Konglomerate des Suodenniemi-Laviagebiets »do not show the character of real basal conglomerates separating two different orogenic cycles because they do not contain migmatite boulders, indicatoıs of a deep erosion.» Die subbothnische Diskordanz wäre nur ein Resultat von lokaler Erosion während der orogenen Entwicklung. Nach Simonens Meinung betonte Sederholm allzu viel die Bedeutung von gewissen Diskordanzen! Etwas später spricht Simonen (1960a) grösserer Gewissheit halber als seine Auffassung aus, dass »no valid evidence of a sedimentation floor or basement for Svecofennian sedimentation has been found.» — Die Auffassungen von Eskola teilt auch Väyrynen (1954).

Es mag noch erwähnt werden, dass Seitsaari (1951) innerhalb des Schiefergebiets von Tampere, auf der Ostseite des Sees Näsijärvi, Konglomerate angetroffen hat, die Gerölle von suprakrustalen, hypoabyssischen und auch von plutonischen Gesteinen enthalten. Die letztgenannten (Aplite, Granodiorite) sind verhältnismässig spärlich vertreten. Nach Seitsaari weisen die Konglomerate auf eine bedeutende interformationäre Erosionsperiode hin, nicht aber auf eine Diskordanz im vollsten Sinne des Wortes. Seitsaari hat übrigens in der Schieferzone von Tampere auch ein Quarzitkonglomeratlager angetroffen. Dies ist bemerkenswert, da Quarzite im bothnischen Grundgebirge sehr selten sind.

Nach den obenbeschriebenen Auffassungen fliessen die postsvionischen und postbothnischen Granite zusammen; an ihrer Stelle hätten wir nur »postsvekofennische Granite.» Meiner Meinung nach hat man sich hier jedoch nicht mit erforderlichem Ernst zu den wichtigen Untersuchungsresultaten Sederholms und Erkki Mikkolas verhalten. Die Erklärungen Eskolas und der anderen erwähnten Forscher über die Bedeutung der beschriebenen Konglomerate haben kaum das Richtige getroffen. Es ist erstens ganz unbegreiflich, warum diejenigen Konglomerate, die von einer tiefgreifenden Erosion zeugen, gerade Migmatitgerölle enthalten müssten. Auf die Existenz von grossen Diskordanzen deuten ja ausdrücklich die Gerölle von Tiefengesteinen hin. Der Umstand, dass die erwähnten Konglomerate oft in verhältnismässig geringen Masse Tiefengesteinsgerölle enthalten, weist keineswegs auf eine entgegengesetzte Sachlage hin. Die Zusammensetzung des Geröllmaterials, sowie der Gehalt an Tiefengesteinsgeröllen, variiert begreiflicherweise von Stelle zu Stelle je nachdem, welche Gesteine in dem älteren Felsgrunde jeweilig entblösst waren. Ausserdem enthalten ja die fraglichen Konglomerate stellenweise, wie z. B. in Enklinge, in reichlichem Masse Tiefengesteinsgerölle.

Aller Wahrscheinlichkeit nach kommen also im südlichen und südwestlichen Finnland zwei verschiedenaltrige Hauptgruppen von Tiefengesteinen, wie auch zwei suprakrustale Formationsserien vor, die von einander durch eine Diskordanz erster Ordnung getrennt sind. Ich werde später (S. 287) weitere Beweise vorlegen.

Es ist ein grosses Verdienst J. J. Sederholms, dass er mit seinen Untersuchungen die Aufmerksamkeit auf die Existenz der subbothnischen Diskordanz gerichtet hat. Aufgabe künftiger Untersuchungen ist es, die geographische Verbreitung der svionischen und der jüngeren Formationen klarzulegen. Es ist sehr beklagenswert, dass diese Umstände nicht mehr Aufmerksamkeit während der letzten Übersichtskartierungen im südlichen und südwestlichen Finnland auf sich gezogen haben. Dies bedeutet m. E. einen Rückschritt in der Entwicklung der geologischen Forschungsarbeit.

DIE VON M. SAKSELA ANGEWENDETE TEKTONISCHE EINTEILUNG DER TIEFENGESTEINE

Wie schon erwähnt wurde (S. 257), teilte ich die Tiefengesteine Mittel- und Südostbothniens in eine synorogene und eine spätorogene Gruppe ein (Saksela, 1930, 1932, 1933, 1934, 1935, 1936, 1953). Die Vertreter der synorogenen Gruppe kommen als konkordante Intrusionen vor. Sie bilden mit der Streichrichtung der umgebenden Schiefer parallele, entweder längliche ausgedehntere Massive oder schmälere lager- und gangartige Massen. Die Gesteine der spätorogenen Gruppe dagegen treten als rundliche oder unregelmässig geformte Gebiete auf, deren Grenzlinien, sowohl in grossen als kleinen Zügen betrachtet, sehr oft quer über die Streichrichtung der älteren Gesteine gerichtet sind.

Die Gesteine der beiden Gruppen verhalten sich verschiedenartig auch in Bezug auf die anderen Parallelstrukturen. So verlaufen die Streckungsrichtungen und die Achsen der Kleinfalten sowohl in den synorogenen Intrusiven als auch in den Schiefern, die sie umgeben oder an die sie grenzen, in auffallendem Masse gleich und sind regional gleichen Schwankungen unterworfen. Es ist interessant, die Kontaktzonen der erwähnten Gesteine zu studieren, wo das synorogene Intrusivgestein gewöhnlich reichlich Schieferfragmente enthält. In diesen Fragmenten sind die Streckungsrichtungen dieselben wie im angrenzenden einheitlichen Schiefergebiet. Dieselbe Orientierung ist auch in den kleinsten Fragmenten sogar ausserhalb der eigentlichen Kontaktzone zu beobachten. Es sei noch erwähnt, dass der obenbeschriebene Parallelismus sogar im »molekularen Bereich», in der Gitterregelung der einzelnen mineralischen Gemengteile, festzustellen ist. Die ostbothnischen Gesteine sind gefügeanalytisch bis auf weiteres nicht untersucht worden, aber Hietanen (1943) hat die synorogenen Intrusivgesteine (die Glieder der Trondhjemitserie und die benachbarten sedimentogenen Gesteine, Kinzigite *etc.)* des Kalantigebiets, SW-Finnland, einer solchen Untersuchung unterzogen.

Die erwähnten Tatsachen zeigen deutlich, dass die Glieder der synorogenen Gruppe sich der Tektonik des umgebenden Felsgrundes geschmeidig anpassen. Ganz anders verhält es sich mit den Vertretern der spätorogenen Gruppe. Die Grenzen der von den Gesteinen dieser Gruppe gebildeten Massive schneiden im Kartenbild sowohl die lokalen Streckungsrichtungen als auch das allgemeine Achsialstreichen schroff ab. Diese Gruppe ist wenig vom tektonischen Bau des umgebenden Felsgrundes abhängig. Es ist interessant, in diesem Zusammenhang feststellen zu können, dass dort, wo die Vertreter der spätorogenen Gruppe als grössere Massive vorkommen, in ihrer nächsten Umgebung Störungen in der Tektonik des älteren Felsgrundes zu beobachten sind. So beobachtet man u. a. merkbare Abweichungen der Streckungsrichtungen von den vorausgesehenen und denjenigen, die weiter entfernt herrschend sind.

Wie man schon auf Grund des oben Erwähnten schliessen kann, besitzen die beiden Tiefengesteinsgruppen eine verschiedene Art mit den älteren Gesteinen Mischgesteine, Migmatite zu bilden. Die Schiefer in der Nähe der synorogenen Intrusionen sind oft adergneisartig ausgebildet. Sie enthalten zahlreiche schmälere oder breitere konkordante Adern der letztgenannten Gesteine. Ebenso enthalten die Intrusive zahlreiche längliche Schieferfetzen. Wo wieder die Gesteine der spätorogenen Gruppe an die Schiefer grenzen, kann ein Eindringen des Intrusivgesteins in die Schiefer beobachtet werden, aber die Intrusivgänge sind jetzt nicht, wie im vorigen Falle, konkordant, sondern sie schneiden die Schiefer in beliebigen Richtungen ab. Wenn die Gänge in reichlicheren Mengen vorkommen, bildet sich in der Kontaktzone ein Mischgestein, dass seinem Charakter nach agmatitisch, intrusivbreccienartig ist.

Die obenerörterten Züge in der Erscheinungsweise der beiden Tiefengesteinsgruppen sind so charakteristisch und lassen sich im Felde so deutlich feststellen, dass die Einteilung der Tiefengesteine gut auf sie gegründet werden kann. Die Verteilung der beiden Gruppen geht aus der Karte in Tafel I hervor. Wie schon erwähnt, gehören zu den beiden Gruppen Gesteine von Peridotiten und Gabbros bis Granite und Granitpegmatite. Die synorogenen Plutonite sind oft deutlich parallelstruiert, aber auch massige Varietäten kommen vor. Die Gesteine der spätorogenen Gruppe sind im allgemeinen ihrer Struktur nach mehr oder weniger massig, obschon schiefrige Abarten angetroffen werden.

Wie sind nun die Verschiedenheiten in der Erscheinungsweise der beiden Tiefengesteinsgruppen zu erklären. Es ist offenbar, dass sie in kausalen Zusammenhang mit der tektonischen Entwicklung des Felsgrundes zu stellen sind. Nach meiner Ansicht beruhen die Verschiedenheiten auf dem ganz verschiedenartigen Intrusionsmechanismus der Tiefengesteine, der wiederum davon abhängt, dass die erwähnten Gesteinsgruppen während orogenetischer Phasen verschiedener Art in den sich deformierenden Komplex eingedrungen sind. Die frühere dieser Hauptphasen ist durch kräftige tangentiale Bewegungen nach Art der alpinen Deckenüberschiebungen charakterisiert, die spätere wieder ist vom Typus der sog. Grundfaltung. Während einer orogenetischen Phase erstgenannter Art sind die synorogenen Tiefengesteine hervorgedrungen. Zahlreiche Überschiebungsflächen (Scherflächen) haben hierbei als natürliche Kanäle fungiert, längs welchen die Granite *etc.* leicht und in grossen Mengen haben eindringen können. Solange sich die einzelnen Decken gegeneinander bewegten, und die Scherflächen einen kleinen Winkel mit der Druckrichtung bildeten (also sich in ziemlich flachgeneigter Lage befanden), war der Teildruck, der danach strebte, die Intrusionskanäle zu verschliessen, verhältnismässig gering, und die Intrusion konnte fortsetzen. Die Intrusionskörper gestalteten sich in dieser Phase begreiflicherweise mehr oder weniger lamellenartig; sie bildeten Ophiolite oder Ophiolithoide. Wenn die Überschiebung lange genug fortgesetzt hatte, begann das ganze sich deformierende Segment des Felsgrundes kräftiger zu reagieren. Die Decken reagierten — um Wegmanns (1928) Worte zu gebrauchen — als eine einheitliche Masse und konsumierten jetzt einen Teil der Energie in einer oder mehreren Aufwölbungen des ganzen Gebäudes. Den zur Zeit der so beginnenden Grundfaltung hervorbrechenden Tiefengesteinen — den Gliedern der spätorogenen Gruppe — standen nicht mehr die obengenannten durch die

Überschiebungsflächen dargebotenen Kanäle zur Verfügung, sondern sie suchten sich anfangs in die nach oben gewölbten Teile der steifen Kruste.

Auf Grund der obigen Erklärungsweise wird es auch leicht begreiflich, dass die verschiedenen Glieder der synorogenen Differentiationsserie in den Überschiebungsbewegungen von einander getrennt wurden. Die Glieder der spätorogenen Serie wieder sind gewöhnlich in Fühlung mit einander geblieben, die basischen Glieder sehr oft an den Randpartien der Massive liegend.

Die beschriebene tektonische Einteilung der Tiefengesteine war, wie aus dem folgenden hervorgeht, keineswegs unbekannt, als ich mit den Kartierungsarbeiten in Ostbothnien anfing. Ich habe sie nur — als erster in Finnland — konsequent in weiten Gebieten angewendet und daneben die Züge, die die beiden Tiefengesteinsgruppen charakterisieren, präzisiert. Schon Törnebohm (1880) beobachtete, dass die Granite hinsichtlich ihrer Relation zu den anderen, älteren Gesteinen sich verschiedenartig verhalten können. Die massigen »jüngeren Granite» bilden Massive mit rundlichen oder elliptischen Konturen, die die Schichten der umgebenden Gesteine bald schroff abschneiden, bald wieder diese seitwärts biegen. Die älteren »Urgranite» können nach Törnebohm ihrer Struktur nach massig sein, sie gehen aber oft in gneisartige Varietäten über. Die Urgranite treten oft als mehr oder minder linsenförmige Massen in den geschichteten oder schiefrigen Gesteinen auf. Sie enthalten, speziell in den Kontaktzonen, Fragmente von den benachbarten Schiefern, diese sind aber nicht, wie die Fragmente in den jüngeren Graniten, unregelmässig geformte scharfkantige Bruchstücke, sondern abgeplättete Streifen und Fetzen. — Später sind in Schweden zahlreiche Beobachtungen gemacht worden, die in dieselbe Richtung deuten (vgl. z. B. Geijer, 1916; Asklund 1921; Sundius 1921a, 1921b, 1923; A. Högbom, 1928, 1931).

In Ostbothnien bemerkte schon Mäkinen (1916) Verschiedenheiten in der Erscheinungsweise der Tiefengesteine. Nach ihm sind in den suprakrustalen Komplex zuerst Magmen eingedrungen, die im allgemeinen Massive konkordanter Art bildeten. Nach dieser Phase folgen, wie Mäkinen schreibt, batholitische Intrusionen. Diese sind scharf und deutlich abgegrenzt sowohl gegen den suprakrustalen Komplex als auch gegen die früher intrudierten infrakrustalen Gesteine. Mäkinen kam also für Ostbothnien zu ziemlich gleichartigen Resultate wie ich. Er unterscheidet jedoch im mittleren Ostbothnien nur eine einzige Tiefengesteinsgruppe, die er dann auf Grund rein petrographischer Gesichtspunkte in Untergruppen einteilt.

H. Cloos (1928) hat die Beziehungen der Kontakte von Plutonen zur Tektonik des Nebengesteins beschrieben und folgende drei Fälle unterschieden:

1. Konkordanz — infolge Benützung von Schicht- oder Schieferungsfugen, meist unter tektonischen Bewegungen längs denselben.

2. Diskordanz — schneidet die vorhandenen Strukturen infolge:
   a) mechanischer Durchbrechung,
   b) Benutzung diskordanten Fugen.
3. Akkordanz — scheinbare, zufällige Konkordanz sonst diskordanter Kontakte.

Nach Cloos sind diese Unterscheidungen meist leicht zu treffen. Sie sind gute Hilfsmittel für die Ermittelung der Beziehungen zur tektonischen Bewegung (vgl. auch Cloos, 1923).

Tröger (1929) unterschied in der variskischen Gebirgsbildung, ausser einer frühtektonischen Phase, an die sich die eruptiven Vorläufer anschliessen, eine haupttektonische und eine spättektonische. Als unbestrittene Repräsentanten der haupttektonischen Phase können nach ihm die meisten der mitteldeutschen Orthogneismassive angesprochen werden.

Es mögen noch die Untersuchungen von Wegmann und Kranck (1931) im Schärenhof von Helsinki erwähnt werden, die Interesse auch für die magmatektonischen Fragen erweckten.

Nach dem Vorschlag von Eskola und Wegmann (Eskola, 1930) wurden die Bezeichnungen »synorogen», »syntektonisch» usw. durch »synkinematisch» usw. ersetzt, und in der dritten Auflage von Ramsay's Lehrbuch (Ramsay, 1931) hat Eskola die Begriffe synkinematisch und postkinematisch definiert. Die synkinematischen Tiefengesteinsmassen sind während der eigentlichen gebirgsbildenden Bewegungen intrudiert worden, worauf die primär gneisige Struktur und die konkordante Erscheinungsweise dieser Gesteine hindeuten sollen. Die postkinematischen Tiefengesteine sind ihrer Struktur nach richtungslos und sie verhalten sich diskordant zu den durchsetzten Schiefern. Sie sind während einer späten Phase der Entwicklung der Gebirgskette eingedrungen, als diese schon in hohem Grade konsolidiert war, und die Bewegungen aufgehört hatten. Es scheint, als ob Eskola die spätkinematischen Tiefengesteine vergessen hätte. Diese sind ja auch deutlich orogen, obschon sie oft massig sind und sich diskordant zu ihrer Umgebung verhalten.

Ich habe zu beweisen versucht (Saksela, 1953), dass die von mir vorgelegten tektonischen Tiefengesteinsgruppen deutlich auch anderswo in Finnland vorkommen. Ich möchte auch in diesem Zusammenhang den Umstand betonen, dass die in Südfinnland reichlich vorkommenden mikroklinreichen Granite, die man »jüngere Granite» oder »Hangögranite» genannt hat, und die in der letzten Zeit ohne weiteres zu der spätkinematischen Gruppe gerechnet worden sind, zum bedeutenden Teil auch synkinematische Granite vertreten. So geht aus der Karte (Abb. 4) von Tuominen und T. Mikkola (1950) hervor, dass der mikroklinreiche Granit derselben Intrusionsserie wie der Oligoklasgranit angehört. Er kommt nämlich ebenso deutlich synkinematisch wie der letztgenannte Granit vor, zwischen der Oligoklasgranitintrusion und der Leptitformation eine schmale konkordante Masse bildend, die das Bestreben hat, an den Biegungstellen anzuschwellen oder kleinere getrennte phakolitartige Partien zu bilden. Auf den synkinematischen Charakter des Mikroklingranits deutet auch der Umstand hin, dass dieser Granit mit den älteren Gesteinen adergneisartige Migmatite bildet, die wenigstens zum beträchtlichen Teil arteritisch sind. Diese Arterite sind, wie Eskola bemerkte, »folded and contorted in detail, but, on the whole, showing the general strike of the leptite belt» (Eskola, 1914, S. 35). Später erwähnt Eskola (1956), dass die kalireichen Granite des Küstengebiets Südfinnlands teilweise die Streichrichtungen der älteren Gesteine schroff abschneiden, teilweise als konforme Intrusionen in den umgebenden Schiefern vorkommen und mit diesen Adergneise bilden. Diese Beobachtungen weisen deutlich darauf hin, dass die erstgenannten Mikroklingranite zu der spätkinematischen Intrusionsserie, die letztgenannten zu der synkinematischen Serie gehören, obwohl alle von Eskola spätkinematisch genannt worden sind.

Bemerkt sei in diesem Zusammenhang, dass noch nicht geklärt worden ist, in welchem Masse die obengenannten Granite postsvionisch oder postbothnisch sind. Auf Sederholms Felsgrundkarte von Finnland (Sederholm, 1930) sind die Gneisgranite des Orijärvigebiets als postsvionisch und die Mikroklingranite als jünger postbothnisch bezeichnet worden. Ältere postbothnische Granite kommen auf der erwähnten Karte nicht südlich von der ostwestlichen Linie Orimattila—Riihimäki—Somero—Nousiainen vor.

Die von mir angewendete tektonische Einteilung gründet sich also, wie schon hervorgegangen sein dürfte, auf keinerlei Vorstellungen oder Annahmen, sondern nur auf im Felde beobachtete Tatsachen, welche ich genau beschrieben habe. Die Einteilung kann sich auch auf ziemlich grosses Beobachtungsmaterial stützen. Es kann also gar nicht von der Richtigkeit oder Unrichtigkeit des Einteilungsprinzips die Rede sein, sondern höchstens vom Grade seiner Zweckmässigkeit. Zweckmässig ist es m. E. schon deshalb, weil die magmatektonischen Beobachtungen in sehr hohem Masse helfen, ein wirklichkeitsgetreues Kartenbild zustandezubringen. A b e r a l l e e n t s c h e i d e n d e n K o n t a k t b e o b a c h t u n g e n z e i g e n, d a s s d i e s e t e k t o n i s c h e E i n t e i l u n g g l e i c h z e i t i g a u c h e i n e A l t e r s e i n t e i l u n g i s t, d i e z u z w e i s e l b s t ä n d i g e n v e r s c h i e d e n a l t r i g e n E r u p t i o n s s e r i e n (D i f f e r e n t i a t i o n s s e r i e n) f ü h r t. Die synkinematische Serie ist die ältere, die spätkinematische Serie die jüngere (im Kreise jeder orogenen Zone). Dies steht in Übereinstimmung mit der obengegebenen Erklärung, worauf die Verschiedenheiten in der Erscheinungsweise der Tiefengesteine beruhen.

Was die e n t s c h e i d e n d e n Kontaktbeobachtungen betrifft, mögen zuerst diejenigen b a s i s c h e n G ä n g e erwähnt werden, die in der Untersuchungen Sederholms im südfinnischen Küstengebiet eine ganz besondere Bedeutung erlangten, weil sie die Gneisgranite durchschneiden, aber älter als die »jüngeren Granite» sind. Derartige Gänge, die also eine sichere Unterscheidung der beiden Altersgruppen der Tiefengesteine voneinander ermöglichen, wurden in Ostbothnien schon von Mäkinen (1916) angetroffen. Später haben auch andere Forscher solche in Ostbothnien gefunden. Väyrynen (1936a) beobachtete im Kirchspiel Kärsämäki, dass der ältere Granit von einem geraden, ca. 100 m langen Gang eines basischen Gesteins quer zur Schieferung durchschnitten ist, und dass dieser Gang wiederum von jüngeren granathaltigen Aplitadern, die augenscheinlich mit den jüngeren Graniten der Nachbarschaft zusammengehören, durchsetzt ist. — Es scheint eine allgemeine Auffassung zu sein, dass die basischen Gänge eine Unterbrechung in der Orogenese, eine ziemlich ruhige interorogene Phase bedeuten (vgl. z. B. Väyrynen, 1936; Metzger, 1945; Edelman, 1949, 1960; Berthelsen, 1955).

Auf die Existenz von zwei nichtkomagmatischen Eruptionsserien weisen deutlich auch diejenigen Kontakte hin, wo ein basisches Tiefengestein durchschneidende Adern in einen saureren Plutonit sendet. So z. B. dringen im Kirchspiel Teuva in den synkinematischen Gneisgranit Gänge von basischen Tiefengesteinen ein, die ihrerseits als Bruchstücke in den spätkinematischen Graniten vorkommen (Saksela, 1935). Im Kirchspiel Sievi durchsetzt ein Hornblendegabbro einen Biotitgneisgranit (Saksela, 1933).

Es ist klar, dass die erwähnten basischen Intrusionen nicht als komagmatisch mit den durchsetzten saureren Gesteinen sein können, denn in einer regelmässigen magmatischen Entwicklung kann ein solcher Rückschritt eines sauer gewordenen Magmarückstandes in ein basisches Magma stattfinden (vgl. Väyrynen, 1936).

Auf die Existenz von zwei verschiedenen Intrusionsserien weisen zum Schluss auch diejenigen Beobachtungen hin, nach welchen die normalen Granite saure Pegmatite oder arteritische Adergneise, deren Gangmaterial aus Granitpegmatit besteht, durchsetzen (Abb. 6). Derartige Beobachtungen sind in Ostbothnien, sowie auch anderswo in Fennoskandia gemacht worden (vgl. Sundius, 1930; Lundegårdh, 1960, Abb. 21).

Die von mir angewendete tektonische Einteilung der Tiefengesteine ist immer noch aktuell. Es mag erwähnt werden, dass z. B. Turner und Verhoogen (1960) in ihrem Lehrbuch eine genau gleichartige Einteilungsweise beschreiben. Einige finnische Geologen haben jedoch nicht dieses Einteilungsprinzip angenommen. So soll sich nach Väyrynen (1936) die Einteilung auf zwei irrtümliche Vorstellungen gründen, erstens auf die Annahme einer einzigen orogenen Periode, und zweitens darauf, dass die Bewegungen während dieser Periode alpinen Charakter gehabt haben. Was die erste Bemerkung Väyrynens betrifft, war ich vor dreissig Jahren noch der Meinung, dass

Abb. 6. Pegmatitschlieriger Gneisgranit, der vom Stockholmsgranit schroff abgeschnitten ist. Rechts oben ein kleineres Bruchstück von Gneisgranit. Hässelby. Massstab 1: 23. Sundius (1930).

Mittel- und Süd-Ostbothnien, sowie SW-Finnland zum Kreise eines einzigen orogenen Zyklus gehören. Aber derselben Auffassung war ja offenbar auch Väyrynen, der sie noch in seinem Lehrbuch vorlegt (Väyrynen, 1954, S. 50). Wie aus dem früher Gesagten hervorgegangen ist, können wir jedoch im Tamperegebiet und in Südfinnland zwei orogene Zyklen unterscheiden. So verhält es sich auch in Ostbothnien, wie später gezeigt werden soll. Dies hindert jedoch auf keine Weise die Anwendung der tektonischen Einteilung. Wir müssen nur des Umstandes eingedenk sein, dass wir es j e t z t in den genannten Gebieten mit vier Tiefengesteinsgruppen, mit zwei postsvionischen und zwei postbothnischen, zu tun haben!

Was die zweite Bemerkung von Väyrynen betrifft, hatte schon Wegmann (1928, 1929a, b, c; Wegmann und Kranck, 1931) gezeigt, dass sich die Betrachtungsweisen der alpinen Geologie auch auf die Erklärung der tektonischen Entwicklung sowohl der Kareliden, als auch der »Svekofenniden» anwenden lassen. Zu gleichartigen Resultaten sind Kranck (1933) im Schärenhof von Ekenäs und ich in Ostbothnien gelangt (Saksela, 1932, 1933, 1935). Metzger (1945, 1947) steht auf dem Standpunkt, dass für die erste Deformation der »Svekofenniden» ein alpiner Stil angenommen werden kann. Die zweite Phase dokumentiert sich nach ihm im Parainengebiet durch eine Art von tangentialer Zusammendrückung der gesamten Bildung. Diese Phase der Faltung wird schliesslich durch eine allmähliche Versteifung der Gebirgsmasse abgelöst, und es kommt zur eigentlichen Grundfaltung im Sinne von Argand.

Nach Väyrynen (1936b) reagierte alles wie eine plastische Masse bei der Intrusion der älteren »svekofennischen» Tiefengesteine. Die Gesteinsmassen wurden nur plastisch zusammengedrückt. Man kann nach Väyrynen keine Spur von starreren Schollen erblicken, die eine Aktivität gegen ihre Umgebung gezeigt hätten. Bei der Intrusion der jüngeren Reihe der Tiefengesteine reagierten schon die früher intrudierten Gesteine starr, und jetzt sind nach Väyrynen die flachen Translationsbewegungen in alpinem Stil möglich. In SW-Finnland haben nach Väyrynen (1956) stellenweise kräftige Überschiebungen stattgefunden, sie können aber von den karelidischen Bewegungen verursacht sein! — Simonen (1960a) erwähnt, dass die grossen Achsialkulminationen und -depressionen, die die karelische Zone charakterisieren, nicht im »svekofennischen» Raum gefunden worden sind!

Die Auffassungen von Väyrynen können nicht Stich halten. Wenigstens während der »bothnischen» Orogenese waren »die starren Schollen», die svionischen Bildungen, von Anfang an vorhanden. Die Unhaltbarkeit der Behauptung von Simonen geht schon aus den in Ostbothnien ausgeführten Untersuchungen hervor (Saksela, 1932, 1933, 1935). — Die Entwicklung im »svekofennischen» Raum ist in grossen Zügen von alpinem Typ.

Nach Simonen (1960b) ist die Zweckmässigkeit der von mir angewendeten tektonischen Einteilung fraglich, weil nicht bestätigt worden ist, dass in Ostbothnien zwei verschiedenaltrige Differentiationsserien vorkämen. Zur Stütze seiner Behauptungen führt Simonen eine von Salli (1956) im mittleren Ostbothnien gemachte Beobachtung an, nach welcher ein mehr saures plutonisches Gestein ein mehr basisches durchsetzt. Und dies soll auf die Existenz von nur einer Differentiationsserie hinweisen! Die von Salli gemachte Beobachtung ist im Zusammenhang mit der vorliegenden Frage vollkommen nichtssagend. Die sauren Glieder der älteren und der jüngeren Differentiationsserie können die basischeren Glieder ihrer eigenen Serie durchsetzen. Die Granite der jüngeren Differentiationsserie können aber sowohl die basischeren Glieder dieser Serie, als auch alle Gesteine der älteren Differentiationsserie durchsetzen. Entscheidend sind nur diejenigen Beobachtungen, nach welchen ein mehr basischer Plutonit einen saureren durchsetzt (vgl. S. 272). Das durchsetzte Tiefengestein gehört in diesem Falle begreiflicherweise zu der älteren Serie, und in Ostbothnien ist dieses Gestein dazu seiner Erscheinungsweise nach in der Regel synkinematisch. Übrigens weise ich darauf, was schon früher gesagt worden ist.

Nach Simonen (1960b) sollen die synkinematischen Plutonite in Ostbothnien innerhalb, die spätorogenen Plutonite wieder ausserhalb der einheitlichen Schieferzone vorkommen, und demnach deutete die tektonische Einteilung des Verfassers nur auf die verschiedenartige Erscheinungsweise der Plutonite innerhalb und ausserhalb der Schieferzone hin. Diese Behauptung Simonens ist völlig aus der Luft gegriffen, wie schon aus der Karte (Taf. I) hervorgeht. Noch rätselhafter ist die Behauptung Simonens, derzufolge es eine festgestellte Tatsache sein soll, dass die Erscheinungsweise der plutonischen Massive verschiedenartig in den verschiedenen Teilen eines und desselben Plutons sein kann. Auf die Untersuchungen von Salli (1956) aufs neue hinweisend, berichtet Simonen, dass die synkinematische Phase des Plutonismus von Gabbros, Quarzdioriten und Granodioriten, die spätkinematische wieder von Graniten vertreten wäre. Ich habe mich sorgfältig, und unter Mitwirkung des betreffenden Geologen selbst, mit den Untersuchungen Sallis vertraut gemacht. Es zeigte sich dabei, dass die magmatektonischen Fragen in den erwähnten Untersuchungen ganz im Hintergrund geblieben waren, und dass die Einteilung der Plutonite n u r a u f G r u n d i h r e r p e t r o g r a p h i s c h e n M e r k m a l e ausgeführt worden war. Es wäre m. E. bedauerlich, wenn man in den künftigen Kartierungsarbeiten den petrographischen Tiefengesteinsgruppen ohne weiteres magmatektonische Namen beilegte.

DIE GRANITEINTEILUNG VON W. WAHL

Nach Wahl (1936a, b) können wir in jeder Gebirgskette, wo die Denudation so weit fortgeschritten ist, dass tiefliegende Schnitte entblösst worden sind, rein empirisch und unabhängig davon, welchen Standpunkt wir der Entstehung der Granite gegenüber einnehmen, das Vorkommen von zwei Arten von in ihrem Auftreten verschiedenen granitischen Gesteinen konstatieren. Einerseits haben wir, wie Wahl behauptet, meistens granodioritische Gesteine, die jetzt in Gestalt von »Gneisgraniten» vorliegen und die während eines relativ frühen Stadiums der Orogenese intrudiert worden sind. Sie bilden gewöhnlich konforme Massive, und es schliessen sich an sie in geringerem Masse sowohl basischere (dioritische, gabbroidische) als alkalireichere, granitische Gesteine an. Anderseits haben wir Mikroklingranite, die eine ausgeprägt pegmatitische Beschaffenheit besitzen und intim mit »Migmatitbildung» verbunden sind. Diese letzteren, die Wahl serorogen nennt, durchbrechen die Granodiorite mit Gneisgranithabitus (die primorogenen Gesteine) und sind daher jünger als diese. Beide Gruppen können nach Wahl als orogen betrachtet werden. (Wahl unterscheidet ausserdem eine dritte Granitgruppe, die postorogenen Granite, die erst nach der eigentlichen Bergkettenfaltung emporgedrungen sind). Wahl bemerkt zum Schluss, dass die von Sederholm vorgelegten Granitgruppen in der Tat seinen drei Granittypen entsprechen, und dass diese sich in Zusammenhang mit jeder Gebirgsfaltung wiederholen.

Wahls Einteilung der Tiefengesteine ist insofern geeignet, Staunen zu erregen, als man darin den Granitgruppen Namen, die auf die orogene Entwicklung hinweisen, beigelegt hat, obschon die Einteilung in Wirklichkeit auf Grund petrographischer Merkmale geschehen ist. Aus Wahls Publikationen geht deutlich hervor, dass die magmatektonischen Fragen nur in sehr geringem Masse Aufmerksamkeit auf sich gezogen haben. So spricht auch Wahl selbst den Wunsch aus, dass die Verhältnisse der Granite zu den orogenen Vorgängen in dem künftigen Untersuchungen festgestellt würden (Wahl, 1936a)! Wir können also sagen, dass Wahl, in grossen Zügen betrachtet, nur den Gruppen, die die verschiedenaltrigen Granittypen Sederholms vetreten, neue, tektonische Namen beigelegt hat. Dieses Verfahren hat keine glückliche Folgen gehabt. Viele Geologen sind nämlich zu der irrtümlichen Auffassung gekommen, dass Wahls Graniteinteilung tatsächlich und in ersten Linie eine tektonische sei. Diese Einteilung ist dann als solche angewendet worden, ohne dass man auch jetzt grössere Beachtung den magmatektonischen Fragen gewidmet hätte. So berichtet Eskola (1954), dass diese tektonische Einteilung in den Kartierungsarbeiten der geologischen Forschungsanstalt angewendet worden ist, und zwar in der Weise, dass auf den neuen Kartenblättern (Massstab 1: 100 000) die synkinematischen Tiefengesteine mit den Gesteinsnamen Granodiorit, Quarzdiorit usw. und die spätkinematischen Gesteine mit Granit bezeichnet werden. Bei diesem Verfahren gelangt man, wie Eskola betont, gleichzeitig zu einer tektonischen Einteilung. Simonen (1960b) erwähnt, dass die neuen Kartierungsarbeiten in S-Finnland die hauptsächlichen Ideen von Wahl bestätigt haben! Er bemerkt auch, dass die von Wahl vorgelegten Granitgruppen jetzt synkinematisch und spätkinematisch genannt werden.

Ich möchte auch in diesem Zusammenhang bemerken, dass die petrographische Einteilung der Tiefengesteine nicht ohne weiteres zu einer tektonischen Einteilung führt. Die petrographischen Gruppen werden nicht dadurch in tektonische Gruppen umgewandelt, dass man ihnen tektonische Namen beilegt. So einfach ist die geologische Forschung doch nicht!

DIE PETROGRAPHISCHEN PROVINZEN VON A. SIMONEN

Simonen (1960b,c) hat die im »svekofennischen» Raum vorkommenden Plutonite in selbständige Provinzen eingeteilt, die ihre eigenen charakteristischen chemischen und mineralogischen Züge besitzen. Die Provinzen sind nach ihrem sauersten Glied folgendermassen benannt worden: die Granodioritprovinz, die Trondhjemitprovinz, die Charnockitprovinz, die Granitprovinz und die Mikroklingranite. Diese petrographische Einteilung soll nach Simonen gleichzeitig eine tektonische Einteilung sein. Diejenigen Provinzen, an die sich keine granitische Endglieder anschliessen, sind im Ganzen synkinematisch, die Mikroklingranite wieder spätkinematisch. Die granitischen Glieder der Granitprovinz sind spätkinematisch, alle anderen synkinematisch. Aber noch obendrein ist Simonens Einteilung im gewissen Sinne auch genetisch. Die Mikroklingranite sollen nämlich zum überwiegenden Teil Produkte einer granitisierenden Metasomatose, alle anderen Provinzen meistens magmatisch sein! Es sieht wirklich so aus, als ob man hier nicht nur zwei, sondern sogar drei Fliegen mit einer Klappe geschlagen hätte. Ich habe mich darum veranlasst gesehen, einige Gedanken anlässlich der Provinzeinteilung von Simonen vorzulegen.

DIE GRANODIORITPROVINZ

An die Granodiorite schliessen sich nach Simonen basischere Gesteine (Gabbros, Diorite), aber dagegen keine granitischen Gesteine an. Es kommen allerdings im Gebiete der Granodioritprovinz Mikroklingranite vor, diese aber gehören nach Simonen nicht zu der betreffenden petrographischen Provinz, weil zwischen der mineralogischen Zusammensetzung der Granodiorite und Mikroklingranite eine »deutliche Lücke» existieren soll. Anderseits teilt Simonen mit, dass zwischen den genannten Gesteinen vermittelnde mikroklinhaltige Abarten vorkommen. Diese wären jedoch Produkte einer später vorsichgegangenen Metasomatose, d. h. einer Granitisierung, die zum Schluss zur Bildung von normalen Mikroklingranit führen soll.

Eine typische Granodioritprovinz ist nach Simonen von dem Orijärvigebiet vertreten (Eskola, 1914). Aber gerade in diesem Gebiet schliessen ja sich, wie schon früher beschrieben worden ist (S. 271), an die synkinematischen Granodiorite lückenlos synkinematische Mikroklingranite und Granitpegmatite an. Keinerlei Zeichen von »Granitisierung» sind beobachtet worden. Das gänzliche Fehlen von granitischen Gliedern wäre ja überhaupt sehr unwahrscheinlich, weil es sich um eine kräftig differentierte magmatische Gesteinsserie handelt. Schon Wahl (1936) bemerkte, dass sich an seine »primorogenen» Granodiorite alkalireichere Granite anschliessen.

Was die von Simonen erwähnte Lücke zwischen den Mineralzusammensetzungen der Granodiorite und Mikroklingranite betrifft, beruht diese offenbar darauf, dass die genannten Gesteine zu ganz verschiedenen Eruptionsserien gehören. Eine scheinbare Lücke kann natürlich davon bedingt sein, dass die verschiedenen Glieder der synkinematischen Gruppe in den tektonischen Bewegungen von einander getrennt wurden. D i e v o n S i m o n e n vorgelegte »Granodioritprovinz» kommt gar nicht vor. Es handelt sich um eine vollständige, normale magmatische Differentiationsserie.

DIE TRONDHJEMITPROVINZ

Die Glieder dieser Provinz bilden nach Simonen eine kontinuierliche Serie von Gabbros bis saure Trondhjemite. Die eigentlichen granitischen Glieder fehlen. So einfach ist die Sachlage jedoch nicht, denn sowohl im Kalanti — als im Turkugebiet kommen, wie Hietanen (1943, 1947) bemerkt, allerlei Übergänge von Trondhjemiten bis Granite vor. Nach Hietanen »the addition of microcline may be primary or secondary». Nach Simonen ist es in vielen Fällen schwer, genaueres über den primären Mikroklingehalt zu sagen. Verfasser hat seinerseits nirgends Beschreibungen über unwiderlegbare metasomatische Umwandlungen von Trondhjemiten in Granite beobachtet.

In diessem Zusammenhang sind die im Kalantigebiet reichlich vorkommenden migmatitbildenden turmalinführenden Granitpegmatite nicht zu vergessen. Stellenweise schliessen sich an diese Gesteine normale Granite an. Die Granitpegmatite sind ihrem Charakter nach deutlich synkinematisch und sie sind während derselben Phase der Orogenese, wie auch die oben erwähnten Glieder der Trondhjemitserie, emporgedrungen, und vertreten zusammen mit den Trondhjemitpegmatiten die jüngsten Glieder der Eruptionsserie.

Zu der Trondhjemitprovinz Simonens gehören also granitische und granitpegmatitische Glieder, wie auch zu der Granodioritprovinz. Dazu sind die gabbroidischen bis dioritischen Glieder in beiden Provinzen gleich (Hietanen, 1943). »Die Selbständigkeit» der Trondhjemitprovinz ist zweifelhaft. Das Vorkommen der trondhjemitischen Abarten ist nur durch eine lokale Anreicherung des Wassers in den empordringenden Magmen bedingt.

DIE CHARNOCKITPROVINZ

Die Charnockite schliessen sich nach Simonen eng an die Trondhjemite an. Nach Hietanen (1947) gehören zu ihnen auch granitische Gesteine. Es dürfte kaum zweckmässig sein, von einer selbständigen Charnockitprovinz zu sprechen.

DIE GRANITPROVINZ

Die Granitprovinz soll im Tamperegebiet, im Mittelfinnland und im mittleren Ostbothnien vorkommen. Zu ihr gehören Tiefengesteine von Peridotiten und Gabbros bis Granite und Granitpegmatite. Die Vorlegung einer derartigen »Granitprovinz» dürfte an sich keinen Meinungsaustausch verursachen, wenn Simonen nicht alle in den erwähnten Gebieten vorkommenden Plutonite zu einer und derselben Differentiationsserie gerechnet hätte. Wie ich schon früher gezeigt habe, kommen hier zwei postbothnische Eruptionsserien vor. Dazu treten noch postsvionische Tiefengesteine auf. Noch mehr befremdend ist, dass Simonen die Plutonite in zwei tektonische Gruppen einteilt, und zwar so, dass zu der spätkinematischen Gruppe alle Granite, und zu der synkinematischen alle übrigen Plutonite gerechnet werden. Es geht schon aus dem früher Erwähnten hervor, dass ein derartiges Verfahren vollkommen verfehlt ist. Simonen teilt mit, dass die tektonische Einteilung im Tamperegebiet und in Mittelfinnland nicht leicht durchzuführen ist. Ich möchte daher bemerken, dass die tektonische Einteilung auch in den genannten Gebieten leicht durchzuführen ist, wenn ernsthaftere Aufmerksamkeit auf die magmatektonischen Umstände gerichtet wird. Die unüberwindlichen Schwierigkeiten fangen erst dann an, wenn man eine tektonische Einteilung nur auf Grund petrographischer Eigenschaften und Zusammensetzung der Tiefengesteine zustandezubringen versucht.

DIE MIKROKLINGRANITE

Dieser Provinz kommt in Südfinnland vor. Nach Simonen sind die Mikroklingranite von folgenden speziellen Zügen charakterisiert:

1. An die Mikroklingranite schliessen sich keine basischere Glieder an.
2. Die chemische Zusammensetzung der Mikroklingranite entspricht nicht der Zusammensetzung der aus den Restmagmen kristallisierten eutektoidischen Granite, sondern sie sind kalireicher als diese.
3. Die Mikroklingranite sind zum überwiegenden Teil Produkte einer metasomatischen Granitisierung. Die besten Beweise in dieser Beziehung sollen die Reliktstrukturen (»die alten Zeichnungen»; Wegmann und Kranck, 1931; Wegmann, 1935) darstellen, die darauf hindeuteten, dass an der Stelle der jetzigen Mikroklingranite früher ein anderes Gestein gewesen ist.
4. Die durch Granitisierung gebildeten Mikroklingranite sollen kräftige Faltungserscheinungen in ihrer Umgebung verursacht haben, die auf diapiritischer Aufwölbung derselben beruht. Jedoch soll die Granitisierung ruhig den Schiefrigkeitsflächen der prägranitischen Kruste entlang stattgefunden haben.
5. Die Mikroklingranite bilden oft konkordante Massive.
6. An die Mikroklingranite schliessen sich allgemein migmatitische Gesteine an.

Wass das Moment 1 betrifft, ist die Angabe Simonens nicht stichhaltig. Es geht z. B. aus den Untersuchungen von Hietanen (1947) hervor, dass sich an die deutlich spätkinematischen Mikroklingranite in der Regel basischere

Tiefengesteine (Granodiorite, Diorite, Gabbros) anschliessen. Simonen (1956) selbst führt an, dass die massigen, hypidiomorphen Quarz- und Granodioriten in der Gegend von Kärkölä allmählich in porphyrische Mikroklingranite übergehen. Die Annahme Simonens, dass es sich hier um Granitisierung handelte, ist aus der Luft gegriffen. Keinerlei Beweise für derartige metasomatische Umwandlungen sind auch vorgelegt worden. — Die Mikroklingranite vertreten deutlich die sauersten Endglieder einer gut entwickelten Differentiationsserie.

Auch die Behauptung im Moment 2 ist nicht stichhaltig. Wie aus Simonens (1960b) eigenen Diagrammen hervorgeht, liegen die Punkte, die die normative Zusammensetzung (Q:Or:Ab) der Mikroklingranite, der Granite der Granitprovinz und sogar der »granitisierten» Granodiorite *etc.* vertreten, von zwei Ausnahmen abgesehen, überraschend genau innerhalb des Gebietes der eutektoidischen Granite. In der Abbildung 7 sind folgende von den von Simonen vorgelegten Tiefengesteinen repräsentiert.

Die Granitprovinz des Tamperegebietes: Granit, Lörpys, Kuru; Granit, Peurnajärvi; Granit, Vuorenmaa, Kankaanpää; Granit, Muurainen, Kuru; Granit, Paarlahti, Teisko.

Die Granitprovinz des mittleren Ostbothniens: Granit, Kalajoki; Granit, Raahe.

Die Mikroklingranite von folgenden Orten: Pajback, Parainen (2 Stücke); Räntämäki; Alajärvi; Kakola, Turku (2 Stücke); Sillanpää, Kisko; Mattnäs, Nauvo; Metsämaa; Skarfkyrkan, Hanko; Manngård, Karjaa; Hanko; Lauttasaari, Helsinki; Kumlinge, Inkoo; Pernaja; Lahdenperä, Kisko. Der »Mikroklingranit» von Lövböle, Kemiö (Pehrman, 1945) ist unberücksichtigt geblieben. Er ist nämlich eine schriftgranitische Verwachsung von Mikroklinperthit (79.4 %) und Quarz (20.6 %).

»Granitisierte» plutonische Gesteine: granitisierter Trondhjemit, Orimattila; granitisierter Oligoklasgranit, Virkkala; granitisierter Granodiorit, Suomusjärvi; granitisierter Granodiorit, Aulanko; granitisierter Quarzdiorit, Hanko; granitisierter Trondhjemit, Järvenpää; granitisierter Trondhjemit, Kerava.

Es gibt keinen Anlass zur Annahme, dass die Gesteine im Diagramm (Abb. 7) in irgendeiner anderen Weise als durch Kristallisation aus den Magmen gebildet wären. Die »alten Zeichnungen» deuten auch nicht auf eine metasomatische Entstehung der Granite, sondern im Gegenteil auf eine magmatische Entstehung von diesen hin. Die »alten Zeichnungen» vertreten in den Granitmassen vorkommende, kräftig metasomatisch umgewandelte Partien von älteren schiefrigen Gesteinen. Die metasomatischen Umwandlungen sind durch die leichtflüchtigen Bestandteile, die aus den umgebenden Magmen emaniert worden sind, verursacht worden. Die Kontakteinwirkung der Granite kommt überzeugend zum Vorschein in den ausgedehnten Migmatitgebieten, wo jede von den zahlreichen granitischen Adern bald kräftigere, bald schwächere metasomatische Umwandlungen in den an-

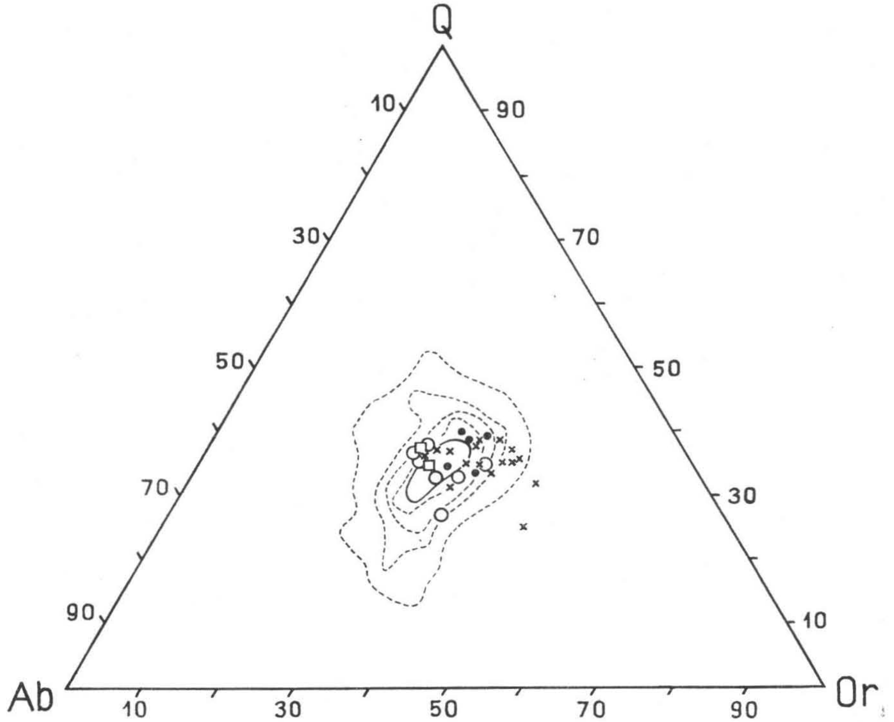

Abb. 7. Die normativen Zusammensetzungen (Q : Or : Ab) der Mikroklingranite Südfinnlands (Kreuze), der Granite der Granitprovinz des mittleren Ostbothniens (Quadrate), der Granite der Granitprovinz des Tamperegebietes (Punkte) und der »granitisierten» Granodiorite *etc.* (Kreise). Nach den Diagrammen von Simonen (1960b) zusammengestellt. Aus dem Diagramm geht auch die Verteilung der entsprechenden normativen Zusammensetzungen der 571 analysierten eutektoidischen Granite und Syenite hervor (Adams, 1952; Bowen, 1954). Die Gebiete der Prozentzahlen >1, >2, >3, >4 und >5—7 sind bezeichnet.

grenzenden Schieferstreifen verursacht hat. Begreiflicherweise kann eine durchgreifende Metasomatose leicht auch die Strukturen der Schieferpartien zerstören. Es ist jedoch festgestellt worden, dass die Schieferpartien ausdrücklich in den synkinematischen Graniten eine bestimmte Orientierung besitzen, dessenungeachtet, dass die metasomatischen Umwandlungen so kräftig gewesen sind, dass von den Schieferpartien nur schattenhafte Reste übrig sind. D i e s e l b e O r i e n t i e r u n g ist in dem umgebenden Granit, sowie in den unumgewandelten Schieferpartien zu beobachten. Gerade diese Umstände führten Wegmann zu der Annahme, dass die ursprüngliche Lage der Schieferstrukturen beibehalten ist. Die Entstehung der obengenannten Parallelorientierung ist m. E. nur in dem Fall möglich, dass der Granit, die metasomatisch unumgewandelten älteren Gesteine und die »gra-

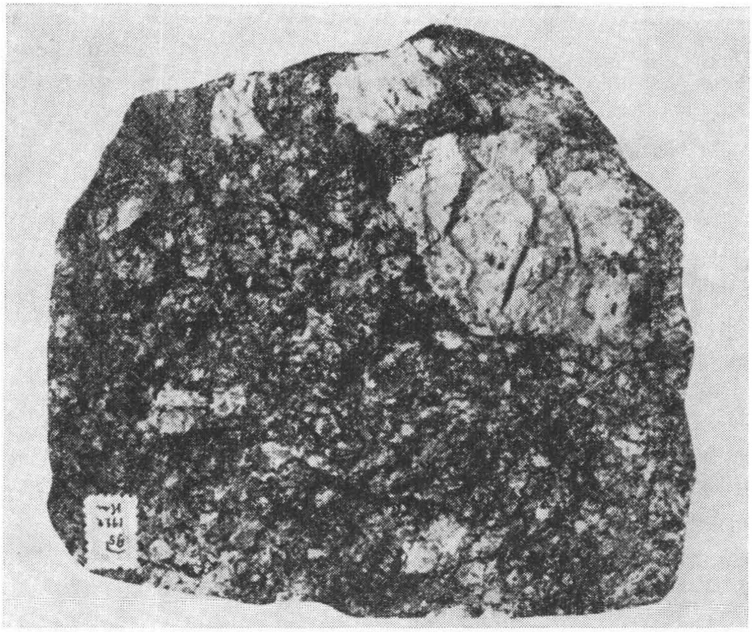

Abb. 8. Der sog. Vaasagranit. Särkimoskär, Maksamaa (Maksmo). 5/7 der nat. Grösse. Saksela (1935).

nitisierten» Teile der letztgenannten ungefähr gleichzeitig, d. h. während der Hauptphase (Überschiebungsphase) der Orogenese und dank denjenigen Gleitbewegungen, die diese Phase charakterisieren, ihre jetzige metamorphe Textur erhalten haben. Es kann sich also um keine a l t e Zeichnungen handeln.

Ich möchte speziell den Umstand betonen, dass der homogene Granit, in welchem sich die genannten umgewandelten Schieferpartien befinden, keinesfalls ein Endresultat der metasomatischen Prozesse vertritt, sondern dass er diese Prozesse verursacht hat. Meiner Meinung nach wäre der Begriff Granitisierung möglichst bald zu streichen, und zwar ganz einfach deshalb, weil die Metasomatose aller Wahrscheinlichkeit nach niemals zur Bildung eines solchen Gesteins geführt hat, das sowohl seiner Zusammensetzung als seiner Struktur nach einem normalen Granit ganz ähnlich wäre. Ausnahmsweise können Gesteine gebildet werden, die der chemischen Zusammensetzung nach den Graniten nahekommen, ihrer Struktur nach aber weit von diesen stehen. Als Beispiel dafür führe ich den sog. Vaasagranit (Abb. 8) an (Saksela, 1935, 1953; Helovuori, 1949).

Was das Moment 4 betrifft, bilden die spätkinematischen Granite stellenweise diapirartige Intrusionen (Wegmann, 1930), die tektonische Störungen in ihrer Umgebung verursacht haben. Aber wie haben die Granitmassen kräftige Faltungserscheinungen verursachen können, wenn sie Produkte metasomatischer Prozesse darstellten, und die Metasomatose dazu ruhig vorsichgegangen wäre? Vielleicht stützt sich Simonen auf die irrtümliche Auffassung (Wegmann und Kranck, 1931), dass das Volumen der Gesteinsmasse durch die Metasomatose vergrössert würde (vgl. Simonen, 1960c).

Die Erwähnung Simonens, dass die Mikroklingranite oft konkordante Intrusionen bilden, weist deutlich darauf hin, dass zu seinen Mikroklingraniten auch synkinematische Vertreter gehören. Auf dieselbe Richtung deuten »die alten Zeichnungen» hin.

Schon früher wurde gezeigt (Saksela, 1953), dass die eigentliche Adergneisbildung ein Ereignis ist, das ziemlich früh, vor dem Empordringen der jüngeren Granite stattgefunden hat. Die Annahme Simonens, sowie auch Wahls (1936a), nach welcher die jüngeren Granite mit kräftigerer Migmatitbildung in Zusammenhang ständen, hat bezüglich adergneisartiger Migmatite nicht das Richtige getroffen.

Nach allem Gesagten wäre es bedauerlich, wenn man die von Simonen vorgelegte Provinzeinteilung in den künftigen Kartierungsarbeiten auf die eine oder andere Art anwendete.

»SVEKOFENNIDEN UND KARELIDEN». KARELOBOTHNIDEN

Schon lange sind die finnischen Geologen mit der Frage beschäftigt gewesen, wo die Grenze zwischen den Kareliden und Bothniden (oder »Svekofenniden») verläuft. Die Grenze ist im Laufe der Jahre bald hier, bald dort gezogen worden, ohne dass man irgendwo Zeichen von einer grösseren Diskordanz zwischen diesen Formationen beobachtet hätte. Kürzlich teilte Simonen (1960c) mit, dass »no valide evidence for a marked unconformity between the Svecofennidic and Karelidic belts has been found in Finland». In der letzten Zeit ist auch der Gedanke vorgelegt worden, dass die karelischen und »svekofennischen» Zonen nur verschiedene Phasen und Sedimentassoziationen desselben Zyklus der Sedimentation verträten. Die »jatulischen» Sedimente der Kareliden repräsentieren kontinentale bis epikontinentale Bildungen, während die »kalevischen» Schiefer und die sedimentogenen Gesteine im »svekofennischen» Raum Bildungen geosynklinaler Art sind (Toivo Mikkola, 1953; Metzger, 1959; Simonen, 1960c).

Nur Eskola scheint immer noch an dem verschiedenen Alter der Kareliden und »Svekofenniden» festzuhalten. Er stützt seine Ansicht u. a. damit, dass die Streichrichtungen der erwähnten Formationen im finnischen Karelien von einander beinahe in senkrechter Richtung durchschnitten wären

(Eskola, 1959). Dies stimmt aber gar nicht, wie schon aus den Übersichtskarten hervorgeht. Zuletzt hat Paarma (1959) auf diese irrtümliche Auffassung Eskolas hingedeutet. Ferner bemerkt Eskola (1960), dass die Unterlage der Svekofenniden nirgends beobachtet worden ist, und dass in ihnen keine ultrabasische ophiolitische Massen enthalten sind. Auch diese Behauptungen sind nicht stichhaltig. Der Umstand wieder, dass Quarzite in geringerem Masse im »svekofennischen» als im karelischen Raum vorkommen, deutet keineswegs, wie Eskola zu glauben scheint, auf das verschiedene Alter der genannten Formationen hin. Die Sachlage wird ja schon aus all dem klar, was oben über das Vorkommen der verschiedenen Sedimentassoziationen gesagt worden ist. Zum Schluss sind nach Eskola (1960) Verschiedenheiten in der Bewegungstektonik der karelischen und »svekofennischen» Zonen zu beobachten. In den erstgenannten wären die Richtungen mehr einheitlich, und in den »svekofennischen» Formationen sollen im allgemeinen keine Überschiebungen vorkommen. Die Unhaltbarkeit dieser Behauptungen dürfte aus dem früher Gesagten hervorgehen.

Die absoluten Altersbestimmungen, die bei uns hauptsächlich von Kouvo (1958a, b, 1960) ausgeführt worden sind, haben in Bezug auf die gegenseitigen Altersverhältnisse der »Svekofenniden» und Kareliden interessante Ergebnisse gezeitigt. Nach ihnen sind nämlich die karelischen und »svekofennischen» Granite gleichaltrig, ungefähr 1800×10^6 Jahre alt. Dies stimmt natürlich sehr gut mit der Auffassung von der Gleichaltrigkeit der karelischen und »svekofennischen» Formationen überein. Bei den von absoluten Altersbestimmungen gegebenen Resultaten ist aber bis auf weiteres Vorsicht am Platze. Erstens sind die untersuchten Proben allzu gering an Zahl, und dazu scheint es mir, als wären die Proben in gewissem Masse planlos genommen worden. Zweitens müssen die Ungenauigkeiten, die sich an die Bestimmungen anschliessen, berücksichtigt werden. Nach Kulp (1960) kann man heutzutage die chemischen Analysen, sowie die Bestimmungen der Isotopverhältnisse mit einer Genauigkeit von 1—5 % ausführen (vgl. auch Faul, 1960). Es ergibt sich aber noch die Frage, in welchem Masse die untersuchte Mineralprobe als ein geschlossenes chemisches System verblieben ist. Diese Frage erschöpfend zu beantworten ist jedoch schwer. Nach Kouvo (1960) sind die Alter, die man mit verschiedenen Methoden für ein Mineral erhalten hat, als konkordant zu betrachten, wenn sie im Rahmen von 10 % mit einander übereinstimmen. Dies bedeutet offenbar, dass die Genauigkeit der Altersbestimmungen unter günstigen Verhältnissen von dieser Grössenordnung ist. Wenn es sich also um so hohe Alter wie z. B. 1800×10^6 Jahre handelt, können die Alter, die 180×10^6 Jahre von einander abweichen, als konkordant betrachtet werden. Nach Edelman (1960) sind die radioaktiven Methoden demnach allzu »grob», um mit deren Hilfe entscheiden zu können, ob die »Svekofenniden» und Kareliden von gleichem oder verschiedenem Alter sind. Edelman bemerkt, dass in der postkambrischen Zeit eine orogene und eine anorogene Periode zusammen durchschnittlich 140—150 $\times$ 10^6 Jahre dauerten. Nach Kautsky (1959) wäre es nicht möglich, mittels absoluter Altersbestimmungen die in Västerbotten und Norrbotten auftretenden Granite altersmässig zu trennen. Er verficht seine Ansicht u. a. damit, dass man in Finnland für die »svekofennischen» und karelischen Granite dasselbe Alter erhalten hat, obschon diese geologisch altersmässig deutlich getrennt sind!

Nach Eskola (1960) ist die Gleichaltrigkeit der karelischen und »svekofennischen» Granite schon als so sicher anzusehen, dass es gewagt wäre, sie zu bestreiten. Dies steht ja aber im völligen Widerspruch mit seiner Auffassung, nach welcher die »svekofennischen» Granite in Zusammenhang mit der svekofennischen Gebirgsbildung und vor der Karelidenzeit emporgedrungen wären. Um sich aus der Klemme zu ziehen, hat Eskola (1958, 1959, 1960) eine seltsame Hypothese von der »Verjüngerung» der svekofennischen Granite vorgelegt. Toivo Mikkola (1959a, 1959b) und Marmo (1960) haben die Bedeutungslosigkeit dieser Hypothese schon nachgewiesen.

Meiner Meinung nach können wir zumindest sagen, dass die von den radioaktiven Altersbestimmungen gegebenen Resultate mit den geologischen Befunden n i c h t im Widerspruch stehen. Die karelischen und »svekofennischen» Formationsserien fliessen unwiderstehlich zusammen. Es ist m. E. hohe Zeit, der so erhaltenen ausgedehnten Formation einen neuen Namen zu geben. Weil es um »Zusammenfliessen» der ehemaligen karelischen und bothnischen Formationen handelt, schlage ich den Namen K a r e l o b o t h n i u m (Karelobothniden) vor. Der Name »Svekofenniden» ist m. E. zu streichen.

In diesem Zusammenhang mag festgestellt werden, dass in Finnland Sedimentformationen vorkommen, die man mit den Molassebildungen vergleichen könnte, und die in der Endphase des karelobothnischen orogenen Zyklus sedimentiert worden sind. Zu diesen gehören die Sandsteine von Pori, die Sedimentformation von Muhos und die Kumpuquarzitformation in Lappland (vgl. Hackman, 1927; Erkki Mikkola, 1941; Väyrynen, 1954; Toivo Mikkola, 1960; Simonen, 1960c; Brenner, 1941).

DIE SVIONISCHE UNTERLAGE DER KARELOBOTHNIDEN

Es dürfte klar sein, dass der grösste Teil dess sog. Gneisgranitgebiets von Ostfinnland die ältesten Bildungen Finnlands vertritt. Altersmässig sind sie zu dem Svionium zu rechnen, und sie bilden die Unterlage der »karelischen» Schieferformation. Nach den radioaktiven Altersbestimmungen sind sie ungefähr 2500—2600 $\times$ 10^6 Jahre alt (Kouvo, 1958). Das Gneisgranitgebiet

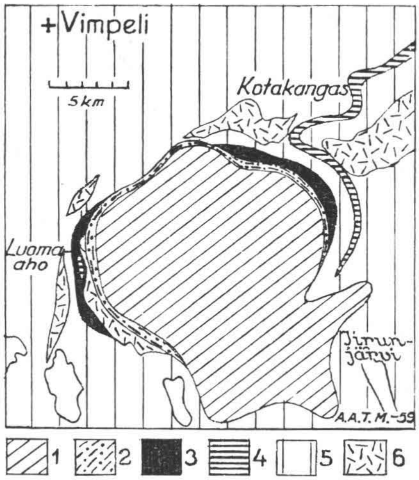

Abb. 9. Die Gneisgranitkuppel von Luoma-aho—Iirunjärvi. 1. Gneisgranit; 2. Metabasit; 3. Quarzit; 4. Kalkstein; 5. Pelitschiefer; 6. Pegmatit. Nach Metzger (1959).

ist sehr heterogen. Neben den Gneisgraniten kommen auch massige »jüngere Granite», sowie auch von diesen durchdrungene suprakrustale Bildungen vor (vgl. Frosterus und Wilkman, 1920; Wilkman, 1921; Matisto, 1958). Es kommen aber in gewissen Masse auch Tiefengesteine vor, die deutlich jünger als die »karelischen» Schiefer sind. So hat man am Westrande des Gneisgranitgebiets Tiefengesteine angetroffen, die die »jatulischen» Quarzite durchschneiden.

Wenn man von dem Gneisgranitgebiet nach SW wandert, kann man feststellen, dass d i e s e l b e alte Gneisgranitunterlage an vielen Stellen unter den karelobothnischen Sedimentformationen hervorsticht. Die Gneisgranitkuppeln von Kontiolahti, Sotkuma und Liperinsalo sind allgemein bekannt (vgl. Frosterus und Wilkman, 1920; Saksela, 1933b). In der Gegend von Kuopio tritt der svionische Gneisgranit in mehreren Fenstern zu Tage (Wilkman, 1923; Preston, 1954). Zwischen Vimpeli und Alajärvi im mittleren Ostborhnien haben wir, wie Metzger (1959) neulich festgestellt hat, eine fensterartige Aufbuckelung des svionischen Gneisgranits (Abb. 9). Hier beginnt die Sedimenthülle, wie auch im Kuopiogebiet, mit Quarziten, die teilweise von Kalksteinen begleitet sind. Über den Quarziten folgen pelitische Gneise. Wie in Kuopio, sind in die Kontaktzone zwischen dem Gneisgranitfenster und den suprakrustalen Bildungen sowohl basische als saure Gesteine (Metabasite, Pegmatite) eingedrungen. Es ist sehr interessant festzustellen, dass die beschriebene Gneisgranitaufbuckelung genau an der von

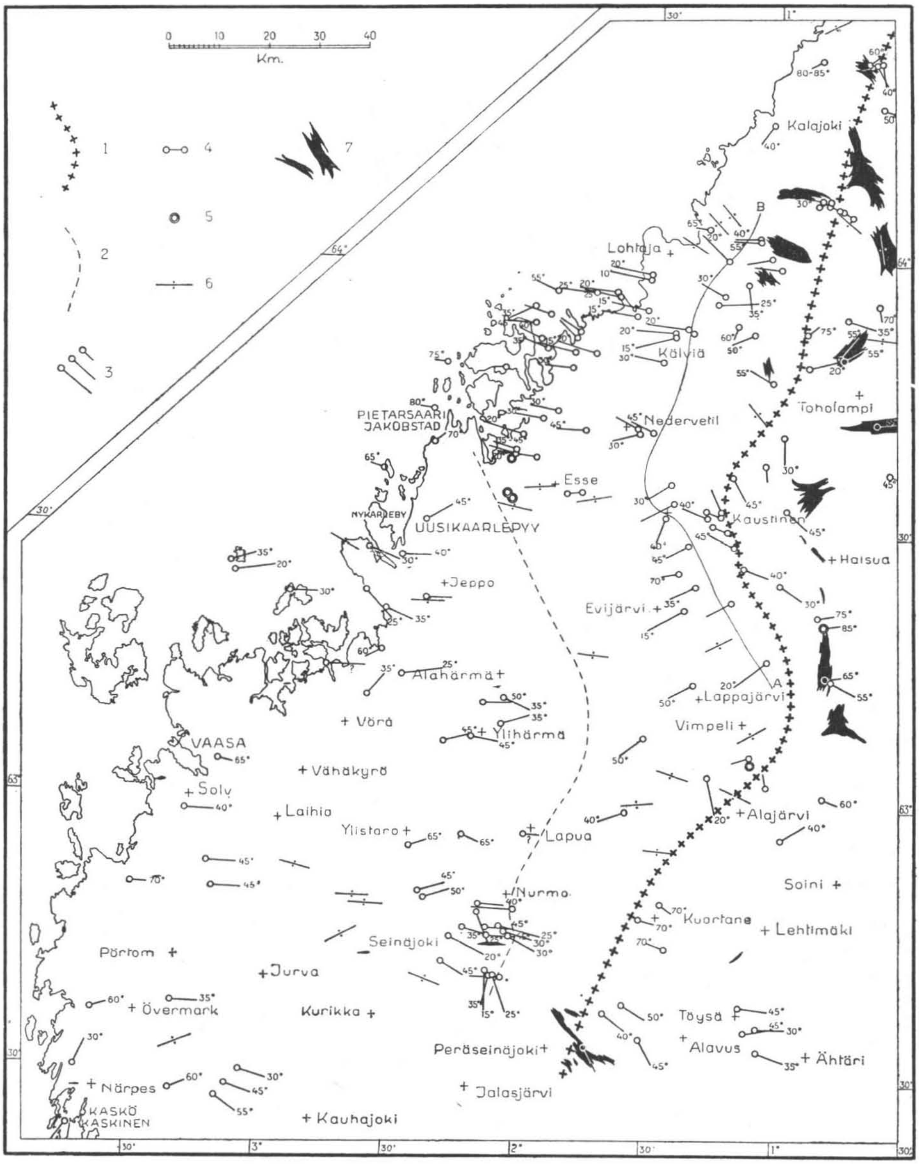

Abb. 10. Die durch die Streckungsrichtungen und Kleinfältelung bestimmten lokalen Achsialrichtungen in Ostbothnien. Aus der Karte geht auch die Verbreitung der Leptite hervor. 1. Die Kulminationszone; 2. Die Depressionszone; 3—5. Die lokalen Achsialrichtungen; 3. Das Achsialfallen 1°—29°, 30°—59° und 60°—89°; 4. Das Achsialfallen 0°; 5. Das Achsialfallen 90°; 6. Das Einfallen 90°; 7. Die Leptitgebiete. A—B ist die sog. Grundlinie. Nach Saksela (1935).

mir vorgelegten ostbothnischen Kulminationszone gelegen ist (Abb. 10), wo gerade das Vorkommen der älteren Bildungen zu erwarten ist (Saksela, 1935).

In diesem Zusammenhang dürfen die von Wilkman (1931) beschriebenen Konglomerate in Ostbothnien nicht vergessen werden. Besonders interessant ist das Konglomerat von Olkkosenmäki, Kirchspiel Haapajärvi, dessen Gerölle zum bedeutenden Teil aus Tiefengesteinen (Quarzdiorite, aplitische Granite) bestehen.

Das Gneisgranitfenster von Alajärvi—Vimpeli, sowie die Konglomerate von Haapajärvi deuten darauf hin, dass die Unterlage der Karelobothniden auch anderswo in Ostbothnien entblösst sein kann. Wie aus meiner früheren Beschreibung hervorgegangen ist, tritt dieselbe svionische Unterlage auch im Tamperegebiet und in Südfinnland zu Tage. Neben dem, was ich schon früher über die svionischen Bildungen in den letztgenannten Gebieten erwähnt habe, mag noch folgendes angeführt werden.

Härme und Seitsaari (1950; Seitsaari, 1951) haben das in der Schieferzone von Tampere vorkommende runde, etwa 8 km im Diameter messende Granodioritmassiv von Värmälä (Abb. 11) untersucht, und sie sind auf Grund der Streckungs- und Schieferungsbeobachtungen zu der Auffassung gekommen, dass es sich um einen geneigten Dom handelt, dessen Empordringen Störungen in der Tektonik der Umgebung verursacht hat. Das Granodioritmassiv hat in den tektonischen Bewegungen sehr steif reagiert, worauf die in den Randteilen des Massivs vorkommenden zahlreichen Breccien- und Bruchzonen hindeuten. Es ist auch interessant festzustellen, dass in den Kontaktzonen des Massivs keine Mischgesteine gebildet worden sind. Der Granodiorit sendet keine Adern in die umgebende Schiefer, und er enthält keine Schieferfragmente. Statt dessen sind in die Kontaktzone, die eine deutliche Schwächezone repräsentiert, gabbroidische, dioritische und granitische Plutonite, längliche Massive bildend, eingedrungen. D a s G r a n o d i o r i t m a s s i v v o n V ä r m ä l ä i s t e i n T e i l d e r U n t e r l a g e d e r S c h i e f e r f o r m a t i o n v o n T a m p e r e u n d d e m A l t e r n a c h p o s t s v i o n i s c h.

Aus den Untersuchungen von Huhma geht deutlich hervor, dass das dreieckige Granodioritmassiv von Kolmisoppi in Kankaanpää eine gleichartige Basalformation wie das Massiv von Värmälä vertritt (Huhma, *et al.*, 1952).

Mit E. Heiskanen bin ich in den Untersuchungen im Valkeakoskigebiet zu dem Resultat gekommen, dass das hier vorkommende kleine Quarzdioritmassiv einen steifen Klumpen vertritt, der in den gebirgsbildenden Bewegungen von seinem natürlichen Zusammenhang, d e r S e d i m e n t a t i o n s u n t e r l a g e d e r b o t h n i s c h e n S c h i e f e r f o r m a t i o n, losgerissen worden ist und der in der Regionalmetamorphose nicht imstande war, sich der allgemeinen tektonischen Orientierung anzupassen (Saksela und

Abb. 11. Das Granodioritmassiv von Värmälä in der Schieferzone von Tampere. Nach Seitsaari (1951) umgezeichnet. 1. Leptite und Feldspatporphyre; 2. Postkarelobothnische Quarzdiorite, Granodiorite, Porphyrgranite und Mikroklingranite; 3. Konglomerate; 4. Phyllite und Glimmerschiefer; 5. Das postsvionische Granodioritmassiv von Värmälä; 6. Postkarelobothnische Diorite und Gabbros; 7. Uralit- und Plagioklasporphyrite, basische Tuffite. Die Streckungsrichtungen und Faltenachsen sind mit Pfeilen bezeichnet.

Heiskanen 1952; Saksela, 1952). Der Quarzdiorit gehört aller Wahrscheinlichkeit nach zur Gruppe der postsvionischen Tiefengesteine.

In diesem Zusammenhang mögen noch die Untersuchungen von Kranck (1933) im Schärenhof von Ekenäs (Tammisaari) erwähnt werden. Speziell wird die Aufmerksamkeit auf die Migmatitzone der inneren Schären gerichtet. Die Hauptkomponente ist hier ein Gneisgranit. Ausserdem kommen verschieferte Gabbrogesteine, Quarzitschiefer, Karbonatgesteine und andere Schiefergesteine vor. Wie Kranck bemerkt, hängt das uneinheitliche Aussehen des Felsgrundes dieser Zone nicht nur von den primären Verschiedenheiten der Gesteinsgemengteile, sondern ebenso sehr von den äusserst komplizierten und durchgreifenden Deformationen ab, die den Gesteinen gleicher Herkunft oft ein sehr verschiedenartiges Aussehen geben können. So geht der Gneisgranit vielerorts in ganz feinkörnige leptitähnliche Gneise oder in kräftig schiefrige, dichte hälleflintähnliche Gesteine über. Die alten grobgneisigen Strukturen werden durch eine vollständige Auswalzung in Schieferstrukturen verwandelt. Diese werden wieder in Miniaturfalten gefaltet, wobei die alten Scherflächen durch jüngere ersetzt werden. Derartige Vorgänge sind nicht in den Gneisgraniten der äusseren Schären beobachtet worden. Nach Kranck kann man von Polymetamorphose des Gneisgranits sprechen. Es ist auch interessant festzustellen, dass die Achsialrichtungen innerhalb der Migmatitzone der inneren Schären in grossen Zügen beinahe senkrecht zu den in den südlicheren Zonen beobachteten Achsialrichtungen stehen. Das Achsialfallen ist immer recht steil (45°—80°). Es scheint mir nicht unmöglich, dass die obenbeschriebene Migmatitzone svionische Bildungen vertreten, die während der karelobothnischen Orogenese aufs neue metamorphosiert worden sind.

Es ist sehr interessant, dass die svionische Unterlage der Karelobothniden jetzt auch im Skelleftegebiet in Schweden festgestellt worden ist. Kautsky (1957) hat hier zwei Suprakrustalserien von wesentlich verschiedenem Alter ausscheiden können, die durch eine orogene Periode mit Faltung und der Bildung des sog. Jörngranits voneinander getrennt sind. Die jüngere, Elvabergserie, liegt auf der tiefverwitterten älteren Maurlidenserie (Grauwacken, graue und schwarze Schiefer, saure und intermediäre Laven und Tuffe, effusive Grünsteine usw.). Die Elvabergserie beginnt mit Basalkonglomeraten (die Vargfors- und Mensträskkonglomerate), die gegen oben allmählich in feinkörnigere Sedimente und schliesslich in graphit- und kiesreiche Schiefer übergehen. Der Zeitraum zwischen der Ablagerung der beiden Suprakrustalserien muss nach Kautsky als bedeutend angesehen werden, da der Jörngranit die Maurlidenserie durchschlägt, jedoch als Gerölle in den Konglomeraten der Elvabergserie auftritt. D i e B a s a l l a g e n d e r E l v a b e r g s e r i e l i e g e n s t e l l e n w e i s e a u c h d i r e k t a u f t i e f v e r w i t t e r t e m J ö r n g r a n i t.

Auf Grund der von Kautsky erhaltenen Resultate ist der weitverbreitete Revsundgranit, wie Gavelin (1958) bemerkt, als »karelisch» und gleichaltrig mit den Adak-, Palja-, Arjeplog- und Linagraniten zu betrachten (Grip, 1946; Ödman, 1957). Die erwähnten Granite vertreten m. E. spätkinematische postkarelobothnische Tiefengesteine. Für den Revsundgranit hat sich ein Alter von 1800×10^6 Jahre ergeben (Quensel, 1956; Gavelin, 1958; Kautsky, 1959).

Es mag noch erwähnt werden, dass nach Metzger (1959) die »Svekofenniden» und »Kareliden» auf demselben alten Gneisgranitsockel liegen, der im Osten weitgehend zu Tage tritt. In dem »svekofennischen» Raum wäre die alte Unterlage jedoch durch Anatexis zu den primorogenen Graniten mobilisiert, und auf diesem Wege in die überlagernde Sedimentdecke eingedrungen. Wie wir gesehen haben, ist der alte Sockel auch im »svekofennischen» Raum der Nachwelt erhalten geblieben.

SCHLUSSWORT

Nach meiner Auffassung ist es klar, dass die jetzige Alterseinteilung des Felsgrundes von Finnland als solche nicht mehr aufrechtzuerhalten ist. Die von Sederholm vorgelegte grosse subbothnische Diskordanz ist eine Tatsache, die man nicht unberücksichtigt lassen kann. Dies bedeutet, dass die svekofennische Formation in zwei Teile zerfällt. Diejenigen Bildungen, die stratigraphisch unterhalb der subbothnischen Diskordanz liegen, sind zu einer gesonderten, älteren s v i o n i s c h e n Formationsserie zu rechnen. Die oberhalb der Diskordanz liegenden bothnischen Bildungen schliessen sich wiederum lückenlos an die Kareliden an und bilden mit diesen die ausgedehnte k a r e l o b o t h n i s c h e F o r m a t i o n s s e r i e. Die svionischen Bildungen sind in weiten Gebieten in Ostfinnland entblösst, sie kommen auch anderswo, so auch in Savo, Ostbothnien und Südfinnland, vor. Die Aufklärung der geographischen Verbreitung der genannten Formationsserien gehört zu den wichtigsten Aufgaben der finnischen geologischen Forschung.

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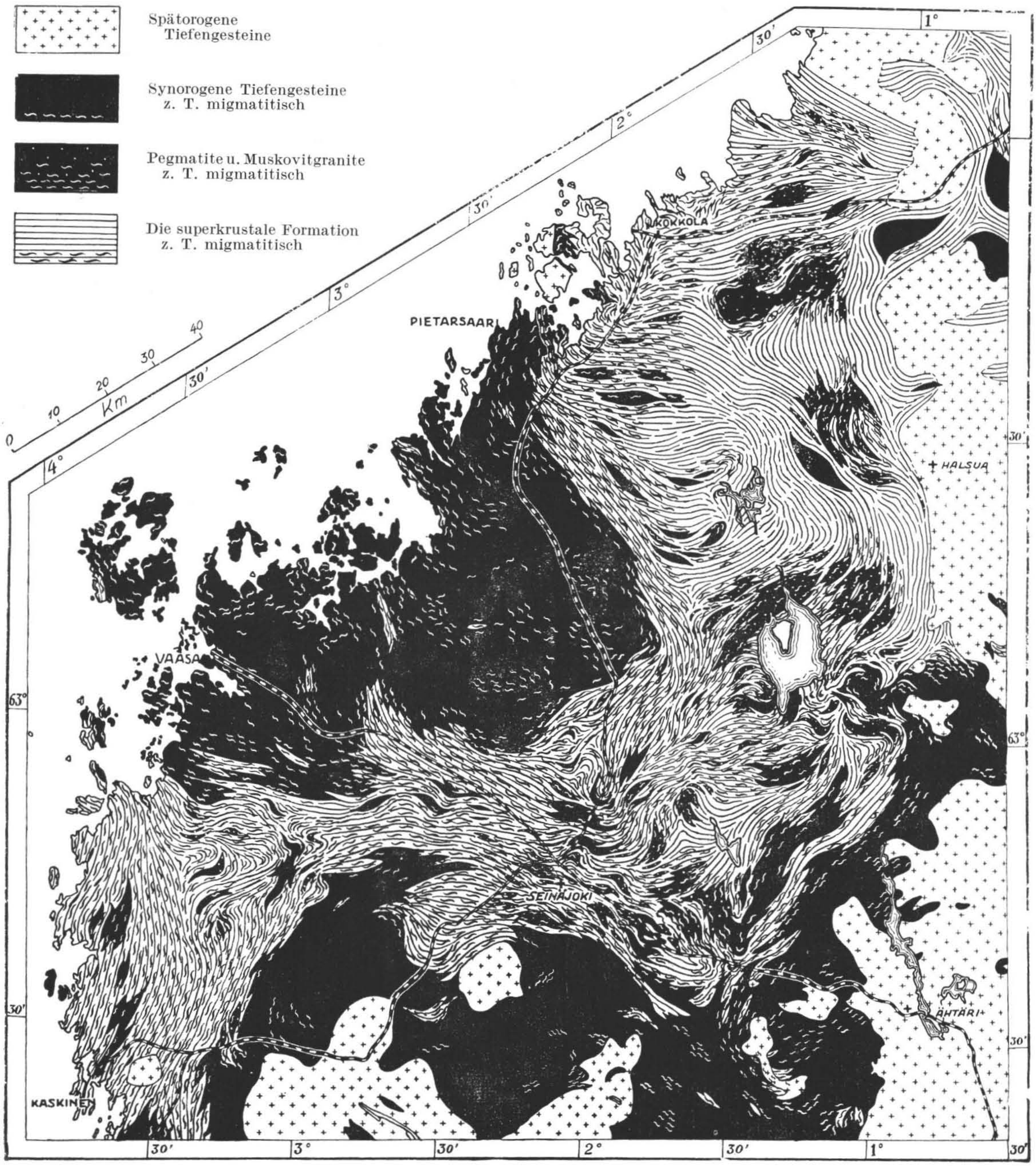

Strukturgeologische Karte des südlichen und mittleren Ostbothniens.

Martti Saksela: Aspekte der finnischen geologischen Forschung.

13

THE TECTONIC POSITION OF THE ANORTHOSITES OF EASTERN CANADA [1]

BY

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ABSTRACT

A brief review of present knowledge about the anorthosites of eastern Canada is presented with a discussion of the reason why this rock type is so abundant in the Grenville province.

All observations indicate that the formation of the anorthosites took place in several stages, and most of them have been repeatedly recrystallized. The author suggests that the time factor may be of importance, and the concentration of anorthosite in eastern Canada may depend on the length of time during with the Grenville geosyncline developed.

CONTENTS

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INTRODUCTION

This report on the anorthosites of eastern Canada summarizes some observations which may explain why this particular portion of the Canadian Shield is richer in anorthosites than any other similar region.

Knowledge about the geology of northern Quebec and Labrador is still incomplete, but it has increased considerably during the last ten years, mainly because of the discovery of rich iron ore deposits. The growing interest in titanium ores has stimulated prospecting and mapping of anorthosite areas. This work has given at least an approximate picture of the distribution and shape of the larger anorthosite bodies. The information about them is still scanty, however, and only two of the major areas are fairly well mapped: namely, the Morin pluton north of Montreal and the Nain pluton in Labrador. Portions of other anorthosite bodies have been mapped where they have occurred within the areas covered by the »1 inch to 1 mile» mapping of the Quebec Department of Mines. Much of the information on the petrology of the anorthosites is hidden in unpublished theses on the library shelves of McGill and other universities.

It is not surprising that little attention is given to the Canadian anorthosites in the petrographic literature.

Furthermore, the magnificent work done by Buddington, Balk and their co-workers in the Adirondacks has completely overshadowed what has been accomplished on the Canadian side.

The Adirondacks have naturally been the Mecca for students of anorthosites; so far, only a few know the grandeur of the northern areas. One has to travel day after day in terrain where the bedrock consists of almost uniformly pure anorthosite in order to appreciate fully the anorthosite problem, and to realize that the great Adirondack area is small compared to many of the northern plutons.

The total area occupied by anorthosites in eastern Canada is more than 200 000 square km. The area of the Lac St. Jean body alone is over 60 000 square km. In the zone marked on the map (Fig. 1), 15—20 per cent of the bedrock consist of anorthosites and related rocks, and therefore they have an importance comparable to that of the granites in most Precambrian terrains.

The reason for this peculiar concentration of an otherwise rather rare rock type is the main topic of this discussion. The petrology and the origin of the rocks will be dealt with only so far as necessary to answer this question.

THE ANORTHOSITE PROBLEM

The origin of the anorthosites is almost as controversial a subject as the granite problem.

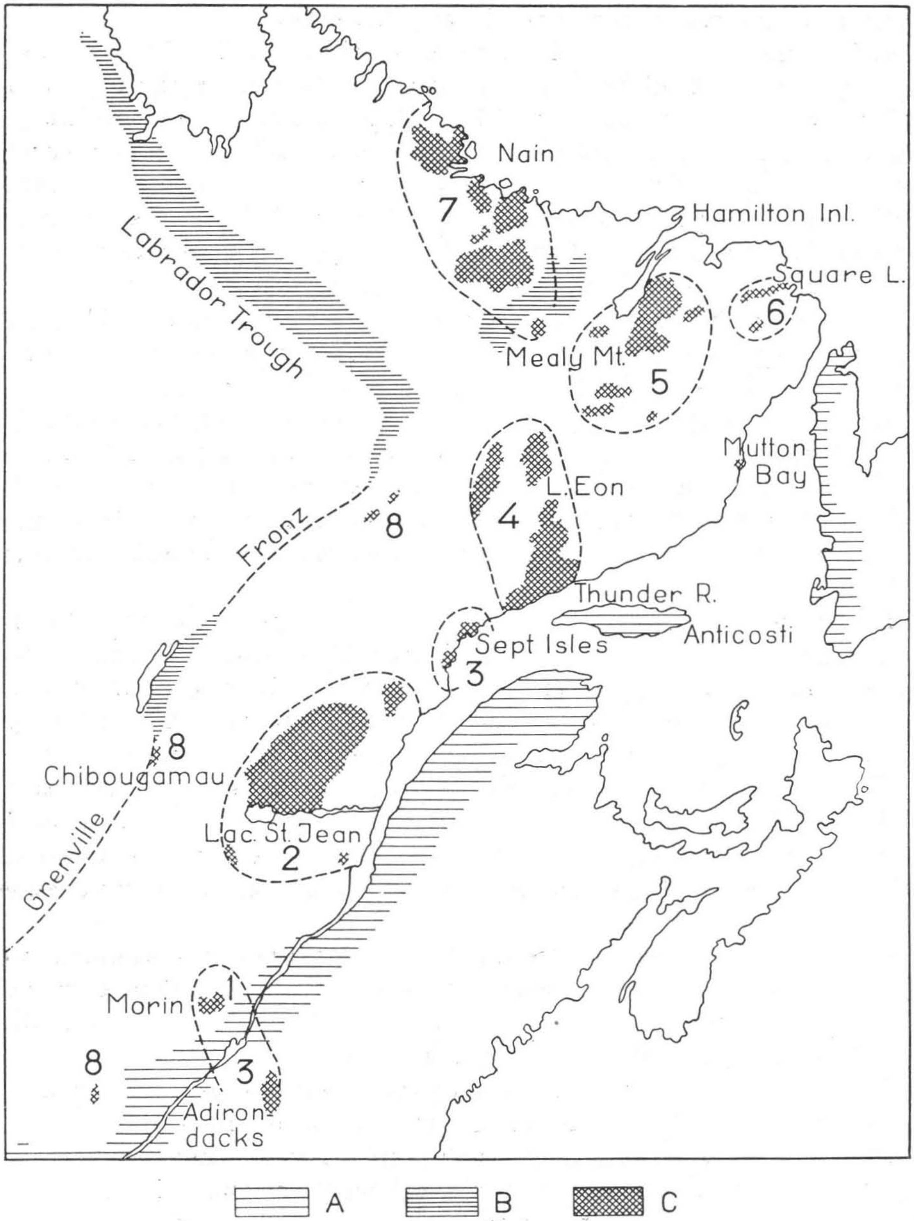

Fig. 1. The distribution of anorthosites and related rocks in eastern Canada. 1. Morin-Adirondacks, 2. Lac St. Jean, 3. Sept Isles, 4. Romain River, 5. Mealy Mountain, 6. Square Island, 7. Nain, 8. Anorthosites of the western margin of Grenville. A. Paleozoic, B. »Proterozoic», C. Anorthosite.

In the current literature, the origin of the anorthosites has been variously described as magmatic, metasomatic, and anatectic.

1. The magmatic theory is represented above all by Bowen's (1917) classic paper on the origin of anorthosites by crystal accumulation, still the basis for all discussions of the problem. This theory has been modified by later authors insofar as they admit the importance of crystal accumulation, but posit a magma of nearly anorthositic composition (gabbroic anorthosite) in which it takes place. The mechanics of the process has been discussed particularly by Buddington (1939), Balk (1931) and von Eckermann (1938). Buddington regards the anorthositic gabbro as the mother magma and believes the syenitic rocks of the complex to be later independent intrusions. Balk, who emphasizes the importance of flow-movement in a partly crystallized melt at the time of the crystal accumulation, believes in a dioritic parental magma, and von Eckermann thinks that the original magma was olivine basalt and that both the anorthosite and also the syenitic and associated granitic rocks were formed by differentiation from this melt. The majority of papers published on anorthositic rocks follow some of these lines of thinking, which are well supported by the rock associations found in the areas studied.

2. A metasomatic origin for at least certain types of anorthosites has been suggested by Michot (1957), Berthelsen (1960), Hietanen (1956) and others. Michot's important papers about the Norwegian anorthosites lay stress on the high temperature ultrametamorphic habitat of the anorthosites and the complicated folding structure in them. In his excellent tectonic studies of the gneiss-anorthosite terrains of southwestern Greenland, Berthelsen describes the anorthosites in this area as sheetlike bodies highly folded together with limestone and other meta-sedimentary rocks. They are explained as formed from limestone by a replacement process.

3. Winkler (1960) and Barth (1936) have suggested that anatectic processes contributed to the formation of anorthosites. In a number of very interesting laboratory experiments in anatexis, Winkler has shown that after partial melting of lime-rich clay sediments, the residue, upon removal of the first molten granitic part, has a composition more or less nearly identical to that of anorthosites. Although the application of this discovery to the formation of the large anorthosite bodies may involve some difficulties, it may give the key to the formation of anorthositic magmas. Barth has also pointed out that anatectic processes may account for the formation anorthosites, but has not specified these processes.

All these discussions contain useful ideas. Even though not one of the hypotheses covers the whole problem, each one explains certain phases of the history of the anorthosites. As will be seen, one essential quality of this history is its complexity. The anorthosites were formed not by a single simple process, but by a sequence of events more complex than for other plutonic rocks. Read comments that there are granites and granites: in this case, the temporal factor is vitally important; the anorthosites may pass through an anatectic stage, a magmatic stage and a metamorphic stage.

The Canadian anorthosites appear to be essentially magmatic, but this does not answer the question of the origin of the magma nor of transformations during and after the consolidation of the rock. Both anatectic and metasomatic processes must have played a role during the emplacement, and also during the millions of years between the original emplacement of the rock and its present position.

SOME GENERAL POINTS OF VIEW CONCERNING THE OCCURRENCE OF THE ANORTHOSITES OF EASTERN CANADA

Some of the anorthosites of eastern Canada are located in the Grenville province, but others are found in western Labrador, which is not generally regarded as part of the Grenville. The Grenville is a broad belt consisting largely of high-grade metamorphic gneisses and many plutonic intrusives. To the northwest, a belt of unmetamorphosed or slightly metamorphosed supercrustal rocks border the area which in Ontario includes the so-called Huronian formations. In the Grenville Problem, F. F. Osborn (1956) discusses the nomenclature of the Grenville province.

The present nomenclature of these formations is confusing. The name Grenville was originally suggested by Lawson (1863) as a stratigraphic name for the comparatively well-preserved sedimentary portions of the gneiss formation in southern Ontario and Quebec. They are characterized by a great abundance of limestone and fairly well-preserved argillites and arenites. Northward these rocks are largely replaced by gneisses and have been increasingly altered with the introduction of much granitic material. Limestone occurs less and less frequently. The name Grenville has, however, been used wherever sediments can be recognized, although there is little evidence that they belong to the same formation.

Recent radio-isotope age determinations (Hurley, Beall, 1960) indicate that the boundary sediments of Huronian type (Labrador Trough, *etc.,)* are much older than believed earlier, probably almost as old as the sediments from which the gneisses were made. Consequently they could be regarded of Grenville age and included in the Grenville province. Tectonically these rocks form as large a part of the same mountain belt as do the so-called Grenville gneisses, a strong argument for their inclusion in the Grenville province.

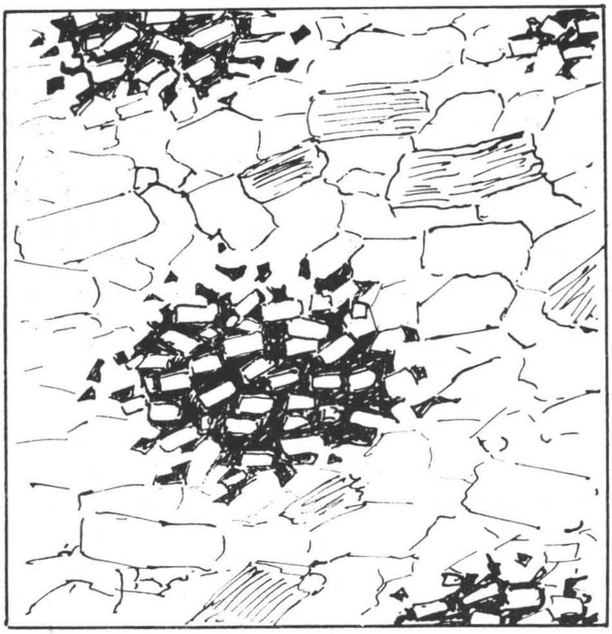

Fig. 2. Anorthositic gabbro from St. Alma, Quebec. Supergrains of ilmenite-magnetite with poikilitically included crystals of plagioclase. The inclusions probably represent the first generation of plagioclase in the rock. At a later stage the supergrains of iron ore were formed and, at the same time and still later, the surrounding mass of coarse plagioclase and pyroxene were formed by recrystallization. One fifth natural size.

A name in agreement with international usage should be found for the folded belt — the orogen or orogens — of the eastern portion of the Canadian Shield, because the use of stratigraphic-geographic names which mean so many different things at the same time can be quite misleading. Perhaps such a suggestion will some day be forthcoming; it would greatly facilitate the reading of geological papers about eastern Canada.

This Grenville, however, and its continuation in the U. S. A. is the home of the anorthosites, and its general petrological features are therefore of interest in this discussion (Fig. 2). This brief review is restricted to the Quebec-Labrador portion; the southern part has been treated in great detail by Engel (1953) and Robinson (1956).

Predominantly gneissic rocks, many pure granites, and other intrusives characterize the bedrock in the northern continuation of the Grenville, and some of the gneisses are high-grade metamorphic. The basic rocks are mainly represented by anorthosite and related gabbros. Although meta-sedimentary rocks with preserved primary features are extremely rare, the gneisses are certainly of partly sedimentary origin. In general, gneisses derived from quartzitic sediments are more common than those derived from limy sediments. Characteristically trondhjemitic and granodioritic rocks are minor constituents. Evidence of volcanic constituents is extremely rare, unless they occur in highly metamorphic disguise.

One often-mentioned peculiarity of the Grenville is the lack of economic mineralization. The only ores of importance are titaniferous iron ores connected with the anorthosites. Another peculiarity is the high sodium content of the gneisses relative to the potassium content.

These features hold good for the typical Grenville gneiss zone. The Huronian-type sedimentary rocks in the west form a mio-geosynclinal zone, or an epicontinental fringe of a folded belt; but, at least in the northernmost and southernmost areas, the sedimentary rocks grade into gneisses with progressive degrees of metamorphism and granitization.

It has been suggested that the northernmost part of Labrador northwest of Seal Lake represents a complex older than the Grenville province. Anorthosites occur in great abundance here also, and generally seem to be less tectonized than in the Grenville province.

A number of geologists working in the Labrador trough area (Gastil, 1960) have presented evidence of two different stages of mountain-folding, not necessarily representing two orogenies, in the Grenville province. The older has caused the folding of the main portion of the trough; the younger appears responsible for the complicated crossfolding in the southernmost section (Mount Read area, *etc.)* and also for the folding of the Seal Lake sediments.

If this be the case, it is conceivable that the anorthosites may have been emplaced between the two foldings. Both foldings have affected the gneisses in the Grenville province, and segments of considerably different age in this complicated group of rocks must be considered, including both slices of the original basement and granitized sediments of at least two other ages.

The general distribution of the anorthosites (Fig. 1) is peculiar because of its apparent lack of correlation with the general structure. At most, one can distinguish a crude alignment of the plutons in a northeasterly direction, perhaps related to the postulated older foldings. The occurrences in northern Labrador are obviously unaffected by the assumed boundary between the northernmost part of Labrador and the Grenville province proper.

This irregularity may be partly the result of incomplete knowledge of the area, but not entirely. It agrees with the general shape of the different anorthosite bodies themselves. Judging by present data, the bodies have no characteristic form; each one has its individual pattern. Most geologists who have worked in anorthosite areas seem to regard them as sheetlike bodies or lopoliths; but, with the exception of parts of the Adirondacks and possibly the Morin pluton, little evidence for this assumption exists. The comparatively well-known Nain area, for instance, seems to have a highly irregular form and it could equally well be called a batholith.

Obviously, however, most of the anorthosite bodies occur along the central axis of the whole province, extending in the southern portion in a northeasterly direction, and in Labrador in a northwesterly direction. It is interesting that the greatest concentration of anorthosites is found along an axis extending from Nain to the north shore of the Straits of Belle Isle at Anticosti, interrupted only east of the south end of the Labrador trough. The tectonic significance of this line will be clearer when the structure of the area has been studied in detail.

The map (Fig. 1) shows that there are two areas in Labrador and northeastern Quebec which are devoid of anorthosites. This may be the result of incomplete knowledge of the area, but may also have some structural significance. These areas are in the coastal section of Labrador from Domino Run to Hoedale, and from the Straits of Belle Isle on the coast to the Johann Beetz area inland from the north shore of the Gulf of St. Lawrence. In the first area the bedrock is characterized by the presence of quartzitic rocks (type locality, Aillik) in a more or less migmatized state. A similar formation is also found at Battle Harbour north of the Straits of Belle Isle. At Mutton Bay (Little Mecantina) farther south, however, syenitic rocks occur, possibly of the same age as the syenites associated with anorthosites, indicating that the southern block may have anorthositic intrusions not yet discovered.

THE ANORTHOSITES PLUTONS OF QUEBEC-LABRADOR

For the sake of convenience, the anorthosites have been divided according to their geographic distribution in the following regions: (See map, Fig. 1)

1. The Adirondack-Morin region
2. The Lac St. Jean region
3. The Sept Iles region
4. The Romain River region
5. The Mealy Mountain or Hamilton Inlet region
6. The Square Island region
7. The Nain region
8. The anorthosites of the western border of the Grenville (Haliburton-Bancroft, Chibougamau, Little Manicouagan, *etc.*).

In addition to the main plutons, each region comprises several smaller plutons.

This chapter contains data about these regions important for the following discussion, and also some notes relative to the work done in the areas mentioned.

THE ADIRONDACK-MORIN REGION

The Adirondack and Morin anorthosite bodies are situated respectively on the south and north sides of the St. Lawrence lowland. Between these areas at Oka on the north shore of the Ottawa River, there is a window of Precambrian rocks partly consisting of anorthosite surrounded by Paleozoic sedimentary rocks. Anorthosite has also been found in drill holes northeast of Montreal. Therefore it is possible that the anorthositic rocks continue under the younger sedimentary rocks all the way from the Laurentian area to the Adirondacks. Even if they do not reach the surface of the basement of the sedimentary cover, there is good reason to believe that they occur at depth as one continuous, fairly narrow pluton over 200 km long.

The Adirondack region is so well known that a further description of it is unnecessary (Buddington, 1939.)

The Morin pluton was described in a monograph by Adams (1893). This paper is still the best source of information about the petrology of the area as a whole. F. F. Osborn (1936, 1949) published important studies of the rocks of the New Glasgow area on the southeast side of the massif, and described one of the few known dykes of anorthosite. Within recent years the Quebec Department of Mines has published several preliminary map sheets with brief descriptions of the rock types (Beland, 1960; Coté, 1948; Klugmann, 1957.) A geochemical study of anorthosites particularly from the Morin area and also dealing with the trace element distribution in the rocks is under preparation by J. Papezic at McGill University, and will soon be ready for publication. Furthermore, Dr. C. Düsing (Research Fellow at McGill University) is carrying out a petrotectonic study of the area.

The Morin pluton consists of two parts of different character. The nearly circular western portion consists of homogeneous coarse-grained, dark gray anorthosite, with little gabbroic material. The eastern portion forms an elongated trunklike appendix, pointing southward in the direction of Oka. The anorthositic rocks are here interbedded with acidic gneisses and are considerably deformed, in certain zones almost mylonitic and fine-grained (»Chertsey type»). As Osborn has pointed out, the rock types may be of different ages.

The same deformation has also affected the gneisses and mangeritic (quartz monsonitic) rocks of the contact zone. As Düsing has clearly shown, this deformation belongs to a movement which occurred later than the general folding of the migmatite gneisses in the area. This discovery is important because a similar, secondary deformation has occurred in many of the other anorthosite regions of Canada. The later deformation has left very few traces in the eastern portion of the Morin pluton, although much of what has been called a cataclastic texture may have been caused by it. As in most anorthosite areas, a certain layering always occurs in the undeformed rock which probably originated during the consolidation. The banding is particularly conspicuous in the gabbroic ilmenite-rich portions of the complex. Several showings of low-grade ilmenite ore are known. Most of the high-grade enrichments occur as veins or pockets deposited during the last phase of consolidation of the rock.

The shape and downward extension of the anorthosite body is still a matter of speculation. The primary layering is usually steep and has not been sufficiently studied to give any definite information about the structure and the emplacement of the body. Probably it has been considerably affected by later tectonic movements. The information to date appears to indicate the existence of a deep-seated funnel-shaped body, which has been strongly deformed along its west and south sides. This agrees with Balk's (1931) description of the Adirondack pluton. Like this pluton, the Morin body is surrounded by pyroxenebearing acidic rocks usually classed as mangerites (Osborn, 1936). They consist of plagioclase, some microperthitic alkali feldspar and pyroxene as main components, and usually hornblende and quartz as well. Along the east border these rocks are partly converted to augen gneisses. In certain zones they are strongly tectonized and grade into mylonites. In some localities they intersect the anorthosite.

The rocks in the drill holes north of Montreal (St. Vincent de Paul) belong to the highly granulated, light green variety found in the New Glasgow area at the southeastern border of the Morin massif.

The anorthosite of Oka is briefly mentioned in the description of the Lachute Map Area by F. F. Osborn (1936) and in the preliminary report on the Oka area by Maurice (1957). A special study of its occurrence is presently being carried out by Mr. Vaughan at McGill University.

The outcropping rock of the area resembles the anorthosites of the eastern section of the Morin area. It is pale green to white and fairly unhomogeneous, and it grades into gabbroic anorthosite. A certain foliation produced by tectonic deformation is generally visible. The pyroxene is often drawn out into stringers, indicating that the deformation here, as in the Morin area, took place at a fairly high temperature. The eastern side of the anorthosite body is more deformed than the western.

About 1 500 feet from the western contact, a drill hole [1] has been sunk through the Paleozoic sedimentary rock in a search for oil. The hole was

[1] I had the opportunity to study the drill core through the courtesy of the Quebec Department of Mines and Dr. T. H. Clark of McGill University.

continued for about 1 400 feet through Precambrian rocks. The upper part of the drill hole intersects several hundred feet of a coarse porphyritic rock with large idiomorphic andesine crystals in a groundmass probably composed of an anorthositic gabbro. The dark minerals are largely altered to chlorite and epidote. This rock grades into pure anorthosite. It is an excellent example of an anorthosite formed by crystal accumulation, and it strongly supports the theory of a magmatic origin for the anorthosite. Unfortunately, the porphyritic rock is not found in the outcropping anorthosite and it is still too early to draw any definite conclusion from the discovery.

THE LAC ST. JEAN REGION

The enormous pluton of Lac St. Jean is one of the largest known anorthosite bodies in the world. The mapping of the area is still mostly in the reconnaissance stage (Dresser, 1914; Denis, 1932). The southern portion of the area has been described by Jooste (1949) (unpublished McGill Ph. D. thesis).

The region contains all the common types of anorthositic rocks and among them are some beautiful examples of coarse-grained, pure anorthosite with small quantities of pyroxene, which have been widely used as ornamental stone, under the trade name of black granite. Particularly outstanding is the rock from Peribonca because of its homogeneity, colour, and perfect igneous texture. The texture is partly subophitic, and lacks any trace of cataclasis. The rocks are often very coarse-grained. Pegmatitic varieties with gigantic grains have been found in the area north of Lac St. Jean (Berrangé, 1959). Layered anorthosites with low to steeply dipping foliation are found along the southeast border of the area connected with bodies of ilmenite ores.

Monzonites and syenitic granites have been intruded into the anorthosites and anorthositic gabbros.

Several smaller areas of anorthosite occur in the vicinity of this large pluton, one of which, the St. Urbain massif, is known because of its ilmenite showing, often mentioned in the literature (Mawdsley, 1927). Other anorthosite areas occur in the region between Lac St. Jean and La Tuque.

The structure of the region, however, remains relatively unknown.

THE SEPT ILES REGION

In this region, several plutons occur along the north shore of the Bay of St. Lawrence, and probably join under the sea. They have been described briefly by Faessler in the text accompanying his geological map of the Sept Iles area (Faessler, 1942). The anorthosite is particularly well-exposed on the off-shore islands at Sept Iles. The rock is a coarse-grained, dark gray pure anorthosite with little or no trace of deformation. Inland, north of the town of Sept Iles, banded impure types occur.

The new iron ore harbour at Shelter Bay, southeast of Sept Iles, has been excavated in a very coarse anorthositic gabbro with a perfect igneous gabbroic texture. It is rich in titanomagnetite and apatite. Still farther southwest at Pentecôte Bay there is an interesting association of anorthositic and more alkaline feldspathic rocks, the latter represented by a pink Rapakivi-like granite and a greenish black spotted hastingsite syenite; both are undeformed. On the Caouis Islands, off the coast at Pentecôte Bay (not visited by Faessler), the anorthosite is unusually coarse and porphyritic with perfectly idiomorphic crystals of plagioclase (over three inches in diameter) in a coarse-grained groundmass of anorthositic composition. The semi-parallel arrangement may be due to crystal sedimentation under undisturbed conditions. The rocks is traversed by dykes of syenitic pegmatite, which are interesting because of the abundance of crystals of a pale yellowish-brown garnet.

THE ROMAIN RIVER AREA

This very large area contains some of the greatest concentrations of anorthosite in the world, possibly greater than the Lac St. Jean pluton. Only the southernmost part along the coast is comparatively accessible and partly mapped; the inland is very little known. The Romain River anorthosite was first mentioned by Low (1895), who travelled in its northern parts; the coastal part was briefly described by Richardson (1853—56); Retty (1944) studied the coastal section east of Mingan and discovered the famous ore deposits of Allard Lake. Klugmann (1954) mapped the region northeast of Thunder River for the Quebec Department of Mines. During the same summer, I spent some months in the little known interior of the northern part of the same large mass of anorthosites. On published maps this massif sometimes appears as two, sometimes as three different bodies of anorthosites. So far this seems to be the result of conjecture and not of actual mapping. Much of what has been regarded as gneiss may be banded, foliated, anorthositic gabbro. Moreover, the two large anorthosite massifs farther northeast are only partly mapped, and the outline marked on the general maps is approximate.

Some petrographic studies, however, were carried out on a portion of these areas by Klugmann (1957) and Emo (1958), (Ph. D. and M. Sc. theses at McGill University) which show that the rock type are of the usual kind with both coarse-grained anorthosites and more fine-grained granulated anorthosites and gabbroic anorthosites.

The area east of the upper Romain River (south of Lake Eon) is interesting because of the enormous extension of rather fine-grained light gray or pure white anorthosites and also much foliated gabbroic anorthosite grading into hypersthene plagioclase ilmenite gneisses. Near the coast, foliated rocks are common. Good exposures are found along the seashore east of Sheldrake River. At Thunder River there is a very coarse gray anorthosite with pockets of ilmenite and hypersthene.

Klugmann suggests that the fine-grained anorthosites of the outer fringe of this body may represent a chilled contact. A petrofabric study, however, would be necessary to determine whether the finer grain is not the result of mechanical deformation. Much movement has taken place in the contact zone, evidenced by the folding in the foliated gabbroic anorthosites. Of petrological inteiest are the titano-magnetite showings along the coast, in which ilmenite veins represent late magmatic concentrates (Klugmann, 1957).

Although only tentative conclusions may be drawn from the scanty material available, it seems that at least the southernmost of the plutons of the area have been relatively more affected by a later tectonic movement, which probably took place after the final emplacement of the rock. This may explain why comparatively plentiful amphibole and relatively low-grade metamorphism (amphibolite facies) is found in some of the gneisses in the surrounding areas.

Klugmann has discussed in some detail the association of syenitic quartz-monzonite rocks with the anorthosites, and he agrees with Buddington that they probably represent later independent intrusions rather than differentiations from the same magma. The whole region is characterized by much coarse-grained granite and granitized rock and practically no preserved supercrustal rocks, with the exception of some small seams of limestone, micaceous gneisses, and amphibolites. The rocks of the area are of the high-grade plutonic type.

THE MEALY MOUNTAIN REGION

This region consists of the great massif of Mealy Mountain east of Lake Melville and several smaller areas on the west side of the bay. The only available geological information has been provided by mining companies particularly the British Newfoundland Company. Some more detailed petrographic data come from McGill students who have worked with these companies. One M. Sc. thesis, by Gillett (1956), describes the rock types at the southern end of the Mealy Mountain massif. The anorthosite here, as well as elsewhere in the pluton, of which I have seen samples, is a coarse-grained, mostly dark gray rock of a normal type. Gabbroic varieties and ilmenite pockets are also mentioned.

An interesting and very wide fringe of monzonitic rocks exists around the southern part of the anorthosite. The gneissic rocks north of the massif in the contact region north of Mealy Mountain (Kranck, 1939) also tend to have a monzonitic rather than normal granite-gneissic composition.

Along the seaboard north of this area, between Domino Run and Hamilton Inlet, there are a number of large dykes or sillshaped intrusions of coarse gabbro-diabase approximately following the coastline which may derive from the same magma as the anorthosite (Kranck, 1939, 1954). These sills have been broken up and partly sheared by later movements in connection with a granitization younger than the gabbro. This seems to indicate that a phase of regional metamorphism may have occurred here more recently than in the southern areas. More observations, however, are necessary before any definite conclusions can be drawn.

THE SQUARE ISLAND REGION

This region consists of an anorthosite body which extends from Square Harbour on the Atlantic coast of Labrador some distance inland, and some other bodies forming the high mountains southeast of Sandwich Bay. I have only seen the seaside outcrops and studied the rock types. To my knowledge, no mapping has been done in the area. These occurrences have a certain importance because they are the easternmost of the anorthosite bodies of Labrador and because the gneisses both south and north of them have undergone conspicuously strong deformation.

THE NAIN REGION

The anorthosite complex of Nain is a classic area for anorthosite studies, famous because it is the type locality of Labradorite and because the »chatoyant» semiprecious stone varieties of the mineral occur there. Early reconnaissance expeditions along the coast discovered the area in the last century (Packard; Coleman; Daly, *etc.*).

The most up-to-date information on the Nain region comes from Wheeler (1937, 1955, 1960), who has worked there indefatigably for many years and who has published several important papers about the area. He has provided the best maps of an anorthosite area in Canada.

His map of 1960 shows that the body has a very irregular and complicated shape in which anorthosites, adamellites, and quartz monzonite rocks are intermingled; the acidic rocks grade into the anorthosites in certain parts of the area and intersect them in others. This observation is of very great interest and gives convincing evidence of the close relationship between both rock types. Generally speaking, there seems to be a gradation from more basic gabbroic anorthosites into more feldspathic anorthosites and finally into alkali-bearing rocks westward. Wheeler suggests the possibility of a layered pluton, but regards the evidence so far as insufficient. Only in the eastern section in the Kighlaghpait mountains is there a more or less independent smaller pluton which definitely belongs to this type. It consists of anorthositic gabbros, troctolites, and also syenitic rocks, forming one of the most impressive layer plutons known. It has been studied in some detail by Morse (1961) who is preparing a paper on the formation (Ph. D. Thesis, McGill University).

The basement rocks of the Nain area are gneisses mostly of the granulite facies.

The rocks of the Nain anorthosite complex are remarkably little affected by later tectonization, although traces of possible cataclastic structures can be seen. Some rocks show semi-ophitic textures and, in general, definite primary magmatic textures. The buff anorthosites, in particular, display these features, and, as Wheeler's studies prove, have a typically high-temperature composition.

THE ANORTHOSITES OF THE WESTERN BORDER OF THE GRENVILLE

In this group belong a number of smaller anorthosite bodies which all appear to have been emplaced in more or less strongly folded gneisses, usually of elongated shape. All show traces of strong tectonization.

These border anorthosites are similar to the rocks of the Haliburton-Bancroft area described by Adams and Barlow (1910) in their classic study, which contains a wealth of interesting observations. Adams's opinion that the banding in the anorthosites may be comparable to the banding formed in flowing glaciers seems worth consideration and probably explains the internal structure of the more granulated types. This concept closely resembles Balk's theory of flow-banding owing to the inertia of the larger crystals in a partly crystallized magma, although Adams assumes a completely solid flow.

Describing several small bodies of anorthosites from southern Ontario, Harrison (1944) states that they belong to the same group. They are characterized by relatively abundant gabbro and are usually accompanied by acidic monzonitic rocks. The accessory minerals — amphiboles, scapolite, zoisite, klinozoisite — indicate low temperature alteration. All are considerably tectonized, and the elongated bodies are conformable with the structure of the surrounding gneisses.

Farther north there are several bodies of this type associated with granulite-facies gneisses in the region of Little Manicouagan and Mount Read.

In the Chibougamau area a fairly large body occurs in which the plagioclase has been markedly epidotized and sericitized.

Many undiscovered smaller bodies no doubt exist in the unmapped portions of Grenville.

THE EMPLACEMENT OF THE ANORTHOSITES

The general geographic distribution of the anorthosite plutons shows a large-scale alignment within the Grenville province and its continuation in the U. S. A. and Labrador, but no particular concentration along any structural lines. Some bodies, particularly the Sept Iles, Mealy Mountain, and Romain River plutons, show a degree of rectilinear arrangement, which, however, may be coincidental or dependent on major faults of later date.

The form of individual plutons, particularly the large ones, gives no evidence of structural control. Tectonization is common in many of the anorthosite areas, but it generally seems to be caused by superimposed later movements. Furthermore, such a deformation seems to be strong only in certain parts of the province. It is very conspicuous on the Quebec side and also in the western marginal plutons, but less so in Labrador from Mealy Mountain northwestward.

Petrographically, the anorthosites are surprisingly similar from Nain down to the American border. The distribution of the rock types, however, varies considerably in different plutons, a fact revealed by a detailed petrographic investigation (Wheeler, 1960). Wherever the general trend of differentiation is known, a relative enrichment of iron in the mafic minerals seems to have taken place in the later fractions of the magma. Papezic's geochemical study, which will be published shortly, discusses this problem in detail. The constant presence of quartz-monzonitic rocks is one of the principal problems in the petrology of the anorthosites. A close genetic relationship between these rock types seems evident, although the fact that the monzonites sometimes occur as later intrusives in the anorthosites and sometimes grade into them makes the interpretation of this relationship difficult. The conditions are similar to those described by J. J. Sederholm (1907) in his classic papers on migmatites: the same granite is sometimes cut by a basic dyke but, near by, intrudes the same dyke. Anatexis may be a factor in the case of the anorthosites, at least as far as the origin of the magmas is concerned. No plutonic rock approaches the average composition of sedimentary rocks and common clay as closely as a quartz monzonite.

Finally, the scarceness of granodiorites and trondhjemites in the Grenville province must be considered. Osborn (1936), quite correctly, has pointed out that the anorthosites in Quebec seem to occupy in a certain respect the same position as the granodiorites and related rocks of the core batholites of British Columbia.

In order to discover the reason for this, information must be sought concerning the physico-chemical conditions in which the anorthosites were formed.

Some conclusions can already be drawn from a consideration of the textures. The most conspicuous feature of the anorthosites is the abundance of very coarse-grained varieties; the grain size is probably twice that of normal gabbros and granites (with the exception of rapakivi-type granites and some other similar rocks which, by the way, often occur with anorthosites). According to the usual explanation, this depends on slow crystallization and or many volatiles. Yoder (1952) and, later, Klugmann (1957) have pointed out the possible importance of high water-pressure during the formation of anorthosites. This theory is contradicted by the anhydrous mineral composition of the rocks under discussion.

In most discussions of anorthosites, much emphasis has been placed on cataclastic textures with broken larger crystals and granulation in the interspaces. In some cases a granulous and even mylonitic texture has been described.

One feature characteristic of gabbroic anorthosites, but seldom mentioned, is a coarse poikilitic texture with large agglomerations of entwined mafic minerals with inclusions of feldspar. These may have a diameter of 10 cm in coarse-grained anorthositic gabbro. The distribution of ilmenite is particularly conspicuous in a polished surface of the rock. The mineral is not evenly distributed, but concentrated in accumulations (»intercumus crystals», Wager, *etc.*, 1960) which can be described as supergrains. Figure 2 illustrates an example from the Lac St. Jean area, Alma, Quebec. More or less well-developed similar structures can be observed in every polished piece of undeformed gabbroic anorthosite. This feature is usually overlooked because it cannot be observed in unpolished specimens.

The structure typifies the crystallization of the normal anorthosite. Obviously there was an unusually wide range of diffusion during the process. The supergrains probably were formed during a slow recrystallization in the presence of small quantities of intergranular liquid after most of the rock had been consolidated, and quite possibly even when it was completely solid. The rock was probably heated repeatedly until slow ionic movement could take place.

Similar conditions may partly account for the large size of the plagioclase grains. Further crystal growth probably took place around the original plagioclase crystals in accordance with observations of »accumulates» in the Skergaard pluton (Wager, *etc.*, 1960). A repeated crystallization is also indicated by the abundance of coronite structures and similar reaction textures in anorthositic rocks and by the occurrence of garnet and other minerals not normally formed.

The mineral association in the anorthositic rocks indicates high-temperature conditions during the main phase of the crystallization. The pyroxenes commonly show complex exsolution structures with one or several generations of exsolution lamellae in a host which is commonly an orthopyroxene or close to it. A typically unzoned calcic plagioclase, often combined with antiperthite, is associated with the pyroxene. Spectacular examples of such pyroxenes are found in the anorthosites from the Lac St. Jean, Mealy Mountain, and Sept Iles regions, and also from the Kighlaghpait massif. The first crystallized pyroxene was therefore an undifferentiated high-temperature variety which, during the long process of consolidation, did split up into more simple types. The common occurrence of hornblende and even biotite and chlorite in many anorthosites, and the occasional complete lack of pyroxene does not necessarily mean that the rock was originally formed at a lower temperature — but rather that earlier pyroxenes and olivines had been altered during later stages in the history of the rock.

The occurrence of high-grade metamorphic rocks together with anorthosites is significant. The anorthosites are associated with an ultra-metamorphic facies as Michot (1957) rightly has stressed.

Although this evidence appears convincing, it does not satisfactorily explain why eastern Canada is so rich in anorthosite. The conditions suggested by all these observations could prevail equally well in any deeply eroded portion of the earths crust, and the relative scarcity of anorthosites in the still older portions of the Canadian Shield and in later orogenic zones remains unexplained.

The stem-magma of the anorthositic rocks may have had a composition not found in the normal basaltic magmas in the subcrust (Buddington, 1939); but, if this was the case, the occurrence of enormous quantities of such a magma in Grenville time must be explained.

Obviously, one remaining factor has not been taken into account, the time factor. Admittedly, this is a highly tentative hypothesis but it explains many of the problems of the anorthosites.

The development of the orogenic belt (or belts) of the Grenville province from the time of the formation of the geosyncline to the diastrophy may well have been much slower than later orogenies. Possibly this epoch in the history of the earth was the first in which geosynclines in the modern sense were formed (Holmes, 1933). In any case, the continents before that time may have been low and the deposition of sediments in the subsiding geosynclinal basins exceedingly slow. The chemical disintegration and the exogenic differentiation, however, was complete, causing the occurrence of carbonates and pure silica. The quartzites in the Grenville are conspicuously glassy and some may have been cherty rather than normal sandstones. The enormous amount of cherty iron formation is characteristic. The great thickness of the gneisses may have been formed essentially from clay sediments.

The long period of quiet accumulation of sediments favoured a gradual increase in temperature as the geosyncline subsided, resulting in the formation of magma. There was ample time for contamination of sediments in the basaltic subcrustal material. The crystallization process of the magma continued for a very long time.

During the early stages, the upward movement of the magma was also quite slow, and, so far, little trace of volcanic activity in the gneiss terrains of the province has been observed.

This explanation accounts for the type of differentiation found in the anorthosite complexes and also for the structural and textural features already described. Disregarding the variable compositions of the anorthosites themselves, their differentiation is similar to the differentiation of layered plutons of basic composition, but the later fractions have been enriched in iron. F. F. Osborn's studies (1959, 1960) show that this type of development in a basic magma can be expected if the oxygen pressure decreases during the early stage of the crystallization. This, according to him, is the normal development in magmatism under anorogenic conditions. Most anorthosites, however, have gone through a complicated process of deformation and recrystallization after this first emplacement. This is clearly shown by the microtectonic studies of Düsing.

The few radio-isotope age determinations currently available from the Grenville province are insufficient to establish the suspected great length of the orogenic cycle. Nevertheless, the age of the trough sediments is at least 2 500 million years, and the age of the later granitization in eastern Canada is less than 1 000 million years, indicating that the earlier orogenic cycle in the province may be twice as long as that of modern orogenies. Holmes (1933) pointed out the possible greater age of the oldest orogenic cycles compared to the younger ones, and this suggestion has been mentioned several times in the literature.

Tentatively speaking, the time factor seems a fundamental element in the explanation of the tectonic position of the anorthosites in Quebec and Labrador. The emplacement of the anorthosites in time and space was controlled by the extremely slow development of the geosyncline in which the Quebec-Labrador orogenies took place about 1 400—800 million years ago.

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14

IDEAS ON THE INTERRELATION BETWEEN IGNEOUS AND SEDIMENTARY ROCKS [1]

BY

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ABSTRACT

The rocky crust of the continents has repeatedly been put through the geochemical cycles, for the total amount of weathered rock corresponds to an erosion of more than 30 km depth. Consequently, all rocks we see today have once been sediments; they have been modified by metamorphism, metasomatism, and at least partly, they have been in the form of magmas or lavas. But some time back they were derived from sediments.

Igneous processes do not lead to significant fractionation of rocks. But sedimentary differentiation is very strong. Therefore, the great multifariousness of igneous rocks is not igneous in origin but inherited from a sedimentary past.

Igneous rocks have been regarded as primary; they form the »juvenile» mother substance of all other rocks (secondary rocks). This has been the idea of all leading petrologists in the past. The average composition of the crust of the earth must correspond to that of igneous rocks; for ultimately all sediments and metamorphites are derived from igneous material. The overall average of the 5 159 analyses of igneous rocks compiled by Washington is still the accepted basis for the crustal abundances of the major elements.

The object of the present paper is to critically examine this assumption and to present evidence for a rather different relation between igneous and sedimentary rocks.

[1] Received February 27, 1961.

OCEAN BASINS

It is now quite clear that the use of the term »crust of the earth» is unfortunate in this connection. The rocks of the continents, from which most of the analyses are taken, differ from the rocks of the ocean floors: The so-called crustal abundances can, therefore, at best only apply to the continents which with their shelves, make up less than one half of the surface of the earth.

Our knowledge of the ocean floors is rapidly advancing, and the study of the sub-ocean petrology has become a very fascinating subject, indeed. Basaltic lavas are ubiquitous. The evidence for a subjacent world-wide basaltic shell is strong; it is probably glassy (Rittmann, 1958). It may have developed by differential melting of the material of the mantel, such as peridotite, eclogite, or still denser rock types.

Enormous effusions of basalts take place mainly through the rifts along the crests of the stupendous mid-ocean ridges twice girdling the globe with their total length of 72 000 kilometers. There are no sediments formed from these lavas; they are overlain, however, by sediments derived from the continents attaining in places considerable thicknesses (several thousands of meters). As a first approximation, I shall regard these sediments as losses from the geochemical cycles of the continents. The sediments will in part become deeply buried and assimilated into new lava flows. If they will ever return to the continent is not known.

CONTINENTS

Three quarters of the exposed surface of the continents are covered by sediments (Pettijohn, 1957). Of the remaining quarter, the largest part is made up of metamorphic and metasomatic rocks and less than 10 percent is igneous. Yet, according to current belief, this small fraction represents the primary rocks. The variation in the chemical composition of the primary rocks is particularly interesting.

They range from dunite and peridotite, with degressions into ijolite, anorthosite, carbonatite, *etc.*, through the whole gamut of monzonitic — syenitic types into the granite clan. How is this astonishing diversity to be explained?

The conventional answer is: By magmatic differentiation, the chief principle of which is fractional crystallization. The primary substance is a homogeneous magma of basaltic composition. By use of a great number of equilibrium diagrams of silicate melts of one, two, three, or even four components, the maternal relation of basalt is established and all types of truly igneous rocks become legalized as basaltic offspring. The last annual report of the director of the Geophysical Laboratory states that Dr. Schairer, one of the most prominent and successful workers of the laboratory, »is investigating the principal systems bearing on the nature of the source magma from which evolve the great diversity of igneous rocks.» The obvious conclusion now generally accepted is: Igneous processes (volcanism and magmatism) are responsible for the great diversification of igneous rocks. This looks like a banal conclusion; what else could differentiate igneous rocks? Further information is gained by scrutinizing the geochemical cycles.

THE WEATHERING OF CONTINENTAL ROCKS

Continental rocks are weathered, transported, and deposited on the shelves; this is the external part of the geochemical cycle. By burying, downbuckling, folding, and orogenesis they again become incorporated into the consolidated rocks of the continents; this is the internal part of the geochemical cycle. A new cycle can start and the rocks are again exposed to the agencies of wheathering. Are the cycles endless?

V. M. Goldschmidt and all geologists following him did not believe in »endless» cycles but thought that sodium accumulated in the ocean. I preferred to look at the problem in a different way and have assumed a balance between input and output of all elements in the sea (see Barth, 1952), and Carstens (1949) has shown that the balanced equilibrium will be attained within a reasonable time.

By thus assuming endless geological cycles, it is easy to calculate that in a million years the weathering corresponds to an erosion of 2.7 kg per square centimeter of the surface of the earth (Barth, 1961); this again corresponds to an erosion to a depth of 10 meters in a million years. Putting the age of the earth at 3×10^9 years, the total erosion will correspond to a depth of 30 kilometers all over the globe— and if we restrict the weathering to the continents, the erosion would correspond to more than 60 kilometers.

This sounds like very much; and it is, indeed, very much more than that calculated by contemporary geochemists (see Wickman, 1954, for further bibliography). But most estimates based on geological and stratigraphical evidence are of the same order of magnitude. They range upward of 30 km referred to the surface of the whole earth. For further discussion, see Kuenen (1950).

There is no reason to reject these estimates. Thus the total effect of weathering corresponds to a great erosional depth. Actually, however, the erosion has never dug so deeply into the crust; the sediments return to the continents and are recycled. The old students of Precambrian geology always searched for a beginning. At one time it was believed that the banded gneisses represented the first crystalline crust of the earth. It had to be abandoned. Time seemed to be without bounds — it was futile to seek confines. Sederholm, the great explorer of the Fennoscandian Precambrian, was seeking a floor upon which the sediments were deposited. He found no floor. The oldest rocks are sediments. Hutton said 200 years ago: »In the economy of the world I can find no traces of a beginning, no prospect of an end». It is still as true as in Hutton's days: no traces of a beginning!

THE RELATION BETWEEN IGNEOUS AND SEDIMENTARY ROCKS

Thus I am forced to the conclusion that our good earth is so old that the rocky crust has been repeatedly put through the geochemical cycles. The cycles are endless as far as the present geology is concerned. All rocks we see today have once been sediments; there is no »juvenile» material. The rocks have been modified by »plutonism», by metamorphism, metasomatism, they have been, partly at least, remelted and have been in the form of magmas and lavas; but some time back they were derived from sediments.

If this is true, we can better understand the diversity of rocks. The diveısity of sedimentary rocks is easily understood. Sedimentary differentiation is strong: quartzites, rocks composed only of silica, are easily made, and so are limestones; ore bodies are formed — 95 percent of all the useful ore in the world is concentrated in sedimentary rocks. But the diversity of igneous rocks has been hard to understand — that is really the problem which has loomed so high in petrological discussions. Now we have an entirely new approach. We used to be confined to igneous rocks starting with a basaltic mother magma, and deriving the other rock types by fractional crystallization. We are no longer limited in our thinking in this way. We can start with any kind of sedimentary material: sandstone, limestone, dolomite, clay, black shales, or bog iron ores. These are our starting materials; we can heat them, mix them, put them through metasomatism, or through differential melting with the formation of pore solutions, ichors, or rock magmas. The diversity now is great, indeed; the starting mateıials are potential sources for all enigmatic rock types; neither anorthosites nor pyroxenites are problems any more.

The idea of remelting sediments is not new. It is constantly being invoked by a great number of geologists to explain the genetic relations of special rock provinces. It has been generally accepted by the transformist school in petrology. And very recently Wyllie and Tuttle (1960) have shown by experiments that most sediments in geosynclines will begin to melt in the depth range 20—25 km; consequently magmas are present in the thick geosynclinal regions of the earth's crust at all times at depths exeeding 20 km. These facts were anticipated by Nieuwenkamp as early as 1948; and in three subsequent papers (1955, 1956a, 1956b) he has elaborated on the consequences. He also makes the following interesting remark: »In the original plutonism of Hutton, the magma consisted of remelted sediments. It would be an interesting historical study to see how it came about that this origin has been rejected.» As a matter of fact, Nieuwenkamp's *»hypothèse persédimentaire»* is almost identical to the ideas presented in the present paper; only Nieuwenkamp did not obtain the value of the total weathering.

SUMMARY

Returning, in conclusion, to the old petrology as it used to be administered, my problem was this: I was taught to look upon igneous activity as a complex of processes which, generally speaking, resulted in differentiation. Starting with a homogeneous magma, the diversity of rock types was derived. Faithfully I have taught the same lesson to my students. But actually it is contrary to common sense. Most people would use the expression »melting pot» as a figure of speech conveying the idea of homogenization, not diversification. We call America the melting pot of the nations of Europe; certainly it is and stands for Americanization, for the making of a new homogeneous nation, not for differentiation. My new idea in petrology is very simple: I believe that the melting pot of the igneous processes will lead to homogenization. If the igneous processes were allowed to go on without interruption, there would be very little rock variation in spite of the many silicate systems investigated in the laboratories. An example to the point is the petrography of Hawaii. Great quantities of basaltic rocks are extruded, igneous activity is intense, and the processes of magmatic differentiation are fully active; chemical analyses and thin section examinations clearly show that crystal fractionation, mostly following the rules of Bowen, operates continuously. And what is the result? Just homogeneous basaltic rocks covering nearly the totality of the islands! As *rarae aves* small amounts of *e. g.* trachyte or melilite basalt may be found; but volumewise such differentiates are pitiably small. The melting pot does its work, and by the lack of sedimentary processes interfering, the result is nearly only basaltic rock of remarkable homogeneity.

The point is that in the continents the igneous processes are not allowed to keep the show for themselves; they are interrupted by sedimentary processes, and sedimentation goes for differentiation.

Thus I shall end with a sentence which I think contains the truth, but which conservative geologists would call a paradox:

The diversification of igneous rocks is caused by sedimentary processes.

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15

VERSUCH EINER GENETISCHEN DEUTUNG DER LAPPLÄNDISCHEN GRANULITE [1]

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INHALT

EINLEITUNG

Der sächsische Granulit, der Prototyp der Granulite überhaupt, ist innerhalb seines geschlossenen Körpers mineralisch und strukturell so gleichförmig ausgebildet, dass man nur wenige palimpsestische Züge finden kann, die sich auf seine unmittelbare Vorgeschichte beziehen lassen. Man kann aber auf Grund dieser reliktischen Bestände eine sedimentogene Vorphase migmatitischer Gneise wahrscheinlich machen (Scheumann, 1961b).

Um diese Erkenntnis auch mit genetischen Verhältnissen anderer Granulite zu konfrontieren, wurden 1960 die finnisch-lappländischen Granulite studiert, die wir 1955 nur flüchtig kennengelernt hatten. Dass P. Eskola im Frankfurter Kolloquium 1959 die sedimentogene Abkunft der lappländischen Granulite nochmals ausführlich erörtert hatte, war die unmittelbare Veranlassung für uns, auch im lappländischen Komplex nach den vermuteten

[1] Eingegangen den 28. März, 1961.

migmatitischen Übergangsformen zwischen dem sedimentären Edukt und dem granulitischen Endprodukt zu forschen. [1]

Es zeigte sich, dass diese genetisch wichtige Vorphase in dem lappländischen Granulitgebiet nicht nur vorhanden ist, sondern sogar grosse Massenanteile dieses Areals umfasst, denen gegenüber Granulite im eigentlichen Sinne zurücktreten. Es ist also das Verhältnis zwischen den Granuliten *sensu stricto* und deren Vorformen ein anderes als bei den massigen mitteleuropäischen Granulitkörpern, bei denen die Granulite *s. str.* unbestritten die Vorherrschaft haben.

Hier muss überdies bemerkt werden, dass in Finnland seit Jernström (1874) der Begriff »Granulit» in einem erweiterten Sinne gebraucht wird, so dass er von der sonst üblichen Definition (vgl. Scheumann, 1961a) abweicht. Er umfasst auch gneisartig ausgebildete Formen, sofern sie Granat enthalten.

Th. Sahama diskutiert dies 1936 und äussert gewisse Bedenken gegen die Bezeichnung als Granulite. Er bleibt aber bei dem alt eingeführten Namen, da er von sich aus eine neue Bezeichnung nicht einführen möchte. Dabei betont er jedoch ausdrücklich, dass in seiner Arbeit der Begriff »Granulit» eine Aussage über strukturelle Eigenschaften der Gesteine nicht enthalten soll.

Unseres Erachtens sind nun aber die strukturellen Eigenschaften für den Gesteinsbegriff »Granulit» so wesentlich, dass wir für die finnischen Gesteine mit abweichenden Strukturverhältnissen diesen Ausdruck nicht empfehlen, sondern den Vorschlag machen möchten, sie nach C. E. Tilley als Granatgneise zu bezeichnen.

Tilley hat 1921 von der Eyre-Peninsula, Südaustralien, Paragneise beschrieben, die diesen grobkörnigen »Granuliten» in Bezug auf das Formbild entsprechen, und die auch in den Mineralkombinationen, sogar in der Zusammensetzung einzelner metabasitischer Einlagerungen mit ihnen übereinstimmen. Er hat die australischen Paragneise »Granatgneise» genannt.

In unseren Ausführungen werden wir den Hauptanteil der grobkörnigen finnischen Granulite ebenfalls als Granatgneise [2] bezeichnen. Mit dieser Bezeichnung werden im grossen ganzen auch solche Gesteinstypen umfasst, die bei Eskola (1952) zu den »field granulites» gestellt sind.

Da jetzt eine neue Spezialaufnahme Lapplands durch die Petrographen des finnischen Landesamts durchgeführt wird und unsere Kenntnis des grossen lappländischen Raumes nur unvollkommen ist, empfiehlt es sich für uns, die in Betracht kommenden Gesteine nur insofern petrographisch zu kennzeichnen, als es nötig ist, um auf genetische Züge in ihnen hinzuweisen, mit denen wir unsere Auffassung begründen können.

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[1] Die Durchführung dieser Arbeiten wurde uns durch die Deutsche Forschungsgemeinschaft ermöglicht, wofür wir unsern Dank aussprechen.

[2] Dieser Begriff enthält keine Aussage über die Zusammensetzung der Granate.

DER MIGMATITISCHE STRUKTURTYPUS DER GRANATFÜHRENDEN GNEISARTIGEN GESTEINE

Die Scholle der in Frage kommenden Gesteine bildet einen nach NO offenen Bogen und scheint in der allgemeinen Richtung nach SW auf das Kristallin des Vorlandes aufgeschoben zu sein, das im wesentlichen von mesozonalem Typus ist.

Die Grenzscheide zum Vorland ist keine glatte Trennungsfläche, sondern zeichnet sich durch unregelmässige Verschuppungen oder Verzahnungen aus. Wir übernehmen von den finnischen Geologen auch die Vorstellung, dass eine Zerbrechung des Granulit-Massivs in einzelne Teilschollen erfolgt ist (Mikkola und Sahama, 1936).

In einem bis 10 km breiten Streifen der äusseren Grenzzone des Granulitbogens spielen Granatamphibolite eine besondere Rolle (Sahama, 1936). Es ist vielleicht nicht ohne Bedeutung, dass in diesen amphibolitischen Metabasiten auch Olivingesteinsderivate, jetzt Serpentinite, in einzelnen linsenförmigen oder knollenartigen Akyrosomen auftreten (Mikkola und Sahama, 1936, S. 361 ff.; Sahama, 1936, S. 14). Diese Gesteinsassoziation der randlichen Metabasite mit »Ultrabasiten» erinnert an die Gesteinsgesellschaft der Grenzzonen der sächsischen Granulite und der Granulite des Waldviertels.

Manchmal gehen die Randamphibolite des lappländischen Gebietes durch Hinzutreten von Quarz und durch Verminderung des An-Gehaltes der Plagioklase in granatführende Hornblendegneise über, die mit den Amphiboliten und Granatamphiboliten wechsellagern.

Ein grosser Teil der Gesteine des lappländischen Granulitkomplexes selbst ist ein ziemlich grobkörniger Granatgneis mit auffällig grossen Granaten und einer sehr wechselnden Cordieritführung. Diese Granatgneise, namentlich die der inneren Zone des Granulitbogens, zeigen die Strukturverhältnisse von Migmatiten. Auf das Auftreten von migmatitischen Strukturbildern hatte schon Eskola (1952) hingewiesen und diese Tatsache auch in seiner Arbeit über Laumontit in Finnland (1960) nochmals erwähnt.

Wir haben das Material, das Prof. Eskola in seiner Arbeit von 1952 verwendet hat, im mineralogischen Institut in Helsinki studieren können. Für die grosse Bereitwilligkeit, mit der er uns seine Handstücke und Dünnschliffe dort zum Studium überliess, sowie für die leihweise Überlassung einer Auswahl dieses Materials durch den Direktor des Instituts, Prof. Saksela, sprechen wir beiden Herren unsern Dank aus. Ebenso bedanken wir uns für die wertvolle Unterstützung durch die Finnische Geologische Landesanstalt bei deren Leiter, Dr. Marmo.

Im allgemeinen sind bei diesen metatektischen Chorismiten (vgl. Scheumann, 1955) Metastere und Metatekte als distinkte Anteile unterscheidbar. Gelegentlich treten die metasteren Restbestände zurück und die granitähnlichen metatektischen Mobilisate überwiegen; es kommen auch Typen von metatektischem Charakter mit innigerer Durchmengung der hellen und dunklen Elemente zur Vorherrschaft.

In den chorismitischen Haupttypen aber sind die Metastere sehr auffällige dunkle Bestandmassen. Sie bilden wolkige Aggregate oder massige Klumpen, auch rohparallel lagig angeordnete, gestreckte oder geschwungene, durch die reiche Führung von grossen Granaten häufig knotige Partien. Zuweilen zerfallen sie in Einzelstücke oder teilen sich in Lagen auf, zwischen die sich das helle Material einschiebt. Manchmal verlieren sie sich als Granatreihungen in die metatektische Umgebung. Typen mit engerer Parallelscharung, in denen Züge dunkler und heller Elemente miteinander wechseln, leiten über zu Lagengneisen, deren alternierende helle und dunkle Lagen grobkörnig ineinandergreifen (Tafel I, Abb. 1—2; Tafel II, Abb. 1).

Die Granate bilden die auffälligste Komponente. Sie haben rundliche (idiomorphe bis subidiomorphe) Formen bis weit über Zentimetergrösse und sind häufig von hellen Gemengteilen durchbrochen oder schliessen solche ein. Wenn sie in Gruppen liegen, sind sie meist in eine quarzreiche Umgebung eingebettet.

Zuweilen ziehen sich Quarzaggregate in Streifen durch eine feinerkörnige Zwischenmasse. Diese enthält auch einzelne grössere Kalifeldspate, die manchmal ebenfalls in Quarz eingebettet sind.

Die Biotitführung wechselt stark. Häufig tritt der Biotit sehr zurück. Er ist in den Metasteren hauptsächlich auf die Granatränder beschränkt. In granitoiden Metatekten liegt der Biotit auch in Einzelkristallen im Quarzfeldspat-Mosaik.

In biotitreicheren Typen bilden metastere Biotitlagen Gleitfolien zwischen massigen Metatektpartien, die gegeneinander bewegt sind. Zuweilen entstehen auf diese Weise grobe Augengneise, in denen Biotitlagen sich zerteilen und grosse Granatlinsen umfliessen (vgl. Eskola, 1952, Abb. 7).

Auch im Gelände, das wir unter der dankenswerten Führung von Dr. K. Meriläinen besuchten, war das Bild des migmatitischen Verbandes für uns sehr eindrucksvoll. Die grösseren, nicht oder nur teilweise verdauten Reste des Altbestandes sind unregelmässige Schlieren, linsenförmige, gerundete oder sich ausspitzende Einschaltungen oder plattenförmige, z. T. boudinierte Lagen. Sie werden von der umgebenden Gneismasse in Einzelstücke aufgeteilt, durchadert und am Rande streifig aufgelöst. Vielfach sind in den durchgreifenden Metatekten grosse Porphyroblasten von Granat oder Cordierit (Tafel II, Abb. 2; Tafel III, Abb. 1). Häufig sind in den Metatekten weisse Mikrokline schwarmartig zusammengedrängt.

In diesen gemengten Gneisen treten Partien auf, die einigermassen homogen durchmengt sind und oft eine Paralleltextur zeigen (Tafel III, Abb. 2). Sie sind von etwa granitischer oder granodioritischer Zusammensetzung und führen manchmal Cordierite. Zuweilen sind die Mafite in ihnen auch Hypersthene, die gelegentlich biotitisiert sind. Die letztgenannten Typen werden in Finnland auch als Charnockite bezeichnet und werden, wie die Enderbite, als magmatische Bildungen betrachtet.

In Gängen und Adern kommen jüngere Granodiorite, Granite, Quarzsyenite und Pegmaplite vor, die die grobgemengten Gneise diskordant durchschneiden (Tafel IV, Abb. 1).

In den migmatitischen Gneisen liegen an manchen Stellen (z. B. an der Strasse zwischen Inari und Nellimö) dunkle schollen- und linsenförmige Akyrosome oder dicke Einzellagen von metabasitischem Typus. Es handelt sich hauptsächlich um amphibolitische und granatamphibolitische Einlagerungen mit grüner Hornblende und akzessorischem Titanit, Epidot und Biotit. Diese Amphibolite enthalten auch manchmal etwas Hypersthen.

Der ganzen Gesteinsgruppe sind, wie Eskola in Frankfurt ausführte, Züge der Amphibolitfazies eigen, wobei er ausdrücklich auf diese Einschaltungen von Amphiboliten mit grünen Hornblenden hinwies.

Die Führung von (z. T. pinitisiertem) Cordierit und von Biotit in den Gneisen selbst kann als ein weiteres mineralisches Kennzeichen einer amphibolitischen Mineralfazies betrachtet werden.

GRANULITISCHE ZÜGE IN DEN GRANATGNEISEN

Dass diese Gesteine trotz ihrer grob-chorismitischen Struktur auch als Granulite angesprochen werden, ist darauf zurückzuführen, dass sie neben einem auf die granulitische Fazies hinzielenden Mineralbestand in ihrem mikroskopischen Verband auch Strukturzüge enthalten, wie sie typischen Granuliten zukommen.

Häufig sind die Mafite der Metastere fast ausschliesslich Granate, und dem Biotit kommt nur die Rolle eines Nebengemengteils zu, der an die Ränder der Granate gebunden ist, so dass die Metasterpartien eine wirksame Gleitkomponente nicht abgeben. Sie sind vielmehr durch den Granatreichtum besonders unbildsam geworden und fungieren nicht in der Weise als Gleitmassen, wie es im überwiegenden Masse bei Migmatiten sonst zu beobachten ist.

Deshalb übernimmt in solchen Gneistypen mit vorwaltendem Granat und zurücktretendem Biotit vorerst die metatektische Quarzfeldspatmasse die Gleitfunktion, wobei der Quarz in der entstehenden Paralleltextur als Gleitkomponente in Erscheinung tritt. Einzelne grosse Quarze (oder Quarzaggregate) dehnen sich dabei zu langen, gebogenen Lamellen aus. In quer zur Parallelrichtung geführten Schnitten erscheinen sie als Quarzschlangen mit glatten (seltener mit körnelig gebuchteten) Rändern, die sich in schlanken Bögen um die Metastermassen und auch um einzelnde Granate winden. Auch kleinere Quarzindividuen des metatektischen Gesteinsanteils beginnen in kleinen Linsen zu fliessen (Tafel IV, Abb. 2; Tafel V, Abb. 1).

Die in Finnland als Noritgranulite bezeichneten Metabasite gleichen in ihrer Mineralzusammensetzung den mitteleuropäischen Pyroxengranuliten, wie sie auch in einzelnen Zwischenschaltungen der Eyre-Peninsula-Granatgneise vorkommen (Tilley, 1921).

Als granulitisches Charakteristikum können wir auch den von Eskola (1952) erwähnten »Antagonismus zwischen Lamellenquarz und Cordierit» ansehen. Die von ihm aufgefundene Ausnahme von diesem Antagonismus zeigt im Schliff Nr. 4396, den wir aus Helsinki mitbrachten, Cordierit neben Plattenquarz. Wir sind allerdings nach diesem Schliffbilde nicht überzeugt, dass der Cordierit die granulitische Deformationsphase hätte überstehen können; wir nehmen vielmehr an, dass er in dieser Umgebung zu den retrograden Bildungen gehört, die der Granulitbildung folgten (vgl. S. 133). Denn in dem pinitisierten Cordieritkristall in der Abb. 2, Tafel V ist eine Quarzlamelle deutlich eingeschlossen, was für eine Neubildung des Cordierits nach der granulitischen Deformation des Quarzes spricht.

Einer späteren Phase gehören ausser solchen Cordieritbildungen offenbar auch Umbildungen von Granat in Biotit, Sillimanit und Herzynit an, ebenso der gelegentlich zu beobachtende Zerfall der Quarzlamellen. Er geht mit einem Rearrangement der feinkörnigen Grundmasse parallel, wobei grössere Kalifeldspate metablastisch einwandern.

Die Sillimanitpakete sind zuweilen in langen Schnüren angeordnet und ihre Stengel manchmal mit Muscovit überzogen, in den sie pseudomorph übergehen.

Diese Muscovitbildung sowie die Pinitisierung der Cordierite kann man vermutlich als eine hydrothermale Mineraldiaphthorese betrachten, die der letzt-unterscheidbaren Durchaderung folgte.

Das Auftreten von Cordierit bei einer nachgranulitischen Durchaderung entspricht den vom sächsischen Granulitgebiet beschriebenen metatektischen Cordieritbildungen (Scheumann, 1961c). Dort ist der nachgranulitische Charakter solcher cordieritführenden Derivate an Bruchstücken mit erkennbaren granulitischen Strukturen und Resten der alten Mineralfazies einwandfrei nachweisbar.

In Lappland ist das Vorkommen von Cordierit in den überwiegenden Fällen allerdings anderer Natur als die nur an vereinzelten Stellen auftretende Cordieritführung in sächsischen Granulitderivaten. Im lappländischen Granulitgebiet gehört der Cordierit zum ursprünglichen Mineralbestand und stellt einen normalen Bestandteil der Granatgneise dar.

DIE GENESIS DER GRANULITE

Fassen wir alles zusammen, was sich aus den Mineralkombinationen ergibt, so müssen wir den Cordierit-Sillimanit-Granatgneisen eine Zwischenstellung zwischen der amphibolitischen Fazies (P. Eskola) oder Hochgneisfazies (I. Rosenquist) und der granulitischen zuweisen, weil in diesen Gesteinen metamorphe Minerale nebeneinander vorkommen, die sonst nur in der einen oder in der andern Fazies auftreten.

Es ist naheliegend, dies als Kennzeichen eines Wechsels aus einer Fazies in die andere aufzufassen, bei dem die Einstellung endgültiger Gleichgewichte nicht erreicht wird.

Dieser Wechsel kann sich in zwei Richtungen vollzogen haben, in einem progressiven oder in einem regressiven Gang der Metamorphose. Im letzten Falle wäre eine tiefergelegene (ältere) Granulitphase von einer höher gelegenen (jüngeren) Amphibolitphase abgelöst worden. Es wären dann in der amphibolitischen Phase Elemente der granulitischen mit verarbeitet worden, die in der amphibolitischen Phase reliktisch erhalten sein könnten.

Die demonstrativen Mineralisationen mit metatektisch gebildeten Biotiten, Cordieriten und Amphibolen, wie sie z. B. bei Törmänen zu beobachten sind, hatten uns 1955 bewogen, eine Folge Granulitfazies — Amphibolitfazies anzunehmen. (Für die Überlassung von reichlichem Untersuchungsmaterial sind wir Prof. A. Laitakari zu Dank verpflichtet.)

Unsere neuen Beobachtungen veranlassen uns aber, solche Erscheinungen, wie sie bei Törmänen vorliegen, nur als eine Schlussphase der Gesteinsumwandlung anzusehen. Die Extrapolation von diesen Phänomenen auf die metamorphe Vorgangsreihe des gesamten Komplexes ist unseres Erachtens nicht stichhaltig.

Vielmehr nehmen wir eine Verschiebungstendenz in umgekehrter Richtung an, eine Tendenz von der amphibolitischen nach der granulitischen Fazies hin, d. h. eine progressive Form der Metamorphose, die nur in einer subsequenten Umbildungsphase von einer rückläufigen Faziesänderung teilweise überdeckt wird.

Übergangsformen zwischen Gneisen und Granuliten *s. str.* finden sich hauptsächlich in stressbetonten Zügen etwa parallel der äusseren Randzone, aber auch im Streifen zwischen den Einzelschollen der übergeschobenen Masse.

In diesen Zonen stärkerer Gleitbewegung sind die Cordierit oder Sillimanit führenden Granatgneise vielfach zu gestriemten oder parallelplattigen Varianten ausgeschiefert. Diese plattigen Derivate können wir in eine Serie ordnen, in der sich eine immer stärker werdende Vorherrschaft der granulitischen Mineralfazies bemerkbar macht. Es können im engen Raume vergesellschaftet verschiedene Übergangsformen auftreten, die vermitteln zwischen den grobflaserigen migmatitischen Granatgneisen, die noch Züge der Amphibolitfazies aufweisen, und deren Myloniten oder Blastomyloniten mit Deformationsstrukturen, in denen die grossen gewundenen Quarzlamellen sich schon ausrichten. Diese leiten schliesslich zu echten Granuliten über mit straff eingeschlichteten Quarzlamellen. Als »echte Granulite» bezeichnen wie die typisch feinkörnigen, stromatitisch geschieferten Plattengesteine, aus denen alle amphibolitischen Züge verschwunden sind und die in Mineralbestand und Struktur den Granuliten *s. str.* entsprechen (Scheumann, 1961a). Sie dürften mit den »granulites proper» Eskolas identisch sein.

Wir können nach allem von der Granulitbildung in Lappland folgendes Bild skizzieren: Ein sedimentäres Ausgangsmaterial, das dunkle Einlagerungen oder Einschübe von diabasischen Typen enthalten haben mag, wie wir es auch für die sächsischen vorgranulitischen Sedimente annehmen, passierte bei einer orogenetischen Verlagerung zunächst die Zone der amphibolitischen Mineralfazies und nahm Mineralkombinationen dieser Fazies an. Es erreichte in der tiefsten Zone der katazonalen Umwandlung den Bildungsbereich der granulitischen Mineralfazies.

Die Ummineralisierung zu Gesteinen dieser tiefsten Fazies wurde durch die Grobkörnigkeit der Migmatite verzögert. Nur die sehr faziesempfindlichen Granate (Matthes, 1961) scheinen in dem Prozess der Granulitisierung vorausgeeilt zu sein, während z. B. die Hornblenden der Amphibolite eine grössere Resistenz bewiesen, so dass sie als Reste der amphibolitischen Mineralfazies erhalten blieben. Die Hypersthenbildung in ihnen, auch wenn sie häufig nur untergeordnet ist, deutet die Richtung der nachfolgenden Umbildung bereits an. Einige Metabasite erreichten in den Noritgranuliten schon den Mineralbestand der typischen Pyroxengranulite des mitteleuropäischen Raumes (mit Orthopyroxenen $\pm$ Klinopyroxen).

Die vollkommene Einformung der Migmatitgneise zu Granuliten bedurfte der aktivierenden Mitwirkung einer intensiven Durchbewegung, die sich in einer tiefliegenden Zone vollzog. In diesen Fliesszonen wurden die Reste der amphibolitischen Mineralfazies schliesslich ganz ausgelöscht; damit entstand nicht nur die saubere granulitische Mineralfazies, sondern es entwikkelte sich in dieser dynamischen Umformung schliesslich auch die Gefügefazies echter Granulite.

SUMMARY

In the granulite area of Lapland great parts of the rock association are formed by sedimentogeneous migmatic garnet gneisses with more or less cordierite. They contain simultaneously minerals or mineral combinations usually occurring either in the amphibolite facies or in the granulite facies. The amphibolite facies is indicated by green amphibole, cordierite, and bio-

tite; the granulite facies by pyrope-rich pyralspite, hypersthene, and hercynite.

This discrepancy in mineral facies suggests that the migmatic gneisses have crystallized under conditions changing from one facies into the other. We may imagine that the original sediments first submerged into the zone of amphibolitic characteristics and became transformed into gneissic migmatites. Sinking further as low as the depth of the granulitic conditions, they also acquired the fundamental features of the granulitic mineral facies.

In garnet-rich and mica-poor metasters of the migmatites, movements can hardly become effective on account of the rigidity of those parts. This they do, on the other hand, in the metatectic parts of such migmatites, where the plasticity of quartzes permits gliding deformations. The quartzes are lengthened into curved lamels, indicating the incipient formation of granulitic structures.

In certain spheres of granulitic depth, these rocks intermediate between the amphibolitic and granulitic facies have been strained by shearing movements. Here they graduate into true granulites, »granulites proper,» which have acquired the complete equilibrium of granulitic mineral facies.

It may be suggested that a shearing or flowing deformation may be an important factor well able to induce the stabilization of a homogeneous facies, straining the lattices of the minerals, and moving the mineral individuals against and into one another.

In addition to the granulitic mineral facies, the rocks also took on the characteristic stromatitic structure of true granulites: in the zones of deformation they grew fine-grained, and the quartzes straightened out into thin plates.

Comparing the genetical results obtained in the various European granulite areas, we find a striking difference between them.

In the rather small and conformant granulite areas of Central Europe, the specific conditions producing granulites have been efficacious enough to transform nearly all the migmatites into true granulites, leaving only some palimpsestic features of the antecedent phase.

In the enormous region over which granulites are scattered in Lapland, however, we can observe all graduations of this granulite-forming efficacy by means of the abundance of the antecedent phases of the granulites.

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ERLÄUTERUNGEN ZU DEN TAFELN

TAFEL I

Abb. 1. Grosses, etwas aufgelöstes Metasteraggregat mit viel Granat (grau), der z. T. von Biotit umsäumt ist. Sotajoki-Mündung, Handst. P. E. 7910, 9/10 nat. Gr.

Abb. 2. Granatreicher Metasterbestand, der sich in parallele Züge aufteilt. Schürfgrube b. Laanila. Handst. P. E. 801, 9/10 nat.Gr.

TAFEL II

Abb. 1. Lagiger Wechsel von Metaster und Metatekt. In den hellen Lagen einzelne zarte Biotitzüge. Kevojoki—Tal b. Kabbirmoras. Handst. P. E. 2945, nat. Gr.

Abb. 2. Unregelmässig gestaltete, schlierig ausgezogene, metatektisch durchaderte metabasitische Linse mit grossen Granaten. Der umgebende helle Gneis enthält ebenfalls grosse Granate. Anbruch an der Strasse Kaamanen—Karigasniemi sw d.Ahvenjärvi.

TAFEL III

Abb. 1. Cordierit und Granat führendes Metatekt durchgreift dunkle Lage ($^1/_2$ m mächtig), von der Bruchstücke unter Auflösungserscheinungen im hellen Material schwimmen. Anbruch an der Strasse Leutolahti—Kaamanen b. Pitkävuono.

Abb. 2. Fleckige Anordnung grosser Granate in einem gleichmässig gemengten massigen Cordierit-Sillimanit-Granatgneis mit leichter Paralleltextur. Block an der Strasse Leutolahti—Kaamanen b. Vuontisjärvi.

TAFEL IV

Abb. 1. Migmatitischer Bestand mit unregelmässigen metatektischen Lagen. Ein diskordanter Gang (etwa 10 cm breit) schneidet durch die migmatitische Textur. Anbruch an der Strasse Leutolahti—Kaamanen b. Pitkävuono.

Abb. 2. Anfang der Parallelausrichtung im Metatekt. Quarze z. T. ausgelängt, z. T. noch undeformiert, namentlich in den Zwischenräumen zwischen den grossen Kalifeldspaten und zwischen den Granaten. Leäksakoaddioaivi, Handst. P. E. 2908, Schl. 8959, Vergr. 6 mal.

TAFEL V

Abb. 1. Einschlichtung des Metatekts. Zwischen den grossen, z. T. durchbrochenen Granaten viel Quarz, der sich in langen Schlangen ordnet und die Granate z. T. umfliesst, z. T. sich an ihnen staut. Die feinkörnige Zwischenmasse des Metatektes (grau bestäubt) besteht überwiegend aus Feldspatkörnchen. Kevojoki—Tal b. Kabbirmoras. Handst. P. E. 2945, Schl. 8963, Vergr. 6 mal.

Abb. 2. Neben plattigen Quarzen pinitisierter Cordierit, in dem eine Quarzlamelle enthalten ist. Joukaisautsi, Schl. P. E. 4396, Vergr. 20 mal.

Abb. 1.

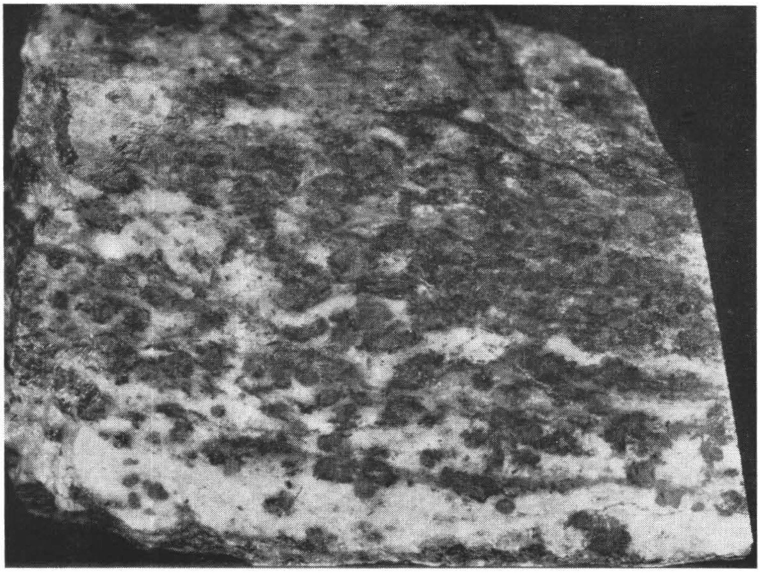
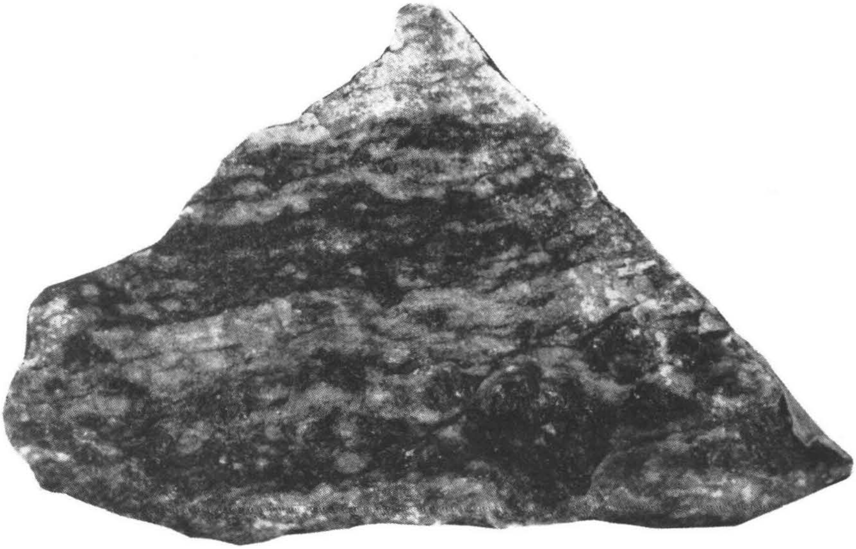


Abb. 2.

K. H. Scheumann, R. Bossdorf und Th. Bock: Versuch einer genetischen . . .



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K. H. Scheumann, R. Bossdorf und Th. Bock: Versuch einer genetischen . . .

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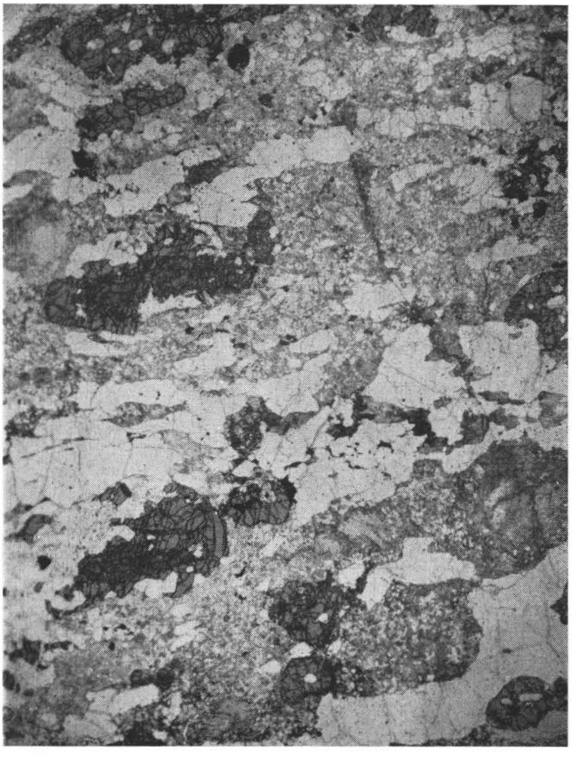

Abb. 2.

K. H. Scheumann, R. Bossdorf und Th. Bock: Versuch einer genetischen . . .

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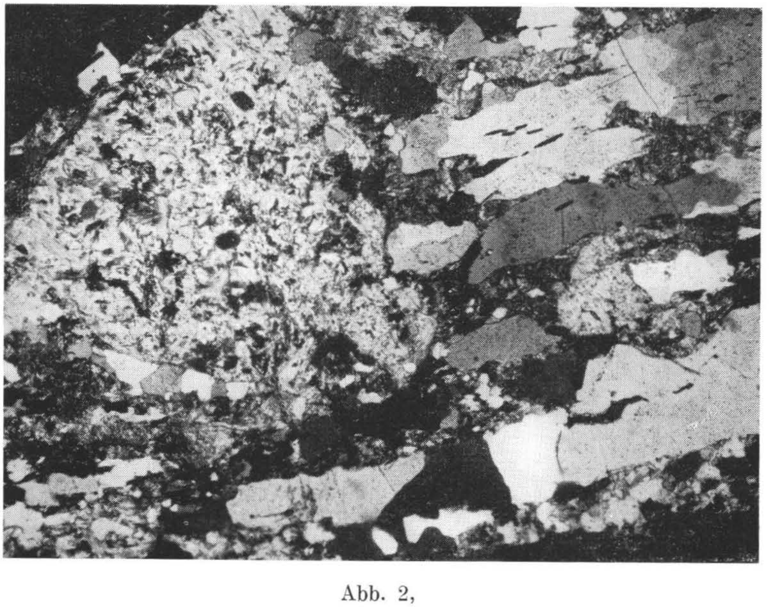

Abb. 2,

K. H. Scheumann, R. Bossdorf und Th. Bock: Versuch einer genetischen . . .

16

GEDANKEN ZUR DEUTUNG GEOLOGISCHEN GESCHEHENS [1]

VON

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ZUSAMMENFASSUNG

Nach Besprechung des Verhältnisses der Geologie zur Geophysik und der Bedeutung des menschlichen Faktors bei der Beurteilung der Probleme, werden einige Fragen der Gebirgs- und Wurzelbildung behandelt. Der Verfasser findet einen wesentlichen Kontrast zwischen den präkambrischen und jüngeren Gebirgen. Der Gedanke einer expandierenden Erde wird betont.

ABSTRACT

After an account on the relations between geology and geophysic and on the importance of the human factor in solving problems, some questions on the formation of mountains and roots are dealt with. The authors finds an essential difference between Precambrian and younger mountainchains. The idea of an expanding earth is pointed out.

RESUMÉ

D'après avoir décrit les relations entre la géologie et la géophysique et l'importance du facteur humain pour la solution des problèmes, quelques questions de la formation des chaînes de montagne et des racines sont traitées. L'auteur trouve des différences essentielles entre les chaînes précambriennes et les plus jeunes. L'idée d'une terre expansive est accentuée.

INHALT

[1] Eingegangen am 7. März, 1961.

EINLEITUNG

Wenn man von dem mineralogisch-petrographischen Rüstzeug des Geologens absieht, so muss man zugeben, dass unsere Wissenschaft sich grosser gedanklicher Freiheiten erfreut. Während die sogenannten exakten Wissenschaften mit ihren Themen und Ergebnissen streng begrenzten und gegebenen Wegen folgen, die von der Innen- und Aussenwelt der einzelnen Forscher unbeeinflusst bleiben, können geologische Probleme von verschiedenen Forschern sehr verschieden beurteilt und in Angriff genommen werden. Hierbei spielt der wissenschaftliche Lebensraum des Forschers und seine innere Veranlagungen eine entscheidende Rolle. In gewissem Sinne bedeutet dies eine Bereicherung der Geologie. Gewiss werden oft die Probleme nur von einer Seite beleuchtet, und gerne der Versuch gemacht, alle Tatsachen einer zuvor erdachten Ordnung der Dinge einzuordnen. Andererseits aber werden die hierdurch leicht entstehenden Differenzen der Ansichten zum befruchtendem Anstoss zur Erkenntniss der best möglichsten Erklärung.

Wenn ich nun hier von geologischem Geschehen schreibe, so meine ich damit die Entwicklung und die Verhältnisse im Grossen. Nicht die Herstellung geologischer Kartenblätter oder deren petrographische Beschreibung soll uns hier sonderlich interessieren, sondern wollen wir eine kurze Stunde bei den weiteren Zusammenhängen verweilen.

GEOLOGIE UND GEOPHYSIK

In den letzten Jahrzehnten haben geophysikalische Forschungen, soweit sie die Erdkruste betreffen, immer grössere Bedeutung für die Geologen bekommen. Dieser Umstand hat zu einer Wechselwirkung geführt, in deren Ablauf einerseits Geologen gerne mit geophysikalischen Daten arbeiten, und andererseits die Geophysiker geologische Begriffe übernommen und angewendet haben. Dieser gegenseitige Begriffsaustausch betrifft in der Hauptsache den Bau der Erdkruste und dessen Beziehung zum Mantel, sowie den Bau und die Entstehung der Faltengebirge. Eine weitere Problemgruppe umfasst die Entstehung der Kontinente und Ozeane.

Bei Verfolgung der oft recht divergierenden Meinungen auf diesem Gebiet habe ich oft stark empfunden, wie viel es noch an Verständnis zwischen den beiden Forschergruppen, den Geologen und Geophysikern, fehlt. Viele der oft phantastisch anmutenden Theorien und Hypothesen könnten bedeutend verbessert oder auch als irrelevant abgetan werden, wenn man auf beiden Seiten mehr von einander wüsste. Oberflächliche Kenntnisse müssen notgedrungen zu Fehldeutungen führen, welches natürlich niemals die Absicht gewesen ist. Vor Allem ist man sich wenig klar über die Begrenzungen, die den Verfahren des Gegenpartners anhaften. Man nimmt oft als gesicherte

Tatsache an, was ursprünglich von dem ausübenden Forscher nur mit Vorbehalt angenommen wurde, einem Vorbehalt, der vielleicht, weil er dem Fachmanne selbstverständlich, nicht ausdrücklich hervorgehoben wurde.

Es besteht ein grundsätzlicher Unterschied zwischen Geologie und Geophysik. Das Beobachtungsmaterial, welches dem Geologen zur Verfügung steht, ist stets an der Oberfläche gelegen und auf dieser punktförmig verteilt. Gelegenheiten zu Beobachtungen in der Teufe in Bergwerken und Bohrlöchern sind zu gering, um an dieser Tatsache etwas zu ändern. Das geologische Bild muss durch Inter- und Extrapolation gewonnen werden, sowohl was das Kartenbild wie auch das Profil betrifft. Die Genauigkeit hängt von der Dichte der Beobachtungspunkte ab. In vielen Gegenden der Erde liegen die Felsaufschlüsse sehr spärlich verteilt.

Die geophysikalischen Instrumente werden von den potentialen Kraftfeldern des Erdballs beeinflusst. Hierbei sind nicht die Verhältnisse im Messpunkt allein ausschlaggebend, sondern der gesammte Raum der Umgebung trägt zum Ergebnis bei. Wie weit der einwirkende Raum gestaltet ist hängt in erster Linie von Empfindlichkeit des Instruments ab. Bei gravimetrischen Messungen muss die Topographie der weiteren Umgebung in Rechnung gestellt werden, und bei seismologischen Beobachtungen kann, je nach Entfernung der Warte zum Epizentrum, der ganze Erdball von Bedeutung werden. Die geophysikalischen Messungen führen also zu flächenförmigen Ergebnissen. Die Konstruktion der Karten und etwaiger Profile ist also eine Aufgabe, die exakter gelöst werden kann, als das Interpolieren in der Geologie.

Weiter muss man sich auch darüber klar sein, dass der Geologe mit gänzlich anderen Einheiten arbeitet als der Geophysiker. Seine Bauelemente sind Gesteine, die neben bestimmten petrologischen Eigenschaften öfters auch paläontologisch-stratigraphische Charakteristika besitzen. Die räumliche Anordnung der Gesteine, die Mikro- und Makrotektonik, gibt dem Geologen das Muster, den Schlüssel zur Geohistorie. Der Geophysiker misst nur physikalische Qualitäten, die Magnetisierung, die Dichte oder die Elastik in der Kruste und die Verteilung dieser Eigenschaften, deren obige Auswahl ein Beispiel ist. Hierbei berücksichtigt er tunlichst die Verteilung sowohl der Oberfläche, wie auch der Teufe nach. Es ist nun *a priori* nicht gesagt, dass diese physikalischen Qualitäten mit bestimmten Bauelementen oder Gesteinen in Bezug gebracht werden können. Dass dies in gewissen Fällen doch möglich ist, und gewissen geologischen Horizonten bestimmte geophysikalische Eigenschaften zugeschrieben werden können, hat zu den zahlreichen Erfolgen geführt, die die Anwendung angewandter geophysikalischer Methoden in sowohl der praktischen wie auch wissenschatlichen Geologie immer wieder gezeitigt hat. Worauf es immer ankommt, ist das Vorhandensein genügender physikalischer Eigenschaftskontraste. Sind solche in genügendem Ausmass und von genügender Stärke vorhanden, so ist es möglich durch geophysikalische Methoden gewisse geologische Horizonte und Bauelemente zu verfolgen. In gewissen Fällen sind aber geophysikalisch feststellbare Leithorizonte nicht mit dem Geologen geläufigen Horizonten identisch. Ferner ist der Geophysiker öfters bereit, gewissen geophysikalisch erkennbaren Horizonten bestimmte petrologische Eigenschaften beizulegen, ohne dass es immer möglich ist, eine solche Korrelation durch direkte Beobachtung zu belegen.

Es erscheint also selbstverständlich, dass es auf diesem Gebiete zu Schwierigkeiten kommen kann, wenn es sich um die Deutung der Ergebnisse handelt. Zu einem nicht geringen Teil beruht dies aber immer noch auf dem oft herrschenden geringen Verständnis zwischen Geologen und Geophysikern, eine Tatsache, die zum grossen Teil auf der ganz getrennt verlaufenden Ausbildung der beiden Forschergruppen beruht. Hier wäre noch viel zu verbessern und zu gewinnen.

DER MENSCHLICHE FAKTOR

Ehe ich an meinen Faden weiterspinne, möchte ich noch auf folgenden Umstand aufmerksammachen. Es ist gewiss das Bestreben eines jeden Forschers sich der Objektivität zu befleissigen. Nur durch eine solche Haltung meint man, sich der absoluten Wahrheit nähern zu können. Man täte Unrecht, wenn man diese angestrebte Ehrlichkeit der Forscher verneinen wollte. Trotzdem muss man sich doch recht skeptisch allen solchen Bestreben, auch unseren eigenen, gegenüberstellen. Solange wir uns an der Oberfläche bewegen und Gesteine oder Versteinerungen zum Vorwurf unserer Betrachtungen machen, ist es durchaus möglich, dass wir uns innerhalb gegebener Grenzen der Objektivität hingeben können. Ich sage innerhalb gegebener Grenzen, denn wir sind wohl selten frei von einem gewissen gerichteten Interesse für unsere Aufgabe, das in unserem Unterbewusstsein leicht unser Handeln beeinflusst. Meine Behauptung sei durch ein Beispiel aus meiner Praxis beleuchtet. Es galt die technische Qualität eines Gesteinsvorkommens zu bestimmen. Ein Schurfgraben wurde ausgehoben. In diesem wurde auf jedem fünften Meter eine Handprobe gewonnen, die dann chemisch analysiert wurde. Das Ergebnis war absolut positiv. Eine erneute Probennahme, diesmal auf jedem Meter, gab schon ein viel schlechteres Ergebnis. Es wurde nun flach unter dem Schurfgraben ein Diamantbohrloch vorgetrieben und der gesammte Bohrkern zur Analyse gegeben. Das Ergebnis war rein negativ. Letzteres wurde durch eine technische Generalprobe, zu welcher der Schurfgraben in seiner gesammten Länge ausgesprengt wurde, voll bestätigt. Trotz des unbestrittenen Willens des Geologen zur Objektivität, wurde die erste Probenahme unzuverlässlich. Die Auswahl der zu analysierenden Proben war unbewusst durch den Wunsch, ein aufbaufähiges Vorkommen gefunden zu haben, beeinflusst. Ähnlich kann es aber auch leicht gehen, wenn es sich um Aufgaben handelt, wo die Probenahme ein auschlaggebender Faktor ist. Das gilt namentlich allen statistischen Untersuchungen, bei denen eine vorausgefasste Ansicht leicht das Resultat diktıert. Wenn es nun schon schwierig ist, an der Oberfläche objektiv zu bleiben, wie muss es umso schwerer werden, wenn wir uns in die Teufe begeben, von der wir nur indirekte und lückenhafte Vorstellungen haben können. Hier werden die bewusst oder unbewusst bereits vorhandenen Auffassungen gern eine Rolle spielen, die unsere Gedankengänge nicht unbeeinflusst lassen können. Dies wird umso mehr der Fall werden, wenn wir uns mit unbeobachtbaren Dingen befassen. Es besteht die grosse Gefahr, dass unsere geophysikalisch-geologischen Gedankenbilder uns in die bodenlose Metaphysik hinabirren lassen.

So hat jeder Gedankengang, jede Auffassung seinen gegebenen Hintergrund, der zwar oft völlig unbewusst und unerkannt unsere Auffassungen und Erkenntnisse des geologischen Geschehens beherrscht. Dieser Hintergrund ist stets im wissentschaftlichen Erleben des Forschers verankert. Das Vermögen zur Objektivität hängt davon ab, wie tief diese Verankerung realisiert ist. Objektivität und Intoleranz sind unvereinbar. Erkenntnis eigener Fehler ist eine absolute Forderung auf dem Wege zur Objektivität. Die Meister unserer Wissenschaft, die Gründer der Schulen, sollten sich stets dessen erinnern, dass auch sie dem menschlichen Faktor, ihrem Hintergrund zugeordnet sind. Absolute Dominanz gehört nicht zur Rüstung eines wahren Geistesfürsten.

KRUSTE, GEBIRGE UND WURZELN

Unsere Ansichten über den Aufbau der Erdkruste geht auch heute noch in vielen Teilen auf die Vorstellungen von Eduard Suess zurück, der mit seinem Antlitz der Erde die erste grossstilige Analyse unseres Erdballs unternahm. Ein Werk gleichen Masses ist seitdem nicht erschienen. Die Gegenüberstellung von leichtem Sial contra schweres Sima ist seitdem eine dauernde Vorstellung geblieben. Eine solche Dualität des Krustenmaterials war schon durch die Airysche Vorstellung vom isostatischen Ausgleich vorausgesagt. Das heutige Bild vom Bau der tieferen Kruste verdanken wir dem erdumspannenden Netz seismologischer Observatorien, unterstützt durch die ebenfalls weltumspannende gravimetrische Vermessung.

Nach diesen Forschungen haben wir mit einer horizontalen Gliederung der Erdkruste zu rechnen. Unter einer durchschnittlich 5 km mächtigen Decke von Sedimentgesteinen, die stellenweise bis 15 km Mächtigkeit anwachsen kann, haben wir den kristallinen Untergrund. Auf Grund der

Beobachtungen auf den riesigen sedimentfreien alten Beulen und Kernen wissen wir, dass granitische Gesteine einen hervorragenden Anteil im Aufbau haben, wenn auch nicht sedimentogene Schiefer fehlen. Der Anteil an simatischen Gesteinen ist verhältnismässig gering. Das Gesamtgepräge ist durchaus sialisch. Die Mächtigkeit dieses Siallagers, von den Geophysikern auch als Granitlager bezeichnet, ist nach den vorliegenden seismischen Daten 10—20 km. In Mitteleuropa wird es gegen die Teufe hin durch die Konrad-Diskontinuität begrenzt, an welcher eine Erhöhung der P-Wellengeschwindigkeit von 6—6.5 auf 7—7.5 km sek^{-1} stattfindet. Neuere Untersuchungen haben nun gezeigt, dass dieses granitische oder Siallager nicht unter den Ozeanen vorkommt, sondern rein auf die eigentlichen Kontinentalblöcke beschränkt bleibt. Die Untersuchungen von Hess, Worzel und Ewing haben deutlich gezeigt, dass unter den Ozeanen die Sedimente unmittelbar auf dem zweiten, unteren Lager der Erdkruste ruhen.

Dieses zweite Lager ist basischer Natur und wird im allgemeinen als basaltisch bis gabbroid (Piezogabbro) angesprochen. Es gehört also dem Sima an. Gegen die Teufe hin wird es von der seismischen Mohorovičić-Diskontiuität begrenzt. Diese liegt in recht wechselnder Tiefe, unter den Kontinenten im Durchschnitt bei 30—35 km und unter den Ozeanen bei 12—15 km. Unter den Gebirgen liegt die Moho-Fläche 50—60 km tief. Es bleibt also im Allgemeinen um 15 km Mächtigkeit für die Basaltschicht der unteren Kruste über der Moho-Fläche.

Die Mohofläche wird als die untere Begrenzung der Kruste gegen den Mantel angesehen. Unter derselben liegt der äussere Teil des Mantels, von welchem man im Allgemeinen annimmt, dass auch er basaltisch bis peridotitisch-eklogitisch zusammengesetzt ist. Diese Zone gehört also ebenfalls dem Sima an. Man hat simatisches Material über und unter der Moho-Fläche. Es wird doch angenommen, dass der Aggregatzustand über der Moho kristallin sei, während derselbe unterhalb der Diskontinuität glasig sei. Um der so entstehenden Zweideutigkeit der Nomenklatur zu entgehen, hat man vorgeschlagen, das Sima oberhalb der Moho als Salsima oder Sialma zu bezeichnen. Nach dem Vorschlag von Adams ist es jedoch besser, für den oberen Teil den alten Namen Sima beizubehalten und die Unterlage, das Substrat mit Ultrasima zu benennen.

Heiskanen unterscheidet nun zwischen der eigentlichen Moho-Kruste, d. h. die Kruste in dem Sinne, wie wir es oben dargelegt haben, und der Gesamtkruste, die stellenweise unter die Moho hinabreichen soll. Er geht dabei von den Berechnungen des isostatischen Ausgleichs aus. Nach seiner Deutung der Airy-Theorie müssen die Wurzeln der Gebirge seitlich durch Gegenwurzeln (Antiroots) unter den Ozeanen kompensiert werden. Diese Gegenwurzeln rechnet Heiskanen noch zur Erdkruste, deren Dicke er durchschnittlich auf 30 km berechnet. Für Heiskanen liegt also die Moho nur unter den Kontinenten zwischen Kruste und Mantel. Unter den Ozeanen liegt sie zwischen dem basaltischen Sima und dem Ultrasima. Ich werde unten auf diese Frage zurückkommen.

Die Kontinentalmassen sind also sialisch und bestehen aus alten präkambrischen Kernen mit angebauten jüngeren Faltengebirgen. Ich möchte schon hier einen Unterschied zwischen den alten präkambrischen Faltenbauen und den jüngeren Kettengebirgen machen. Es scheint mir trotz gewisser formalen Ähnlichkeiten recht grosse Gegensätze zwischen den beiden Typen zubestehen, besonders was die Dimensionen betrifft.

Die postkambrischen Faltengebirge sind stets ein zentrales Problem in der Geologie gewesen, wobei besonders die jüngsten, die Alpiden wegen ihrer guten Erhaltung eine besondere Rolle gespielt haben. Altmeister Suess fand den einfachen und effektiven Schlüssel zum Problem der Tektogenese durch die Kontraktionstheorie. Die tangentiellen Spannungen in der schrumpfenden Erdkruste gaben nicht nur die formale sondern auch die energetische Ursache der Gebirgsfaltung. Die Entdeckung der Deckenfalten der Alpen durch Lugeon führte zu der Analyse von Argand, nach der die Alpen die hochgepresste Tethys-Geosynklinale zwischen Afrika und Europa ist. Dabei wurde vorausgesetzt, dass die beiden Kontinente sich tangential genähert hätten. Argands Profet im Norden wurde Eugen Wegmann, der nach seinen Jahren hier sich getreulich bemüht hat, das alte und junge Gebirge geistig zu vereinen.

Der Deckenbau als Ergebnis reinen tangentialen Druckes sollte nicht lange unangefochten bleiben. Ampferer war der erste, der auf die Möglichkeit hinwies, dass Bewegungen im tiefen Untergrund die Ursache zur Tektogenesen sein könnte. Ernst Kraus baute diesen Gedanken weiter aus durch seine Narbentheorie, R. W. van Bemmelen sieht in Strömungen des Untergrundes die Ursache zur Störung des hydrostatischen Gleichgewichts und E. Haarmann sucht die Ursache zu den Strömungen des Untergrundes in kosmischen Einflüssen. Alle drei geben eine neue Analyse der Tektogenese. Wer je Gelegenheit gehabt hat diese Forscher über ihre Theorieen und Ansichten sprechen zu hören, der wird sich stets gerne dieser Stunden wahrer Freude und höchsten geistigen Genusses erinnern. Es ist hier nun weder Zeit noch Raum, die einzelnen Theorien näher zu besprechen sondern begnüge ich mich mit der Hervorhebung einiger spezieller Züge.

Nach der Airyschen Deutung der Isostasie müssen die Gebirge Wurzeln haben. Die Tiefe der Wurzeln ist eine Frage ersten Ranges. Nach den seismologischen Untersuchungen gehen die Daten für die Moho-Fläche auf 50—60 km hinab unter den Gebirgen. Man kann also mit Wurzeln in dieser Grössenordnung rechnen. Heiskanen und seine Mitarbeiter haben den isostatischen Ausgleich durch die Gebirgswurzeln aus gravimetrischen Daten berechnet. Die Annahme ist, dass die Kruste durchweg, also jede Gliederung unbeachtet, eine Dichte von 2.67 habe und auf einem Substrat, dem Mantel, mit einer Dichte von 3.27 ruhe. Auf diese Weise kommt Heiskanen bei einer normalen Krustenmächtigkeit von 30 km auf eine Wurzelmächtigkeit von 26.7 km. Das bedeutet eine Wurzeltiefe von 56.7 km. Es sei nun zunächst darauf hingewiesen, dass alle gravimetrischen, wie auch magnetischen Ergebnisse der exakten Tiefenwirkung entbehren. Für jede Kurve aus einer Serie Messresultaten zusammengesetzt gibt es an und für sich unendlich viele mathematische Lösungen. Wenn man also trotzdem derartige Messungen zu Tiefenbestimmungen ausnutzt, so geschieht dies nur mit Hilfe gewisser Annahmen, deren Wahrscheinlichkeitswert nicht immer im Bereich unserer Kontrolle liegt. So sind auch die von Heiskanen gemachten Annahmen nicht unwidersprochen geblieben. Dass seine Ergebnisse trotzdem bewundernswert gut in unser Bild von dem Bau der Gebirge passen, beruht letzten Endes wohl darauf, dass Heiskanen und seine Schüler den allgemeinen Hintergrund erkannt haben und ihre Annahmen geschickt diesem inneren Umstand haben angleichen können. Ich will mit dieser Aussage durchaus nicht die Grösse und Bedeutung Heiskanens und seiner Schule verringern. Das Werk wird immer den Meister rühmen. Es sei nur darauf hingewiesen, dass es im Falle eines anderen allgemeinen geologisch-geophysikalischem Hintergrundes durchaus möglich wäre, die entsprechenden Annahmen zu finden, die demselben entsprechen würden. Mit dieser Tatsache müssen wir uns abfinden.

Wenn man versucht die verschiedenen Theorien und Annahmen auf einen Nenner zu bringen, so kommt man zu folgendem Gesammtbild, dessen Hauptzüge überall wiederzuerkennen sind. Die Kruste besteht innerhalb der Kontinente aus einer von wenig mächtigen Sedimenten bedeckten Siallage, der Granitischen Schicht der Geophysiker und den alten kristallinen Kernen der Geologen. Unter dieser etwa 15—20 km dicken Lage folgt der simatische Anteil der Kruste, die basaltische Schicht, die sich auch als dünnes Lager unter den Ozeanen befindet. Die untere Grenze dieser basaltischen Lage gegen das glasige Ultrasima ist die Moho-Diskordanz. Das Ultrasima befindet sich offenbar in einem mobileren Aggregatzustand, der Bewegungen grösseren Ausmasses erlaubt. Es herrscht eine gewisse Schwäche in dieser Asthenosphäre. Schon der Gedanke Airys von der Isostasie setzt voraus, dass das Absinken der Gebirgswurzeln seitlich durch ein Emporsteigen des Ultrasimas kompensiert wird. Die neueren Theorien vom Gebirgsfaltungsvorgang setzen direkt Bewegungen im Ultrasima voraus. Wie diese Bewegungen im einzelnen vorgestellt werden ist für unsere Zwecke zunächst minder von Bedeutung.

Die für die Tiefe der Wurzeln von Heiskanen gegebene Lösung fusst auf der Annahme eines einheitlichen Gewichts der gesammten Kruste. Die seismologischen Daten sprechen jedoch für eine Gliederung der Kruste in zumindestens eine granitische und eine basaltische Schicht, oder in Sial und kristallines Sima. Wenn man die die üblichen Zusammenstellungen über die Natur der Gebirgswurzeln verfolgt, findet man zumeist, dass unter den Gebirgen der Zuwachs an Mächtigkeit hauptsächlich auf das Sima fällt. Danach müssten also die Wurzeln simatischer Natur sein. Es ist dies natürlich auch nur eine Annahme, deren Wahrscheinlichkeit bislang ausserhalb des Bereichs des Beweisbaren liegen muss.

Nach unseren Vorstellungen von den Geosynklinalen als Geburtsstätten der Tektogene werden relativ mächtige Sedimentstapel in diesen unter Absinken befindlichen Räumen aufgetürmt. Die Mächtigkeit dieser Ansammlungen hat die Grössenordnung von 12—18 km. Damit eine solche Folge sich auf einem sinkendem Untergrund aufstapeln kann, wird vorausgesetzt, dass ebenderselbe Untergrund gegen die Teufe hin ausweichen kann. Wie war nun dieser Untergrund beschaffen?

Es sind hier anscheinend prinzipiell zwei Möglichkeiten vorhanden. Einmal kann man annehmen, dass die Sialschicht den weichenden Untergrund ausmachte. Im Falle einer Geosynklinale des Cordilliera-Typs im Sinne van Bemmelens wäre es also der Kontinentalrand, welcher durch die Belastung hin abgedrückt wird. Im Falle einer Geosynklinale des indonesischen Typus ist die sinkende Sialpartie zwischen zwei Sialhochs gelegen. Andererseits könnte man auch mit Sima als Unterlage am Rander der Sialschollen rechnen. Dieser Gedanke scheint jedoch bisher nicht beachtet zu sein.

Ist die Unterlage der Geosynklinale sialisch, so muss die Sialschicht ja tief hinabgedrückt werden und käme in eine Tiefenlage von 40—45 km unter der Geosynklinale. Dann bleibt für das Sima nur noch 10—15 km in den Wurzeln. Unter solchen Umständen würde also eine Wurzel schon von Beginn an mehr sialisch als simatisch sein.

Nun wird stets angenommen, dass die Abwanderung sowohl der Sedimentpakete als auch des Sials in den Wurzeln mit durchgreifenden Vorgängen der Metamorphose und Metasomatose verbunden war. Die Anschauungen von Kraus und van Bemmelen setzen überhaupt derartige Umsetzungen des Materials als aktive Kräfte der Tektogenese ein. Eine Differentierung des absinkenden Materials in Neosial, welches nach oben steigend den sialischen Teil der Wurzeln verstärkt, und in Neosima, welches in die Tiefe absinkt, ist zum grossen Teil verantwortlich für den Auftrieb des Tektogens.

Die hier erwähnten Vorstellungen von der Differentiation in Neosial und Neosima unter dem aufsteigenden Tektogen haben, trotz deren spekulativer Natur, eine anregende Bedeutung für weitere Probleme der kristallinen Schiefer. Die erste Differentation grossen Stiles muss in den ältesten Zeiten unserer Erde stattgefunden haben, deren Ergebnis eben die Scheidung von Sial und Sima vom Ultrasima war. Durch den Zyklus der Geosynklinalen und Tektognesen wird dieser Vorgang sozusagen wiederholt. Nicht fertig differenziertes Material wird weiter veredelt. Druck und Temperatur lockern die chemischen und kristallographischen Bindungen, die Schwere differentiert das Material weiter. Ich habe einmal früher solche Prozesse als metamorphe Segregation bezeichnet. Das ansteigende Neosial wird in der Hauptsache granitischer und granodioritscher Natur sein. Auf seinem Wege gegen Oben wird es die umgebenden Massen, auch die tiefhinabgedrückten Sedimente infektieren. Ob wir diesen Prozess in durchgreifender Metasomatose mit Stoffzufuhr, hauptsächlich an Alkalien, oder in sogenannten Granitisierungserscheinungen abgebildet finden beruht sicher auf lokalen und regionalen Umständen. Granite scheinen jedenfalls das Hauptprodukt dieser Differentiation zu sein. Ich bin sehr geneigt, die Granite mit der Bildung von Neosial in Verbindung zu stellen. Sie wären dann in gewissem Sinne Produkte eines zeitlich unbegrentzten Differentiationsprozesses nach der Schwere.

Ehe wir die Wurzeln verlassen sei noch einiges über die Gegenwurzeln in Sinne von Heiskanen gesagt. Ich will nicht die Darstellung bei unserm prominenten Geophysiker an sich beanstanden. Es sei jedoch darauf hingewiesen, dass ein isostatischer Aufstieg des Ultrasimas vielleicht garnicht eine ultrasimatische Kruste als solche erfordert. In diesem Falle wäre die Kruste natürlich nicht vollständig über den Erdkörper geschlossen, eine Möglichkeit die gewisse geologische Konsequenzen zuliesse, auf die ich hier nicht näher eingehen will.

PRÄKAMBRISCHE GEBIRGE

Es ist die allgemeine Auffassung, dass die präkambrischen Tektogene die durch den isostatischen Ausgleich zu Tage geförderten Wurzeln uralter Gebirge darstellen. Der metamorphe Gesteinsinhalt spricht seine deutliche Sprache. Die Mineralparagenesen sind das Produkt hoher Druck- und Temperaturverhältnisse. Wenn auch retrograder Metamorphismus in gewissem Sinne eine sekundäre Umprägung gestattet hat, so finden wir doch zumeist die eingefrorenen Ungleichgewichte tieferer Faziesverhältnisse im Sinne von Eskola. Von petrologischen Standpunkt aus ist also die Annahme, es lägen hochgetriebenen Wurzeln vor, wohl berechtigt.

Der petrologische Inhalt der präkambrischen Gebirge ist jedoch nur ein Teil des Problems. Wesentlich ist vor Allem die Ausdehnung der präkambrischen Tektogene. Früher sprach man von der ubiquitären präkambrischen Faltung. Dies beruhte auf der Beobachtung, dass das Kristallin der alten Schollen überall stärkstens durchgefaltet war. Heute können wir zwar auch im Präkambrium verschieden alte, räumlich umgrenzte Tektogene unterscheiden, aber die Tatsache, dass diese ausserordentlich weiträumig die primären Schollen durchziehen bleibt auch heute noch eine beachtliche Tatsache.

Nach den heute geltenden Annahmen sind die Wurzeln der jungen Gebirge verhältnismässig schmale Zonen, die jedenfalls schmäler sind als die oberflächlichen Faltungsräume. Bei der ausserordentlichen Breite der präkambrischen Tektogene, für die Svekofenniden-Kareliden können bald 1 000 km veranschlagt werden, stellt man sich die Frage, wie müssen die Gebirge des Oberbaus bemessen gewesen sein? Sie müssten im Vergleich zu den heutigen Gebirgen Riesen gewesen sein.

Soweit wir heute die Tektonik der primären Gebirge überblicken, finden wir viel flache Faltung, oft an die Kofferfalten des Juras erinnernd, oft auch zu isoklinalen Serien mit flacher Achsenlage zusammengedrückt. Die weitgehende Granitisierung scheint oft wenig an den tektonischen Verhältnissen geändert zu haben. Nur die postorogenen Granite, oft mit Rapakivistruktur, durchschlagen die älteren Formen.

Der Mineralbestand der präkambrischen kristallinen Schiefer deutet auf durchaus sialische Verhältnisse, d. h. eine Tiefe um 10—15 km und Temperaturen, die nur selten sich 1 000° genähert haben dürften, in vielen Fällen aber bedeutend niedriger gelegen haben. Wenn wir dazu die räumliche Ausdehnung berücksichtigen, so müssen doch erhebliche Zweifel an der üblichen Deutung als Gebirgswurzeln entstehen.

Diese Meinung ist im gewissem Sinne eine Abweichung von der aktualistischen Auffassung, die unsere Wissenschaft beherrscht. Gewiss können wir feststellen, dass der sedimentogene Anteil der primären Gebirgsbaue nach uniformitaristischen Anschauungen verstanden werden muss, ebenso die vulkanitischen Bildungen. Was aber den damaligen Tiefgang der Gebirge betrifft, erscheint mir doch Alles für abweichende Verhältnisse zu sprechen. Ich sehe in den Tektogenen der alten Kerne keine Wurzeln, sondern weiträumige Faltungen mit relativ geringen Tiefgang.

Eine der ausgeprägtesten Diskordanzen in der Geohistorie ist zwischen Kambrium und Präkambrium gelegen. In den meisten Gegenden haben wir hier nicht nur eine tektonische, sondern auch eine zeitliche Lücke von grossem Ausmass. In gewissen Gegenden handelt es sich 800—1 000 Millionen Jahre. Oberhalb dieser Diskordanz haben wir teils völlig unkonsolidierte Sedimente mit reicher fossiler Fauna, unterhalb Kristallin mit spärlichen organischen Resten. Es gibt wohl keinen grösseren Gegensatz in der Geologie. Es erscheint mir daher durchaus nicht unberechtigt, wenn man für das Präkambrium einen etwas anderen Massstab anwendet als für die jüngeren Epochen. Einmal nähern wir uns zeitlich dem Urzustand unseres Planeten, und andererseits müssen wir damit rechnen, dass die allgemeinen geophysikalischen Verhältnisse nicht dieselben waren wie heute. Derartige Gedanken sind nicht fremd. So hat z. B. von Eckermann für die Zeit der

Entstehung der Rapakivigranite einen steileren Temperaturgradienten angenommen als der, der heute herrscht. Schon allein ein solcher Umstand muss sich in den ältesten Zeiten noch mehr ausgewirkt haben. Dies sei nur als ein Beispiel erwähnt. Manche andere Faktoren werden aber auch in präkambrischer Zeit nach einem von dem heutigen verschiedenen Massstabe gewirkt haben.

Man hat in den alten Schilden die Reste der ersten Sialbildung der Kruste. Das älteste Sial wird überwiegend granitisch gewesen sein, entstanden durch die Grossdifferentiation der Schwere nach. Dieser Prozess führte allmählich zur Bildung wahrer Sedimente, die aber auf der wenig dicken Sialkruste weiträumig gefaltet wurden unter dauerndem Anstieg sialischen Materials, teils als magmatische Massen, teils in metasomatischen Prozessen. Es scheint als ob dieses Geschehen, welches in einer allmählichen Versteifung der Schollen resultierte, zu Ende des Präkambrium allmählich abklang. Danach trat die Erde in ein zweites Stadium ein, das durch die Bildung der eigentlichen Geosynklinalen an den Rändern der älteren Kerne ausgezeichnet ist.

DIE GEOFRAKTUREN

Die älteren Schilde sind durchwoben von Bruchspalten, deren Dichte in Bruchteilen von Metern und im Meilen gemessen werden können. Die meisten dieser Spalten sind immer und immer wieder in Bewegung gesetzt worden. Das isostatische Sinken und Steigen des fennoskandischen Raumes unter und nach der letzten Eiszeit ist durch die Differentialbewegungen längs solcher Spalten zu verstehen. Ein grosser Teil dieser Spalten finden wir auch in den Küstenlinien wieder. Der bottnische Meerbusen, in welchem jotnischer Sandstein, nach T. Mikkola die Molasse der Svekofenniden, unter kambrischen Sedimenten bewahrt liegt, ist sicher schon präkambrisch angelegt. Oft werden derartige Bruchspalten als verhältnismässig junge Erscheinungen angesehen. Sollte man sich die Verhältnisse ein Mal systematisch vom Standpunkt der gesammten Erde ansehen, so erscheint es mir glaubhaft, dass man zu den jüngeren Spalten stets Parallelen in den alten Schildern finden wird. Man kann dann aber nicht umhin weiter zu untersuchen, ob nicht auch die jungen Spalten nur wiederbelebte solche höchsten Alters sein können. Dieser Frage ist wenig Aufmerksamkeit geschenkt worden.

FIXISMUS UND MOBILISMUS

Man findet in der Gedankenwelt der Geologie heute zwei wichtige Prinzipien. Während man seit Alters her sich der Vorstellung hingibt, dass die alten Schilder in ihrer Lage eine Konstanz aufweisen, kann man sich doch nicht von gegenseitigen horizontalen Verschiebungen im geosynklinalen Raum frei machen. So setzt die Analyse der Alpentektogenese ein Wandern Afrikas gegen Europa voraus. Die Theorie von den Wanderungen der Kontinente von Wegener geht in dieser Beziehung noch viel weiter. Im ersten Falle muss man mit dem Absinken von Landbrücken und auch deren gelegentliches Auftauchen rechnen, die es ermöglichten die paläobiologischen Wanderungsfragen zu lösen. Typisch für die fixistische Auffassungen sind die z. B. die paläogeographischen Karten bei Koken oder Arldt. Aber auch die meisten ganz modernen paläogeographischen Karten, z. B. die von Termier et Termier zeigen deutliche Züge eines, ich möchte sagen gezügelten, Fixismus. Die oben erwähnten modernen tektogenetischen Theorien von Kraus und van Bemmelen stehen sich in der Frage Fixismus oder Mobilismus fremd gegenüber. Während Kraus für seine Unterströmungstektogenese unbedingt mobilistische Ansichten haben muss, kann van Bemmelen für seine hydrostatische Tektogenese nur den Fixismus gebrauchen.

In dieser Frage besteht noch viel Unsicherheit. Die so interessante Theorie von Wegener, auch in der Umarbeitung von Gutenberg, ist hauptsächlich auf die Verhältnisse um den Atlantik aufgebaut. Man stellt sich aber gerne die Frage, was geschah ausserhalb dieses Raumes. Wie verhielten sich die doch nicht ganz unbedeutenden Sialmassen Eurasiens? Und wie wirkte sich das Ganze auf den Pazifik aus? Es ist klar, dass die Verhältnisse im Karbon um den Südatlantik nicht verstanden werden können, wenn man nicht gleichzeitig die Tektogenese der Variskiden in Europa und Asien, das Schicksal der Thetys und die Geschehnisse um den Pazifik berücksichtigt.

Für die Frage des Fixismus contra Mobilismus ist auch unsere Auffassung von Veränderung der Grösse der Erde, und damit ihrer Oberfläche von entscheidender Bedeutung. Wir können entweder mit einer schrumpfenden oder einer expandierenden Erde rechnen, oder gar mit völlig konstanten Verhältnissen.

Rechnet man mit einer schrumpfenden Erde, so kann die gegenseitige Annäherung der Urschollen längs interkontinentaler Geosynklinalen verstehen. Weniger verständlich bleiben die circumkontinentalen Geosynklinalen. Gänzlich unerklärt bleiben aber die Geofrakturen. Die Vorstellung des Zerberstens der einsinken Kruste ist mechanisch recht einfach, das etwaige Einsinken der Kruste aber ein viel zu komplizierter Vorgang, um auf einen so einfachen Nenner gebracht werden zu können. Gewiss entstehen Sprünge und Risse in schrumpfenden Körpern, das sehen wir an erkaltenden Magmagesteinen. Die grossen Geofrakturen sind aber eher Ergebnisse einer Dehnung. Auch in den Erklärungsversuchen der Tektogenese wird heute viel von Dehnungen gesprochen. Dehnungen können nun natürlich auch entstehen, wenn sich zwei Schollen einander nähern. Das Gesamtmuster der Frakturen spricht doch nicht für eine derartige Erklärung, namentlich nicht die Spalten, die senkrecht zu der Zusammenschubrichtung der Geosynklinalen verlaufen.

Lehnt man die Schrumpfungstheorie ab und rechnet mit konstanten Verhältnissen, so kommt der Mobilismus wieder mehr in den Vordergrund, denn eine Beweglichkeit der Urschollen im horizontalem Sinne kann sowohl Tektogenese wie auch die Frakturen erklären. Trotzdem erscheint diese Erklärung wenig befriedigend. Die Vorstellung, dass die grossen Urschollen konstante Einheiten im paläogeographischen Bilde und auch ihrer Lage nach beharrlich sind, leuchtet auch heute noch überall durch, wenn man von den sozusagen extremistischen Ansichten von der Verschiebung der Kontinente im Sinne von Wegener absieht. Wenn dieses Gedankenbild zurecht besteht, so kann man aber auch dasselbe gegen den Hintergrund einer expandierenden Erde betrachten. Ich glaube, dass dies ein Gedanke ist, an den wir uns immer mehr gewöhnen müssen. Die Darlegungen des ungarischen Geophysikers Egyed von dem expandierenden Erdkörper sind sehr beachtungswert. Das von Egyed entworfene Bild eines neuen Erdmodells kann nicht nur Dehnungen in der Erdkruste erklären, sondern gibt auch die nötige Energiequelle an. In dieser Richtung können wir die Brücke zwischen Fixismus und Mobilismus finden.

SCHLUSSBEMERKUNGEN

Durch die obige Darstellung habe ich beabsichtigt mit einigen Streiflichtern gewisse Probleme der Geologie, hier im Sinne wahrer Erdeerforschung zu beleuchten. Gewiss können die in der Einleitung erwähnten Begrenzungen durch den persönlichen Hintergrund, den menschlichen Faktor, auch auf meine Darstellung der Probleme geltend gemacht werden. Der persönliche Hintergrund kann aber in gewissem Sinne auch zu einem positiven Bilde beitragen. Während der langen Jahre, während welcher ich den Vorteil hatte Mitglied unserer Gesellschaft zu sein, haben sich die interessanten Vorträge und Schriften selten mit Fragen des Erdaufbaus beschäftigt. Es kann daher vielleicht berechtigt sein, jetzt zur 75-Jahrfeier, einen Ausblick auf die so mannigfaltigen Probleme und Fragen, deren schliessliche Lösung noch in der Ferne liegen mag, zu geben.

17

THIRTY YEARS' DISCUSSIONS ON THE ORIGIN OF THE LATE SVECOFENNIAN VEINED GNEISSES IN CENTRAL SWEDEN [1]

BY

NILS H. MAGNUSSON

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In the year 1929 the author began investigations in the Kantorp and Stav mines. The dominating ores there are the so-called pegmatite-iron ores of the Kantorp type. These were originally quartzbanded ores of the type common to the ore-bearing region of Central Sweden, but they have later been richly impregnated with feldspar and mica, while the quartz has been driven out and enriched in other parts of the ore masses. In these parts even magnesia has been enriched. Such ores are the quartz-anthophyllite ores of Kantorp. Where limestones, dolomites and larger skarn masses were present, new skarn minerals have been formed through reactions between this material and the pegmatite solutions. These ores therefore show a chaotic mixture of older minerals and new ones and are called skarn-pegmatite ores.

It was quite clear from the very beginning of my investigations in Kantorp and Stav that the alterations of the quartz-banded iron ores in Kantorp were caused by the same processes that had altered the surrounding rocks, the leptites and slates to coarse-grained veined gneisses. The author therefore began to investigate and map the Kantorp region, and in 1932 expressed the opinion that the iron ores of Kantorp and Stav and the surrounding rocks originally had the same appearance as the corresponding ores and rocks in the best preserved parts of the ore-bearing region of Central Sweden. It was also considered that they had undergone the same alterations as these ores and rocks in connection with the first Svecofennian folding and the intrusion

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of the old synkinematic granites before the alteration that gave the whole ore-bearing complex its present coarse, inhomogeneous texture.

That the Stockholm granites are associated with the pegmatitization in the Stockholm region was suggested by Holmquist as early as 1907. The author came to the same conclusion in 1932 and expressed the opinion that new regional investigations of the veined gneisses in the county of Sörmland certainly should prove that younger granites of the Fellingsbro and Stockholm series play a great role in this region and that their co-operation in the processes that gave us the veined gneisses was essential. This does not mean that all the pegmatites appearing in the gneisses came from the young granites. Rather must the material coming from them have been subordinate in comparison with the material that has, through their instigation, been mobilized in the older rocks themselves. This was made quite clear by the fact that the young granites are rich in microcline while in the largest part of the complex white pegmatites rich in plagioclase dominate.

In describing the bedrock in the Kantorp region the author (1936) expressed his opinion as follows:

»The alteration of the Kantorp rocks to veined gneisses was caused by a regional sinking, through which large parts of Central Sweden came into the deep earth zones where the palingenetic processes are predominant. This sinking was caused either by orogenetic or by epeirogenetic movements. Several geologists assume them to be orogenetic but I think epeirogenetic movements to be more probable. We have not found in the non-gneissic region surrounding the area of the veined gneisses such tectonic features that could be expected if the sinking was caused by an intense folding. What we can state is that the parallel structures are always very steep (80—90°) except in local folds of the layers where the axes usually dip gently (20—30°). The steep dips of the parallel structures were caused by a strong tangential pressure, not by a new folding. I therefore think that the foldings are much older and that the strong pressure only caused a compression of the complex in one direction and a stretching out in the other. The rock bodies existing before the alteration into veined gneisses were thus deformed and more elongated.

»The deformation was in some degree plastic in detail. We find, for instance, the greenstone dikes sometimes curved and bent. Small S-shaped curves are found here and there, but these studies of details do not force us to assume more intensive foldings of the whole complex of the veined gneisses.

»The alteration to veined gneisses was caused by the high temperature in connection with the compressional forces and by emanations and solutions from deeper zones of the earth's crust, where the palingenetic processes were stronger.

»The veined gneisses are characterized by an inhomogeneous structure with coarser pegmatitic veins. Here and there appear small gneissic but homogeneous fragments of the original rocks, which have escaped the stronger alteration. In the transition zones between the veined gneisses and the central part of the ore-bearing region in Central Sweden, where normal leptites, graywackes and slates are the characteristic rocks, one can follow the alteration and see the pegmatitic parts develop as small spots or strings, which become more and more numerous. In the veined gneisses the pegmatitic veins dominate the rocks. The whole rocks are usually recrystallized and pegmatitic. We can therefore also call the alteration process a pegmatitization. Among the homogeneous fragments in the veined gneisses of the Kantorp district we have leptites, slates, old granites, quartz-banded iron ores, skarn iron ores, and limestones. Among them the slates always seem to be most easily pegmatitized. Also the leptites are much more easily altered than the old granites which from the very beginning were relatively coarsegrained. Among the leptites the types rich in CaO are altered in a greater degree than those poor in CaO and the banded types more than the more massive ones.

»The alteration process is to be considered as a metamorphic differentiation. The most soluble or fusible constituents have been segregated in the form of more or less irregular spots and veins and bands. That these spots and veins and bands received pegmatitic structure depended upon the emanations and solutions that must have soaked through the whole complex at the beginning of the alteration process. We have excellent proofs for this opinion in the chemical alterations that can be proved. Microscopical investigations and comparative studies of nineteen analyses from the district have shown that both the slates, the leptites, and even the old granites have been altered in such a way that an enrichment of the stablest constituents and a reduction of the most soluble substances have taken place. We find a reduction of silica, alkalies and calcium and an increase of aluminium, iron and magnesium. As this alteration also has occurred in the slates, the increase of the last-named elements can not be explained by assuming transport of material from the slates. In my opinion it is necessary to assume granitic emanations and solutions soaking their way through the strongly schistose leptite-slate-old granite complex with its iron ores and limestones.

»The veins which can be regarded as metamorphic differentiates from the rocks themselves show transitions to more independent, larger, intrusive pegmatites, and from the latter to the younger granites of the district. The pegmatitic veins of the rocks themselves vary according to the original chemical composition. The intrusive pegmatites are more independent of the composition of the surrounding rocks and are richer in quartz and microcline. The same tendency is still more pronounced in the younger granites, which often appear as central bodies in the pegmatitized areas. The granitic emanations and solutions causing the metasomatic alterations have always preceded the pegmatite intrusions and these preceded the younger granites. With the intrusions of these granites the pegmatitization finished.

»The tangential pressure continued also after the alteration process was essentially ended and caused undulose extinction in the quartz grains of all rocks of the district, the veined gneisses, the more or less independent pegmatites as well as the younger granites. This gives us new proofs that the pegmatites and the granites have originated during the pegmatitization process of the region. The pegmatites and the younger granites are concentrated segregation products. The main mass has come from deeper parts of the earth crust. Partly, however, they may have come from the rocks now accessible for our investigations. It is, in fact, often difficult to determine whether we have larger segregations in situ or intrusions before us.»

In a discussion about the origin of these veined gneisses (Magnusson, 1934) P. J. Holmquist suggested that the chemical alterations referred to might be due to weathering at the earth's surface followed by metamorphism in depth. This theory does not, however, account for the connection with the intrusive pegmatites and granites. In the same discussion H. G. Backlund declared that he could see no cogent reasons for assuming the metasomatic alteration of the veined gneisses, the cordierite, andalusite, and sillimannite-bearing rocks, all of which, in his view, represented argillaceous sediments. The occurrence in the gneissose old granites of areas rich in these minerals indicates, according to Backlund, the former presence of such sediments, which were assimilated by the granite magma long before the alteration to veined gneisses. My reply to Backlund was that it is possible that I have exaggerated the role played by the metasomatic alteration; but that the alteration took place and gave rise to rocks richer than before in alumina and femic oxides cannot be denied. In the same discussion N. Sundius tried to explain the veined gneisses containing garnet, andalusite, cordierite and sillimannite to be magmatic differentiates in situ in a gigantic magma according to the theory of H. E. Johansson. He said that the garnet gneiss has to a large extent a peculiar composition corresponding neither to the eruptive rocks we know from younger formations nor to any sedimentary rocks. Sundius also said that what he called in the discussion »the migmatite theory» — as applied not only here but everywhere in the world where it was used — could not solve the problem. If this theory is correct, it should be impossible to make stratigraphic studies of the gneisses. Against this I should like to emphasize that the irregularities on both a large scale and in detail cannot be explained by the »magmatic theory» of Johansson and that the chemical alterations in the veined gneisses are only seldom so intensive that it is impossible to distinguish the original sediments from the volcanic rocks and from the old granites. In the slates the minerals rich in alumina and femic oxides should have originated even if no metasomatic alteration had occurred. Through these alterations, however, the amount of such minerals has been larger than without the metasomatic alterations.

After collecting the existing analyses of the garnet gneisses in the Sörmland region, W. Larsson (1932) has expressed the opinion that all these gneisses are granites that have assimilated leptitic and sedimentary rocks. Even this theory cannot solve the essential problems. There are normally no difficulties in distinguishing between old granites and the supracrustal rocks but naturally there may have been such an assimilation here and there, though only on a small scale.

N. Sundius later changed his mind. In 1947 he declared that the veined gneisses of the Sörmland region have all originated from femic leptites of volcanic origin; that the alumina-rich minerals have come into existence through late-magmatic metasomatism, connected with the pegmatites; that the pegmatites have come from the old granites; and that the pegmatites have no connection with the younger granites. Against these opinions I have (1948) stated that the old granites outside of the veined gneisses are never followed by pegmatites; that the old granites are altered in the same way but not so strongly as the supracrustal rocks; and that there is an intimate connection between the pegmatitic spots and veins of the veined gneisses and the more independent, more intrusive pegmatite masses, and that from these there are all transitions to the younger granites of the Stockholm—Fellingsbro series.

In 1936 and 1937 H. G. Backlund published his well-known papers »Der Magmaaufstieg in Faltengebirgen» and »Die Umgrenzung der Svekofenniden.» In these papers Backlund expressed his opinion that the Svecofennian granites and gneisses are granitization products in situ deriving for the main part from sediments. In the discussion that began after publication of these papers the author (1937 and 1938) emphasized that the veined gneisses could not be coordinated with the old granites. Special granites have, however, arisen as products of the alteration process that have given us the veined gneisses. These granites show only small variations. They are instead nearly always red or grey porphyritic granites (Fellingsbro granites), rich in quartz and microcline and showing all transitions to even-grained, medium- or fine-grained granites (Stockholm granites). The granite series older and younger than these granites and not combined with veined gneisses are all differentiated series with gabbros, diorites, tonalites, granodiorites and granites proper. The granites appearing together with veined gneisses tend more and more, as they leave the region of these gneisses, towards an eutectic composition.

Between these eutectic granites and the concentrations of granitic material inside the veined gneisses there are all transitions. As two diagrams published in 1938 illustrate, the granite concentrations inside the veined gneisses have highly varying compositions. Mostly the amounts of potassium are high as are also the amounts of alumina and femic oxides. Two conclusions may be drawn from these facts. The first is that these granites have not released themselves from the supracrustal material they have soaked through and from which they have taken a good deal of material. Second, the high amount of potassium in several of the analyses indicates a concentration of potassium in a higher degree than soda. We are thus justified in speaking of potassium metasomatism.

The corresponding granites outside the veined gneisses have, as the diagrams show, lower amounts of alumina and femic oxides and about the same amounts of potassium and of sodium. The compositions thus very nearly approximate the eutectic composition.

On a map published in 1938 I tried to illustrate the distribution of the veined gneisses and the late Svecofennian granites and pegmatites. The central area of the veined gneisses is surrounded by a zone rich in pegmatites and granites. Outwards from the central area these pegmatite and granite concentrations become more and more sharply defined as dikes and stocks.

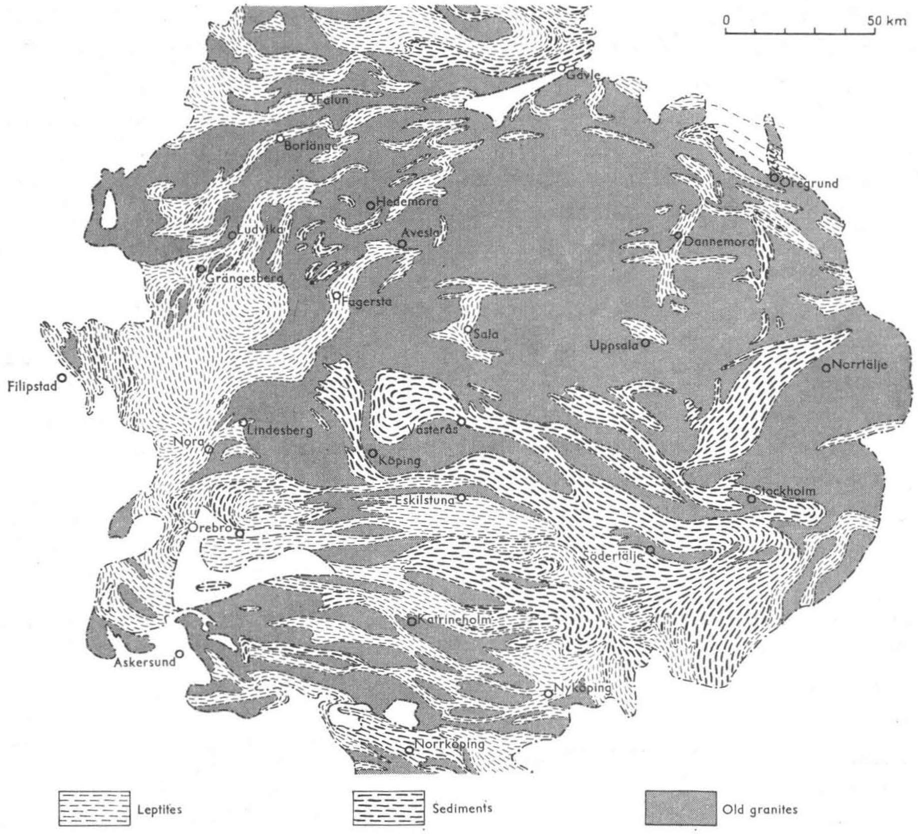

Fig. 1. Map showing the bedrock of Central Sweden after the Svecofennian folding.

From this zone a broad belt runs from the region between Örebro and Västerås in a northerly direction over Ludvika and Grängesberg up to Falun and from Falun in an easterly direction to Öregrund and Gräsö. In this belt rounded massifs of granites of nearly eutectic composition occupy large area. They are always accompanied by a great many pegmatite dikes.

Where concentrations of pegmatite dikes occur without massifs of such granites it seems to be quite clear that such massifs exist in the depth beneath the pegmatites. It is also quite clear that these massifs have the form of diapirs and, as there are all transitions between these granites and the granite concentrations in the veined gneisses, it also seems to me quite clear that the diapirs belong to the roof of the veined gneisses and that the granite material in them and in the pegmatites surrounding them must have come from the zone of the veined gneisses.

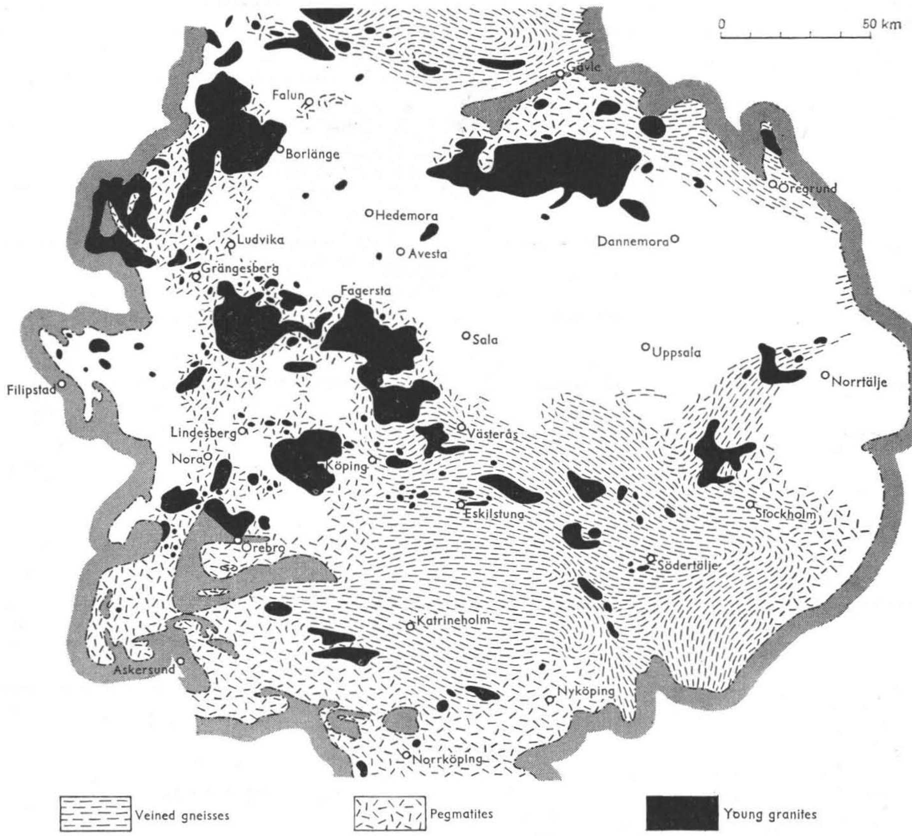

Fig. 2. Map showing the distribution of the late-Svecofennian veined gneisses, pegmatites and granites in Central Sweden.

According to the opinion expressed by the present author in discussions with Backlund in 1937 and 1938 as well as earlier, granitic material must have passed through the zone of the veined gneisses. This zone acted as a filter, and in the roof of this zone the granitic material streaming upwards became ever more nearly eutectic in composition. At the same time, more basic material was left behind.

Since 1938 I have had opportunities of studying several other iron ores situated within the veined gneisses and the zone rich in pegmatites surrounding the area of these gneisses and I have found the same impregnations with feldspar and mica, the same formation of new skarn associations in limestones and dolomites occurring together with the iron ores. The carbonate rocks free from iron ores and situated within the veined gneisses are more or less altered to rocks rich in »schlieren,» veins and irregular spots of silicate-

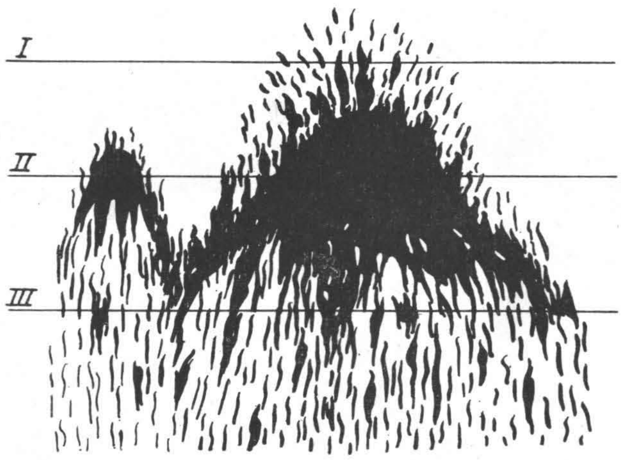

Fig. 3. The diagram illustrates the author's views on the relationship of the veined gneisses and the younger granites and pegmatites associated with them. The black veins in the lower part of the diagram (level III) represent granitic material soaking the veined gneisses. This material collects later at a higher level (II) into more and more homogeneous granite masses from which pegmatitic material is later injected into the overlying bedrock where such pegmatite accumulations (level I) are apparently unrelated to any granite massif; such a massif is certainly to be found, in the author's opinion, at depth. Beneath this granite, the occurrence of veined gneisses is to be expected.

minerals such as diopside, tremolite, forsterite, spinel and serpentine. These minerals must have been formed through wandering emanations and solutions driven before the pegmatites and granites.

These observations in the mines and in the limestone quarries confirm the view that metasomatic alterations through emanations and solutions were essential for the formation of the veined gneisses in Central Sweden and that there has been a transport of material on a large scale through the filter of the veined gneisses.

In the Kantorp region the original leptites were dominant among the supracrustal rocks and the real sediments were subordinate elements. When the author continued his investigations into the region between Kantorp and the Baltic, he soon found that the sediments belonging to the Mälar series were the dominant rocks and that most of the so-called garnet gneisses had originally been graywackes and slates. The mineral constituents of these veined gneisses are: quartz, plagioclase (An_{15}—An_{30}), perthite microcline and biotite in varying proportions, and together with them the characteristic alumina-rich minerals almandite, cordierite and sillimannite, and, more rarely, andalusite. In these rocks graphite is rather generally present, especially in association with iron sulphides, as sparse impregnations.

Studying the details, the author found that the layers have, through metamorphic differentiation, been more sharply limited than they were in the original sediment. The biotite has to a great extent accumulated in one layer and the light minerals in another. Through recrystallization, the quartz and the feldspars have been essentially more coarse-grained and the veins pegmatitic and irregular. Studying the layers, the author also noted that the alumina-rich minerals referred to in the foregoing are very irregularly distributed. In many cases the almandites are collected in small spots here and there within the layers. The same irregular distribution often characterizes the other alumina-rich minerals. The quartz and feldspar often collect into true veins in the layers or cut the bedding. Finally, the microcline has been concentrated together with quartz and subordinate plagioclase in »schlieren» and veins varying from very thin lens-shaped bodies to mighty intrusive dikes, grading over to pegmatites and granites.

The author has gained a strong impression of a disintegration process at high temperature acting in conjunction with emanations and solutions which caused the irregularities described.

Also in the old granites and especially in types containing essential amounts of mica or hornblende we can study such irregularities in the distribution of original minerals: quartz, plagioclase, microcline perthite, biotite and hornblende. Even here lensshaped concentrations or veins appear. The youngest are red concentrations rich in quartz and microcline in the form of veins and intersecting dikes and it is also here, especially in the last-named concentrations, that the impression is given of material on the point of releasing itself from the gneisses.

The pegmatite material impregnating the quartz-banded iron ores in the Kantorp mines often contains sillimannite and cordierite together with a high content of biotite. The remnants of the quartz-banded ores indicate that the ores were originally pure jaspilites. There must therefore have been a considerable transport of material into the ores. As mentioned, the quartz is driven out of the ores impregnated with pegmatite material. In the cordierite- and anthophyllite-gneisses in these mines the distribution of the minerals has become very irregular. There are streaks and spots rich in cordierite and others rich in anthophyllite. We have here a new example of the same disintegration process as in the slates described in the foregoing.

The cross-cutting pegmatites and granites in the outer zone must be considered as material that has passed through a zone of veined gneisses at depth, loosening and assimilating material from these rocks. In the pegmatites of the outer zone we can find such minerals as tourmaline, beryl, orthite and apatite, sometimes even topaz and here and there uraniferous minerals. In the iron ores cut by these pegmatites the same minerals also appear. The pegmatites are further more often followed by alterations of the leptites to mica-schists in which the same minerals are again enriched.

In the intrusion zone outside the central area of the veined gneisses there are also sulphide impregnations clearly connected with the pegmatites and the mica-schists. In this zone, south, southwest, and southeast of the central area pyrite, pyrite-pyrrhotite, pyrrhotite, copper pyrite-cobaltite, and sphalerite-galena occurrences are found. These occurrences must either be new formations in connection with the processes that gave us the veined gneisses, or older occurrences removed and strongly altered by these processes. In the central zone there exist no sulphide occurrences worth mentioning, only sparse impregnation. In the intrusion zone in the roof of the veined gneisses, on the other hands the concentration to real occurrences could take place. The concentration occurred in the relatively well-preserved parts of the very irregular — from the metamorphic point of view — outer zone. Often we can see a zoning of the sulphide minerals against local migmatite fronts. Finally, the iron ores of this zone are usually more or less impregnated with varying sulphide minerals.

In other words, the author's opinion is that in connection with the formation of the veined gneisses the metals have been driven out from the sediments and from older sulphide ores and concentrated outside the migmatite front in the intrusion zone. In this way, among other metals, also zinc and lead have been precipitated, especially where the migmatite front is relatively sharp, as in the Åmmeberg field and in the northern part of Utö in the southern archipelago of Stockholm.

On the island of Utö, in the continuation of the zone from Åmmeberg to Tunaberg, there are zinc-lead ores as fahlband-impregnations in a banded complex of limestone layers and hälleflint-layers. The latter layers are ash tuffs often mixed with carbonate material. The ores consist of sphalerite, galena, and pyrrhotite in varying proportions and in connection with the sulphide invasion the limestone and the limestone-bearing hälleflintas were altered to skarn essentially consisting of diopside, tremolite, and scapolite. The migmatite front is situated northwest of the iron ore-bearing zone, and this front has, as in the Åmmeberg Field, been stopped by the boundary of a limestone-rich, markedly banded complex, into which the solutions and emanations that carried material for the formation of skarn and sulphides continued. The sulphide-impregnations have, with the outermost offshoots, reached the skarn-bearing jaspilite-iron ores of this island. Above all, in these ores and near them, the sulphides referred to have been accompanied by chalcopyrite, pyrite, chalcocite, arsenopyrite, and bornite. Fahlbands with sparse impregnations of pyrite and pyrrhotite appear north of the zinc-lead ores.

In the area of the diapiric Fellingsbro-Stockholm granites and the pegmatites around them, we have here and there found chalcopyrite, chalcocite and bornite in very small concentrations. Where pegmatites have entered carbonate rocks, as in the Yxsjö mines, scheelite has been formed, together with sulphide minerals and such skarn minerals as hedenbergite, grossularite and an amphibole rich in alkali and iron. These minerals appear in a chaotic mixture of quartz and feldspars, both microcline and plagioclase, remnants of limestone and a high amount of fluorspar.

In the Hörken mines nearby, there was found, in connection with exploration, the same hedenbergite-garnet-amphibole skarn. This is mixed with pegmatite minerals (quartz, microcline, and plagioclase), a high amount of fluorite and pyrrhotite, as well as small amounts of pyrite and chalcocite. Besides, these masses contain small amounts of both scheelite and molybdenite.

It is interesting to note that two older ore types were found in Hörken, namely, a skarnbanded-iron ore and a sulphide ore of the Falun type with galena and sphalerite in limestone, ophicalcite, and tremolite-diopside skarn. Both these ore types are intersected by several amphibolites in such a way that they and their skarn minerals must be considered older than the amphibolites. These are, on the other hand strongly influenced by the scheelite-molybdenite-bearing skarn-pegmatite masses, which must be younger.

The material for these characteristic mineral associations bound to the diapiric young granites and their pegmatites must have come from a zone of veined gneisses in the depth. The material had been loosened in the veined gneisses or deeper and then soaked its way up to the diapiric magma chambers and from there direct or through the pegmatite dikes into the surrounding rocks.

The author is convinced that it will be possible in the future to distinguish which minor constituents have been enriched in the iron ores of Central Sweden in connection with the late Svecofennian alteration processes described here and which minor constituents appear in about the same amounts as they had in the primary iron ores or after having been enriched through magnesia-metasomatism.

H. von Eckermann (1923), in a paper on the rocks and contact minerals of Tennberg, has described how, in the limestone of this occurrence at the contact of the young diapiric granite, an inner diopside-wollastonite zone and an outer garnet-vesuvianite zone have been formed. In the manganiferreous skarn and carbonate iron ores of Bastkärn, wollastonite and vesuvianite have been formed in the limestones outside the ores in the parts of the mines situated near the young diapiric granite. It is here worth mentioning that in the Åmmeberg mines, already referred to, vesuvianite and wollastonite are characteristic minerals in the skarn, appearing together with the zinc-lead ores of these mines.

The Geological Survey of Sweden has, as a special concession, received a number of absolute age determinations made at two geological institutes in the Soviet Union. The author has published and discussed these determinations in a paper (1960) in »Geologiska Föreningens förhandlingar».

46 2192—61

For the area of the veined gneisses of Central Sweden, the Geological Survey has obtained five new age determinations according to the potassium-argon method, the dates of which fall between 1700 and 1780 million years. Of these determinations three are of sediments, one of an old granite and one of the Stockholm granite. The age of the last-named rock is 1760 million years.

A. Parwel and F. E. Wickman published in 1954 some preliminary results of age determinations on Swedish pegmatite minerals according to the uranium-thorium-lead method. The result indicates that the pegmatites in the Svecofennian areas of Stockholm and Falun are about 1800 million years old. This result corresponds very well with the determinations according to the potassium-argon method.

The age determinations have thus given new and valuable support to the view that the pegmatites in the Svecofennian area and the granites of the Fellingsbro—Stockholm series are closely connected with the veined gneisses as products of the same palingenic processes.

Two diapiric granites outside the region of the veined gneisses have given an age of 1800 and 1850 million years. The explanation must be that these granites were formed at a higher level above the veined gneisses and that the palingenic processes at depth continued longer.

Veined gneisses occupy large areas also north of Central Sweden. In good outcrops along the coast southeast of Nordmaling in the broad geosyncline of Central Norrland, it has been possible for S. Gavelin (1955) to study in detail the relation between the mica schist sediments, the veined gneisses and the young granites. The conclusion is that the Revsund granites originated through alteration and mobilization of the sediments and that the veined gneisses with their abundant pegmatitic veins and biotite-plagioclase-rich material imply a metamorphic differentiation during which the more mobile quartz-feldspar material separated and could even intrude into the surrounding rocks and form hybrids. If the veined gneisses have been formed through metamorphic differentiation, the granites have, according to Gavelin, been formed through homogeneization processes at higher temperatures.

Two age determinations of the Revsund granite have given ages of 1700 and 1750 million years. The alteration of the geosynclinal sediments of Central Norrland into veined gneisses is thus of the same age as the corresponding alteration process in Central Sweden.

According to O. Ödman (1957), the Karelian Lina granites in Norrbotten county in northernmost Sweden are typical late-orogenic granites accompanied by aplites and pegmatites and closely related in time to the late Karelian veined gneisses. In many cases the Lina granites are believed to have passed through a magmatic stage, having unquestionably intruded into their wall rocks; but, on the other hand, there often occur gradual transitions from sediments to granites containing ghost-like remnants of the former. In such cases the granite does not show any intrusive features and is considered to have been formed by granitization in situ.

The serorogenic alterations in the northernmost part of Gävleborg county are summarized by Lundegårdh (1960) as recrystallizations of older microcline and sometimes of other minerals, migrations of potassium and formations of new microcline granoblasts, intergrowths and porphyroblasts locally assembled into veins (migmatitization), mobilizations in the first instance of silica and feldspar, either as veins and small dikes of pegmatite and sometimes aplite or masses, in part real intrusions, of pegmatite and fine medium- to fine-grained serorogenic rocks.

These short summaries of the views of S. Gavelin, O. Ödman and P. H. Lundegårdh indicate that we now all agree in the most essential respect, *viz.*, that there is a close relationship between the formation of the veined gneisses and the occurrence of special granites and pegmatites. There are, however, several divergencies as to what has really happened by and large and in detail, especially in regard to the transport of material.

In a very interesting paper S. Gavelin has recently (1960) discussed some problems concerning the relationship between the kinetometamorphism and metasomatic transformation during granitization and the formation of veined gneisses. He has suggested that the migration of chemical elements and diffusion connected with granitization and pegmatitization can be profoundly facilitated through kinetometamorphism under certain conditions and that kinetometamorphism must therefore be considered an important factor in interpretations of granitization and pegmatitization phenomena.

The present author can agree with these suggestions, as he had already touched these problems in 1936 in his description of the Kantorp region. I reckoned, as has been pointed out, with high temperature in connection with the directing pressure and with emanations and solutions coming from deeper zones of the earth's crust. The difference seems to be that Gavelin assumes more local migrations of material while my observations have forced me to assume more intensive transport of material inside the veined gneisses and from them upwards into the intrusion zone with its granites and cross-cutting pegmatites. This difference stems perhaps from the fact that I made my observations in the central part of the Sörmland area of veined gneisses, Gavelin in the outer parts.

The author's opinions may be summed up in the following points:

1. The formation of the veined gneisses in Central Sweden is a disintegration process at high temperature, which has caused irregularities even in details.

2. Metasomatic alterations caused by granitic solutions and emanations played a prominent role in the formation of the veined gneisses, and have soaked their way upwards through the zone of the veined gneisses and caused chemical alterations in them.

3. High pressure has also played a conspicuous role in the alteration processes that formed the veined gneisses, but it is not necessary to assume a new folding apart from smallscale local foldings. The essential folding processes were older.

4. There are all transitions between the pegmatitic veins and »schlieren» of the veined gneisses, the pegmatitic and granitic concentrates inside the gneisses and the more and more independent pegmatites and granites outside them.

5. No pegmatites occur in conjunction with the old synkinematic granites in Central Sweden. All the pegmatites in this region belong to the late Svecofennian palingenetic processes.

6. The late Svecofennian granites outside the area of the veined gneisses are diapiric concentrations of granitic material in the roof of the veined gneisses, through which the material has soaked its way upwards.

7. The granitic concentrations inside the gneisses have quite varying compositions and are often especially rich in potassium. As the granites have become more and more independent of the gneisses, they have come nearer and nearer the eutectic composition common to the diapirs.

8. The zone of the veined gneisses has acted as a filter. The solutions and emanations passing through this filter have driven out several metals and other elements from the sediments and from older sulphide occurrences and concentrated them outside the local migmatite fronts or in the diapirs from which they have gone out into the surroundings.

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18

FELDSPAR-EQUILIBRIUM TEMPERATURE OF SOME FINNISH ROCKS [1]

BY

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ABSTRACT

The feldspar-equilibrium temperature of some Finnish Precambrian granites is given using the method developed by Barth. The crystallization temperature of feldspars in large granite pegmatite bodies ranges from 580°—630°C. Microcline granite and aplite show low feldspar-equilibrium temperatures, 450°C and 390°C respectively. The highest temperature, 730°C, was found in anorogenic rapakivi granite.

Analytical data on the feldspars have been presented from only a few Finnish Precambrian rocks and in only one case has the composition of feldspar been used to estimate the temperature of crystallization (Kaitaro, 1953). The present author has collected available analytical data of Finnish feldspars and k-values have been calculated by the equation:

$$k = \frac{\text{mol fraction of ab in alkali feldspar.}}{\text{mol fraction of ab in plagioclase}}$$

The temperature of crystallization has been estimated from the curve presented by Barth (1956) which gives the relation between the temperature and the k-values. The analytical data used, the calculated k-values and the estimated feldspar-equilibrium temperatures of Finnish rocks are presented in Table 1.

Most of the analytical data of Finnish feldspars is from the granite pegmatites in the Tammela area described thoroughly by Mäkinen (1913). The feldspar-equilibrium temperature of these pegmatites is 580°—610°C (*cf.*

[1]) Received February 28, 1961.

Table 1. Chemical data of feldspars, k-values and estimated

	1	2	3	4
Potash feldspar: ...	$Or_{67}Ab_{31}An_2$	$Or_{69}Ab_{30}An_1$	$Or_{70}Ab_{28}An_2$	$Or_{71}Ab_{28}An_1$
Na_2O	3.57	3.60	3.44	3.31
K_2O	11.19	11.59	11.86	12.09
Ab	32.7	32.1	30.6	29.4
Or	67.3	67.9	69.4	70.6
Plagioclase:	$Or_4Ab_{92}An_4$	$Or_4Ab_{92}An_4$	$Or_3Ab_{93}An_4$	$Or_9Ab_{86}An_5$
CaO	0.68	0.68	0.91	1.00
Na_2O	10.69	10.69	10.86	10.02
An	3.4	3.4	4.4	5.2
Ab	96.6	96.6	95.6	94.8
k	0.34	0.33	0.32	0.31
T	610°C	600°C	590°C	580°C

1. Granite pegmatite. Härkäsaari, Tammela (Mäkinen, 1913).
2. Granite pegmatite. Härkäsaari, Tammela (Mäkinen, 1913).
3. Granite pegmatite. Heponiitynmäki, Tammela (Mäkinen, 1913).
4. Granite pegmatite. Pakkalanmäki, Tammela (Mäkinen, 1913).

Table 1, anal. 1—4). This agrees with Mäkinen's conclusion, obtained from the twinning of quartz, that the main phase of crystallization of the granite pegmatite has taken place above the inversion temperature (573°C) of low and high quartz. Furthermore, the temperature of crystallization of the pegmatites is the same as that of large pegmatites in the Precambrian of Norway which is, according to Barth (1956), around 600°C. Recently Neuvonen and Vesasalo (1960) have concluded that the crystallization of a small lithium pegmatite in the Tammela area took place at a temperature around 450°C. This temperature has been roughly estimated, however, by means of hydrothermal alteration products of petalite and pollucite and therefore it does not correspond to the temperature of the main phase of crystallization.

Microcline granite (Table 1, anal. 5) and albite aplite (Table 1, anal. 6) from the Tammela area show lower feldspar-equilibrium temperatures than granite pegmatites. The microcline granite represents migmatite-forming late-kinematic granites of the Precambrian, Svecofennidic mountain chain and its origin is due to granitization (*cf.* Simonen, 1956 and 1960). The feldspar-equilibrium temperature of this granite is only 450°C and it is very similar to that of anatectic granites in the Precambrian of Norway (about 450°C according to Barth, 1956). Furthermore, mention must be made that the feldspar-equilibrium temperature of aplite, 390°C, is very close to that of aplites (410°—430°C) determined by Barth (1956).

Barth (1956) has stressed a rather constant crystallization temperature around 600°C for large granite pegmatite bodies. This is also suggested by feldspar-equilibrium temperatures of some Finnish Precambrian rocks.

5	6	7	8
$Or_{84}Ab_{15}An_1$	$Or_{86}Ab_{12}An_2$	$Or_{67}Ab_{32}An_1$	$Or_{66}Ab_{30}An_5$
1.80	1.26	3.69	2.99
14.11	14.54	11.12	9.88
16.2	11.6	33.5	31.5
83.8	88.4	66.5	68.5
An_5	An_5	$Or_{10}Ab_{83}An_7$	$Or_9Ab_{61}An_{30}$
		1.43	5.75
		9.55	6.51
5.0	5.0	7.6	32.9
95.0	95.0	92.4	67.1
0.17	0.12	0.36	0.47
450°C	390°C	630°C	730°C

5. Microcline granite. Häiviö, Kalliojärvi, Tammela (Mäkinen, 1913).
6. Albite aplite. Kietyönmäki, Tammela (Mäkinen, 1913).
7. Granite pegmatite. Varala (Lokka, 1935).
8. Wiborgite. Pikiruukki, Monrepos (Wahl, 1925).

Finnish material when we call attention to the fact that the feldspar-equilibrium temperature of the pegmatite in Varala (Table 1, anal. 7) is 630°C, only slightly higher than that of the granite pegmatites in Tammela. In this connection a mention must also be made of the chemical composition of potash feldspar from pegmatite belonging to the anorogenic, Precambrian Åva granite (Kaitaro, 1953). The content of plagioclase as exsolution perthite is 28 percent, which is about the same as generally found in potash feldspars of granite pegmatites. Kaitaro (1953) concluded that this pegmatite was formed at a high temperature.

Wahl (1925) has published chemical data of the feldspars of the anorogenic rapakivi granite with large potash feldspar ovoids surrounded by plagioclase mantles (Table 1, anal. 8). The feldspar-equilibrium temperature of the rapakivi is as high as 730°C, indicating a truly magmatic range of temperature. A magmatic origin of the rapakivi granites has been suggested by Finnish geologists and a high temperature of crystallization has been pointed out by monoclinic symmetry of potash feldspar rich in perthite. Furthermore, idiomorphic, hexagonal bipyramidal high quartz of the rapakivi shows that the crystallization has taken place above the inversion temperature (573°C) of low and high quartz.

The sporadic and small amount of information presented in this paper shows that it seems to be possible to divide the Finnish Precambrian rocks into separate provinces by feldspar-equilibrium temperatures. The present data are, however, insufficient to characterize the different provinces, and therefore the main purpose of this paper is to call attention to the need for collecting more data on the feldspar-equilibrium temperature of Finnish rocks, because this information is one important way to characterize and to discuss the conditions of origin of granites and gneisses.

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OLIVINE FROM RAPAKIVI [1]

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ABSTRACT

Olivine occurs sporadically in grey or green colored, hornblende-bearing silicic varieties of the rapakivi complex in southeastern Finland. Its manner of occurence and chemical and physical properties are presented. The olivine is an iron-rich member of the forsterite-fayalite series and the forsterite content of the analysed olivine is only 6 mol. percent.

Petrographic study of the specimens collected during the geological remapping of the wide rapakivi area in southeastern Finland (so-called rapakivi massif of Viipuri) has shown that olivine occurs sporadically in grey or green colored, silicic varieties of the rapakivi. The olivine-bearing varieties of rapakivi are not widely distributed, but they have been found, however, as small bodies in different parts of the wide rapakivi area. Olivine as a rock-forming mineral of green-colored rapakivi varieties in Finland has been already reported by various authors (*cf.* Hackman, 1934; Savolahti, 1956), and some mineralogical data have been presented which point to an extremely iron-rich variety of the forsterite-fayalite series. To complete and increase the knowledge of the mineralogy of the Finnish rapakivi, I am here presenting the mineralogical data of the olivine. Some other minerals of the rapakivi will be described in the nearest future.

The mineralogical compositions of olivine-bearing rapakivi varieties are presented in Table 1. The content of olivine is usually very low, only a few percent. A remarkable concentration of olivine has been found only in one case in a lenticular lens or schlieren in a hornblende rapakivi. This

[1] Received February 28, 1961.

Table 1. Modal mineralogical composition of olivine-bearing rapakivi varieties, axial angle and birefringence of olivine.

	1	2	3	4	5	6	7	8	9
Quartz	16.7	15.2	16.2	22.6	20.4	20.4	26.4	32.3	37.7
Plagioclase	12.7	38.0	23.1	14.7	23.7	22.1	21.2	14.8	10.4
Potash feldspar	10.8	30.7	49.8	46.6	44.1	43.7	44.1	42.6	47.4
Olivine	30.0	0.8	1.2	2.3	1.0	5.1	5.2	0.6	0.6
Pyroxene	1.1	0.9	—	—	—	4.6		0.2	0.1
Hornblende	18.3	11.4	8.5	9.1	9.0	3.3	1.7	7.0	1.8
Biotite	4.8	0.9	0.3	3.5	0.5		0.3	1.3	0.6
Chlorite+iddingsite	1.0	0.4	0.4	—	0.3	—	—	0.8	0.8
Ore	3.4	0.8	0.5	1.2	0.6	0.6	1.1	0.4	0.2
Zircon	0.2	—	—		—			—	0.2
Apatite	1.0	0.9	—		0.2	0.1			0.1
Fluorite	—	—	—		—	—		—	0.1
Calcite	—	—	—	—	0.2	—	—	—	—
Olivine:									
Axial angle 2Vα	51°	—	50°	49°	51°	49°	52°	—	—
Estimated composition	6 % Fo	—	4 % Fo	2 % Fo	6 % Fo	2 % Fo	8 % Fo	—	—
$\gamma - \alpha$	—	0.049	—	0.052	0.050	—	—	—	0.050
Estimated composition	—	11 % Fo	—	0 % Fo	6 % Fo	—	—	—	6 % Fo

1. Dark schlieren in hornblende rapakivi. Ristisaari, Pyhtää.
2. Green hornblende rapakivi. Ristiluoto, Haapasaari.
3. Green hornblende rapakivi. Koisonniemi, Taipalsaari.
4. Green contact variety of the hornblende rapakivi. Nurmaanjärvi, Ahvenisto massif (Savolahti, 1956, p. 52).
5. Green hornblende rapakivi. Tullisenlampi, Lemi.
6. Green marginal variety of hornblende rapakivi. Between Siikajärvi and Leppälahti, Ahvenisto massif (Savolahti, 1956, p. 52).
7. Green rapakivi. Savitaipale.
8. Greyish green rapakivi. Koivuluoto, Vehkalahti.
9. Grey rapakivi. Kilpisaari, Haapasaari.

lens, rich in ferromagnesian minerals, contains as much as 30 percent olivine (Table 1, anal. 1).

Olivine-bearing rapakivi varieties contain abundant quartz, plagioclase, and potash feldspar in highly variable proportions. The association of olivine with quartz indicates that it is iron-rich, because magnesian olivine and quartz are incompatible in igneous rocks. Other mafic minerals of the olivine-bearing varieties are pyroxene, hornblende and biotite. Pyroxene occurs only sporadically, but hornblende is always present and in greater amounts than biotite. Texturally the olivine-bearing rapakivi varieties are even-grained rocks of medium or coarse grain size. Some varieties contain large ovoids of potash feldspar. A characteristic feature of the olivine-bearing rapakivis is a greyish or green color. The green color is due to the weed texture produced by the green hornblende (*cf.* Savolahti, 1956).

Chemical compositions of some olivine-bearing rapakivi varieties are presented in Table 2. The composition varies greatly but the silica content of

Table 2. Chemical compositions of olivine-bearing rapakivi varieties.

	1	2	3	4	5
SiO_2	63.72	65.87	69.50	69.44	72.58
TiO_2	0.96	0.78	0.45	0.33	0.34
Al_2O_3	14.29	13.81	12.34	14.01	12.98
Fe_2O_3	1.92	0.98	2.02	0.80	0.86
FeO	5.74	4.98	3.97	3.39	1.83
MnO	0.13	0.11	0.06	0.11	0.11
MgO	0.69	0.77	0.35	0.26	0.25
CaO	3.28	3.10	2.33	2.04	1.01
BaO	—	0.02	0.03	—	0.10
Na_2O	2.68	3.66	2.50	2.69	3.01
K_2O	4.73	5.57	5.80	6.36	5.28
P_2O_5	0.38	0.21	0.08	0.19	0.18
F	—	0.13	0.11	0.05	0.36
CO_2	0.44	—	—	—	—
H_2O+	0.64	0.48	0.43	0.44	0.66
H_2O-	0.14	0.11	0.10	0.20	0.38
	99.74	100.58	100.07	100.31	99.93

1. Green rapakivi (olivine-bearing). Tullisenlampi, Lemi. Anal. P. Ojanperä.
2. Green marginal variety of hornblende rapakivi. Nurmaanjärvi, Ahvenisto massif (Savolahti, 1956).
3. Hornblende rapakivi (olivine-bearing). From between Siikajärvi and Leppälahti, Ahvenisto massif (Savolahti, 1956).
4. Green rapakivi (olivine-bearing). Kaitjärvi (Hackman, 1934).
5. Standard mixture of the East Fennoscandian rapakivi granites. (Sahama, 1945).

olivine-bearing green colored rapakivis is always lower and the contents of CaO, MgO and FeO higher than in common rapakivi types. For comparison, the main constituents of the average chemical composition of rapakivi are also presented (Table 2, anal. 5).

The olivine occurs either as small, separate idiomorphic grains or as grains surrounded by hornblende or sometimes by monoclinic pyroxene. Poikilitic olivine enclosing quartz and potash feldspar grains has also been observed. The olivine is generally rather fresh, but alteration to yellowish brown iddingsitic material has taken place along its margins and fissure cracks. The color of the olivine is light greenish.

The most suitable and accurate way to estimate the composition of small and scattered olivine grains is by measuring the axial angle with the universal stage. The results of measurements and the estimated composition using the 2V determinative curves of Poldervaart (1950) are presented in Table 1. They show that the olivine from rapakivi is rather uniform in different specimens, and is fayalite containing only 2—8 mol. percent forsterite. Measurements of birefringence by means of a Berek compensator and the estimated composition, using the birefringence determinative curve of Poldervaart (1950), are presented in Table 1. They also indicate that olivine from rapakivi is rich in the fayalite component.

Table 3. Chemical composition and physical data for olivine from green colored rapakivi. Tullisenlampi, Lemi. Analyst: Pentti Ojanperä.

Chemical composition						
Weight %		Mol. prop.	Atomic ratio		Calculated composition	
SiO_2	31.26	5 202	Si	5 202	Fe_2SiO_4	92.1 mol. %
TiO_2	0.46	58	Ti	58	Mn_2SiO_4	1.8 »
Al_2O_3	0.00	—	Fe^{+3}	56	Mg_2SiO_4	6.1 »
Fe_2O_3	0.44	28	Fe^{+2}	8 951		
FeO	54.31	8 951	Mn	171		
MnO	1.21	171	Mg	588		
MgO	2.37	588	Ca	11		
CaO	0.06	11	Na	12		
Na_2O	0.04	6	O	20 331		
K_2O	0.00	—				
H_2O+	0.27					
H_2O-	0.02					
	100.44					

Physical data		Estimated composition	
α	1.815	Fe_2SiO_4	95 mol. %
β	1.851	Mg_2SiO_4	5 »
γ	1.865		
$2V_\alpha$	51°	Fe_2SiO_4	94 mol. %
		Mg_2SiO_4	6 »
$d_{(130)}$	2.826 Å	Mg_2SiO_4	10 mol. %

To study more closely the chemical composition and physical properties of olivine, it was separated from the green colored rapakivi variety occurring in Lemi by means of an electromagnetic separator and Clerici solution. The mineralogical as well as chemical composition of the host rock of the separated olivine are presented in Table 1 (anal. 5) and Table 2 (anal. 1).

The chemical composition of the separated olivine is presented in Table 3. The purity of the material used for chemical analysis was tested under the microscope by means of an immersion liquid of approximately the same index of refraction as the mineral itself. No foreign grains were detected, but some olivine grains contain small inclusions of quartz. The total amount of quartz inclusions is estimated to be 0.6 weight percent and the amount of iddingsitic alteration product along the fissure cracks is estimated to be less than 0.3 percent. No corrections for the mentioned inclusions have been made in the analytical data, because they do not change the mutual ratios of the main components of the olivine. The analysed olivine is an extremely iron-rich member of the forsterite-fyalite series containing 92.1 mol. percent fayalite, 6.1 mol. percent forsterite and 1.8 mol. percent tephorite. Small quantities of Ti and Fe^{+3} also enter the investigated olivine.

The optical properties of the analysed olivine are presented in Table 3. The refractive indices were determined by the immersion method in sodium

Table 4. (130) spacing of olivine from rapakivi and estimated composition.

	$d_{(130)}$ [1])	Forsterite content			
		From X-ray data		From optical data	From chemical data
		Curve for natural olivines	Line for synthetic olivines		
1	2.826 Å	10 mol. %	5 mol. %	5—6 mol. %	6 mol. %
2	2.827 Å	8 mol. %	4 mol. %	6 mol. %	—

1. Analysed olivine from green colored rapakivi. Tullisenlampi, Lemi (*cf.* Table 3).
2. Olivine from dark schlieren. Ristisaari, Pyhtää (*cf.* Table 1, anal. 1).

[1]) average of 5 determinations.

light at 20°C. The axial angle was measured on the universal stage. The optical data show that the olivine is fayalite. Using the determinative curves of Poldervaart (1950), the forsterite content of the analysed olivine is estimated to be 5 mol. percent by means of the refractive indices and 6 mol. percent by means of the axial angle. These estimates are in very good agreement with the actual chemical composition of the investigated olivine (*cf.* Table 3).

Two samples of olivine from rapakivi were X-rayed with an internal silica standard using a Norelco Geiger-counter diffractometer. The (130) spacings were measured and the composition was estimated by means of the X-ray determinative curve presented by Yoder and Sahama (1957). The results of these determinations are given in Table 4. The content of forsterite, estimated from the X-ray determinative curve for natural olivines, ranges from 8 to 10 mol. percent and the Mg_2SiO_4-content of the analysed olivine is by this method 10 mol. percent. It must be mentioned that no correction for Mn-content is necessary, because the analysed olivine satisfies all the specifications placed by Yoder and Sahama (1957) on the composition of olivines used to construct the X-ray determinative curve.

Considering the composition of the olivine from rapakivi obtained by the different methods we must first state that the measurements of the axial angles of a great number of samples (Table 1) indicate that the forsterite content ranges from 2—8 mol. percent. The chemical study shows a forsterite content of 6 mol. percent and a good agreement exists between chemical and optical determinations (*cf.* Table 3). The estimation of the composition by the X-ray determinative curve gives slightly higher contents of forsterite than the other methods (*cf.* Table 4), but the differences are within the upper limits of error of the X-ray method (± 4 mol. percent according to Yoder and Sahama, 1957). By using, however, the linear variation of d_{130} for synthetic olivines in the estimation of the composition, a better agreement with both chemical and optical methods has been obtained.

Therefore it may be possible that the X-ray determinative curve for natural iron-rich olivines is a little closer to the linear variation of d_{130} for synthetic olivines than presented by Yoder and Sahama (1957).

The determinations made by optical, chemical and X-ray methods show that the olivine from rapakivi is an iron-rich member of the forsterite-fayalite series containing less than 10 mol. percent of forsterite. A characteristic feature of the rapakivi seems to be that its ferromagnesian minerals are extremely iron-rich.

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20

OBSERVATIONS ON RECENT LAKE BALLS AND ANCIENT *CORYCIUM* INCLUSIONS IN FINLAND [1]

BY

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ABSTRACT

The occurrence of small balls consisting of living blue-green algae on the sandy bottom on the western shore of Lake Kivijärvi in Finnish Lapland is described. On the southern shore there is another kind of balls, greater in size and built up by peat and moss, spruce needles and fragments of grass and twigs. The conditions in which these lake balls have been formed are discussed and compared with other known discoveries of similar globular forms. The possibilities of such aegagropilous balls being preserved morphologically as coal-bearing fossil remains in sandy layers of metamorphosed sediments are discussed. The author comes to the conclusion that the coal-bearing sacks of *Corycium enigmaticum* in the Precambrian phyllites in Aitolahti may have probably originated from ancient aegagropilous forms of algae and that they thus may also be true fossils from the morphological point of view.

CONTENTS

[1] Received May 15, 1960.

INTRODUCTION

The origin and nature of aegagropilous — from Greek *aigagros* (goat) and *pilos* (hair) — forms of algae, so-called lake balls (Smith, 1950) or »Seebälle» (de Lagerheim, 1892) have always aroused the interest of phycologists. de Lagerheim already mentioned that there are two different kinds of aegagropiles: genuine lake balls (Echte Seebälle), which contain radial outgrowing filaments of living algae (*e. g. Cladophora* balls and *Stigomena* balls) and quasi lake balls (»Unechte Seebälle»), formed in a mechanical way by the action of under-water parts of waves and consisting of waterlogged plant fibers and other filamentous material (*e. g. Posidonia* balls and peat balls). True aegagropilous forms af algae have been described from Finland in botanical papers by Häyrén (1944) and Luther (1947) and lake balls, consisting of peat and tangled fibrous plant material from lakes in Finland by Helminen (1957), Häyrén (1947) and Kujala (1944). On the other hand, records of ancient aegagropilous plants as fossils are difficult to find in the literature.

The occurrence of the coal-bearing sacks of *Corycium enigmaticum*, discovered by Sederholm in the Precambrian varved schists of the Tampere area are described by Sederholm (1897, 1911) and also by Rankama (1948, 1950), van Straaten (1949) and Seitsaari (1951). Krejci (1924) and van Straaten (1949) have suggested a tectonic origin for *Corycium*. Rankama, on the contrary, has proved the organic origin of the *Corycium* with his carbon isotope determinations. More knowledge on the earliest life has been presented by Tyler and Barghoorn (1954), who have discovered identifiable fossils of blue-green algae, fungi and possibly a flagellate from the Precambrian black flints in Ontario.

The author owes sincere thanks to Prof. H. Luther for the opportunity to study aegagropiles in the collections of the Botanical Museum, to Dr. C. Cedercreutz for determinations of algae, and to Mr. P. Isoviita for determinations of plant material in the lake balls. The author is greatly indebted to Professors P. Kallio and U. Varjo for kind assistance with the photography.

AEGAGROPILOUS FORMS OF ALGAE

In a shallow cove on the western shore of Lake Kivijärvi, Kittilä parish in August 1959, the author found a great number of small balls composed of algae. They were lying on the even sand bottom at depths ranging from 15 cm to one meter. The greenish brown balls had a diameter of 1.5—4 cm, and all of them lay inside or partly within a zone of *Equisetum limosum*. The algal balls were structurally porous as was the loose felty layer of blue-green algae which covered the bottom within the horsetail zone. When dried, these balls shrank to hard compact pieces only a third their original aize.

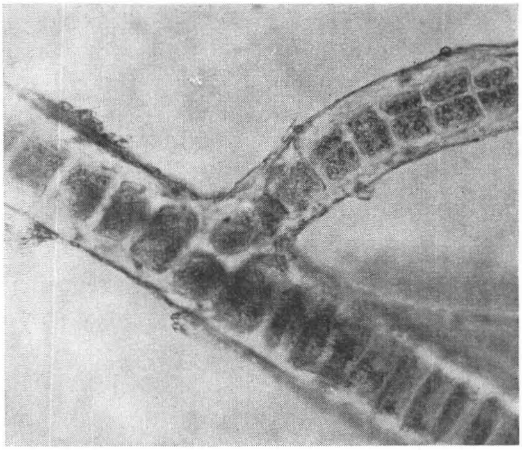

Fig. 1. Branched filament of *Stigonema ocellatum* from a sand-bearing lake ball in Lake Kivijärvi. Photo Matti Sulkinoja. Magn. 445 ×.

Under the microscope the lake balls show a network of a distinctly branched blue-green alga, *Stigonema ocellatum* (Fig. 1) [1]. In addition to the abundant filaments of *Stigomena*, a few diatoms and some pollen (mainly *Alnus*) also occur in the balls. However, the *Stigomena* balls also consist of minerogeneous material, mainly fine to medium sand (grain size 0.02—0.6 mm), consisting of plagioclase, potassic feldspar, green hornblende, quartz and muscovite. The same minerals make up the sand on which the lake balls lie.

Lake balls composed of blue-green algae may not be too uncommon. Häyrén (1947, p. 91) has discovered numerous small *Stigonema* balls with a diameter of 0.5—4 cm in Tvärminneträsk and Sandträsk in southernmost Finland. de Lagerheim (1892) mentioned that balls of *Stigonema ocellatum*, 0.5—1.5 cm in diameter were found by Nordstedt in a lake in Småland in south Sweden. Neither Häyrén nor de Lagerheim mentioned whether these *Cyanophyceae* balls contain minerogenic material as do the sand-bearing lake balls from Kivijärvi.

Best known among the genuine aegagropilous forms of algae are the *Cladophora* balls, belonging to *Chlorophyceae*. Häyrén (1944) has described such balls of *Aegagropila Martensii* in brackish water from seven places on the southern coast of Finland, and H. Luther (1946) from the Ekenäs (Tammisaari) region. According to de Lagerheim *(op. cit.)* Lorenz discovered lake balls formed of *Aegagropila Sauteri*, which appear in several lakes

[1] Determined by Dr. Carl Cedercreutz.

in Sweden, Great Britain, Germany, Austria and Italy. In the United States *Cladophora* balls seem to be of rare occurrence (Smith, 1950, p. 216).

The balls described by Luther had a maximal diameter of 6 cm and appeared at depths of 0—1.5 meters on clay or mud bottoms inside a swell-moderating *Phragmites* or *Scirpus lacustris* belt.

Prof. Hans Luther has kindly given the author an opportunity to study some algal balls, which are kept in the botanical museum of the University of Helsinki (Helsingfors). Among these are some *Cladophora* balls from Hederviken in Sweden. They are greyish green in colour and have a diameter of 20—25 cm. A sample from the spongy surface shows a sparse impregnation of silty clay.

Besides *Stigonema*, de Lagerheim *(op. cit.)* mentioned *Hapalosiphon pumilus* and *Conferva chtonoplastes* as examples of ballforming blue-green algae. Häyrén (1947) has, in the inner archipelagos of south and west Finland, discovered another species, *Tolypothrix tenuis*, which often forms irregular balls with radial growth. There are also species among the *Phaeophyceaes*, which form aegagropiles as *e. g. Sphacelaria racemosa*, described by Häyrén (1947) from Finland and *Sphacelaria cirrhosa*, described by Wittrock (1884) from Gotland in Sweden. The latter balls, which occurred on sand bottom in a shallow gulf, had a diameter of 1—4 cm and showed concentric growth rings. de Lagerheim *(op. cit.)* also described *Rhodophyceaes* which form true aegagropilous balls, *e. g. Fastigiaria furcellata* f. *aegagropila* from the firth of Kiel in the Baltic sea and *Lithothamnium* species on the sea bottom from greater depths.

LAKE BALLS COMPOSED OF PEAT WITH INCORPORATED PLANT FRAGMENTS

The author found lake balls of entirely another type on the sand bottom of a cove in the SE corner of Lake Kivijärvi. About 80 globular or elliptic dark brown peat balls occurred here, lying in irregular trains for a distance of abt. 40 meters, 4—8 meters from the shoreline at a depth of 20—50 cm. Each ball of elongated shape lie with its longitudinal axis parallel to the shoreline, as was also the case with the numerous sunken cones of spruce in the same horizon. On the narrow beach above the waterline, there was plenty of organic material (spruce needles, leaves, moss, peat fragments and pieces of rind) transported by the surge on the both sides of the mouth of a little brook.

The dark brown lake balls have a firm consistency; when dried they only shrink a little and preserve their shape. Their diameter varies between 3 and 13 cm; the smaller balls are generally globular, while the larger are often elliptical in shape (Fig. 2). The balls show a core consisting of *Carex-Bryales*-peat with a mantle layer of brown moss and incorporated fragments of other

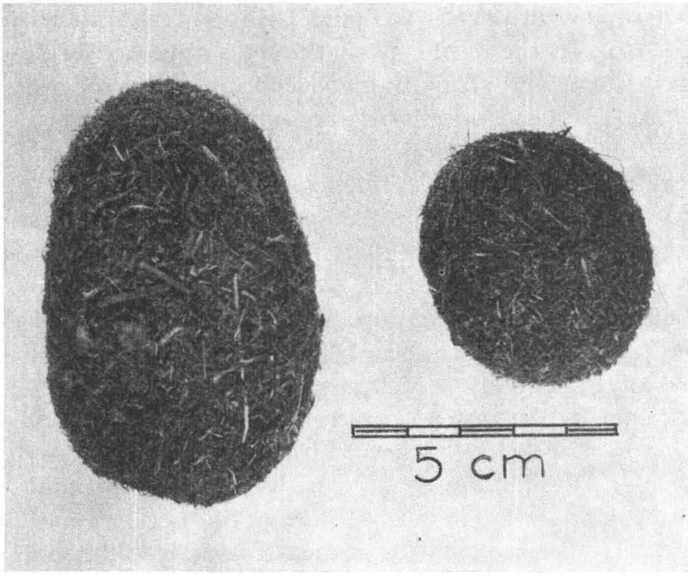

Fig. 2. Lake balls consisting of peat and fragments of straw, twigs and spruce needles. Lake Kivijärvi. Photo Matti Sulkinoja.

plants. Under the microscope [1] the peat core shows, in addition, fragments of *Sphagnum* leaves, blue-green algae *(Stigonema ocellatum)*, and also some pollen, mainly from *Pinus* and *Betula*. The surface layer consists of well preserved brown moss such as *Campylium stellatum*, *Calliergon stramineum*, *Drepanocladus* spp., *Hygrohypnum* sp. and *Aulacomnium palustre*. Enclosed in the thready binding substance, which consists of roots, straws, and fibres of leaves, are offal materials from plants, such as spruce needles, fragments of grass and dwarf birch and scales of cones. In some of the balls some reindeer hair and bird feathers were found among the enclosed material. Though these lake balls are principally built of dead organic substance, some contain living fresh green tops of *Drepanocladus* species with characteristic curved leaves.

The mantle as well as the core is highly impregnated with fine sand, which has been incorporated while the balls were rolling on the sandy bottom of the lake. The minerals in the balls are the same as in the sand on the bottom: plagioclase, green hornblende, quartz, microcline and muscovite.

The peat balls described above belong to the so-called quasi lake balls (»Unechte Seebälle», Häyrén, 1944; Luther, 1947). Similar balls have been described in Finland from Uusikaupunki (Nystad), Rutajärvi (Häyrén, 1928

[1] The determinations were performed by Mr. Pekka Isoviita.

and 1947), Herajärvi in Carelia (Kujala, 1944) and Vääräjärvi in Sääminki (Helminen, 1957). At the former place the balls appeared in brackish water and contained leaves and straw fragments of grass and sedges with needles of scotch pine. The other localities were on lake shores.

The author had an opportunity to study a lake ball from Rutajärvi at the Botanical Museum of the University of Helsinki. The diameter of this ball was 13 cm. It was made up of filaments of grasslike plants with fragments of moss, and was also impregnated with fine sand which was especially rich in muscovite.

Among similar balls of foreign origin from the collection of the Museum may be mentioned some globular and oval balls, composed of leaf nerves of *Posidonia oceanica*, a plant, which, according to de Lagerheim *(op. cit.)* generally appears in the entire Mediterranean. These sea balls (formerly named *Aegagropilae marinae)*, collected from sea shores in Italy, the French Riviera and Spain, are all impregnated with fine sand in similarity with the lake balls from Kivijärvi. There is also a ball of another type found on the shore of the Lake Silsersee in Switzerland. This lake ball has a diameter of 12 cm, is built up of a core of peat with a mantle of densely incorporated needles of larch and is also impregnated with sand.

ORIGIN OF THE LAKE BALLS

In Lake Kivijärvi, *Stigonema ocellatum* forms unattached brown, felty, in places incoherent, mats within a wave-moderating horsetail belt. Detached masses of the alga rug, which are too light, are thrown up on the shore by the swell and become decomposed, while fragments, which through a suitable impregnation of sand have achieved the necessary stability, are rolled slowly to and fro on the firm sand bottom by the gentle movement of the waves. According to de Lagerheim *(op. cit.)* the true lake balls get their globular shape and increase in size in the following way: because of the undulatory movements of the water they turn changeable new sides to the sunlight, and the result is a radial growth of branches of the living algae in all directions. Under the microscope several of those beautyfully branched blue-green algae could be observed in the sand-bearing algal balls from Kivijärvi. Many algal balls often show a radial or concentric structure (de Lagerheim, 1892, pp. 91, 92 and Häyrén, 1947, p. 91). Simultaneously with the growth of the algae, sand from the bottom is continously incorporated in the rolling ball. Luther *(op. cit.)* mentioned some *Cladophora* balls, which were at a depth of 3 m, outside a reed belt. Because the swell at this depth was not strong enough to roll the fairly heavy balls around, one half of each ball was living and the other half dead and decaying. According to Smith (1950, p. 216), there is often a death and decay of filaments at the center of the growing ball, and gases given off during photosynthesis may accumulate in this cavity and cause the balls to rise to the surface of the lake. Smith *(loc. cit.)* stated, that the formation of *Cladophora* balls is not due to the mode of growth of the alga but to an undulatory movement of the water at the bottom, set up by wave action at the surface. The continued abrasion of the plant mass, as it swishes back and forth across the sandy bottom, then results in a globular form.

The growth of the quasi lake balls in the southern end of Lake Kivijärvi seems to have occurred in a somewhat different way. The organic material of which they consist (peat ore moss in the core and moss, needles and straw in the mantle layer) is secondary water-logged material, that has been transported to this locality. Here the swell brings sand as well as the organic substance on the bottom into simultaneous movement. Peat fragments and tassels of moss are for the most part thrown on the shore or dispersed by the swell. However, where small peat or moss fragments through impregnation of sand have received a suitable stability, the gentle undulatory movement of the waves at a certain depth is able to give them only a slowly rolling in place motion. Then the growth of the balls begins in a mechanical way through incorporating of straw, twig, needles and other material lying on the sandy bottom. The impregnation of sand from the bottom continues during the rolling, when the lake balls with optimal specific gravity continue to grow.

The lake balls found in Kivijärvi thus represent two different groups: the *Stigonema* balls belonging to the genuine aegagropilous forms of living algae, and the peat balls belonging to the so-called quasi lake balls. The formation of the algal balls is as well a passive result of the action of underwater parts of waves as of the active radial outward growth of living alga filaments. The peat balls are formed in a mechanical way from plant fibers and other filamentous material by gentle undulatory movements of the water on shallow sandy bottom. Greenish living tops of moss specimens on the surface show that an increase in size of those quasi lake balls may in places also occur through a modest vegetative growing.

When such aegagropilous formations rich in carbonaceous material gradually become embedded in sand, silt or clay on the bottom of the sea or lakes, they have obvious presuppositions to retain their character in relation to the sediment in which they are embedded, in spite of processes such as diagenesis, recrystallization and moderate pressure. The fact that the balls generally also contain minerogenous material similar to the sediment in which they are embedded, may further increase their resistence. Kryshtofovich (1944, p. 69) points out that on the sea or lake bottom the mud particles are especially apt to penetrate through infinitely small interstices and to fill, in this manner, all cavities in the object buried on the bottom. The filamentous structure of aegagropilous algal agglomerations makes it easy for the bottom sand to penetrate and fill the cavities. Considering the scarcity and capricious occurrence of lake balls in recent times, it is understandable that the possibilities of discovering fossils of ancient aegagropiles are very small.

DISCUSSION OF THE ORIGIN OF THE CARBONACEOUS INCLUSIONS IN BOTHNIAN PHYLLITES ON THE SHORE OF LAKE NÄSIJÄRVI, AITOLAHTI PARISH

In this connection the coal-bearing sacks from the phyllite beds close to the eastern shore of lake Näsijärvi, first discovered and described by Sederholm, attract attention. Sederholm, who gave them the name *Corycium enigmaticum*, was convinced of their organic origin and considered them to be sack-like remains of primitive algae. Doubts about the primary fossilic nature of *Corycium*, have later been expressed. Referring to similar sacks of mergelic or pelitic material in Tertiary layers in Rumania, Krejci (1924, p. 59) suggested a tectonic origin for *Corycium*, and v. Straaten (1949, p. 14) did not consider them to be of fossil origin at all. According to v. Straaten they might represent squeezed protrusions of carbonaceous layers formed during folding by sliding movements. Rankama (1948) has, with his carbon isotope investigations, stated that the carbon in Corycium is of organic origin and that this is consequently a real fossil. On the contrary, the question of in how high a grade the carbonaceous sacks represent the original form and structure of the ancient organisms is left open by Sederholm. Rankama (1950, p. 76) refers to the studies of Kryshtofovich on phytoleims, which explain the manner of formation and preservation of both open and closed *Corycium* rings.

In his paper on sulphide-bearing schists, Marmo (1956, p. 85) points out the important part which sapropelic sediments have in building up the Bothnian sediment series of the Tampere area. The great amounts of sulphur, chiefly fixed in pyrrhotite in the sulphide-schists which in places contain very sparse graphite, bear witness to an abundant occurrence of primitive living organisms such as sulphur bacteria in early Precambrian time in basins with slight water circulation, and an oxygenpoor water.

There are, on the contrary, no sulphides in the carbonaceous inclusions in the phyllites of Aitolahti. This may be due to quite different conditions during the formation of these carbon-bearing bodies.

In the author's opinion, there are many features in the shape, structure and occurrence of the *Corycium* sacks that help prove that we are here dealing with fossils of ancient aegagropilous forms of algae. They might have been formed during the Precambrian era chiefly by threadlike algae (eventually together with other fibrous organic material) with enclosed sand, similar to the way they form in shallow water on present-day sea and lake shores. They have then been embedded in flood sediments and the primary structure of the algae is later totally destroyed during diagenesis and metamorphism, with fine disseminated carbon in the balls the only remaining organic evidence.

SHAPE AND STRUCTURE

The carbon often occurs most compactly in the periphery of the inclusions, where it causes the distinctive dark ring-shaped outlines of *Corycium enigmaticum* on weathered and polished rock surfaces. According to Rankama (1948, p. 397), the rock inside the carbon circles differs from the layer of phyllite in which the *Corycium* occur as inclusions, only by its greater content of finely disseminated carbon. Rankama also discovered in a metapsammitic layer, rounded or irregular elliptic balls which except in the circular or elongated periphery, also had a concentration of compact carbon in their centre. Furthermore, a third type of carbonaceous inclusion which is evenly impregnated with fine disseminated carbon is found in the phyllite layers in Aitolahti. In 1945, the author found a bullet-like fragment of a carbonaceous body of this latter type in a loose, quarried phyllitic material on a jetty on the eastern shore of Näsijärvi in Aitolahti. This elongated ball with evenly impregnated carbon lies parallel to and only half a centimeter from the contact between the metapsammitic and metapelitic layers (Fig. 3). Lenticular or rounded carbon-bearing inclusions of a similar type were found in Aitolahti by Sederholm as well as Rankama. They are considerable larger than both firstmentioned types of carbonaceous sacks and sometimes they show concentric layers of varying carbon content. Calcite can often be microscopically observed in them. Sederholm considered them to be organic remains, while Rankama (1948, p. 401) supposed that their origin might be similar to that of the calcareous concretions in Archean varved schists.

According to Rankama (1950, p. 76) the concentric internal structure of some *Corycia* may be interpreted as being the result of diffusion. In the authors opinion, the concentric texture of the carbon-bearing inclusions could well be primary and caused by the concentric structure common in genuine aegagropilous algal balls. The calcite content again may have its origin from calcareous water plants taking part in the formation of the ancient lake balls. In this connection it must also be mentioned that blue-green algae are able to secrete lime (Stirton, 1959, p. 111).

According to Stirton *(op. cit.)* Proterozoic rocks have yielded innumerable fossils, especially calcareous algae. Carroll and Mildred Fenton (1957)

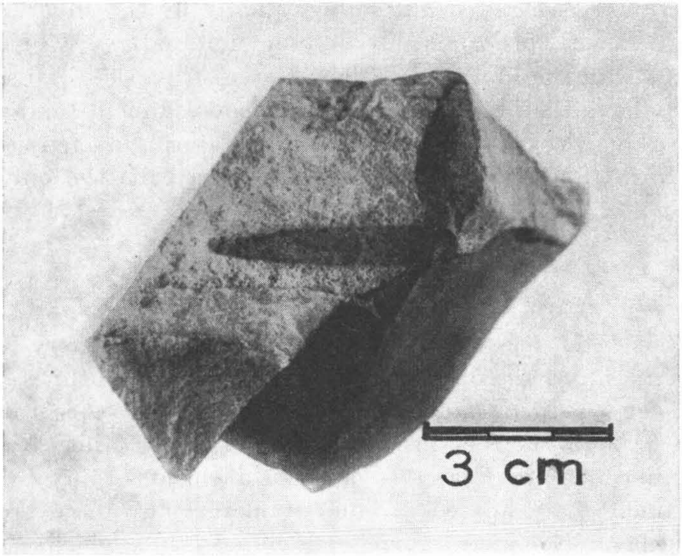

Fig. 3. Carbonaceous lenticular inclusion in a metapsammitic layer of the varved schist in Aitolahti. The contact between the metapsammitic and metapelitic layers is visible below the horizontal dark inclusion. Collection of the Biological Museum in Oulu.

have described some of the known algal fossils from the Belt Mountains, Montana. These calcium carbonate precipitates are arranged in globular or hemispherical masses with concentric laminations, but no plant structures have been observed. Discoveries in late Precambrian layers from eastern and western Siberia also bear witness to the existence of a flora of blue-green algae in Precambrian time. The fossil remains discovered in Siberia resemble calcareous concretions in their form. Also of great interest is the discovery of fossil spores by Russian paleontologists (esp. Timofejev) in calcareous, argillaceous and marl schists belonging to the younger Precambrian (Marmo, 1959, p. 153).

According to Stirton *(op. cit.)* the oldest known identifiable plants are blue-green algae and fungi from the black flints of Ontario, Canada, which are thought to be 2 billion years old. According to Beerbower (1960) the fossil-bearing flint formation may belong to the middle Huronian and be abt. 1.5 billion years old. Tyler and Barghoorn (1954) have there found at least two kinds of fungi and two different blue-green algae represented by spores and longer filaments.

v. Straaten (1949, p. 16) called attention to the fact that freshly embedded algae contain very small amounts of carbon, while the *Corycium* sacks contain a high quantity of this element. Owing to the present diameter of the compacted wall of *Corycium* (1—5 mm) v. Straaten considers that a very improbable wall thickness has to be assumed for the living organism. If the *Corycium* inclusions could be thought to originate not from solitary alga individuals but from aegagropilous agglomerations of filamentous algae, there should be carbon enough in the algal balls to bring about such compacted carbonaceous walls.

v. Straaten *(op. cit.)* also finds the concentric internal structure of some *Corycium* balls hard to explain. In recent genuine aegagropilous formations of algae, on the other hand, such a concentric structure is usual. de Lagerheim *(op. cit.* p. 91) described in detail such structures in balls of *Sphacelaria cirrhosa* from a shallow sandy bay in Gotland in Sweden: »Die Ballen, die einen Durchmesser von 1—4 cm besassen, waren von ziemlich faster, fast filzartiger Consistenz und bestanden aus einer ungeheuren Menge mehr oder weniger radiären Fäden, die mit ihren zahlreichen Zweigen wie zusammengefilzt waren. Bei dem Durchschneiden einer solchen grösseren Kugel beobachtete man 2—3 zuweilen sehr deutlich concentrische Schichten, deren jede 4—5 mm dick war. Eine Schicht entsprach wahrscheinlich den Zuwachs eines Jahres, und war also als ein Jahresring zu betrachten.»

On the other hand, concentric structures are not uncommon in late Precambrian fossils of calcareous algae (Stirton, *op. cit.*, p. 155).

SIZE

The great amount of sand in some recent lake balls ($\frac{1}{4}$—$\frac{1}{3}$ of the volume of the *Stigonema* balls in Kivijärvi) may further increase their resistance during embedding, diagenesis and metamorphism and may also guarantee them a minimum size which could well correspond to the size of the discovered *Corycium* sacks. The length of the carbonaceous sacks described by Sederholm is generally 2—3 cm, but can rise to 30 cm. The diameter of the sacks investigated by Rankama varied from 1 to 11 cm, and generally between 1 and 3 cm. Considerable larger are the lenticular or rounded inclusions in which carbon seems to be evenly disseminated or to appear in concentric compacted layers. These balls, which are calcite-bearing according to Rankama, have a maximum length of 55 cm and also appear embedded in coarse-grained flood layers in the phyllites. Owing to the occurrence of recent aegagropilous balls with the size of a head of man *(e. g.* sea balls of *Posidonia oceanica* and *Aegagropila Sauteri;* de Lagerheim, *op. cit.*, pp. 91 and 93) the possibility of finding fossil remains of embedded aegagropiles of the above-mentioned size does not seem improbable.

MODE OF OCCURRENCE

In the mode of occurrence of the carbonaceous inclusions in the Aitolahti phyllite formation, there are some features which further support the hypothesis of their aegagropilic origin. The fact that *Corycium* mainly occurs in coarse-grained metapsammitic beds, which may represent flood layers, is easy to understand if they could be thought to originate from ancient globular plant agglomerations, similar to the recent lake balls on the sandy shores of Kivijärvi. In addition, the larger carbonaceous *Corycium* balls occur not only embedded in the same horizon in sandy flood layers, 10—120 cm in thickness, but are also generally aligned in some irregular trains, similar to the appearance of the lake balls in Lake Kivijärvi. The variations in shape and size of the carbonaceous inclusions in the same trains also resemble the recent circumstances in Kivijärvi. Another special feature is the short length of the *Corycium* trains; according to Rankama (1948) only 10—20 meters. This also is in conformity with the conditions at known occurrences of recent lake balls which seldom occur in sections longer than 40 m.

The carbon-rich slate fragments of varying form described by Rankama (1948, p. 400) from metapsammitic as well as metapelitic layers, usually differ from *Corycium* proper in shape and structure. Their origin can well be tectonically loosened fragments of mud originating from ancient felty algal layers on the bottom, similar to them, which have supplied organic material to the recent lake balls in Kivijärvi. Common to all of the *Corycium* balls is their sharp limitation from the surrounding sediment, caused only by the abundance of carbon in the inclusions and their often rounded or elliptical shape.

CONCLUDING REMARKS

In places the *Corycium*-bearing varved schists in Aitolahti seem rather undisturbed, and Sederholm (1897) supposed that the Bothnian schists here have not undergone an intensive folding. Later investigations by Seitsaari (1951) and Simonen and Kouvo (1951) show, however, that metamorphism and tectonic deformation of the varved strata can be locally very intensive. According to Simonen and Kouvo *(op. cit.)* the coarse-grained sandy parts of the varves have a predominant graywacke character and show that the depth of the water has not been very great. The continuous sedimentation of thick varved strata has been possible through the downsinking of the geosynclinal basin during the course of accumulation. The high boron content of the varved schists suggests marine conditions of deposition.

In places the varved schist layers in Aitolahti show indisputable folding by sliding movements similar to those described by v. Straaten *(op. cit.*, pp. 11—13) from varved glacial clays in Jakola, SE of Imatra. Aurola (1960,

p. 34) later discovered a similar folding by sliding in coarser sediments of glacifluvial character in a sand pit in Turku (Åbo).

Against the background of the violent and intense compression which the sliding movements caused in the sediments in Jakola and Turku, the hypothesis of a tectonic origin for all of the free embedded *Corycium* inclusions in slightly disturbed arenaceous flood layers does not seem probable. On the other hand, a hypothesis that *Corycium* are fossil remains of ancient sand-bearing aegagropilous forms of algae, or eventually other fibrous plant material, can explain the mode of occurrence of the carbonaceous inclusions in the coarse-grained sandy parts of the varved schists, in spite of the destruction during metamorphism of every primary plant structure in the balls. The latter hypothesis can also explain their great amount of carbon and their resistance in morphologic respects.

According to age data presented by Kouvo (1958, p. 64) there is a lower limit of 1 800 million years on the age of formations in Finland in which probable signs of ancient life have been found. The findings of identifiable filaments of at least two different blue-green algae described by Tyler and Barghoorn *(op. cit.)* from the Gunflint formation of Ontario, with a probable age of abt. two billion years, could give support to the assumption that filamentous blue-green algae have taken part in the formation of the ancient marine algal balls which are buried and preserved as fossil remains in the varved schists in Aitolahti.

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21

ON THE ALBITE OF GRANITIC ROCKS [1]

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INTRODUCTION

In his recent paper, Schermerhorn (1960) advocated the idea that all the albite in granites is secondary. He says *(op. cit.*, p. 123): »In granites it is common to find lower-temperature minerals with varying amounts of higher-temperature relics in a higher-temperature fabric on which lower-temperature textures have been superposed to varying extents». This is perfectly true, and is a relatively often observed phenomenon in many granitic rocks. But when Schermerhorn extends his generalization to include the actual albite granites with granoblastic texture and separate grains of microcline, albite, and quartz without any undisputable signs of »decalcification» as he also claims for them, his statements, in the opinion of the present writer, are highly unwarranted.

This statement by Schermerhorn will be dealt with in this paper, in particular.

THE DECALCIFICATION OF PLAGIOCLASE

The essential statement of Schermerhorn *(op. cit.*, p. 129) is that »the primary plagioclase is now pseudomorphosed by secondary albite». This means, according to him that the anorthite component of the plagioclase has been decalcified, and in extreme cases, Ca has been completely removed from the rock.

Before going into the petrographic evidence of this feature, which is evident in the initial stages in many rocks, but certainly not in all, it may be of use to discuss this problem from the chemical point of view as well.

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If starting from the components of a plagioclase, the »decalcification» is actually bound to the anorthite component. Under the microscope we can see that this phenomenon results in contemporaneous »albitization» of the feldspar, and sericitization or saussuritization. Chemically, the former implies an introduction of potassium, and the latter implies an introduction of water, or of sodium silicate, silica and water.

If the »decalcification» of a plagioclase is caused by the introduction of potassium, then at least the following possibilities exist:

(1)
$$\underset{\text{anorthite}}{4(CaO.Al_2O_3.2SiO_2)} + \underset{\text{K-feldspar}}{K_2O.Al_2O_3.6SiO_2} + Al(OH)_3 \rightarrow$$
$$\underset{\text{epidote}}{4CaO.3Al_2O_3.6SiO_2.H_2O} + \underset{\text{muscovite}}{K_2O.3Al_2O_3.6SiO_2.2H_2O} + 2SiO_2.$$

(2)
$$\underset{\text{anorthite}}{4(CaO.Al_2O_3.2SiO_2)} + K_2O.SiO_2 + 3SiO_2 + H_2O \rightarrow$$
$$\underset{\text{epidote}}{4CaO.3Al_2O_3.6SiO_2.H_2O} + \underset{\text{K-feldspar}}{K_2O.Al_2O_3.6SiO_2}.$$

(3)
$$\underset{\text{anorthite}}{3(CaO.Al_2O_3.2SiO_2)} + K_2CO_3 + 2H_2O + 2CO_2 \rightarrow$$
$$\underset{\text{muscovite}}{K_2O.3Al_2O_3.6SiO_2.2H_2O} + \underset{\text{calcite}}{3CaCO_3}.$$

Only third equation allows the removal of calcium (in aqaeous solution) from the system, and then muscovite will be the sole product of the »decalcification» of anorthite. Even then the calcite would most probably stay in the rock body, if of a considerable size, as has sometimes been observed in albitized rocks.

The removal of epidote and muscovite from the system, on the other hand, is entirely unlikely.

If the »decalcification» of plagioclase is caused by the introduction of water or sodium-bearing compounds, at least the following possibilities exist:

(4)
$$\underset{\text{anorthite}}{4(CaO.Al_2O_3.2SiO_2)} + H_2O \rightarrow \underset{\text{epidote}}{4CaO.3Al_2O_3.6SiO_2.H_2O}$$
$$+ Al_2O_3 + 2SiO_2$$

(5)
$$\underset{\text{anorthite}}{4(CaO.Al_2O_3.2SiO_2)} + Na_2O.SiO_2 + 3SiO_2 + H_2O \rightarrow$$
$$\underset{\text{epidote}}{4CaO.3Al_2O_3.6SiO_2.H_2O} + \underset{\text{albite}}{Na_2O.Al_2O_3.6SiO_2}$$

$$(6) \quad \underset{\text{anorthite}}{CaO.Al_2O_3.2SiO_2} + Na_2CO_3 + 4SiO_2 \text{ (in water solution)} \rightarrow \underset{\text{albite}}{Na_2O.Al_2O_3.6SiO_2} + CaCO_3$$

Of equations (4) through (6), only the last is such that the total removal of Ca would be theoretically possible. In the other reactions, epidote would be formed and also undoubtedly remain in the rock.

Consequently, even theoretical chemical consideration of the possible decalcification of the plagioclase does not support the complete removal of Ca from the system, but only the reconstruction of minerals.

$CaCO_3$ formed in reactions (3) and (6) is, of course, to some extent water-soluble, but even then it would undoubtedly precipitate very early in the rock as calcite, unless there are fissures along which the hot water could carry it out from the rock. However, there would still be places in which $CaCO_3$ would be retained by the rock. Equation (3) implies potassium- and (6) implies sodium-metasomatic features.

In this connection, the experiments carried out by Assarsson (1960) may be of interest. He found that at a temperature of 120° to 220°C, microcline reacts with $Ca(OH)_2$ in the following way *(op. cit.*, p. 63):

$$\text{potash feldspar} + Ca(OH)_2 \rightarrow KOH + CaO.SiO_2.H_2O + 2CaO.SiO_2.H_2O + 3CaO.Al_2O_3\,(SiO_2/H_2O)_3 + m\,CaO.\,n\,SiO_2.\,o\,Al_2O_3.\,x\,H_2O.$$

From this reaction one may conclude that the potassium is released in the form of water-soluble KOH which may be removed from the system and/or act elsewhere producing potash metasomatism. CaO, on the other hand, forms insoluble silicates, some of which under somewhat different conditions, may easily form epidote. Such a chemical composition (clinozoisite: $4CaO.3Al_2O_3.6SiO_2.H_2O$) is actually already represented by the last compound in the above-cited formula.

Petrologically, the end products of the above reactions can usually be seen under the microscope, but, which compounds have caused the reaction involved, cannot be determined. This is displayed by the microphotographs in Plate I:

In Fig. 1, much muscovite and some epidote occur as the reaction products. In Fig. 2, the microcline is replacing plagioclase, and the sericite has been formed contemporaneously. These examples are the products of potassium metasomatic processes, and correspond to equations 1 through 3.

Fig. 3 illustrates well reactions (4) through (6) in which epidote, (or epidote and albite) have been formed. Typically, all the reaction products are either retained by the »decalcified» plagioclase, or they have been transferred only a short distance. In the case of Fig. 3, the plagioclase still has the composition of oligoclase.

Furthermore, all these reactions mainly occur in the post-kinematic orthoclase-granites or in syenitic and monzonitic rocks preferably containing strongly-perthitic potash feldspars, but they are also very common in the synkinematic gneisses and granodiorites (Figs. 2 and 3), in any granitized rocks, and also in the recrystallized granite porphyries (Fig. 1). Therefore, there is a strongly-supported belief, that the saussuritization and sericitization may, in very many cases, be caused by potassium metasomatism. This conclusion is made on both chemical and petrological bases.

As may be seen in Fig. 2, microcline is often replacing strongly sericitized plagioclase, and occupies fields with diffuse margins within the crystals of plagioclase.

The most plausible cause of epidotization of the anorthite component of plagioclase would, however, be water which according to equation (4), would produce alumina and silica in addition to epidote. The reaction takes place in the presence of hot water, the aluminium hydroxide would probably be formed instead of Al_2O_3. The aluminous minerals, however, are rather unusual in connection with saussuritization of a plagioclase; and because muscovite is almost invariably present in such cases, the contemporaneous introduction of potassium must be involved in the reaction. Therefore, equation (2) would also work, even if in some modified form. Thus the following equation may best portray this phenomenon:

$$(7) \qquad \underset{\text{anorthite}}{12(CaO.Al_2O_3.2SiO_2)} + K_2O.SiO_2 + 5H_2O \rightarrow$$

$$\underset{\text{epidote}}{3(4CaO.3Al_2O_3.6SiO_2.H_2O)} + \underset{\text{muscovite}}{K_2O.3Al_2O_3.6SiO_2.2H_2O} + SiO_2$$

Removal of the Ca from the rock by volatile agents is entirely out of the question because although $CaCO_3$ could, of course, be removed by water, it could hardly be completely removed from the rock. HF forms very insoluble CaF_2, which often occurs in the latekinematic granites and is omnipresent in the rapakivi granites.

Furthermore, in all the reactions proposed in the equations (1) through (7), the essential products of such a »decalcification» are epidote and muscovite. These compounds are certainly retained by the rock and will undoubtedly remain as proof of the decomposition of the anorthite component of the plagioclase.

In addition, both these components are usually formed inside of the sericitized or saussuritized crystal of plagioclase, and also inside of a more albitic marginal portion of plagioclase, if present.

In ore geology, on the contrary, there are examples showing that during such an »epidotization», the alkalies expelled producing albite rocks outside, as has happened in the Catoctin formation, Virginia (Grophy, 1960), where the calcium of plagioclase was retained in the lattice of epidote.

FELDSPARS OF THE LATEKINEMATIC MICROCLINE GRANITES

As mentioned above, all observed saussuritization and sericitization of plagioclase (the »decalcification» of Schermerhorn) as described in the foregoing chapter, are typical in synkinematic granitized rocks, in many orthoclase granites, and also in granites, syenites and manzonites containing strongly perthitic microcline (usually of more inferior triclinicity than is the microcline of granitized rocks and of most latekinematic granites).

There is, however, a group of microcline granites which have been termed latekinematic and which generally have an ideal granite composition. In these, usually pink, fine- to medium-grained rocks which tend to be leucocratic or aplitic, the plagioclase is overwhelmingly albitic and often unaltered. The microcline there has formed well separated grains and it is, in most cases, of almost highest triclinicity. Figure 4 of Plate I illustrates a detail of the texture of such aplitic, latekinematic granite which contains perfectly fresh albite and highly triclinic microcline.

Such granites are comparatively common in Precambrian areas such as Finland, Canada, and West-Africa, and no signs of possible large scale decalcification of plagioclase can be seen. They are true microcline-albite granites, and such granite bodies are compositionally rather homogeneous.

As shown in the foregoing chapter, even if plagioclase is »albitized», the calcium is still retained in the rock and especially incorporated in epidote. Therefore, and because microcline-albite granites are usually rich in muscovite and do not contain any amphibole or pyroxene, the modal composition as calculated from the chemical rock analysis would give the »original» composition of the plagioclase. Such a calculation may be easily made. In Table 1, CaO, alkalies and modal feldspars are compiled for a set of Sierra Leonean latekinematic granites (Wilson and Marmo, 1958). There the An-content of plagioclase is 4.5 to 7.5 and the mean of 5 analyses is 5.5.

Table 1. Granites from Sierra Leone (Wilson and Marmo, 1958).

	VM 162	VM 199	VM 226	VM 309	VM 587
CaO	1.1	0.69	1.8	0.8	0.6
Na_2O	4.77	4.08	5.16	4.92	3.63
K_2O	4.95	4.81	3.85	4.77	5.04
Albite	38.3	36.0	43.5	42.0	30.7
Anorthite	2.0	1.7	2.3	2.5	2.5
Microcline	22.9	25.0	22.5	26.1	29.4
% An	5	4.5	5	5.5	7.5

Table 2. Some granites from Finland.

	1	2	3	4	5	6	7	8	9	10
CaO	0.21	1.63	1.56	1.11	1.28	0.84	1.25	1.17	0.74	0.56
Na_2O	2.73	4.42	3.98	3.65	4.12	4.16	4.07	4.75	4.40	3.15
K_2O	5.58	3.82	4.48	4.62	4.82	5.72	3.61	4.97	4.57	4.90

1 = Mätäsvaara (Kranck, 1945).
2 = Parsinmäki, Lapua (Hietanen, 1938).
3 = Ranua (Lokka, 1950).
4 = Inkoo (Lokka, 1934).
5 = Kirjavalahti, Sortavala (Lokka, 1934).
6 = Pitkäranta (Lokka, 1934).
7 = Leppävirta (Lokka, 1934).
8 = Trifona, Petsamo (Lokka, 1934).
9 = Ylivieska (Lokka, 1934).
10 = Virrat, between Kotala and Piili (analyst: A. Heikkinen).

In Table 2, some Finnish microcline-albite granites are shown, and again the composition is very similar. In both cases similar anorthite contents of the plagioclase have also been optically observed.

On the other hands, epidote-bearing latekinematic granites are also known in which the epidote could be interpreted as having been derived from the anorthite-component of the plagioclase. Still, these granites are in exactly similar metamorphic, tectonic and geologic environments, with the CaO still in the rock (as epidote), but the optically determined An-content of the plagioclase is definitely lower than that calculated from the rock analyses. This, of course, is also expected and proves that the removal of Ca from the rock, in connection with the »decalcification» of the plagioclase, is very unlikely. In true microcline-albite granites, the albite, as well as microcline, is primary and has original composition and crystallography.

DISCUSSION

On the foregoing pages an attempt was made to show that in equigranular, fine- to medium-grained microcline-albite granites, the albite is not a product of »decalcification» and of subsequent total removal of Ca from the rock, but, if epidote is sparse or absent, is of primary character. This generalization applies only to these particular granites.

From Schermerhorn's paper *(op. cit.)*, the present author obtained the impression that his aim in constructing his theory, was to force also these granites (which are rather abundant among t r u e granites) into the category of magmatic rocks.

There are, in these granites, at least the following characteristics which do not fit into magmatic theory:

1. The co-occurrence of potash feldspar and albite forming separate grains;
2. The presence of obviously primary microcline which cannot be formed under magmatic conditions *(e. g.* Marmo, 1959).

I have discussed these questions in several papers *(e. g.* Marmo, 1957, 1959), and therefore my arguments and conclusions will not be repeated here. I would only like to say again what I have stressed earlier (Marmo, 1960, p. 303): »There are two ways to interpret certain phenomena occurring in nature: either we adopt some general theory and then attempt to make all observations made in the field fit into this theory, . . ., or each phenomenon observed is attempted to interprete in easiest and most plausible way». I prefer the latter alternative.

In this case, Schermerhorn *(op. cit.,* p. 131) wrote: »There is one class of albite granites, however, in which the texture is granoblastic so that from microscopic evidence alone . . . nothing can be said about its primary or secondary origin».

His simple interpretation for these rocks is that they »have been recast by metamorphism». Why?

To give an unwarranted explanation only in order to make the observation fit a certain theory, is not the correct way to handle phenomena of nature. This is still less so, because there are many other explanations at hand which have been discussed and proposed by several authors. These explanations cannot be rejected just by neglecting and not even mentioning them.

The presence of potash feldspar with albite has been explained as being possible if the emplacement takes place under hydrothermal conditions and at temperatures well below that at which microcline is disordered. At the same conditions, but if the emplacement is simultaneously sluggish, the microcline may be formed primarily.

These statements must be disproved and in addition the statements that: 1. All existing microcline has originally been monoclinic (there is so far no proof of this) and 2. All albite in granites is due to the expulsion of calcium from the whole rock, must be proved before the theory set forth by Schermerhorn begins to be warranted.

To summarize: Certainly there are granites which contain albitized plagioclase (p. 394). The orthoclase granites as well as some syenites and monzonites containing perthitic microcline of inferior triclinicity plus oligoclase, may fullfil the requirements of a magmatic origin, but certainly there are also albite-microcline granites which do not fall into this category.

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EXPLANATIONS TO THE PLATE I

Fig. 1. Granite porphyry. Pihlajavesi, Kolmisoppijärvi, Finland. Resorbed phenocryst of plagioclase containing large plates of muscovite and dispersed fine scales of sericite plus grainlets of epidote. N +, 20 ×.

Fig. 2. Granodiorite. Ähtäri, 350 m W of Rintala. Sericitic plagioclase grain is replaced in patches by microcline. N +, 20 ×.

Fig. 3. Granodiorite. Pihlajavesi, 1 km NW of the Karimojärvi. Saussuritic inset of oligoclase containing coarse crystals of epidote in the centre. N +, 25 ×.

Fig. 4. Aplitic granite. Pihlajavesi, 3 km NE of Lappi. Fresh albite and microcline form separate grains in granoblastic texture. N +, 25 ×.

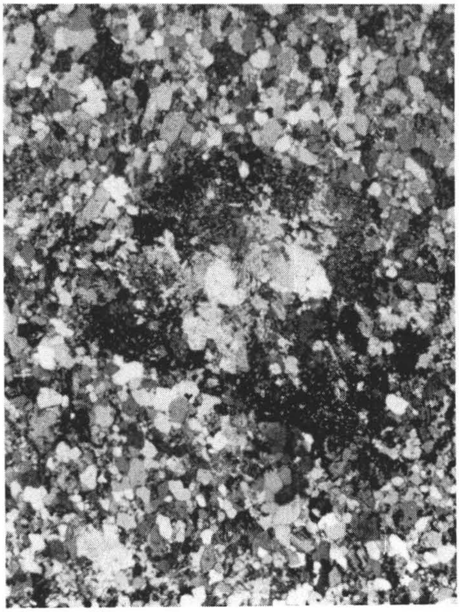

Fig. 1

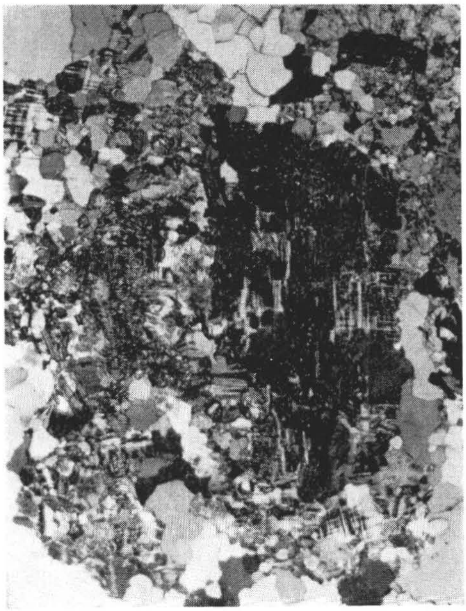

Fig. 2

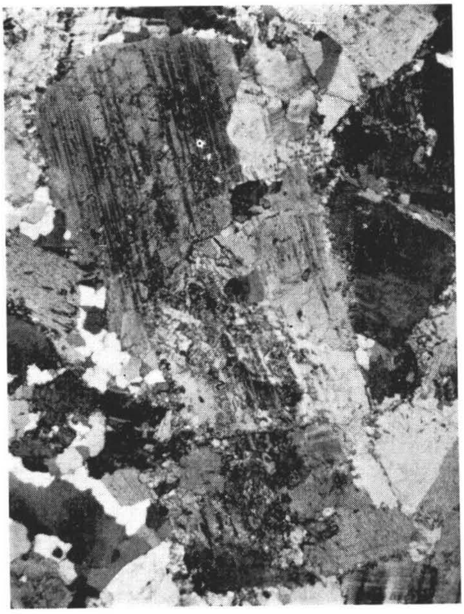

Fig. 3

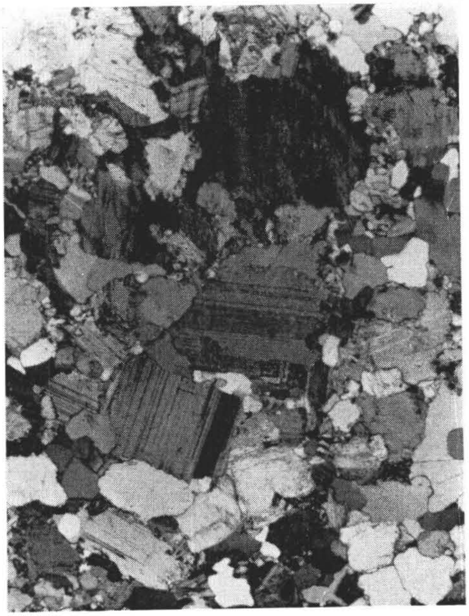

Fig. 4

V. Marmo: On the Albite of Granitic Rocks.

22

A NEW APOPHYLLITE OCCURRENCE IN THE VIIPURI RAPAKIVI AREA [1]

BY

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ABSTRACT

Apophyllite is a rather rare mineral in Finland and has previously been described only from the Viipuri rapakivi area. In this paper a new occurrence of this mineral from the same rapakivi area is presented, including x-ray data, optics, chemical composition, DTA, and GTA.

OCCURRENCE

This apophyllite occurrence was found in 1958 during planning for the XXI International Geological Congress excursions. The locality is situated near Langinkoski park in the NW-part of the town of Kotka, S-Finland. Two hundred meters south of the Langinkoski coffee house there is a new road cut in rapakivi granite. The granite is a gray green, sometimes reddish, porphyritic rock. Orthoclase ovoids often surrounded with plagioclase occur as phenocrysts, but the largest ovoids are often devoid of plagioclase mantles. The average size of the ovoids is 2—5 cm. The groundmass is mainly medium-grained, consisting of reddish orthoclase, greenish plagioclase, grayish quartz, and dark biotite. The granite contains angular fragments of svecofennian granodiorite, microcline granite, pegmatite, mica gneiss, and migmatites, measuring from 20 cm to many meters in diameter. In addition pegmatites consisting of microcline perthite, albite, and quartz and belonging to the rapakivi have been found. The pegmatites sometimes occur as small lenses but usually their shape is irregular and they are only some ten or twenty cm in diameter. Often there is a miarolitic cavity in the middle

[1] Received February 28, 1961.

of these pegmatites, the walls covered with albite, apophyllite, fluorite, and calcite. In addition aggregates of garnet and pistasite have been found in a coarse-grained variety of this granite at the same exposure. These, the author supposes, are due to contamination of the rapakivi granite by migmatite inclusions.

MINERAL DESCRIPTION

The great majority of the pegmatites consist of microcline perthite, albite, quartz, and biotite. In many places microcline is graphically intergrown with quartz. Albite, when occurring on the walls of miarolitic cavities usually forms white idiomorphic crystals up to two centimeters in diameter. The composition of plagioclase has been determined by the x-ray method of Smith (1956) and the result indicated 95 percent albite and 5 percent anorthite.

Calcite, apophyllite, and fluorite seem to be younger than the above-mentioned pegmatite minerals. Dark green aggregates of very fine-grained materials, which proved to be a mixture of chlorite, muscovite and calcite (x-ray analysis), but in some places exclusively chlorite, also belong to this generation or perhaps to a younger one. The fluorite crystals are as large as two cm in diameter. The colour grades from dark violet to colourless.

Table 1. Chemical compositions of apophyllites from Viipuri rapakivi area.

	Wt. per cent					Cation proportions				
	1	2	3	4	5	1	2	3	4	5
SiO_2	51.58	52.64	52.06	52.19	52.99	858	877	867	868	882
TiO_2	0.00									
Al_2O_3	0.6	0.20								
Fe_2O_3	0.07	0.20								
FeO										
MnO	0.01									
MgO	0.54									
CaO	23.78	24.37	24.79	25.20	24.72	424	434	442	449	441
Na_2O	0.20	0.45				7	14			
K_2O	5.20	5.12	5.38	6.12	5.19	110	108	114	130	110
P_2O_5	0.05									
H_2O+	16.1	15.58	16.51	16.44	15.89	894	866	918	913	882
H_2O-	0.94	0.71								
F	1.70	0.80	0.84		2.09	90	42	44		110
	100.77	100.07	99.58	99.59	100.88					
$-O = F_2$	0.72	0.33			0.88					
	100.05	99.74			100.00					

1. Langinkoski, Kotka, Finland. Anal. Heikkinen 1960.
2. Ihalainen, Lappeenranta, Finland. Anal. Volborth, 1953.
3. Pyterlahti, Finland. Anal. Beck, 1862.
4. Pyterlahti, Finland. Anal. Beck, 1862.
5. Theoretical composition of apophyllite.

Table 2. Physical properties of apophyllites from Viipuri rapakivi area.

	N_O(Na)	N_E(Na)	N_O—N_E	Specific gravity
1.	1.5335	1.5365	0.003	2.368
2.	1.5345	1.5375	0.003	2.360
3.	1.5330	—	<0.001	2.375
4.	1.534	1.538	0.004	2.375

1. Langinkoski (measured at 22°C).
2. Ihalainen. Volborth, 1953.
3. Pyterlahti. Volborth, 1953.
4. Pyterlahti. Volborth, 1953.

Calcite occurs as large white grains measuring many cm in diameter. Apophyllite, the youngest of these three minerals, often forms radiating aggregates. In many places apophyllite brecciates the older pegmatite minerals.

Usually the apophyllite is sulphur yellow and individual crystals are 1—4 cm in diameter. In many places a two mm thick brown apophyllite rim has been found between yellow apophyllite and other pegmatite minerals. Sometimes the apophyllite occurs as very fine-grained colourless flakes. In every case the mineral has been identified by the x-ray diffractometer or the Debye-Scherrer camera.

The chemical analysis of the yellow apophyllite is presented in Table 1, and for comparison some other analyses from apophyllites from the Viipuri rapakivi area are also included. Taking only the main constituents into consideration, it is seen that the chemical composition of apophyllite from Langinkoski is very near that calculated from the ideal formula ($KCa_4FSi_8O_{20}.8H_2O$, Taylor and Náray-Szabó, 1931), except that OH substitutes for about 20 percent of the fluorine. Compared with other apophyllites from the Viipuri rapakivi area, the apophyllite from Langinkoski is the richest in fluorine.

The specific gravity and optics were very near those from the other localities. Refractive indices were determined by the single variation method. The results and comparisons are presented in Table 2.

The yellow apophyllite from Langinkoski was also studied by x-ray methods. The powder pattern was recorded with a Philips x-ray G—C diffractometer using silica as an internal standard (run ½°/min., CuKα radiation, Ni filter). The results are presented in Table 3 with the data of »ASTM 7—170». The intensities were measured with a planimeter. The d-spacings and intensities of apophyllite on the ASTM card are similar with those of the investigated mineral, but when 2θ is greater than 59°, there is a difference of one degree of 2θ between the reflections (marked with asterisks in Table 3).

Table 3. X-ray powder data for apophyllites.

	1.		2.	
(hkl)	d (Å)	I/I_0	d (Å)	I/I_0
002	7.88	9	7.85	9
101	7.81	8	7.76	9
103	4.54	15	4.534	20
200	4.47	2	4.473	3
004	3.943	100	3.943	100
211	3.882	5	3.882	6
104			3.606	2
212	3.581	15	3.570	10
114	3.347	8	3.347	9
220	3.168	5	3.168	5
105	2.981	60	2.976	70
301	2.940	4	2.934	6
214			2.812	3
311			2.788	2
312	2.670	2	2.664	3
313	2.500	10	2.493	15
215	2.483	20	2.479	30
116	2.429	8	2.428	10
216	2.204	3	2.199	5
107	2.191	3	2.187	7
402			2.155	3
315	2.114	5	2.109	10
324	2.108	6	2.103	10
420	2.007	4	2.004	6
	1.773	4	1.771	7
	1.768	10	1.764	10
	1.723	3	1.721	5
	1.678	3	1.675	3
			1.655	2
	1.624	1	1.620	2
	1.610	1	1.607	2
	1.579	40	1.578	55
	1.549*	4	1.522*	8
			1.510*	2
	1.494*	3	1.470*	4
			1.453	2
			1.442	2
	1.420*	2	1.398*	4
	1.352*	2	1.334*	4
	1.317*	4	1.300*	5
	1.283	2		
	1.242	2		
	1.177	3		

1. Apophyllite, Langinkoski, Kotka, Finland.
2. Apophyllite, ASTM 7—170.

Apophyllite was also subjected to differential thermal analysis (DTA) and thermo-gravimetric analysis (TGA). The DTA (Fig. 1) was made in a Kanthal REH 7—50 vertical furnace with a nickel sample holder (Leeds & Nordtrupp control unit with a Speedomax G Recorder). Quartz was used as an ineffective comparison material (L. & N, X/Y Recorder G). The rate

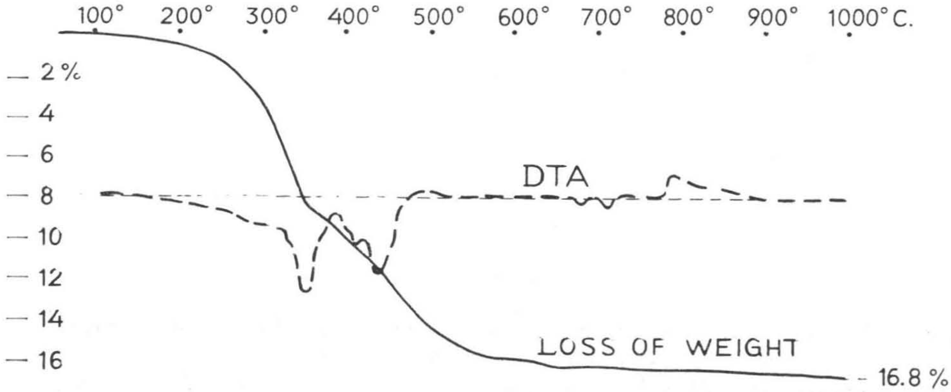

Fig. 1. Differential thermal analysis and dehydration curves of apophyllite.

of temperature increase was 10°C/min. The TGA (Fig. 1) was made with a Stanton Thermobalance HT—D, the temperature increase being 4.4°C/min.

As seen in figure 1, apophyllite has one strong endothermic break at 350°C and one at 440°C. In addition there is an exothermic break at 790°C. Comparing the DTA-curve with the TGA-curve, it is clearly seen that the first endothermic break at 350°C indicates the temperature where about half of the water content is lost. The second endothermic break indicates the point where greatest part of the remaining water is lost.

OTHER APOPHYLLITE OCCURRENCES IN THE VIIPURI RAPAKIVI AREA

Volborth (1953) describes the apophyllite from Ihalainen in the town of Lappeenranta. In the limestone quarry of Ihalainen lenses of pegmatite belonging to the surrounding rapakivi granite are present. The manner of occurrence of the apophyllite in these pegmatites is similar to that in Langinkoski.

From the rapakivi granite quarry of Pyterlahti, apophyllite has been found in coarse-grained, nearly pegmatitic granite (Beck, 1862) and to the south in Virolahti (Nordenskiöld, 1855) there is another small occurrence.

Acknowledgments — This work is a part of the reinvestigation of the Viipuri rapakivi area led by Dr. A. Simonen. The author wishes to thank him for many suggestions during the work, Mr. A. Toivonen, M. A. for the DTA and Mr. J. Hyyppä, M. A. for the TGA. To Miss. T. Åberg the author is indebted for drawing figure 1 and to Mr. R. Ojakangas, M. A. for correcting the English of this manuscript.

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23

AN X-RAY STUDY OF THE PHOSPHATE MINERALS FROM THE ALKALINE ROCK AREA OF SONGO, SIERRA LEONE [1]

BY

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ABSTRACT

Rockbridgeite ($FeFe_4(PO_4)_3(OH)_5$) and aluminian strengite ($(Fe, Al) PO_4 \cdot 2H_2O$) are major constituents of a dike-like occurrence in the ijolite at Songo, Sierra Leone. These minerals, as well as wardite ($Na Al_3 (PO_4)_2 (OH)_4 \cdot 2H_2O$) and cacoxenite ($Fe_4'''((OH)_3 (PO_4)_3) \cdot 12H_2O$) have been investigated by x-ray powder method. Weissenberg photographs have been taken of the rockbridgeite. The space group of this mineral is the same as reported for frondelite and zinkian rockbridgeite (D_2^5—$C222_1$), and the unit cell parameters proved to be; $a_0 = 13.84$ Å, $b_0 = 16.91$ Å, and $c_0 = 5.16$ Å.

The chemical analysis shows the Fe'' in rockbridgeite to be partly oxidized to Fe'''. Also leaching of some P_2O_5 has taken place. The chemical and spectrochemical analyses show that Al''' is substituted for part of the Fe'''.

CONTENTS

[1] Received February 22, 1961.

INTRODUCTION

C. O. Baker, Vladi Marmo, and M. K. Wells have described (1956) the ijolite area of Songo, Sierra Leone. They found that the ijolite stock is cut by a dark dike rock occurring as a heavily weathered outcrop and further traced by the use of boulders. The distribution of the boulders suggests that the ESE trending dike is 1 500 ft or more in length and 10 to 30 ft in width. The boulders are rusty, soft and strongly disintegrated. It was possible to get relative fresh samples from the only outcrop by blasting, but even then only small fragments fresh enough for study were found. The authors suggested at that time, solely on the basis of microscopy, that the rock consisted mainly of kataphorite with some zeolites and other minor constituents. Later Marmo paid more attention to this dike rock and came to the conclusion that the mineral taken for kataphorite, and resembling it in thin sections was some other species. At Marmo's suggestion the present author has tried to resolve the problem of this mineral. Without visiting the locality where the specimens were taken, it is impossible to give a complete description of the manner of occurrence and relations with the surroundings.

PETROGRAPHY

The specimens represent two kinds of rocks. The fresh or only partly altered dark, almost aphanitic, rock consists mainly of aggregates of rockbridgeite, spherulites of Al-strengite, and irregular grains of goethite (Plate I, Fig. 1.) The rockbridgeite is replaced in many places by very fine-grained minerals. In these altered areas, goethite and Al-strengite seem to be the prevailing alteration products. Where the alteration of the rock has proceeded further, these secondary minerals occur as small nodules. Where highly altered, they form a rusty, sometimes white or yellowish porous rock. Wardite and some unidentified minerals also occur in this zone.

In the fresh rock there also occur some small (diam. 1—10 mm) cavities, the walls of which are covered with beautiful radial aggregates of Al-strengite, limonite, and a golden yellow cacoxenite.

According to a rough estimate, the slightly altered rock consists of about 70—80 percent rockbridgeite, some 20 percent goethite, 5—10 percent Al-strengite, and replacement products in varying amounts. Although megascopically the rock seems to be fresh, varying degrees of alteration may be seen microscopically.

The chemical analysis from the best preserved rock is presented in Table 1, column 3. The most conspicuous features are the exceedingly high Fe''', phosphorus, and water contents, and the very small amounts of silica, magnesium, calsium, and alkaline.

Table 1. Analyses of rockbridgeite and phosphatite from Songo, Sierra Leone.

	1.	2.	3.
SiO_2		0.32	0.55
TiO_2		1.91	2.64
Al_2O_3		9.17	15.88
Fe_2O_3	49.20	49.20	42.22
FeO	11.07	5.40	5.20
MnO		0.08	0.12
MgO		0.16	0.13
CaO		0.60	1.27
Na_2O			0.85
K_2O			0.00
P_2O_5	32.80	22.40	22.08
H_2O+	6.93	8.72	10.47
H_2O-		0.98	0.93
Total	100.00	99.02	100.34

1. Theoretical composition of rockbridgeite. Calculated from $Fe''Fe_4'''(PO_4)_3(OH)_5$ (Lindberg, 1949).
2. Rockbridgeite. Songo, Sierra Leone. Anal. P. Ojanperä.
3. Phosphatite. Songo, Sierra Leone. Anal. P. Ojanperä.

ROCKBRIDGEITE, $Fe''Fe_4'''(PO_4)_3(OH)_5$

Rockbridgeite (Plate I, Fig. 2) forms fibrous masses, often with a radial structure. The length of the fibers is usually under 0.2 mm, but occasionally exceeding to 1 cm. According to thin section examinations, the fresh rock is in many places almost monomineralic. Rockbridgeite is extreemly strongly pleochroic with Z showing maximum absorption. The pleochroism varies from deep green to deep brown. Concentric colour banding is often found in the middle of the radial aggregates, where the pleochroism varies from deep blue to pale brown (Plate II, Fig. 1). The indices of refraction seem to vary from grain to grain, being between 1.845—1.895.

To obtain pure material for chemical analysis, attempts were made to separate rockbridgeite by physical methods. The rock was ground to 140—200 mesh after which it was separated by heavy liquids in a centrifuge. After numerous treatments, a fraction with specific gravity of 3.40 was obtained. It was estimated to contain 3—5 percent impurieties, but further separation did not give any better results. Perhaps the greatest part of these impurieties is fine-grained goethite.

The chemical analysis is presented in Table 1, column 2. In column 1, the theoretical composition of rockbridgeite is presented. It is easily seen that a remarkable part of the Fe''' is substituted by Al, and that the material is partly oxidized by conversion of Fe'' to Fe'''. Also it is evident that a leaching of P_2O_5 has taken place. Frondel (1949) describes rather accurately the alteration phenomena of rockbridgeite. Therefore the present author refers only to the above paper.

Table 2. X-ray powder data for rockbridgeites.

	1.			2.		3.	
(hkl)	$d_{obs.}$	$d_{calc.}$	I	d	I	d	I
020	8.37	8.45	10	8.36	10	8.41	10
200	6.93	6.92	20	6.94	20	6.90	20
				6.46	10		
101	4.82	4.83	20	4.85	10	4.83	10
111	4.68	4.65	10	4.67	10	4.64	10
230	4.33	4.37	5	4.34	10	4.34	10
040	4.18	4.23	10	4.20	10	4.19	10
240	3.60	3.61	40	3.58	40	3.58	30
301	3.44	3.44	30	3.431	20	3.43	10
410	3.39	3.39	40	3.391	50	3.37	20
311		3.37					
321	3.175	3.186	100	3.186	100	3.18	100
420		3.202					
250	3.016	3.038	10	3.017	30	3.02	30
430	2.940	2.949	5	2.934	10	2.94	10
						2.85	10
151	2.774	2.772	30	2.763	30	2.76	30
440	2.675	2.675	5	2.670	20	2.67	20
002	2.582	2.580	20	2.589	20	2.59	30
351	2.412	2.411	50	2.415	40	2.42	50
						2.33	10
270	2.269	2.281	5	2.269	20	2.26	20
				2.26	10	2.23	10
171	2.162	2.161	5	2.169	10	2.16	10
080	2.110	2.114	5	2.106	20	2.11	10
412	2.059	2.053	20			2.06	30
640	2.019	2.025	5	2.019	20	2.02	20
252	1.966	1.966	10	1.966	30	1.96	30
				1.930	10	1.94	10
062	1.902	1.903	<5	1.897	10	1.90	20
701	1.840	1.846	20	1.836	30	1.84	30
262		1.835					
480	1.805	1.804	5			1.80	10
800		1.730		1.731	10	1.75	10
272	1.713	1.709	5	1.709	10	1.71	10
0.10.0	1.688	1.691	10	1.688	20	1.69	20
490	1.648	1.651	10	1.637	20	1.64	20
840	1.599	1.601	50	1.592	50	1.59	80
680	1.553	1.558	5	1.551	20	1.55	10
652	1.536	1.533	20			1.53	30
4.10.0	1.511	1.519	<5	1.513	10	1.51	10
292	1.477	1.482	5	1.478	10	1.48	10
353	1.459	1.455	5	1.455	10	1.46	10
163		1.460					
4.11.0	1.395	1.406	10	1.393	10	1.39	10
941		1.392					
2.12.0	1.368	1.380	<5				
1.12.1	1.345	1.352	<5				
4.12.0	1.294	1.305	30			1.29	30
971	1.256	1.258	10	1.255	20	1.26	30
7.10.1	1.245	1.247	10	1.244	10	1.24	10
4.13.0	1.214	1.217	<5				
981	1.208	1.209	<5				
434	1.188	1.182	<5				
11.3.1		1.194					
12.0.0	1.152	1.153	5				
11.5.1		1.150					

	1.			2.		3.	
(hkl)	$d_{obs.}$	$d_{calc.}$	I	d	I	d	I
0.14.2	1.089	1.093	10				
2.14.2	1.078	1.080	<5				
4.15.0	1.067	1.072	<5				
10.11.0	1.027	1.028	<5				
10.9.2		1.023					
7.14.1	1.009	1.010	<5				
13.5.1	0.996	0.996	5				
14.0.0	0.988	0.988	<5				
12.9.0	0.983	0.983	5				

1. Rockbridgeite from Songo, Sierra Leone. (FeKα-rad., Mn-filter).
2. » » Rockbridge, County, Va. (Lindberg, 1949).
3. » » (Frondel, 1949).

Frondel also suggests that a substitution between Fe''' and Al may take place in rockbridgeite. The analysis in Table 1 was made from a slightly impure material. The presence of aluminium was verified spectrochemically in definitely pure rockbridgeite grains. The spectrochemical analysis showed abundant iron and phosphorus and considerable amounts of aluminium. Thus, the rockbridgeite from Songo is aluminian rockbridgeite.

As yet, there is not much x-ray data available for rockbridgeite. Frondel (1949) gives the d-spacings for rockbridgeite and Lindberg (1949) the d-spacings, their indices, the unit cell dimensions, and the space group for frondelite which is the equivalent MnFe''' mineral. Lindberg and Frondel (1950) give the cell dimensions for zinkian rockbridgeite which contains 5.20 percent ZnO. All these investigations prove rockbridgeite to be isostructural and isomorphous with frondelite. The present author has conducted an x-ray investigation of the rockbridgeite from Songo, and the results of the powder method and single crystal work are presented below.

The powder patterns were recorded with a Debye-Scherrer camera (diameter 57.3 mm) using filtered FeKα-radiation (FeKα_1 = 1.93597 Å, FeKα_2 = 1.93991 Å, and FeKα = 1.9375 Å). The reflections were indexed with the aid of rotation and Weissenberg photographs. Those few reflections with l-indices of 3 or 4 were indexed analogically with the reflections of frondelite (Lindberg, 1949). The intensities are estimated visually. The results of the powder method are presented in Table 2 with the d-values for rockbridgeites published by Lindberg (1949) and Frondel (1949). The powder data is for the pleochroic (green to brown) mineral described above, but the powder pattern for the mineral with blue pleochroism was similar. There were only some differences in the intensities of the reflections. The new powder pattern agrees very well with those published by Lindberg and Frondel.

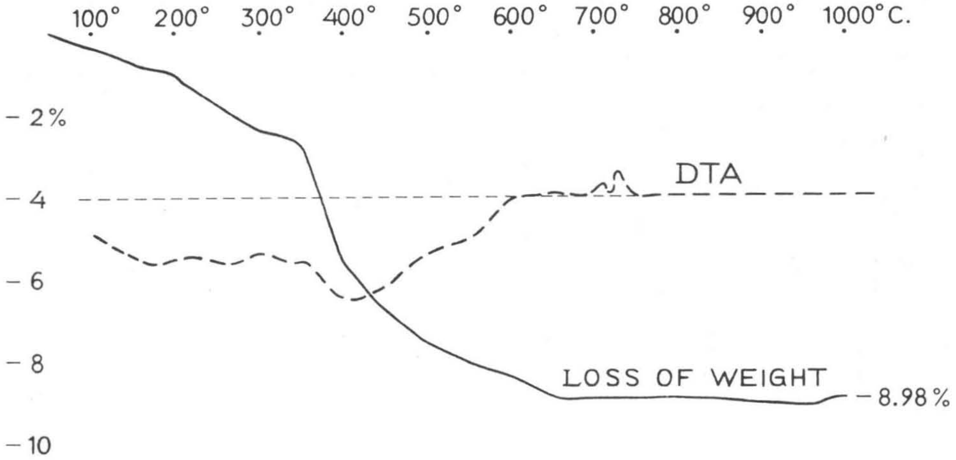

Fig. 1. Differential thermal analysis and dehydration curves of rockbridgeite.

The single crystal photographs were taken with a Weissenberg camera. The crystal was rotated about the axis [001]. The Weissenberg photographs were taken from zero, first, and second layer lines (equi-inclination method). Every layer had C_{2l} symmetry. By analogy with frondelite, an orthorombic mineral is indicated (V_h). The conditions for non extinctions were: on (hkl) h + l even, k even or odd; on (0kl) l even, k even or odd; on (h0l) h + l even; on (hk0) h even, k even or odd; on (h00) h even; on (0k0) k even; and on (00l) l even. Diffraction symmetry is D_{2h} -mmm; diffraction symbol mmm C— —2_1. This refers to the space group D_2—$C222_1$.

a_0 is calculated from the $d_{(14.0.0)}$-spacing, the accurate value of which is 0.9884 Å; b_0 is calculated from the $d_{(4.\ 15.\ 0)}$-spacing (0.9833 Å); and c_0 is measured from the c-axis rotation photograph. The cell parameters proved to be: $a_0 = 13.84$ Å, $b_0 = 16.91$ Å, and $c_0 = 5.16$ Å. The volume of the unit cell is 1208 Å^3. The corresponding unit cell parameters calculated from powder photographs by Lindberg for the rockbridgeite from Rockbridge were 13.73 Å, 16.82 Å, and 5.18 Å respectively.

Rockbridgeite from Songo was also subjected to thermal analyses. The DTA (differential thermal analysis) and TGA (thermo- gravimetric analysis) curves for this mineral are presented in figure 1. The DTA was made in a Kanthal REH 7—50 vertical furnace with a nickel sample holder (Leeds & Nordtrupp control Unit with Speedomax G Recorder). Quartz was used as an ineffective comparison material (L & N X/Y Recorder G). The rate of temperature increase was 10°C/min. The TGA was made with a Stanton Thermobalance HT—D, the temperature increase being 4.4°C/min.

According to the TGA, there is a constant loss of weight between 50°—350°C. At the latter temperature, a sudden increase of dehydration occurs and continues until 600°C. According to the DTA-curve, endothermic reaction occur until 600°C, the clearest endothermic break peaking near 400°C. There are two weak exothermic breaks at a little more than 700°C. A comparison of the DTA and GTA curves shows that they coincide very well.

According to Frondel (1949), rockbridgeite has an exothermic break near 650°C, a weak endothermic break near 350°C, and possibly others. The differences between these breaks and those of rockbridgeite from Songo are small and may be partly due to compositional differences and partly due to strong alteration of the rockbridgeite from Songo.

ALUMINIUM STRENGITE, $(Fe, Al)PO_4 \cdot 2H_2O$

The mineral identified as aluminian strengite occurs as small spherical aggregates between the rockbridgeite aggregates. Occasionally the walls of small cavities are covered by colourless aggregates of Al-strengite. The biggest colourless aggregates consist of radial needles measuring to 0.2—0.3 mm. In addition, the mineral forms larger nodules with a white or yellowish tint. The Al-strengite in these nodules is so fine-grained that it is difficult to distinguish the individual grains under the microscope. From these nodules, there is a gradual change to an altered yellowish, brownish or white porous material that is described below.

During the separation of rockbridgeite, a fraction with a specific gravity under 2.8 was also separated, and Al-strengite proved to be the prevailing species. The minerals from this fraction were investigated by means of x-ray powder photographs. Al-strengite occured as colourless needles, as turbid white grains, and as dirty green almost opaque grains. All the coloured varieties were remarkably softer than the colourless ones. The powder patterns were all similar except that a broad line at 2.74 Å (two adjoining lines at 2.72 Å and 2.76 Å) occurred in the patterns of coloured grains. This line was strongest in more highly coloured grains. At the same time the characteristic lines of Al-strengite became weaker and more diffuse.

The d-spacings of aluminian strengite from the open cavities are presented in Table 3. The d-spacings of variscite, Fe-variscite, and strengite are also presented. The gradual increase of the d-spacings with the increase of Fe-content is clearly seen.

CACOXENITE, $Fe_4'''((OH)_3(PO_4)_3) \cdot 12H_2O$

This mineral occurs in the small cavities with or without Al-strengite, forming tiny acicular [001] crystals 0.1—0.3 mm in length. Usually the

Table 3. Comparison of powder data of the minerals belonging to the variscite-strengite series.

1.		2.		3.		4.	
d	I	d	I	d	I	d	I
6.31	10					6.1	40
5.31	100B	5.42	100	4.45	90	5.5	80
4.81	60	4.86	70	4.88	50	4.9	60
4.54	10						
4.26	100B	4.29	100	4.30	100	4.36	100
3.92	60	3.94	70	3.95	50	4.00	60
3.62	40	3.66	50	3.68	50	3.70	50
				3.50	5		
3.35	10	3.36	20	3.38	5	3.43	40
3.20	20	3.24	30	3.25	30	3.29	50
3.05	90	3.07	100	3.08	90	3.12	80
2.90	80	2.93	90	2.95	40	2.99	70
		2.88	70	2.91	40		
						2.86	70
		2.71	30	2.74	10	2.79	20
2.61	40	2.65	50	2.64	5	2.69	20
2.56	40	2.58	50	2.60	10	2.62	20
		2.53	10				
2.47	70	2.51	50	2.51	70	2.54	80
		2.46	50	2.49	30		
2.36	40B	2.40	30	2.42	10	2.43	60
		2.36	50				
2.28	40	2.30	30B	2.32	5	2.36	40
2.21	10	2.22	20B	2.25	5	2.28	20
2.14	40	2.16	20	2.17	10		
		2.13	10				
2.08	50	2.09	30	2.11	30	2.13	60
2.04	40	2.06	30			2.09	20
2.01	40	2.03	30	2.05	10	2.06	50
1.95	60	1.965	30	1.99	30	2.00	60
1.92	40	1.929	30	1.935	30	1.96	60
1.856	50	1.870	30	1.880	20	1.90	40D
		1.849	30				
		1.817	10			1.83	40
		1.789	50	1.790	20	1.80	50
1.753	60	1.763	70	1.765	20	1.76	50
1.720	20	1.728	30	1.730	10		
		1.704	10	1.710	10	1.72	40
1.677	40	1.682	10			1.69	40
1.642	10						
				1.640	20	1.65	60
1.600	60B	1.611	50	1.615	20 (d)	1.63	70
		1.596	50			1.61	70
		1.579	30	1.580	20	1.60	70
plus 11 lines to 1.195		plus 6 lines to 1.435		plus 22 lines to 0.993		plus 31 lines to 1.02	

1. Variscite, Lucin, Utah (ASTM 8—157).
2. Ferrian variscite, Casa de Pedra, Congonhas do Campo, Minas Gerais, Brazil (ASTM 7—69).
3. Aluminian strengite, Songo, Sierra Leone. FeKα-radiation, Mn-filter, 57.3 mm camera.
4. Strengite, Leonora Mine, Glessen, Hesse (ASTM 2—0244).

Table 4. X-ray powder data for cacoxenite from Songo, Sierra Leone.

d	I	d	I	d	I
22.9	vs	3.56	w	2.030	m
14.0	m	3.43	m	1.968	w
11.9	vs	3.27	m	1.926	m
9.6	s	3.17	s	1.814	m
9.0	s	3.10	s	1.760	m
8.0	m	3.00	m	1.723	w
6.9	s	2.93	m	1.680	w
6.3	w	2.78	s	1.629	w
5.75	w(d)	2.70	w	1.606	w
4.86	s	2.63	w	1.580	m
4.58	m	2.61	w	1.536	vw(d)
4.34	vw	2.44	m(d)	1.507	vw
4.14	m	2.37	vvw	1.473	w
3.97	m	2.30	w	1.318	m
3.82	w	2.24	w	+12 lines	
3.70	m	2.19	m	to 0.979	

crystals form exceedingly beautiful radial aggregates with a silky luster and golden yellow colour (Plate II, Fig. 4). This mineral was identified by the x-ray powder method. The interpretation of the film was made in the Department of Mineralogy at the British Museum with the contribution of Dr. Max Hay. According to this investigation the x-ray powder pattern is identical with that of cacoxenite from the type locality, St. Benigna, Bohemia.

The recording was made on film with a Debye-Scherrer camera (57.7 mm), using FeKα-radiation. The d-spacings are presented in Table 4. The reflections with d-spacings of 22.9 Å and 14.0 Å were recorded with a camera with a diameter of 114.6 mm, using CrKα-radiation.

ACCESSORY MINERALS IN UNALTERED ROCK

These minerals have been identified from the fractions obtained by the separation of rockbridgeite with heavy liquids. The identification has been performed with the help of x-rays, and the manner of occurrence is uncertain because only some few very small grains of the corresponding minerals have been found.

Chlorite has been found as glass-like pink grains and as dark green grains. Quartz occurs as colourless grains in the light fraction. Smoky gray plagioclase is present in very small amounts. Vivianite seem to be present but in such small quantities that there was not enough material for x-ray study.

ALTERATION CRUST

The investigated rock specimens are covered by a soft yellowish or brownish crust. It is difficult to distinguish individual mineral grains under the

Table 5. Interplanar spacings of wardites.

1.		2.		3.		4.	
d	I	d	I	d	I	d	I
8.56	<5						
6.60	20	6.67	10	6.62	20	6.65	20
5.72	10	5.69	10	5.72	20	5.68	20
5.01	20	5.01	20	5.01	40	5.01	30
		4.85	10	4.82	30	4.82	10
4.76	100B	4.78	90	4.77	100	4.76	50
		4.74	100	4.73	100	4.72	70
4.15	5						
		3.95	30	3.96	30	3.95	40
3.93	40	3.93	20	3.93	20	3.93	5
3.500	60	3.49	20	3.47	40	3.48	40
						3.36	5
		3.123	60	3.11	50	3.12	50
3.081	90B	3.094	90	3.09	80	3.09	90
		3.037	40	3.04	50	3.03	40
3.016	90	3.005	90	2.99	70	3.00	90
		2.885	20	2.90	5		
2.842	50	2.835	40	2.83	40	2.83	70
2.757	<5	2.758	10				
		2.691	5	2.68	10		
		2.637	<5	2.63	10	2.62	5
2.594	70	2.597	90	2.59	70	2.60	10
2.544	5	2.544	10	2.54	20	2.55	5
				2.50	5		
				2.39	10	2.39	10
				2.38	10		
				2.37	5		
2.330	<5			2.33	5	2.33	10
				2.30	40		
2.260	<5					2.26	10
2.169	5			2.16	20	2.16	10
				2.15	10		
2.118	30			2.115	30	2.115	20
				2.11	20		
						2.076	5
2.059	5			2.06	10		
				2.03	5	2.036	5
				2.02	5		
2.016	<5					2.006	5
1.963	5					1.969	10
1.931	5			1.93	10	1.93	5
1.905	<5						
						1.87	5
1.820	<5					1.84	5
1.774	40					1.77	30
1.705	5						
1.668	30						
						1.655	10
1.537	20						
1.523	20						
+18 lines until 0.992 Å							

1. Wardite, Songo, Sierra Leone. (film.)
2. Wardite, Songo, Sierra Leone. (counting rate computer.)
3. Wardite, Beryl Mountain, N. H. (diffractometer, Owens, Altschuler, and Berman, 1960.)
4. Wardite, Fairfield, Utah. (diffractometer, Owens, Altschuler, and Berman, 1960.)

microscope. Accordings to spectrochemical analyses, the light coloured portions consist of the following major elements: Al, P, Fe, and Ti (in order of abundance). The brown coloured portions contain the same elements, but iron prevails due to plentiful goethite. Many preparates for x-ray diffractometric investigations were made from these soft materials. The most prevailing mineral proved to be Al-strengite. In some of these preparates, a mineral belonging to the millisite-wardite group was also present. A few preparates contained only this mineral. In addition, an unidentified mineral was noted in some samples. Because of the very high aluminium content in some places where the x-ray analyses showed Al-strengite be the only crystal facies, it is probable that some amorphous hydrated alumina occurs.

The mineral belonging to the millisite-wardite group (yellowish soft fine-grained material) was investigated by x-rays both on film (57.3 mm camera, FeKα-radiation Mn-filter) and with a counting rate computer. The results of measurements are presented in Table 5 with the d-spacings of two published wardites (Owens, Altschuler, and Berman, 1960). The film was unable to resolve the broad lines at 4.76 Å and 3.08 Å. Using the CRC, these lines became clearly separated. According to Owens, Altschuler and Berman, it is possible to distinguish wardite from isostructural millisite. Two reflections (3.40 Å and 3.92 Å) characterize the Homeland millisite, the Fairfield millisite, and the ferrian millisite from Sénégal. These reflections are absent in patterns of wardite from Fairfield and Beryl Mountain. Three other reflections (3.95—3.96 Å, 3.11—3.12 Å, and 2.54—2.55 Å) are present on wardite patterns, but do not appear in patterns of the three above mentioned millisites. Taking the above facts and Table 5 into acount, it is evident that the material from Songo is primarily wardite ($NaAl_3(OH)_4(PO_4)_2 \cdot 2H_2O$) rather than millisite ($(Na, K)CaAl_6(PO_4)_4(OH)_9 \cdot 3H_2O$). Millisite and wardite may coexist but the latter seems to be prevalent.

A short mention about the most common unidentified mineral might be of value. A yellowish sometimes reddish alteration product of aluminian strengite has been described. The powder patterns (recorded on film) include some lines in addition to those of Al-strengite. The strongest were at 2.72 Å and 2.76 Å. The same d-spacings occur in the most diffractometer patterns of Al-strengite from the corresponding alteration zone. In addition, lines at 4.43 Å, 3.76 Å, and 3.48 Å are clear. The relative intensities are always similar, indicating that these lines belong to a specific species and not to a mixture of minerals.

CONCLUSIONS

It has been shown that the dark dike-like rock cutting the ijolite of Songo, Sierra Leone consists of rockbridgeite, aluminian strengite, goethite, wardite, and cacoxenite. The P_2O_5 content (22 percent) is very high, and the SiO_2 content (0.55 percent) very low. Accordingly it is appropriate to call this rock a phosphatite, consistent with the usage of the word carbonatite. Carbonatites often occur with alkaline rocks and many geologists regard them as genetical associates. Phosphatites have also been described from alkaline rock areas, for example from the huge alkaline rock area of the Kola Peninsula in Russia. The author thinks it is possible that the iron-rich phosphatite and ijolite from Songo are genetically related.

Acknowledgements — The writer wishes to thank Dr. Vladi Marmo, the director of the Geological Survey of Finland, for his many valuable suggestions during the work. Thanks are also due to Dr. Max Hay of the British Museum for his contribution in identifying the cacoxenite. Acknowledgments are also due the following Geol. Survey of Finland personel: Mr. P. Ojanperä M. A. for the chemical analyses; Mr. A. Löfgren M. A. for the spectrochemical analyses; Mr. J. Hyyppä M. A. for the GTA; Mr. A. Toivonen M. A. for the DTA; and Mr. E. Halme for taking the photographs. To Miss. T. Åberg the author is indebted for drawing the diagram and to Mr. R. Ojakangas M. A. for correcting the English of this manuscript.

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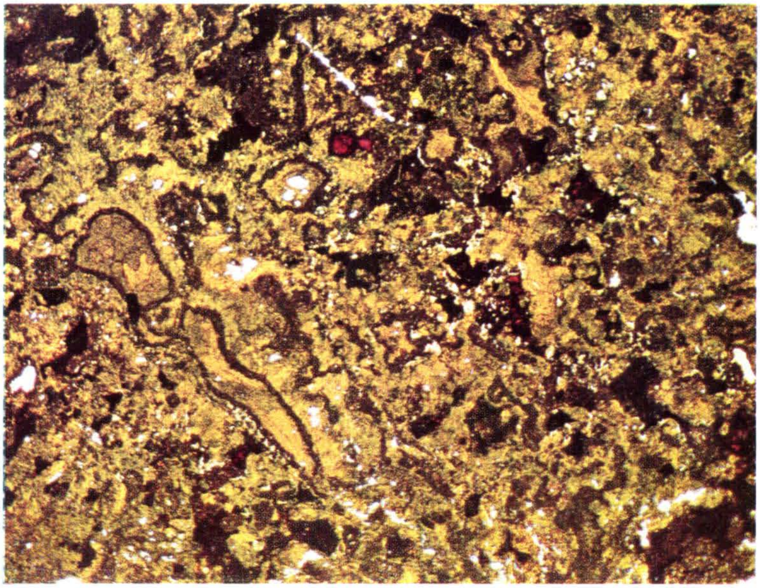

Fig. 1. Phosphatite consisting of rockbridgeite (greenish) that is partly altered to aggregates of goethite and Al-strengite (brownish), and individual grains of goethite (red). Without analyzer. Magn. 18×. Photo Erkki Halme.

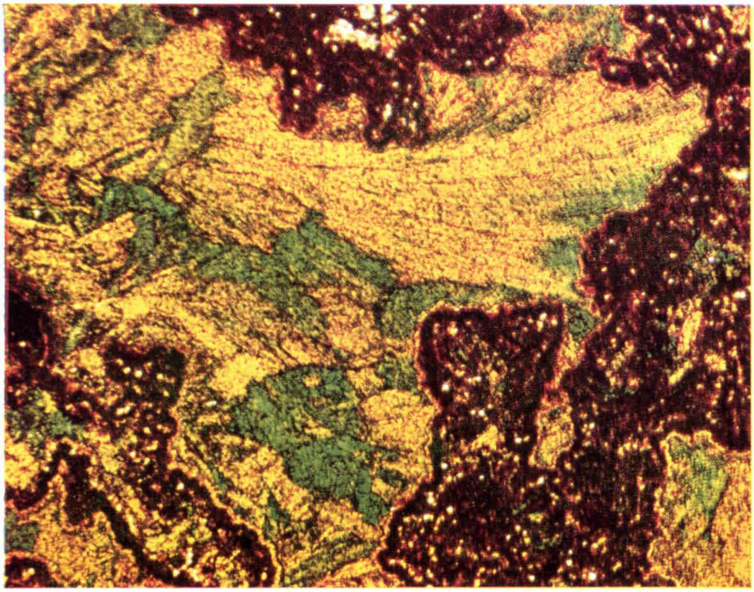

Fig. 2. Aggregates of rockbridgeite, partly altered to goethite and aluminian strengite. Without analyzer. Magn. 77× . Photo Erkki Halme.

Atso Vorma: On the x-ray Study of the Phosphate Minerals...

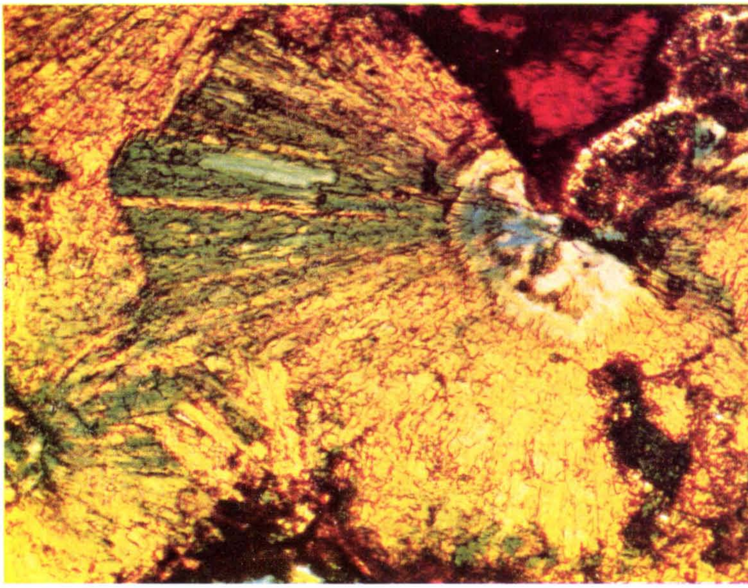

Fig. 1. Spherical aggregate of rockbridgeite showing concentric color banding. Without analyzer. Magn. 162×. Photo Erkki Halme.

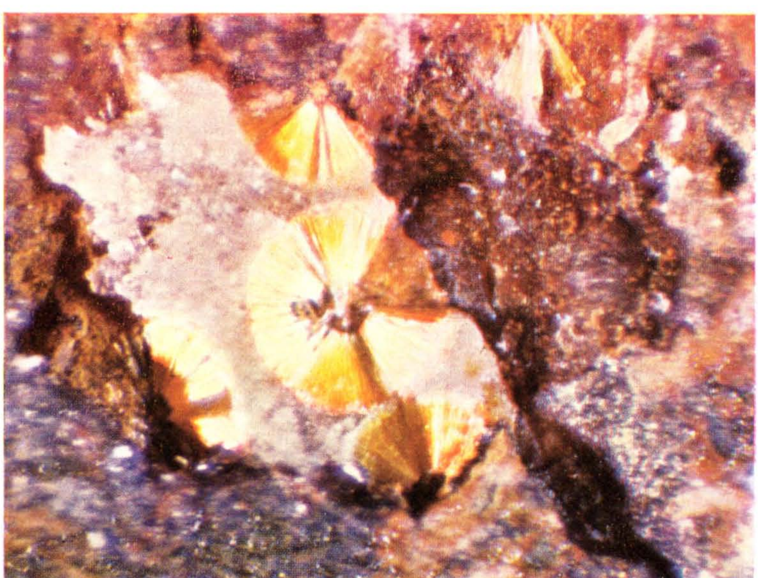

Fig. 2. Radial aggregates of cacoxenite. Magn 92×. Photo Erkki Halme.

Atso Vorma: On the x-ray Study of the Phosphate Minerals...

24

TWO LITTORINA TRANSGRESSIONS IN VIROLAHTI, SOUTHEASTERN FINLAND [1]

BY

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CONTENTS

INTRODUCTION

In the summer of 1953 the author and his assistants carried out peat investigations for the Geological Survey of Finland in the commune of Virolahti, southeastern Finland. The topographic map of the district drew attention to the bog known as Ruokolammensuo (Fig. 1), situated 20—30 meters above sea level. In the light of Hyyppä's (1937) interpretation of postglacial changes in the shoreline of southern Finland, it appeared possible to determine with greater exactitude on the basis of the sediments of Ruokolammensuo the location of the shore of the Littorina Sea in the area. One might even expect, in fortunate circumstances, to find evidence of the transgression caused by the marine stage in question within the bounds of the bog.

[1] Received April 8, 1961.

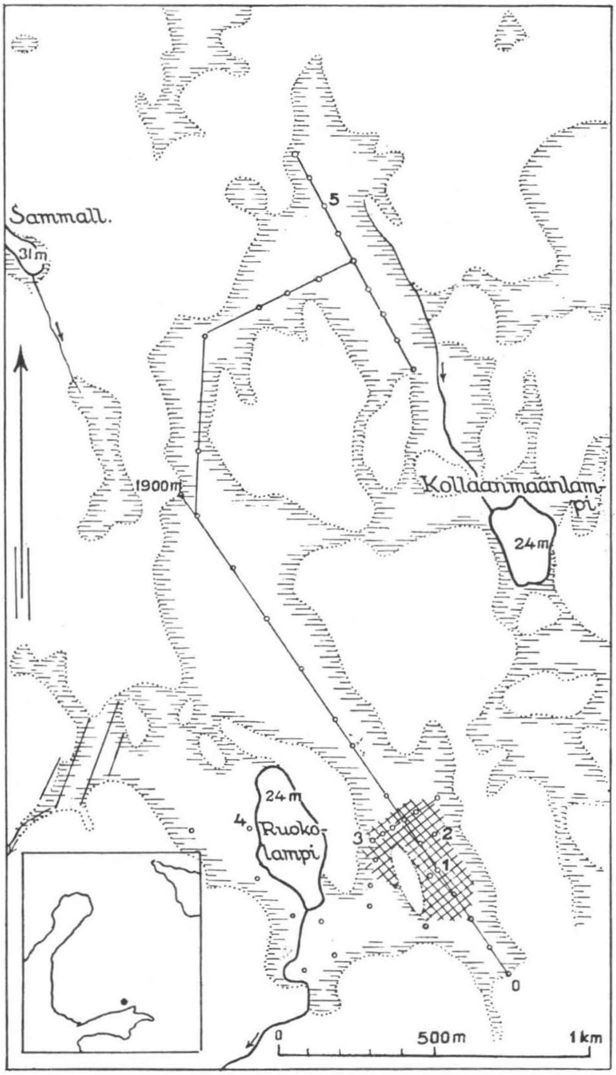

Fig. 1. Map of Ruokolammensuo, including survey lines and sites investigated. Circles indicate drilling sites, numbers 1—5 the spots where samples were collected and crosshatching the area where transgressive sediments have been found.

Mr. A. Leino, one of the research assistants, was assigned the task of carrying out an orientative exploration of Ruokolammensuo. In certain parts of the bog he met with transgressive layers. Subsequently, in follow-up investigations, he took two series of samples, Profiles 1 and 5, which will be presented later. Laboratory studies proved that the bog contained evidence of the Littorina transgressions.

In the summer of 1956 the present writer had the opportunity to explore Ruokolammensuo further, with the assistance of Mr. A. Leino. It was endeavored on this occasion to determine the extent and maximum height of the transgressions within the area covered by the bog. For this purpose numerous drill-holes were made and additional series of samples collected, Profiles 2, 3 and 4 being presented in conjunction with the present study. The drillings were carried out primarily in different spots located in the southern part of the bog. Previously investigations had been conducted along lines drawn lengthwise across the different basins of the bog. Drill-holes were made at 62 different points, the level of each of which was taken.

The pollen analyses included in the study were performed in the laboratory of the Department of Surficial Deposits of the Geological Survey by Miss Ester Uussaari, Ph. Cand., the macrofossil determinations by Mr. V. E. Valovirta, Ph. Cand., and the diatom determinations by Mr. K. Mölder, Ph. D., and Miss Kyllikki Salminen, Lic. Ph. The drawings were done by Mrs. Lyyli Orasmaa and Mrs. Marja-Leena Malmivaara. The English translation is the work of Mr. Paul Sjöblom. I wish to express my appreciation to all the persons mentioned as well as to Mr. A. Leino, the research assistant.

THE BOG AND ITS LOCATION

Ruokolammensuo is situated in territory belonging to the village of Klamila, Virolahti commune, approximately eight kilometers north from the coast and some twelve kilometers east of the town of Hamina (Fig. 1). The bog comprises two basins running NW—SE and connected by a narrow stretch 200—300 meters wide. The area of the bog is roughly 125 hectares.

Ruokolammensuo slopes from north to south. The elevation of the bog in the northern part of the eastern basin is 28—30 m and in the channel joining the basins 28—29 m. At about the middle of the western basin the elevation is 26—27 m, in the environs of Ruokolampi (pond) 24—25 m, while at the end of the southern cove of the basin, where the brook flowing from the aforementioned pond leaves the bog, the altitude of the surface of the bog is 20 m. The last-mentioned represents the threshold level of this basin. The terrain continues to descend and the elevation at the margin of the cartogram at the point of the brook is 15 m and from there about 400 m west, at the edge of the area of clayey soil, ten meters. In this clayey area the brook joins Pitkäkoski river, which flows southeast and empties about six kilometers farther on into the Gulf of Finland. The terrain thus descends steadily from the bog basin into the sea without observable thresholds.

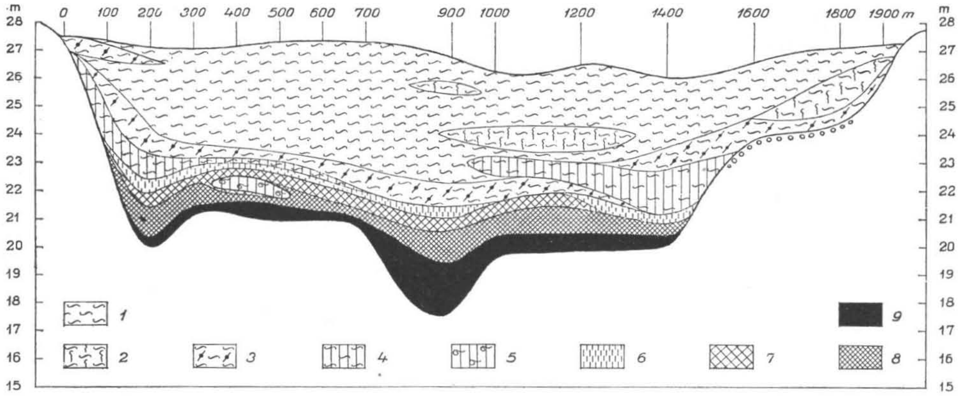

Fig. 2. Longitudial profile from the survey line running across the western basin of the bog. Key to signs: 1 — *Sphagnum* peat. 2 — *Eriophorum-Sphagnum* peat. 3 — *Scheuchzeria-Sphagnum* peat. 4 — *Carex-Sphagnum* peat. 5 — *Bryales-Sphagnum* peat. 6 — *Phragmites* peat. 7 — Coarse detritus ooze. 8 — Fine detritus ooze. 9 — Silt and clay.

The western basin of the Ruokolammensuo consists in its southern part of pine bog with many subshrubs as far, approximately, as the level of the northern end of Ruokolampi, and likewise along the margins and in the channels. Elsewhere the area can be classified as treeless bog, mostly of the *Eriophorum vaginatum* type, but in the eastern parts also of various *Carex* types.

The greatest thickness of the peat layer, 5.7 m, has been met with in the northern part of the eastern basin of the bog. Underlying the peat in the spot referred to is a dense bed of light-colored clay 2.2 m thick, but mostly the peat borders sharply on till or bedrock. In the eastern basin of the bog, the peat is predominantly of the *Carex* variety.

The relative depths of the western basin of the bog and its surficial deposits are represented in the profile drawing of Fig. 2. It was made from the survey line running across the basin and shown in Fig. 1. *Sphagnum* peat forms a thick, quite uniform bed at the surface of the bog. The base of the bog deposits consists of till, which is stony in the northern part of the basin, giving a washed appearance.

RESEARCH MATERIAL

Series of samples have been collected from five sites in Ruokolammensuo for detailed investigation. These sites are marked in the map of Fig. 1 with the numbers 1—5. Three of the series (1—3) are from the southern part of the western basin, where, on the evidence of field explorations, a transgression occurrence could be observed. One of the profiles (4) is from the western shore of Ruokolampi and one from the northern part of the eastern basin of the bog (5). A pollen analysis of all the series of samples was carried out, together with a macrofossil study of profiles 2, 3 and 4, and the diatoms of series 1, 2, 4 and 5 were identified.

PROFILE 1

Sample series 1 was taken from a spot 400 meters from the southernmost cove of the western basin at an elevation of 27.1 m and a depth of 6.0 m. The pollen and diatom analyses made from the series are presented in the form of a diagram in Fig. 3. The bog deposits appearing in the left-hand margin of the figure have been presented in greater detail in this case than in the corresponding portion of Fig. 2.

In this spot in the bog the bottom layer consists of silt to a thickness of 20 cm. Overlying it is, first, 0.5 m of fine detritus ooze and, next, 25 cm of coarse detritus ooze, which changes without any distinct border after a transitional stage of allochtonous peat into telmatic *Bryales-Carex* peat. Its upper border lies 22.3 m above sea level. The peat later became submerged, as a consequence of which it is overlain by allochtonous peat as well as fine and coarse detritus ooze up to the 4.4 m level. From there upward the peat varies from one thin layer to another, remaining predominantly *Carex*, however, up to the 3.8 m level, from where the peat continues upward, dominated by *Sphagnum*, all the way to the surface.

In the classification of diatoms, *Campylodiscus clypeus* and, in general, the so-called clypeus flora (M-B. Florin, 1946) have been included among salt water forms in this study, because they belong as quite an essential part to the diatom flora of, *e. g.*, the Littorina transgressions, as will be subsequently pointed out.

The age of the lowest part of the series of layers, on the level of the clayey silt, is indeterminate, according to pollen and diatom analyses. On the other hand, the overlying fine detritus ooze originated during the *Pinus* maximum during the Boreal time (V), corresponding to the Ancylus in the evolutionary history of the Baltic Sea. *Ceratophyllum* is likewise concentrated in the corresponding horizon.

At this level the diatoms consist of fresh-water species, together with fresh- and brackish-water forms. Included are great-lake diatoms in abundance, typical representatives of Ancylus lake flora. For example, at 5.4 meters the most common is *Cyrosigma attenuatum* (11 %). Among other species might be mentioned *Amphora ovalis* (2 %), *Epithemia hyndmanni* (4 %), *Eunotia clevei* (1 %), *Melosira arenaria* (3 %) and *Stephanodiscus astraea* (1 %). The diatoms of the sediment in question are predominantly the same as the great-lake diatoms observed by the present author (Salmi, 1948) to

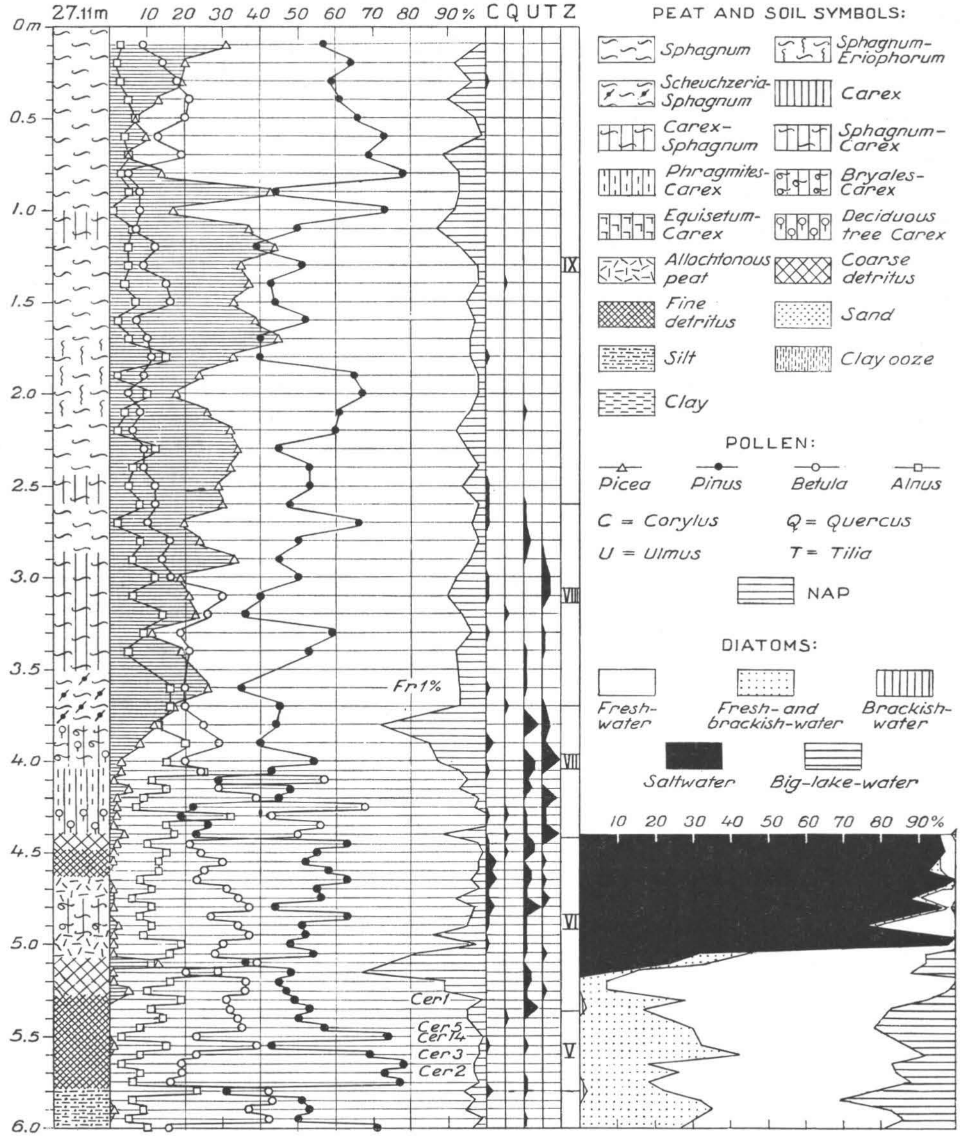

Fig. 3. Profile 1. Pollen and diatom diagram. Zone V = Boreal, Zones VI and VII = older and younger Atlantic, VIII = sub-Boreal, IX = sub-Atlantic.

occur in the sediment of the Ancylus transgression of Hangassuo (bog), the proportions in both cases also being pretty much the same. Nevertheless, Donner (1952) subsequently explained that the sediment mentioned had been deposited in salt water during the final stage of the Yoldia, even though his material does not include diatoms suggesting a salt-water habitat. He bases his argument on the ground that the threshold of Hangassuo was so high during the transgression that the water covering it was at most three meters deep, and therefore the salt water from the sea had not been able to rise over it into the basin of Hangassuo. Since, however, according to Hyyppä (1937), the depth of water in the Ruokolammensuo area during the Ancylus transgression had been 12—13 meters and no diatoms native to a salt-water habitat exist in the sediment dating from that time, Donner's interpretation must be deemed erroneous. The same conclusion has been reached by Mölder, Valovirta and Virkkala (1957). On the other hand, Sauramo drew on Donner's misinterpretation in several connections, most recently in his extensive work dealing with the history of the Baltic Sea, which appeared in 1958.

The bog deposits referred to in the foregoing belong up to the 4.4 m level to Zone VI and consist of transgressive allochtonous peat as well as both fine and coarse detritus ooze. In the evolutionary history of the Baltic Sea they correspond to the first half of the Littorina stage.

At the beginning of Zone VI the diatoms consist for a short distance of fresh water flora like the foregoing, but at the contact of coarse detritus and Phr-allochtonous peat they change character through the introduction also of diatoms from a salt-water habitat. Their proportions are as follows:

	Depth, meters							
	4.4	4.5	4.6	4.7	4.8	4.9	5.0	5.1
Amphora mexicana v. major	3	21	8	15	—	—	—	—
Campylodiscus clypeus	87	68	72	72	12	32	17	2
» v. bicostata	—	—	—	—	80	1	32	4
» echeneis	5	8	13	9	4	5	—	—
Diploneis interrupta	1	—	—	—	—	—	—	—
Total percent	96	97	93	96	96	38	49	6

Campylodiscus clypeus and v. *bicostata* are the first salt-water species. The latter is in the majority at the 4.8 m level, where it abruptly ceases to be present. From there on upward the former is the predominant species. Starting at the 4.8 m level, proceeding upward, a new salt-water species appears, *Amphora mexicana* v. *major*, but *Campylodiscus echeneis* occurs scantily though evenly distributed throughout almost the entire deposit. The salt-water diatoms in these parts of the profile are exceedingly poor in variety of species but so dominant numerically as to reduce the share of other diatoms to comparative insignificance. It should be noted that the same diatoms have been met with also in the telmatic peat inundated by the transgression, though, it is true, more scantily than in the limnic sedi-

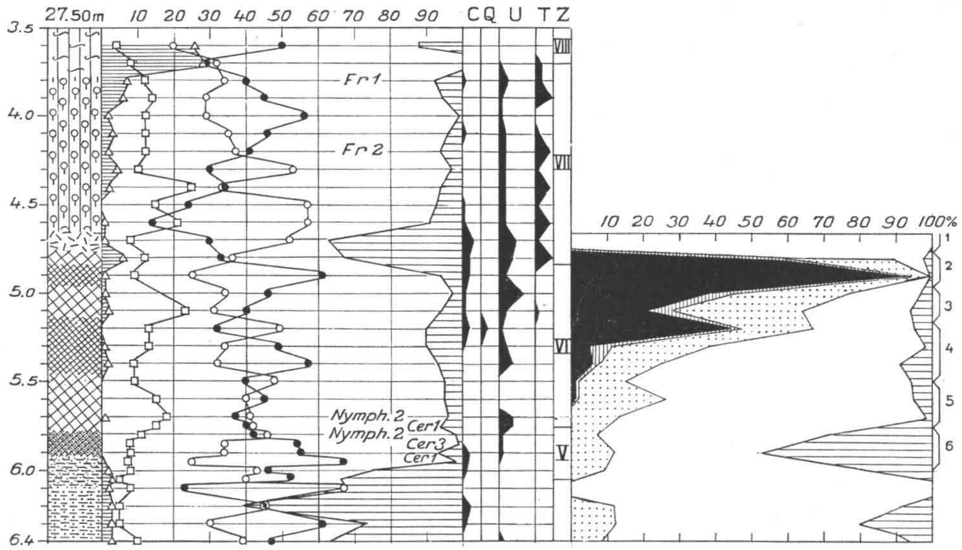

Fig. 4. Profile 2. Pollen and diatom diagram.

ments. It should be further pointed out that the diatoms in the allochtonous *Phragmites* peat at the 4.6—4.7 m depth are mostly tattered; from the 4.4 m depth upward no diatoms have been discovered.

PROFILE 2

The series of samples (Fig. 4) comes from a spot, about 110 meters north of the preceding one, where the surface of the bog is at an elevation of 27.5 m and its depth is 6.4 m.

The succession of layers resembles that described in the foregoing all the way to the 5.5 m level, where fine detritus overlies the coarse detritus ooze at an elevation of 22.0 meters. This suggests a transgression. A corresponding phenomenon is to be observed in the sequence also at the depth of 5 m, marking the 22.5 m elevation. Thus it would appear as if two transgressions had occurred in the area of Ruokolammensuo. It is only from the 4.7 m depth that the peat continues without disturbance all the way to the surface.

The pollen diagrams made from Profiles 1 and 2 parallel each other, and the diatom studies confirm the parallelism. The silt in the lower part of Profile 2 originated, to judge by the diatoms, in a small, shallow, cold-water lake with diatoms scantily present. The silt contains, according to the NAP, *Artemisia* and *Chenopodiaceae* pollens in considerable amounts (10—25 %), the latter occurring at the depth of 6.0 m in the diagram. The circumstances referred to suggest that the silt may greatly antedate the Boreal stage; accordingly, sediments from the interval between the silt and the clay ooze originating during the Boreal stage are lacking. The same indication is given by the sediment of the lower part of Profile 4 in Fig. 6.

At the level of the *Pinus* maximum of the Boreal stage, big-lake diatoms occur in abundance, the same species as at the corresponding depth in the preceding diagram. The end of the Boreal stage is marked here by *Ceratophyllum* thorns and *Nymphaea* pollens.

It was soon after the Boreal stage that the first salt-water diatoms appeared in the Ruokolammensuo area — at the 5.6 m level. The species are *Nitzschia scalaris* and, above it, *Amphora mexicana* v. *major*, the proportions being, to be sure, slight. In the next two samples *Campylodiscus clypeus* and *Nitzschia scalaris* are on a par, jointly contributing 5—6 %, but at 5.2 m the salt-water species reach the figure of 44 %. The following species are represented: *Nitzschia scalaris* 21 %, *Campylodiscus clypeus* 14 %, v. *bicostata* 8 % and *Diploneis smithii* 1 %. At the depth of 5.1 m, represented by coarse detritus ooze, the share of the salt-water diatoms falls to 21 %. The species are *Nitzschia scalaris* (11 %) and *Campylodiscus clypeus* (10 %). The decline in the abundance of salt-water diatoms and the increase in fresh- and brackish-water species at the last-mentioned level and in succeeding samples from higher up serve, along with the coarse detritus ooze at that level, to indicate shallower water than previously. The fine detritus ooze met with at 4.9 m suggests, together with its content of diatoms, a recurrence of deeper water. In the latter layer the diatom flora consists of salt-water forms to the extent of 92 percent, the share of the rest being quite small. The species and their proportions are as follows:

Amphora mexicana v. major	28	percent
Campylodiscus clypeus	52	»
Campylodiscus clypeus v. bicostata	5	»
Campylodiscus echeneis	3	»
Diploneis interrupta	2	»
Melosira westii	1	»
Rhabdonema arcuatum	1	»
Total	92	percent

In the following sample, proceeding upward, the share of salt-water diatoms once more decreases and the flora suggests a sinking of the water level. The diatom flora of the allochtonous peat at 4.7 m consists exclusively of species native to a shallow-water lake habitat, which provides evidence that the basin of Ruokolammensuo had become cut off from the marine influence and evolved into a small isolated lake.

Table 1. Macrofossils of Ruokolammensuo.

Species	Profile 2						Profile 3				Profile 4							
No.	1	2	3	4	5	6	1	2	3	4	1	2	3	4	5	6	7	8
Alisma plantago-aquatica	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	—	—
Alnus glutinosa	—	—	—	—	—	1	—	—	—	—	—	—	—	—	—	—	—	—
Andromeda polifolia ..	—	—	—	—	—	—	2	—	—	—	5	2	—	—	—	—	—	—
Betula alba	—	—	—	—	—	4	—	—	—	—	2	3	—	—	—	—	—	—
Carex lasiocarpa	—	—	5	3	—	—	2	3	27	10	40	1	—	—	—	—	—	—
» pseudocyperus .	—	—	—	—	—	—	—	—	—	8	—	—	—	—	—	—	—	—
Comarum palustre ...	1	—	1	2	—	—	9	—	3	1	—	—	—	—	—	—	—	—
Iris pseudacorus	—	—	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—
Lycopus europaeus ...	—	—	—	—	—	—	1	—	—	1	—	—	—	—	—	—	—	—
Lysimachia thyrsiflora	1	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	—	—
Menyanthes trifoliata	—	—	1	2	1	1	—	9	7	—	—	—	—	—	—	—	—	—
Najas marina	1	60	—	11	—	—	—	9	—	5	—	—	15	—	—	2	—	—
Nuphar luteum	—	—	1	—	2	—	—	—	—	1	—	1	—	—	—	—	—	—
Nymphaea alba	—	—	1	—	—	—	—	—	—	1	—	1	—	—	—	—	—	—
» candida ...	—	—	1	—	7	—	—	—	—	—	—	—	—	—	—	—	—	—
Phragmites communis	r	+	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—
Potamogeton natans ..	—	—	2	—	3	—	—	—	—	9	—	28	2	—	—	2	—	—
Ranunculus sp.	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	—	—
Rubus chamaemorus..	—	—	—	—	—	—	—	—	—	1	—	—	—	—	—	—	—	—
Scheuchzeria palustris	—	—	—	—	—	—	2	—	—	—	13	3	—	—	—	—	—	—
Scirpus Tabernae-montani	18	3	1	—	4	3	—	—	—	1	—	1	—	—	—	—	—	—
Viola palustris	5	2	—	—	—	—	3	—	—	—	—	—	—	—	—	—	—	—
Coleoptera	—	—	—	—	—	—	2	—	—	—	—	—	—	—	—	—	—	—

Macrofossil determinations have also been made from samples 1—6 of the sequence of deposits. The places where they were collected are marked in the right-hand margin of Fig. 4, and the results are to be seen in Table 1.

Najas marina is concentrated in both fine detritus ooze layers, which also contain the most salt-water diatoms. Sample 2 in especial contains seeds in abundance. The species mentioned is lacking in the coarse detritus ooze between the layers. Sample 4, again, lacks the macrofossils present in Sample 5 taken below it, such as *Nuphar*, *Nymphaea candida*, *Potamogeton natans* and *Scirpus Tabernaemontani*. All the species listed are missing, once more, from Sample 2, with the exception of the last. Its most abundant occurrence is in the allochtonous peat (Sample 1), which contains other remnants of flora characteristic of shallow water or a filling-up stage. The abundance of NAP at the corresponding level is due to *Phragmites* pollens. At 5.4—5.0 m in the profile the share of *Sparganium* pollens is 2—5 %.

The impression has been gained that the clypeus flora, represented by fine detritus diatoms, denotes, with its high peaks in this and other Ruokolammensuo profiles, a Littorina transgression. The same conclusion has been reached by, among others, Maj-Britt Florin (1946) on the evidence of its

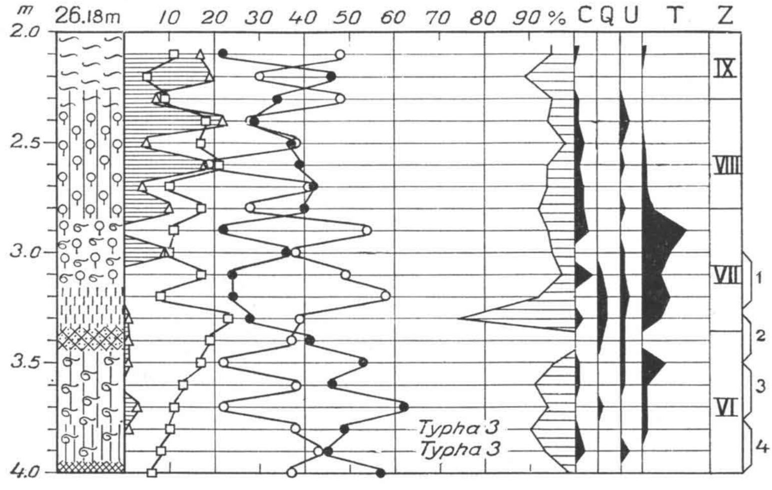

Fig. 5. Profile 3. Pollen diagram.

mass occurrence. According to Salmi (1945), too, *Campylodiscus clypeus* is quite common in the sediment of the L II transgression of northern Estonia; it likewise occurs quite generally, as pointed out by Hyyppä (1937), in the sediment of the Littorina transgressions of southern Finland.

PROFILE 3

The series of layers was collected by drilling from a site in the narrow channel between two islands in the southern part of Ruokolammensuo where the elevation of the surface of the bog is 26.2 m (Fig. 5) and the depth of the bog is 4.0 m.

Overlying the till at the bottom is a 5 cm layer of detritus ooze. Over it is a layer of *Bryales-Carex* peat 0.5 m thick, the elevation of the upper limit of which is 22.7 m. The peat is covered by a thin layer of fine detritus ooze mixed with sand. Higher up occurs allochtonous peat, and from the depth of 3.1 there begins a peat layer composed of various kinds of peat, which continues without interruption all the way to the surface.

The pollen diagram is easily dated in accordance with the preceding ones. Thus, the lowest part of the sequence up to the higher fine detritus ooze and allochtonous peat belongs to Zone VI. Accordingly, the lowest part of the sequence of layers is younger than in the case of the preceding ones.

The profile also involves macrofossil studies, comprising Samples 1—4 (*cf.* Fig. 5 and Table 1). *Najas marina* has been met with in Samples 2 and

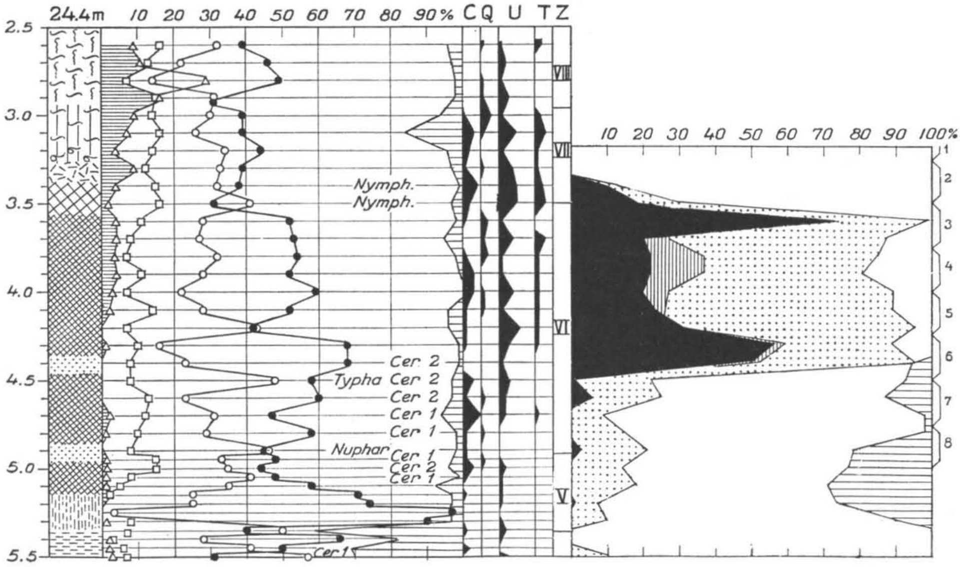

Fig. 6. Profile 4. Pollen and diatom diagram.

4. Both include detritus ooze, from which the seeds of the species evidently derive, indicating their genesis in salt water. On the other hand, the other species from the lowest sample belong to the lower part of the peat, where the *Typha* pollens also belong; the same is true of *Carex pseudocyperus* as well as of the other extremely abundant macrofossils of the sample in question. The species met which in Samples 1 and 3 belong to the peat, but the *Najas marina* belonging to the fine detritus situated in between them, in Sample 2, indicates that the peat had been submerged by salt water as in Profile 1. The abundance of NAP at 3.3 m is due to the *Phragmites* pollens.

PROFILE 4

The sequence of layers was taken from a spot on the western shore of Ruokolampi where the altitude of the surface of the bog is 24.4 m — *i. e.*, considerably lower than in the foregoing case (Fig. 6) — and the depth of the bog is 5.5 m.

In the lower part of the sequence overlying the till base there are first clay and clay ooze. On top of the latter is a thick layer of fine detritus ooze, which is cut at two points by a thin layer containing a mixture of fine sand. Their position seems to suggest a temporary sinking of the water level in the area. At 3.5 m the deposit undergoes a change into coarse detritus ooze and, at 3.3 m, after a transitional form as allochtonous peat, into peat, which, varying in species, continues all the way to the surface.

In the light of the pollen diagram, the time of sedimentation of the clay remains indeterminate, nor does the diatom study offer a clear chronological solution either, for the diatom flora of the sediment consists of species characteristic of small fresh-water lakes, exactly as in the case of the clay of the lower part of Profile 2. The flora includes various species from the families of *Cocconeis, Cymbella, Eunotia, Fragilaria, Hantzschia, Melosira, Navicula, Pinnularia* and *Surirella*. *Artemisia* and *Chenopodiaceae* pollens claim a share of 10—19 % in the clay, and among other NAP mention should be made of *Thalictrum* and *Ranunculaceae*. Zone V can be traced down to the lower part of the clay ooze and the fine detritus ooze as far as the lower layer of fine sand. The upper part of the section includes *Ceratophyllum* once more, as in Profiles 1 and 2, and the species continues to be present in the lower part of Zone VI up to the height of the upper layer of fine sand, the sediment there marking the first rise of salt-water diatoms.

The diatom flora confirms the conclusions to be drawn from the sediment of the Ancylus lake corresponding to Zone V, for it contains an abundance of great-lake diatoms, representing typical Ancylus flora, as follows:

	Depth, meters			
	4.9	5.0	5.1	5.2
Amphora ovalis	2	4	—	4
Campylodiscus noricus	1	1	—	1
Epithemia hyndmanni	4	4	17	1
Gyrosigma attenuatum	7	8	2	6
Melosira arenaria	7	5	10	14
Stephanodiscus astraea	1	1	3	1
Total percent	22	23	32	27

The same great-lake species continue to be present in undiminished numbers up to the 4.5 m depth.

The first signs of salt-water diatoms are met with at 4.9 m, where *Campylodiscus clypeus* occurs, and again at 4.6 m, which contains *Hyalodiscus scoticus* in addition. A remarkable increase in salt-water diatoms takes place at 4.4 m (50 %) and at 4.3 m (56 %), whereas a lower count recurs at a higher level. There fresh-and brackish-water diatoms take precedence, notably *Fragilaria pinnata*, the share of which in several samples ranges between 40 and 60 percent. A new, strong but brief increase in the salt-water diatom count occurs at 3.6 m, where it reaches the figure of 74 %. In the sample referred to, the percentage of salt-water forms is as follows:

	Depth, meters		
	3.6	4.3	4.4
Amphora mexicana v. major	7	5	4
Campylodiscus clypeus	63	35	25
Campylodiscus echeneis	1	3	3
Diploneis didyma	1	—	—
Navicula peregrina	—	—	5
Nitzschia scalaris	1	13	12
Rhopalodia gibberula	1	—	—
Surirella striatula	—	—	1
Total percent	74	56	50

Among the salt-water diatom flora, *Campylodiscus clypeus* dominates in the case of both the lower and, especially, the higher upsurgence. There is the difference to be noted that *Nitzschia scalaris*, *Navicula peregrina* and *Campylodiscus echeneis* play a more prominent role in the lower horizon than in the higher ones.

As for the macrofossils, it will be observed (Table 1) that *Najas marina* again occurs in two horizons and in conjunction with the peaks achieved by the salt-water diatoms. The species occurs more abundantly at the higher peak than lower down, quite as in Profiles 2 and 3. It should be mentioned that not a single seed has been met with in the fine detritus situated between the layers in question and that *Iris pseudacorus* has been found in the sample representing coarse detritus ooze and allochtonous peat.

The results of both the diatom and macrofossil investigations justify drawing a parallel between the peaks in the occurrence of salt-water diatoms observed in Profile 4 and the *Najas marina* seeds appearing in conjunction with them in the same horizons, on the one hand, and the corresponding matters in the preceding profiles, on the other hand. Accordingly, the strong upsurgences of salt-water diatoms in Profile 4 signify two separate transgressions. They cannot, however, be observed on the basis of the sediment, because the series was collected so much lower down than the preceding ones that the site was submerged during the regression between transgressions. The conspicuous occurrence of *Fragilaria pinnata* there also indicates a sinking of the waterlevel at the location of the sediment under consideration.

The dates of the transgressions correspond well in the light of the pollen diagrams Nos. 2, 3 and 4, the earlier falling into the first half of Zone VI and the later quite at its end, close to the contact between Zones VI and VII.

PROFILE 5

The series of layers was taken from the northern part of the eastern basin of Ruokolammensuo, as shown in Fig. 1, the site being higher than the pre-

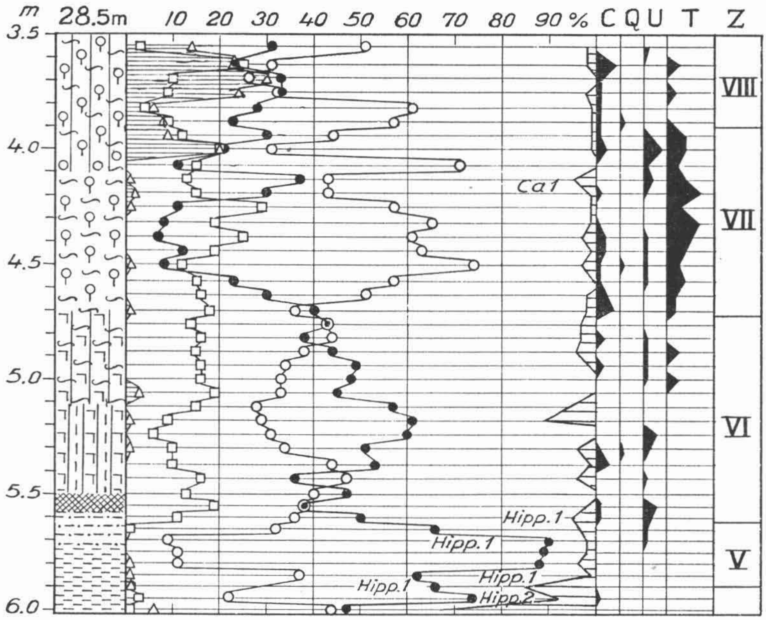

Fig. 7. Profile 5. Pollen diagram from the eastern basin of Ruokolammensuo.

ceding ones, with the elevation of the surface of the bog at 28.5 meters. The depth of the bog at the site is 6.0 m (Fig. 7).

The sequence in the lower part, overlying the till, consists of silt, clay and, higher up, oozy clay, which changes, following a thin transitional layer of fine detritus ooze, into *Phragmites-Equisetum-Carex* peat and this, in turn, into other varieties of peat, which extend without disturbance up to the surface.

This sequence likewise parallels the other ones on the basis of its pollen diagram. Zone V includes part of the clay as well as the oozy clay overlying it. The lowest silt antedates this zone.

The *Pinus* maximum characteristic of the Boreal stage features Zone V in this case too. Deviating from the diagrams, *Hippophaë* pollens are met with here. The absence of the species from the sediments of the western basin in likely to be due to the fact that it was for a long time exposed to the influence of the transgressions, which was apparently detrimental to it (Salmi, 1948). The boundary between Zones VI and VII must be drawn, according to the evidence, principally, of the pollens of *Tilia* and other rare deciduous trees, at the depth of 4.7 m, where the peat of the humid Atlantic stage changes into deciduous tree-*Sphagnum* peat.

According to the diatom analysis, the water sediments of the lower part of the series of layers originated during the extensive fresh-water period, representing between 5.9 and 5.6 meters typical Ancylus flora. The share of great-lake diatoms in the deposit under consideration is at 10 cm intervals, starting from the bottom, 34, 87, 68 and 17 %. *Melosira arenaria* is the most common among the species. In addition, there also occur *Amphora ovalis*, *Epithemia hyndmanni*, *Gyrosigma attenuatum* and *Stephanodiscus astraea*. Higher up in the fine detritus ooze one meets with fresh-water flora characteristic of small lakes. This indicates that the basin had become isolated so as to form a separate lake of small dimensions. Diatoms were sought higher up in four different horizons, but, except for certain *Pinnularia* and *Eunotia* individuals, none were found. Since salt-water diatoms were not met with in the profile in question, it means that even the deepest levels of the northern part of the eastern basin of Ruokolammensuo were outside the sphere of influence of the Littorina transgression and, in general, had no connection with the Littorina Sea.

THE TRANSGRESSIONS AND THEIR ELEVATIONS

Of the transgressions observed in the area of Ruokolammensuo, the lower and older falls in the early part of Zone VI, according to the pollen diagram, while the higher and younger falls toward the end of the same zone, close to the boundary between Zones VI and VII. In the evolution of the Baltic Sea they correspond to the two oldest transgressions of the Littorina Sea, from which times are known the Littorina shores L I and L II. The rest of the stages of the Littorina Sea, according to Hyyppä's (1937) pollen diagrams, are distinctly younger than those met with in the area of Ruokolammensuo.

Profile 5 showed that the eastern basin of Ruokolammensuo was never connected with the Littorina Sea. When an attempt is made to ascertain the highest waterlevel of the Littorina Sea in the region, the height of the threshold of the aforementioned basin takes on great importance.

According to the topographic map, elevations of 25 m and higher keep the basin generally sealed off from the direction of the sea as it exists at present. It will be noted from the longitudinal profile presented in Fig. 2 that the bottom of the bog in the northern part of the western basin lies at an altitude of 24—26 meters. At 24—25 m it is stony and exhibits washing. At this level the peat layer lies directly over the stony base.

Along the line running north across the channel that links the basins (Fig. 1), the altitude of the bog's sandy till bottom is at first 25.1 m and then northernmost point on the line 24.9 m. The altitude of the sandy till bottom is also 25 meters at a distance of 200 m along the line running northeast from the last-mentioned point. Farther on the bottom begins to descend and on top of the till comes oozy clay, as in Profile 5 (Fig. 7). Investigation reveals that the threshold altitude of Ruokolammensuo is approximately 25 meters. Even the highest waterlevel of the Littorina Sea never rose above this altitude in the Ruokolammensuo area.

In the western basin of the bog, the upper limit of the Littorina sediments, according to Profiles 1, 2, 3 and 4, lies 22.6—22.9 m above sea level. Accordingly, the maximum height of the younger transgression is situated somewhere between 23 and 25 meters. According to Profiles 2, 3 and 4, again, the sediment of the older transgression extends to an altitude of 22.1—22.4 meters, which means that the maximum value of this transgression is between 22.5 and 25. The maximum of each transgression must therefore be extremely close to that of the other. The values approach each other further for the reason that the uppermost sediments in each case consist of fine detritus ooze, the deposition of which would have required an overlying depth of water of at least 0.5—1.5 m.

Attention has been drawn in the foregoing to the fact that washed stones underlie the peat in the northern part of the western basin of Ruokolammensuo. Figure 2 shows that at points 1 600 and 1 800 m the elevation of the bottom is 24.0 and 24.5 m respectively. At point 1 900 m, which lies at an elevation of 26.0 m, as at other points above 25 meters, the bottom consists of sandy till. In the light of the material herein presented, it may thus be considered certain that the bottom of washed stones corresponds locally to the height of the Littorina maximum, which is approximately 24.5 m above sea level.

Hyyppä (1937) had previously investigated the postglacial changes in the shoreline at, *e. g.*, Virojoki, in the commune of Virolahti, and in the area of Hamina, quite close to the eastern and the western sides of Ruokolammensuo. The L I maximum obtained by him at Virojoki is 22—23 m and the L II maximum about 20 m, or, according to the relation diagram, approximately 21 m. The maximum of the Littorina Sea in the Hamina area reported by him is 23.5 m. The points investigated at Ruokolammensuo and in the Hamina area lie pretty much on the same isobase, while those at Virojoki lie perhaps slightly lower, although the difference cannot be great.

In carrying out his studies, Hyyppä levelled a large number of ancient shorelines. Certain altitudes of interest from the standpoint of the present investigation warrant mention. H a m i n a: Kettuvuori 24.3 m, Munavuori 23.9 m, 23.6 m and 23.5 m, and Rakila 23.5 m. V i r o j o k i: Pajulahti 24.3 m, 23.6 m and 22.6 m.

The value 24.3 m has been obtained at both Kettuvuori, in the Hamina area, and at Pajulahti. On the basis of the bog sediments studied by him in the latter locality, Hyyppä (1937) considers it too high a Littorina maximum for the area. On the other hand, it nevertheless corresponds well to the value obtained in the present study of Ruokolammensuo as the maximum height of the Littorina Sea. Accordingly, this value may be accepted as signifying the highest shore of the Littorina Sea in the areas where the research material discussed in the foregoing has been collected. The next highest concentrations of shore levels lie at elevations of 23.6 and 23.5 m, which values have been obtained at numerous different points. Comparison of this altitude with the maximum height of the other transgression met with in the Ruokolammensuo region will be made on the basis of diatom and macrofossil analyses in the following, along with discussion of the question as to which of the transgressions reached higher.

The salt-water diatom flora of Ruokolammensuo observed in Profiles 1—4 suggests a lagoon-type, fairly shallow marine stage. However, certain features of the occurrence of diatoms during the respective transgressions yield evidence of the relative depths of the water.

In Profile 1 *Campylodiscus clypeus* v. *bicostata* is the dominant species at the L I level but *Campylodiscus clypeus* at the higher level. After the disappearance of the former, there emerges alongside the latter *Amphora mexicana* v. *major* in abundance, while *Campylodiscus echeneis* waxes strong at a higher level. In Profile 2 also *Campylodiscus clypeus* and *Amphora mexicana* v. *major* occur in greater abundance in connection with the higher than with the lower transgression.

Halden (1929) has observed that *Campylodiscus clypeus* and *Amphora mexicana* v. *major* thrive best in brackish water, whereas *Campylodiscus clypeus* v. *bicostata* seems to prefer saltier water. Similar results have been presented by Cleve-Euler (1935 and 1952), Maj-Britt Florin (1946) and Sten Florin (1948). M-B. Florin's diagrams, furthermore, reveal that *Campylodiscus clypeus* v. *bicostata* generally occurs in slightly deeper water than the main form and the other diatoms just referred to.

Attention was drawn in the foregoing to the fact that *Najas marina* has been met with only in transgressive sediments. The plant thrives, according to various investigators, in lagoons or in shallow, protected bays with oozy bottoms. Backman's (1941) study, which includes data on the requirements of plants in regard to the salinity of water, reveals that this species can stand a fairly broad range of variation. *Najas marina* grows in fresh water but it also thrives in, *e. g.*, Lake Aral, the salinity of which is 10.8 $^0/_{00}$. More interesting are the data on the reaction of the plant to different depths. Ulvinen (1937) has studied the present occurrence of *Najas marina* along the coast of the Gulf of Finland near Kotka, about 35 km west of Ruokolammensuo. He has found the plant at depths of not more than one meter, and the same result has been obtained by Häyrén (1941) in studies carried out in the Sipoo archipelago. On the other hand, H. Luther, as cited by Backman (1941), reported finding the species generally at a depth of at most two meters; in rare cases, he found it in a fertile state at a depth of 2.5 m but, on occasion, in a sterile condition at a depth of as much as 3.5 m. Almquist (1929) reports the species to thrive best in water 0.5—1.0 meters deep.

On the basis of the present occurrence of *Najas marina*, it may be concluded that the optimal depth for it in sea water is approximately one meter. This datum is important in considering the case of Ruokolammensuo.

As noted in the foregoing, *Najas marina* seeds were met with in considerably greater abundance at the level of the higher than that of the lower transgression (Table 1). Profile 2 contains the most seeds. The altitude of the upper limit of the ooze is there 22.6 m. It may be estimated that in the case referred to the depth of water must have been optimal, or approximately one meter. When this value is added to the altitude of the upper limit of the fine detritus accumulated during the time mentioned, the result obtained is such that the maximum of the transgression dating back to L II was approximately 23.5 meters above sea level.

The evidence provided by the diatoms indicated that the water had been slightly deeper during the time of L I than that of L II. Since *Najas marina* seeds occur to a smaller extent in L I sediment than in that of the later transgression, the same conclusion is suggested. Since, however, seeds have been met with in all the profiles investigated, it is reasonable to assume that the water covering the area was of sufficient depth for the plant to exist satisfactorily, or about two meters. When this value is added to the highest limit of the ooze originating during L I time, or to 22.4 m in Profile 3, the maximum height of the L I obtained is approximately 24.5 m above present sea level.

The maximum heights of the Ruokolammensuo transgressions thus obtained correspond to the values previously reported from the region (Hyyppä, 1937). The difference between the maximum heights of the transgressions thus amounts to approximately one meter, and the L I represents the highest waterlevel of the Littorina Sea achieved in the region.

SUMMARY AND REVIEW OF RESULTS

Two Littorina transgressions have been met with in the Ruokolammensuo area, L I and L II. The former represents the highest limit of the Littorina Sea in the region, corresponding to 24.5 meters above present sea level. Investigations have placed the maximum of L II at an altitude of 23.5 meters.

Transgressive sediments have been encountered only in the southern part of the western basin of the bog, in an area of about ten hectares. The area was once part of a lagoon protected by two islands, and it began to turn boggy during the period between transgressions in spots now situated between 22 and 22.5 meters above sea level.

From the L I maximum height the water sank at least to the lower limit (22.0 m) of the peat formation referred to, or about 2.5 meters. This indicates the minimum, though not total extent, of the regression, which nevertheless probably closely approximates the value mentioned. During the L II maximum, the water rose to the 23.5 meter level.

Comparing the estimations of the Ruokolammensuo transgressions with the chronological table published by Hyyppä (1960), the age of L I would approximate 4 500 B. C. and of L II 4 000 B. C. The age of L I in the Askola region, according to the studies of Virkkala (1952), which are based on both bog-geological and archeological material, would also date back to 4 500 B. C.

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ON THE FAULT LINES IN FINLAND [1]

BY

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ABSTRACT

The trends of the pronounced fault lines in Finland are presented on the map. The main eskers follow the trends of the fault lines, and thus the formation of the eskers seems to be in genetic relation to the crustal movements. The net slip of the faults varies locally owing to the combination of crossing fault systems. The fault lines are primarily of Precambrian age, but movements have taken place at different times in different directions and thus these lines are now shear zones. The character of the block movements and faults is briefly discussed.

CONTENTS

INTRODUCTION

The rock crust of Finland was already peneplaned in Precambrian time. In the topography there exist, however, valley systems, in many cases manifested as watercourses, systems of lakes and rivers. More than half a century ago observations were made of their more or less rectilinear mode of occurrence, especially in the archipelago of southwestern Finland and in the lake districts of the continental area, and already at that time they were explained as shear zones (Frosterus, 1902, p. 97; Sederholm, 1910, 1911,

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1913; Tanner, 1911, 1938; *cf.* Hausen, 1940, 1942, 1944; Edelman, 1949, p. 37; Syvänen, 1957; Tuominen, 1957; Marmo, 1959; Härme, 1960, p. 40; *etc.).* Owing to the lack of accurate topographic maps at that time, attention was devoted to the shore lines of the seas and lakes as well as to the river systems, where these topographic forms were easily noted. At present the map material of Finland is much better, and aerial photographs are especially practical in this respect. This paper is a preliminary report on cartographic and photogeologic studies as well as on field observations made in connection with the geologic mapping.

THE MODE OF OCCURRENCE OF THE SHEAR ZONES

It is generally known that the earth's crust is broken into a mosaic of pieces, and this is also the case with the Finnish Precambrian. In the marked and especially in the rectilinear valleys, the crystalline rock has proved to be highly ruptured and sometimes also mylonitized. The very fractured nature of the rock in these valleys has favored weathering and therefore these fracture lines or zones now occur as valleys. Field observations show that fault movements have taken place along these lines.

Earlier it was supposed that the continental ice of the Pleistocene epoch strongly eroded the surface of the Precambrian rocks in Finland. During recent years, there has been general agreement, however, that the glacial erosion was strong only on the highest outcrops and even there probably only some meters or some tens of meters in magnitude. The preglacial valleys are in general well preserved, and in the bottom of the valleys the glacial erosion has often been rather slight. There are grounds to suppose that the more pronounced present valleys are of preglacial age. It has been proved that here and there in the bottom of the valleys occur thin layers of kaolin which gradually pass downwards into fresh crystalline rock. Unfortunately only a few such occurrences have been described *(cf.* Härme, 1949; Virkkala, 1959, p. 11).

The longest known coherent valleys situated on the fault lines are about one hundred kilometers long, as, *e. g.*, one between Porkkala and Lahti (Sederholm, 1913, p. 11; Härme, 1960, p. 41). Generally, however, the valleys are branched, and only some kilometers or some tens of kilometers long. They have been proven to be parts of different fault systems.

The manner of occurrence of the fault lines is usually of the known *en échelon* character. Such an *en échelon* fault series is often crossed by another similar one, and then the main direction of the valley does not follow either of the crossing directions (Fig. 1) but is oriented between them in one resulting direction. Of such character are, *e. g.*, some long lake basins such as those of Näsijärvi and Päijänne *(cf.* Sederholm, 1913, Figs. 1 and 2; Tanner, 1938, Figs. 158, 160 and 164). In such cases, however, the directions of the proper fault lines are demonstrable as jointing or as faults of small scale in the rock outcrops on the edges of the valley. The same is often revealed by the topographic forms of the valley, too.

The net slip of the faults varies greatly, even in the different parts of the same line. This is due to the combination of the crossing fault systems. According to the field observations, it seems likely that in most cases the displacement had not been solely in one direction, but rather that the blocks were moved to and fro at different times in different directions. Further, the displacements did not take place along only one plane but serially along parallel planes. There is usually a gradation from single fractures (shear joints) to true shearing (shear structure). So it is often more expedient to speak of shear zones than of fault lines.

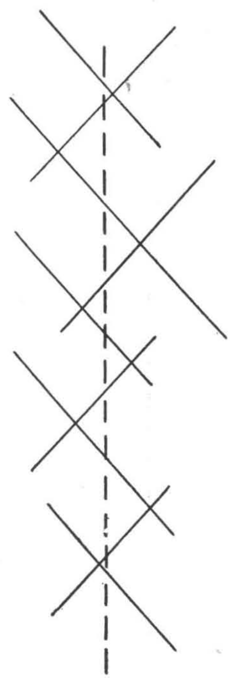

Fig. 1. Two crossing *en échelon* faults. The broken line shows the main direction of the valley formed by weathering. Schemed.

The contacts of two different rock belts or bodies are often crushed and sometimes even mylonitized. It is a quite natural phenomenon that movements take place at the contact of two rocks of different tenacity. It is to be noted, however, that the larger fault lines do not follow only the rock contacts but often cut across the areas of different rocks.

Examples of grabens are the sandstone area of Satakunta (Laitakari, 1924, p. 31), the area of the Muhos formation (Brenner, 1941, p. 548), the Gullkrona block (Edelman, 1949, p. 37) and the Orijärvi area (Tuominen, 1957, p. 23; Härme, 1960, p. 41).

In the appended map (Fig. 2) are presented the trends of the most marked fault lines in Finland. They are mainly compiled from topographic maps, partly from aerial photographs and partly from field observations. It is, perhaps, to be supposed that the continental ice scoured out the shear valleys which were oriented near the direction of its flow, whereas those with different orientations have been filled by the glacial drift and therefore are difficult to observe in the present topography. The field observations show, however, that the most conspicuous fault lines on the map are in fact the most pronounced faults in the field. On the other hand, it is to be noted that locally, on a small scale, the ice to some degree followed the trends of the valleys which were oriented slightly differently from the directions of main ice movement.

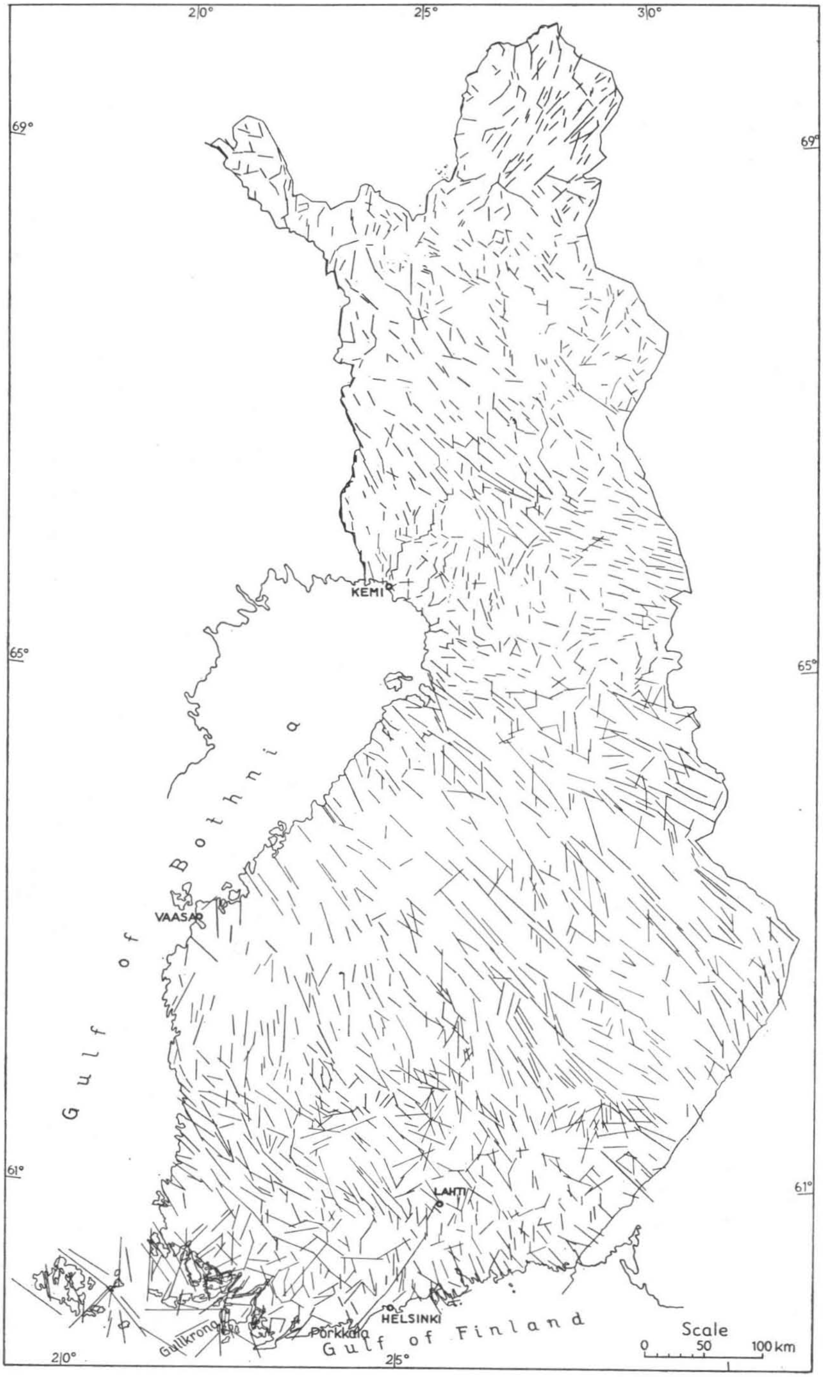

Fig. 2. The trends of the most marked fault lines in Finland.

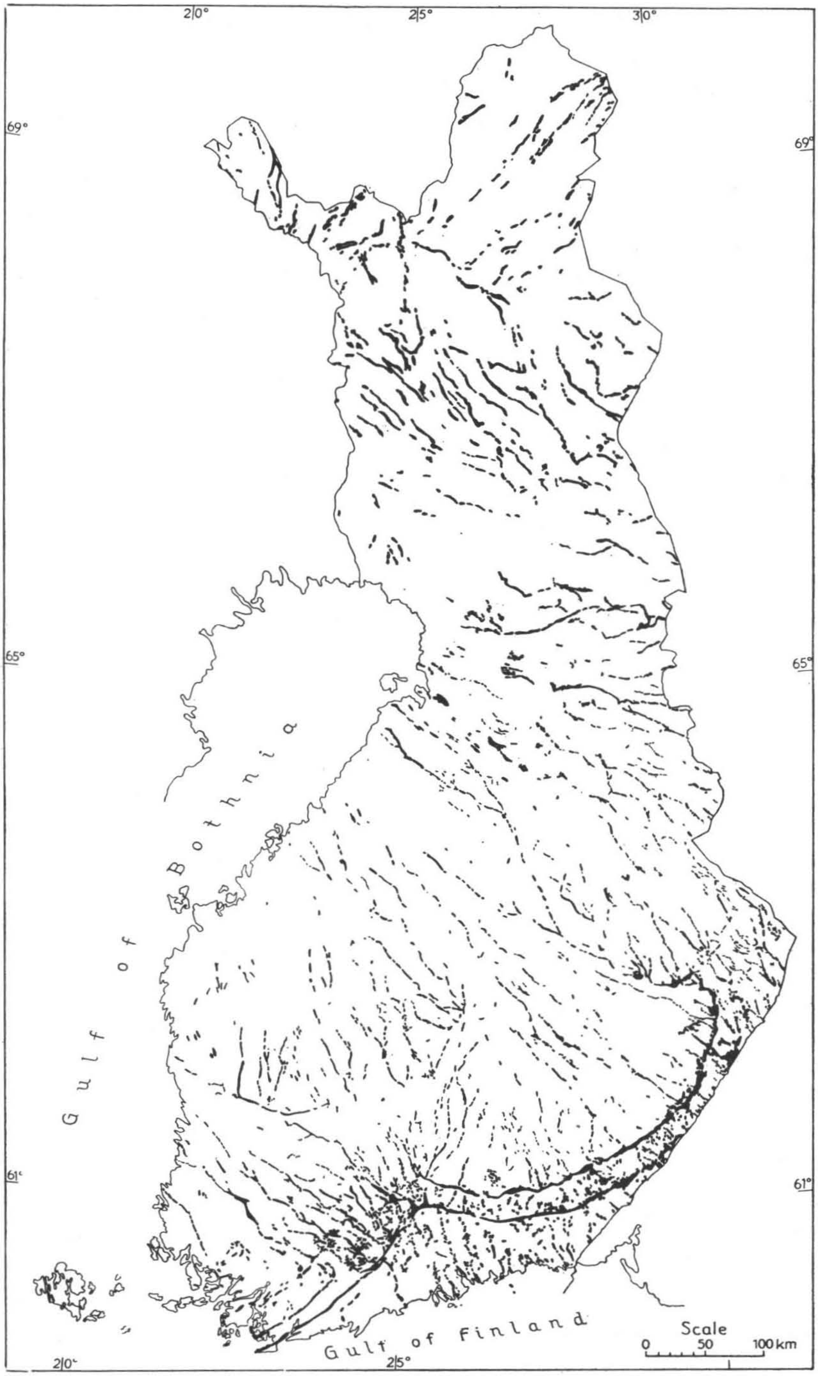

Fig. 3. The trends of the eskers in Finland (according to Okko, 1960).

THE TRENDS OF THE FAULT LINES AND ESKERS

In comparing the trends of the fault lines (Fig. 2) and those of the eskers (Fig. 3; *cf.* Okko, 1960), there are many similarities to be noted. In general the eskers follow the trend of the fault lines where these are prominently oriented. On the other hand, where the fault lines show no well-developed trend, the eskers are absent or at least have no pronounced general orientation (*e.g.*, SE of Kemi and Vaasa as well as in the NE-part of the map). In the southern coastal area the curves of the fault lines should be noted: in the southwestern part the curves are concave to the west and in the southeastern part they are concave eastward. These curves lie on both sides of the area between Lahti and Helsinki.

The relation between the eskers and the fault lines in Finland is dealt with by Hyyppä (1946, 1951, 1954), who has suggested that the eskers have been formed in fractures in the continental ice sheet and that these fractures were caused by the block movements of the earth's crust (*cf.* Sederholm, 1911; see the map). In this connection I have no wish to go into the theories of the formation of the eskers, but only to state that in general the main trends of the eskers follow the orientation of the most marked pre-Quaternary fault lines.

THE AGE OF THE FAULT LINES

There is considerable factual support for the conclusion that at least most of the fault lines are of Precambrian age (*cf.* Sederholm, 1913, p. 46; Edelman, 1949, p. 38; Härme, 1960, p. 42). Sandstone dikes of Cambrian age are met with in the archipelago of southwestern Finland (Tanner, 1911; Sederholm, 1913, p. 2; Simonen and Kouvo, 1955, p. 75). The Jotnian sedimentary formations in Satakunta and Muhos lie in grabens (Laitakari, 1925; Brenner, 1941; Simonen and Kouvo, 1955). The Precambrian granites of Obbnäs and Bodom are obviously genetically related to the afore-mentioned fault line of Porkkala—Lahti (Edelman, 1949, p. 38) but, on the other hand, later shear zones cut the Obbnäs granite (Härme, 1960, p. 42), thus showing that movements repeatedly took place on the same lines.

The late- and postglacial uplift of Fennoscandia is a well-known phenomenon (*cf.* Witting, 1918; Ramsay, 1924; Gutenberg, 1941; Kääriäinen, 1953). The area of the strongest upheaval, the center, is in the northern part of the Gulf of Bothnia, and a notable seismic center is also located in the same area (Sahlström, 1930). It is most likely that the movements related to the uplift have taken place along the existing fault lines (*cf.* Kvale, 1960) and that these movements have been associated with small earthquakes (*cf.* Honkasalo, 1960, p. 119). Since observations show that uplift has taken place during the whole of postglacial time, there are grounds to suppose that at least some movements occurred along the same fault lines during glacial time as well.

It is generally assumed that during Pleistocene time the earth's crust in Fennoscandia was pressed down by the ice load and that the late- and postglacial uplift is a return to isostatic equilibrium. This may be the case even if there is no evidence to indicate subsidence during glacial time. It is to be noted that at the present time vertical movements of the earth's crust take place in many regions where there is no reason to attribute it to the action of an ice load. Recently Kvale (1960) has remarked that the uplift in Norway is not as uniform as presumed and that the local uplift movements are connected instead with the seismicity of the North Atlantic area.

The map of the fault lines (Fig. 2) shows that the pronounced trends of the fault lines are to some degree radially arranged around the northern part of the Gulf of Bothnia *(cf.* Sederholm, 1913, Fig. 3). An exception is the area southeast of Kemi, where no prominent trend occurs. The radial arrangement of the fault lines around the uplift center supports the view that the uplift movements have taken place particularly along these lines.

Only some general features of the fault lines and their trends in Finland are presented in this preliminary report. The new topographic maps offer good possibilities for additional, more detailed studies, and especially in the aerial photographs the probable fault or shear lines are, in spite of loose sediments and vegetation, often easily discernible. The geophysical measurements and their interpretation (especially the magnetic, electric and gravimetric investigations) are also a valuable aid.

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26

THE APPARENT AGE PATTERN OF THE CRUST [1]

BY

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ABSTRACT

Isotopic datings are commonly low owing to the diffusion of the daughter element. The diffusion increases as the temperature rises. At great depths in the earth's crust a level is met with under which all geochronometers show a zero age. Owing to the isostatic rise, the apparent age of the continental blocks will decrease with the tenor of depth. The discrepancy observed between the radiometric measurements and classical field observations in Finland is briefly discussed from this point of view.

CONTENTS

INTRODUCTION

The accuracy of age determinations by radiometric methods has improved greatly during recent years. Chemical and isotopic analyses can be made with such precision that the ages calculated by different chronometric methods might agree within a few per cent. In many cases, however, the mutual concordancy is much poorer. This depends more on the character of the sample material than on the deficiency of the methods employed. It is a

[1] Received March 15, 1961.

well-known fact that potassium-argon datings of feldspar yield lower ages than determinations based on co-genetic mica. As shown by Wetherill, Aldrich, and Davis (1955) as well as by Afanasiev *et al.* (1960) the age calculated for microcline is systematically about 20 per cent lower than that for mica. It is pointed out by Folinsbee *et al.* (1960), among others, that the argon leakage is evidently connected by disintegration of the albite in the perthite. The effect of metamorphism on the different nuclear chronometers has been demonstrated by, *e. g.*, Long, Kulp, and Eckelmann (1959). It is evident that all methods K^{40}-Ar^{40}, Rb^{87}-Sr^{87}, Th^{232}-Pb^{208}, U^{235}-Pb^{207}, and U^{238}-Pb^{206}, commonly used for geochronometry, yield datings which may be low because of the diffusion of the daughter element from the analysed mineral.

THE EFFECT OF THE TEMPERATURE ON THE GEOCHRONOMETRY

To give correct age determinations, the mineral used for dating has to behave as a closed system during its entire history. This requirement is seldom if ever strictly fulfilled. During the vast period of time since the mineral was formed, some of the elements belonging to the decay series must have diffused in or out of the mineral. We are able to obtain nearly exact datings, however, if the diffusion took place slowly enough. The rate of diffusion flow is determined by the concentration gradient (dc/dx), by the surface (q) available for the diffusion, and by the diffusion coefficient (D)

$$s = q \cdot D \cdot dc/dx \qquad \text{(Fick's rule)}$$

During the existence of the mineral, the concentration of the parent element decreases at the rate of an exponental decay function, while the amount of the final daughter element increases at the same rate. The diffusion coefficient remains constant at a given temperature but might change greatly if the construction of the mineral is severely damaged by radioactivity. On the other hand, the coefficient depends on the temperature, on the type of mineral and on the diffusing element. The quantitative values for the diffusion of parent or daughter elements in the minerals used for dating are poorly known. The theory of diffusion in the radioactive minerals has been discussed, however, and a mathematical solution for the problem been given by Wasserburg (1954), Nicolaysen (1957), and Tilton (1960). Considering the effect of diffusion on the daughter to parent ratio, it was possible for Wasserburg (1954) and Fechtig, Gentner, and Zähringer (1960) to correct the measured K-Ar ages. Wetherill (1956) developed a graphical method that evaluates the age for series of discordant Pb^{207}/U^{235} and Pb^{206}/U^{238} measurements. The procedure also yields the time at which the loss of lead took place. As pointed out by Tilton (1960), the application of this method would give the remarkable result that many old minerals all over the world lost lead around 500 million years ago. Tilton showed that the departure from the concordia-curve can be explained by continuous loss of lead rather than by an episodic loss mechanism as assumed by Wetherill. Tilton (1960) calculated the position of the »discordia» curves on which the corresponding diffusion values D/a^2 can be read (a is the active radius of the mineral). The deviation from the concordia line depends on the diffusion coefficient and on the grain size and shape of the mineral.

The temperature dependance of the diffusion coefficient is given by the relationship:

$$D(T) = D_0 \exp(-Q/RT)$$ in which Q denotes

the activation energy of diffusion and R is the gas constant.

Owing to the exponental relationship, the diffusion increases rapidly at higher temperatures. On the other hand, this effect can still be increased by the increase of activation energy with rising temperature. This situation is observed in metal alloys, and Fechtig and co-workers (1960) were able to observe it also in the diffusion of argon in anorthite, augite, fluorite, and margarite. At temperatures above 500°C the diffusion of argon begins to increase at such a rate that practically all the argon is driven out of the mineral in a relatively short period of time, *e. g.*, in from 10 to 100 million years.

Tilton (1960) discussed the temperature dependency of the activation energy for lead diffusion in zircon and is of the opinion that the diffusion should be controlled by temperature-independent vacancies in the crystal and that the activation energy is low. The numerical value estimated by him is surprisingly low.

The diffusion of argon takes place easily because it is foreign to the mineral lattice. In principle all the daughter elements formed by the nuclear decay are foreign to the minerals of the parent element. Consequently, the rules controlling the diffusion should be principally the same for all the isotopic chronometers. Severe losses of the daughter element can be caused by prolonged heating at elevated temperatures. The ultimate state for all the timers at high temperatures will be the same: the leakage of the daughter element out of the mineral is faster than the rate of its creation resulting from the radioactive decay of the parent. As this situation is reached, the daughter-parent ratio will decrease and the apparent age of the mineral will finally approach zero. The chronometer stays at zero as long as it remains at this high temperature but starts again to measure time as soon as the temperature is lowered. Any later heating, if sufficiently high and long, can rewind the chronometer for a new timing. A rock complex containing different types of chronometers gives a heterogeneous age pattern if the formation was reheated at a temperature able to zero-set, say, the K-Ar meters but not sufficient to affect the Rb-Sr dating. This same heating might slightly have changed the U-Pb ratio in zircon. Examples of discordant ages of this type are plentifully found in the recent literature, *e. g.*, Long, Kulp, and Eckelmann (1959). It is evident that the temperature and time needed for zero-setting vary greatly in the different chronometers pairs and depends on the type of host mineral.

As demonstrated by Tilton (1960), ages plotted on the Pb^{207}/U^{235}—Pb^{206}/U^{238} diagram move from the original discordia curve towards a new discordia curve if the mineral lost lead as a result of an episodic increase of diffusion rate. As the lead loss approaches 100 percent, the point approaches the new discordia curve, and the mineral will be dated young.

The common lead or model lead dating differs fundamentally from all the others. It might seem reasonable to assume that this type of chronometer cannot be affected by the elevation of temperature. At very high temperatures, however, the diffusion might attain such values that during prolonged periods of time the isotopic composition of a stable lead mineral changes so as to be the same as that of the surroundings. Consequently, the zero-setting of the common lead chronometer is possible but certainly requires much more time and a higher temperature than any other isotopic timer.

THE TEMPERATURE IN THE CRUST OF THE EARTH

According to findings from seismic waves, the earth is divided into crust, mantle and core. The lower boundary of the crust is defined by the Mohorovičić discontinuity. This boundary is evidently caused by a chemical change from a simatic gabbro layer to an ultra-simatic ultrabasite layer (Adams, 1951). The depth of this boundary, Moho, varies from place to place. Under the ocean bottom, the discontinuity is met with at a depth of less than 15 km but it goes down near the coast and depths of over 60 km are measured under the high mountains in the continental area. In the uppermost part of the crust a granitic Sial layer is assumed. It is separated from the Sima by a so-called Condard discontinuity at a depth of about 5—10 km under the oceans and about 15 km under the continents.

Several facts point to the fact that the temperature rises with the depth in the upper lithosphere. The thermal gradient is observed to vary in different parts of the world, but the value 1 degree/30 meters is generally accepted as an approximate in the continental areas. At ocean bottoms the gradient measurements indicate a slightly higher estimate, about 26 meters/

degree. Many trials have been made to interpret the temperature and thermal gradient at the greater depths (*e. g.*, Verhoogen, 1956). The estimates vary considerably depending upon the basic assumption concerning the origin and generation of the heat in the earth. The temperature of about 650°C, critical at least for the diffusion of argon in the silicate minerals, is met somewhere at the depth of between 20 and 50 km. From the point of view of geochronometry this means that at this depth a line is met under which the K-Ar chronometer is not able to measure time. Theoretically, the same holds for all the nuclear timers. The depth of the »zero-line» varies, however, for the different chronometers and in the different parts of the earth owing to the thermal pattern of the crust. It can be expected that the zero-line for a given chronometer is generally met with at a lesser depth under the oceans than under the continents because of the difference in the thermal gradient. On the other hand, the zero-line for rubidium-strontium dating evidently lies deeper than that for the potassium-argon method.

It should be pointed out that melting and recrystallisation will be able to zero-set the timers is any conditions. Therefore, the zero-line must be considered as an ultimate level below which the age of the mineral is not valid. Recrystallisation can take place above this line and intrusive bodies can rejuvenate chronometer minerals. The minerals in unmetamorphosed clastic sediments originate from different formations and can therefore differ greatly in age. Such a complex cannot display uniform and concordant datings before it is metamorphosed. Since this mainly takes place during diastrophism, the age shown by the metamorphosed sediments will be that of the orogeny.

THE APPARENT AGE PATTERN

Theoretically, it is possible to assume an ultimate zero-line under which all the chronometers point zero. During the geological history of the earth, the zero-line has been under a process of steady evolution. Because of erosion and sedimentation, the continents are isostatically rising and parts of the ocean bottoms are sinking down. Under the process of sinking, new portions of the lithosphere are pressed down below the zero-line. When rising, new parts of the crust are moved above this line and the chronometers are released for a new timing. On an old continent, the formerly deep parts of the crust have reached the top and are exposed for our studies. The age pattern of such a part of the crust, if it could be measured by radioactive datings, would show a decline corresponding to the tenor of the depth. The age pattern would be homogeneous for a given type of chronometer from the top of the crust to the zero-line of this particular mineral, if the rise of the block took place at a steady rate. The geological evidence in-

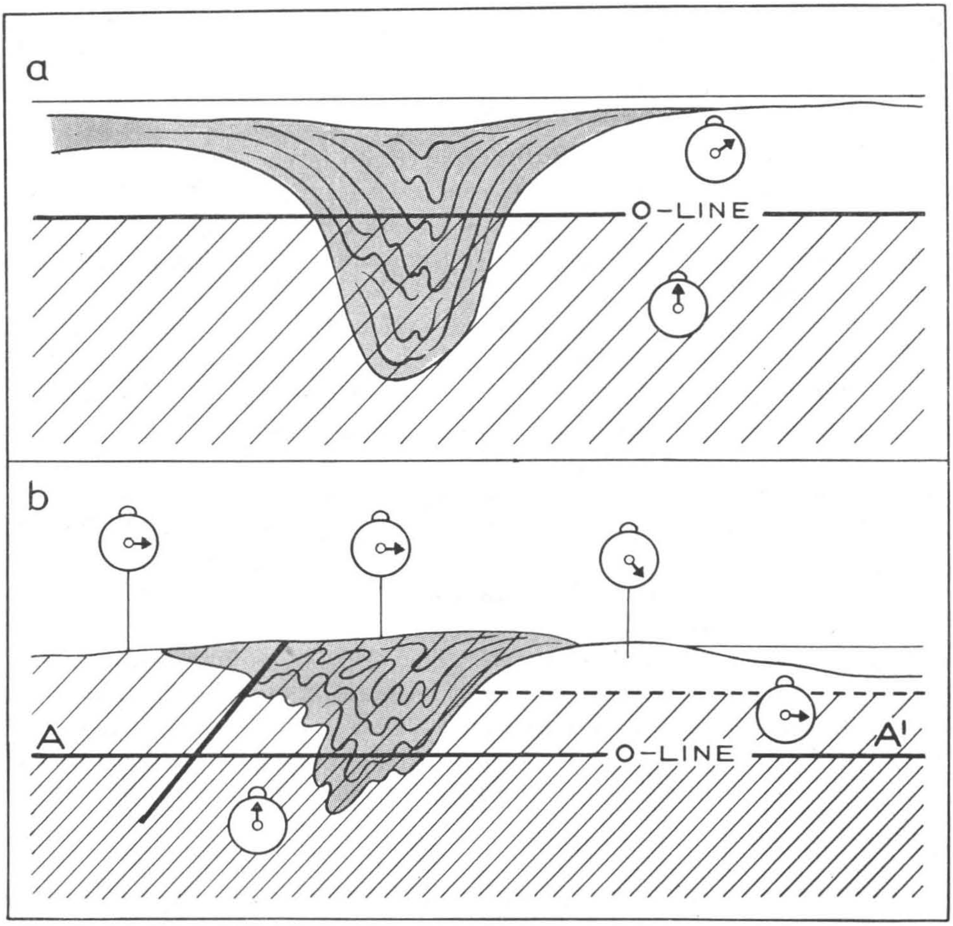

Fig. 1. Evolution of the apparent age pattern in the crust of the earth.
a) The youth of mountain chain and the adjacent basement.
b) The same some hundred million years later.

dicates that this is not usually the case. The isostatic anomalies remain unbalanced until an epoch of diastrophism affects the whole upper lithosphere and the released forces cause vertical movements in continents far from the orogenic belt proper. Consequently, the nuclear age of a continental block will decrease stepwise so that each step will correspond to an epoch or period of universal diastrophism. The distribution of the mineral dates given by Gastil (1960) might be interpreted to display such steps.

During the youth of a geosyncline the sediments are assumed to be pressed down into the simatic depths (*e. g.*, Umbgrove, 1950). The deepness of the present oceanic troughs (Gutenberg, 1959) and the thickness of the sediments in at least some of the ocean trenches (Ewing and Heesen, 1955) suggest that a great part of the geosynclinal sediments reach depths where the zero-line for the common geochronometers is achieved. The same is also clearly indicated by the high grade of metamorphism and by the palingenesis observed in the rocks of the deeply eroded old continents. The situation is schematically shown in Fig. 1a. Part of the continent and the deepest part of the geosyncline sediments are below the hypothetical zero-line. The upper part of the basement is above the line and its age is not affected by the orogeny. The same mountain chain is shown in the schematic section of Fig. 1b some hundred millions years later. The sediments and the adjacent continent have risen up and the denudation has caused deep parts to crop out of the mountain chain.

On the right side of Fig. 1b the old basement is exposed with the unaffected chronometers showing the age of its original creation. It is covered, in the most parts, by metamorphic sediments which display the age of the orogeny. This same age is met with in the geosynclinal zone in the middle of the section. In the left part of the drawing, the vertical movements have raised the lower portion of the old basement to the level of erosion. It will show the age of the young orogeny since it originates from the depth below the zero-line. The part of the crust, sediment or basement which reached only the upper part of the zero-zone during the orogeny, will exhibit discordant datings.

The rocks adjacent to an orogeny may, as seen, display radiometric ages ranging from the age of the old basement to the age of the diastrophism. On the margin of the continent several successive orogenic belts can follow each other (Wilson, 1959). In deeply eroded areas only the roots of the oldest of these are observed by nuclear dating. In an ultimate case, the geochronometry will entirely fail to detect the age difference between these rocks. This situation is achieved when the erosion has reached the line A-A′ in Fig. 1b.

THE AGE PATTERN OF THE ROCKS IN FINLAND

The stratigraphy of the Precambrian formations in Finland was outlined by Sederholm *(e. g.*, 1932) and others during the early decades of this century. Two main cycles of sedimentation have generally been accepted by Finnish geologists. The Karelian system in eastern and northern Finland has been assumed to be younger than the Svecofennian system in the southern and western parts of the country. The nuclear datings given by Kahma (1956), Kouvo (1958), Polkanov and Gerling (1960), and Kouvo and Kulp (1961) do not, however, fit in to the old frames too well. According to these measurements, granitic intrusions in both the eastern and western parts of the country display an age of about 1 800 million years. The age of the basement under the Karelian formations was dated at 2 700—2 800 million years old (Kouvo and Kulp, 1961). The discrepancy between the geological reasoning and nuclear measurements inspired an intensive discussion of the age problem in the pages of the professional Finnish news letter »Geologi» during the two last years (Eskola, Marmo, T. Mikkola, and Paarma). Recently the problem has been thoroughly summarised by Metzger (1959) and Simonen (1960) and to some extent by Eskola (1960). Metzger and Simonen tend to believe that the Karelian and Svecofennian formations represent different sedimentation and tectonic environments during one single cycle, in accordance with an assumption made by T. Mikkola in »Geologi» in 1953. Eskola and some others are of the opinion that geological facts support the existence of two orogenic cycles but that the invasion of potassium and granitization rejuvenated the Svecofennian intrusives during the Karelian diastrophism.

Additional radiometric datings are needed before this predominant problem of the Finnish stratigraphy can be solved. On the other hand, it is necessary to know how the datings should be interpreted. According to the ideas discussed in the present paper, the two opinions referred to can be demonstrated by means of Fig. 1b. The old basement with the age of 2 800 million years under the Karelian sediments corresponds to the basement on the right side of the drawing. If the Svecofennian rocks actually date from a very ancient period, they are represented by the basement situated on the left in the picture. If Metzger and Simonen are right, the Svecofennian rocks correspond to the geosynclinal formations in the middle of the section. Since the geosyncline sediments and the basement on the left both were simultaneously down below the ultimate zero-line, they do not differ from the point of view of geochronometry. Consequently, the age problem in Finland can hardly be solved by mineral datings.

The aspects presented here are highly schematic. It is still possible that parts of the crust now denudated in western and southern Finland never did go down to the ultimate line. Consequently, older, even though discordant, datings will possibly be measured in the Svecofennian area, if it really is older than the Karelian rocks. Some of the common lead measurements performed point, however, in the opposite direction (Kouvo and Kulp, 1961).

Even if the stratigraphy is never solved with the mineral datings for deeply eroded areas, as in Finland, the measurements should not be omitted. As indicated by many earlier studies, discordant datings are quite useful in interpreting the structural evolution of the crust components. The implication of mineral datings for younger orogenic belts might likewise, if the ideas presented in the present paper are tenable, give valuable information concerning metamorphism, mineral equilibrium, and possibly the mechanism of the secret of mountain building.

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27

HEAVY METAL ANALYSIS OF HUMUS IN PROSPECTING[1]

BY

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ABSTRACT

The application of heavy metal analysis of humus to prospecting is studied by testing the method on fields with and without known mineralizations. The accuracy of the method and the dependence of the heavy metal concentrations in humus on the type of vegetation and on the nature of the underlying soil are also studied. In the analysis a semi-quantitative colorimetric titration method is used. The results although they must be taken with reservation, indicate that the method may be useful in locating buried ore deposits.

CONTENTS

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INTRODUCTION

Geochemical prospecting is based on the fact that the metals of most mineralizations have been distributed also into their environment (bedrock, soil, water, organic materials, *etc.*) thus causing higher than average metal concentrations. When anomalous concentrations are found in analyzed material, the related mineralization may not be far away.

For the preliminary geochemical exploration of large areas, a rapid, cheap and sensitive method of analysis is needed. The material to be analysed must be homogeneous, evenly distributed over the region and easy to sample. For this purpose many kinds of materials have been tried. Encouraging experimental results have been obtained by analysing humus (organic litter above mineral soil) in glaciated regions (Norway: Vogt and Berg, 1946 and 1948; Heltzen, 1956; Canada: Ermengen, 1957; Warren, Delavault and Cross, 1957), but also opposite results have been published from other areas *(e. g.* Fulton, 1950).

The first experiments in applying of humus analysis in prospecting in Finland, made by Lindberg at the Kolima ore field, gave promising results — the method was thus discovered to be worth closer examination and taken into routine use in making tests. For a detailed investigation of the discovered anomalies, more accurate methods are needed. It is, of course, possible to use customary materials, even if such means are more expensive. For this purpose vegetation (Rankama, 1940; Marmo, 1953), peat (Salmi, 1955 and 1956), waters (Marmo, 1958) and glacial till *(e. g.* Kauranne, 1958) have been used by the Geological Survey of Finland.

Until now (1959) some 50 000 heavy metal determinations have been made of humus from several exploration fields with diverse types of mineralizations. From these fields some have been chosen as examples for the present study. The exposure of the ore deposit of Petolahti was known before the humus investigation was performed and therefore it was easy to determine the treshold value of the anomaly as to clearly reveal the influence of the mineralization, but on the other fields described here the investigation was made before the outcrop of mineralization was found. Lindberg and Lyytikäinen are responsible for the work done at the Kolima field and Kauranne for the rest of the following study.

METHOD OF ANALYSIS

The determinations were made by using a slightly modified method of semi-quantitative colorimetric titration of total heavy metals (Cu+Zn+Pb) described by Bloom (1955). The method is simple and may used by nontechnical personnel. As a leaching agent the ammonium citrate buffer

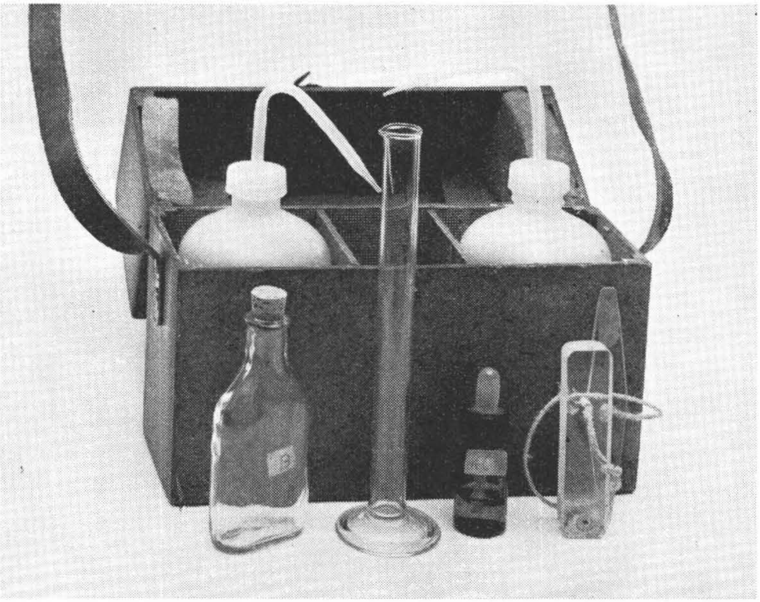

Fig. 1. Kit for field analysis. It contains: 1 measuring cylinder (25 ml), 1 plastic measuring scoop and spoon, 2 plastic squeezing bottles (one for dithizone and the other for buffer pH 8.5), 2 small bottles (for buffer pH 2.0 and KCN solution).

(pH 8.5) is used; it takes the exchangeable heavy metals from humus complexes into solution, but apparently does not attack mineral grains or more stable compounds. The results are given in milliliters of 0.03 percent dithizone xylene solution needed for the titration. Specific tests are used for copper and lead. The presence of lead is tested by adding a few drops of 0.5 percent KCN solution into the measuring cylinder where the analysis is made — if the red colour of dithizone remains, the sample contains lead. The colours of copper and zinc dithizonates differ slightly but the colour of humus solutions may mask the difference. The determination of copper is, however, possible by using a more acidic buffer, pH 2.5 (Holman, 1956) or using a spot test with rubeanic acid (Warren, Delavault and Cross, 1957).

The analysis may be made either at the sampling site or in a field or other laboratory. One person is able to perform 100 analyses a day in the field, or 200 analyses a day in a laboratory. If the weather is cold and wet, it is more convenient to collect the samples in small paper bags and analyse them later on. Sampling along paths is easier than orientation with a compass, one person may take over 200 samples on the average per day. The field-analysis kit used by Geological Survey of Finland is shown in Fig. 1.

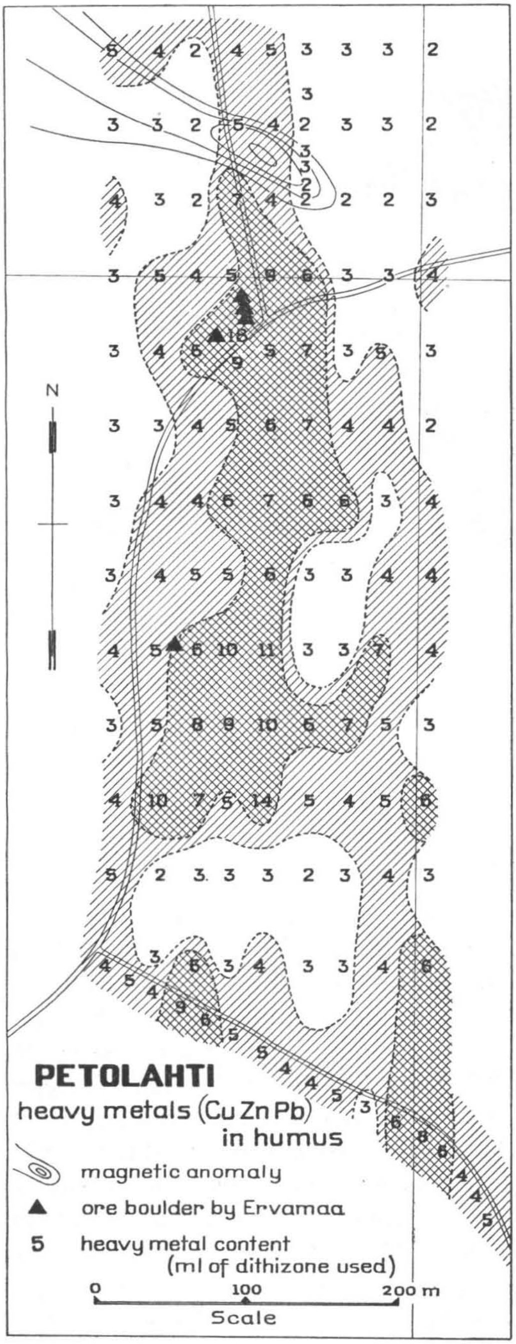

Fig. 2. Heavy metal concentrations in humus, Petolahti, West Finland.

AREAS INVESTIGATED

PETOLAHTI

A new pyrrhotite, pentlandite, chalcopyrite mineralization at Petolahti, western Finland, was recently found in diabase (description according to P. Ervamaa, pers. comm.) and there an opportunity arose to investigate the distribution of heavy metals (Cu) in humus. The region is flat mostly forested land, the surficial deposits being predominantly till but also containing some areas of sorted sediments. It is sparsely settled, with practically no danger of contamination. The only really serious source of contamination is the road running through the field investigated.

The humus samples were taken at intervals of 20 meters along lines running W-E, 50 meters apart. The analytical results are presented in Figure 2. The background value was taken as 4 ml of dithizone. The highest concentrations met with, close to ore boulders, corresponded to 18 ml of dithizone solution. No appreciable anomaly could be found above the outcrop of ore, whereas a fan-shaped anomaly extends from the outcrop in the direction of glacial ice transport (S 5°E).

The fan-like form of the anomaly is disturbed by cultivated fields, where the humus has been mixed with mineral soil, indicating that the heavy metal contents of this kind of humus are

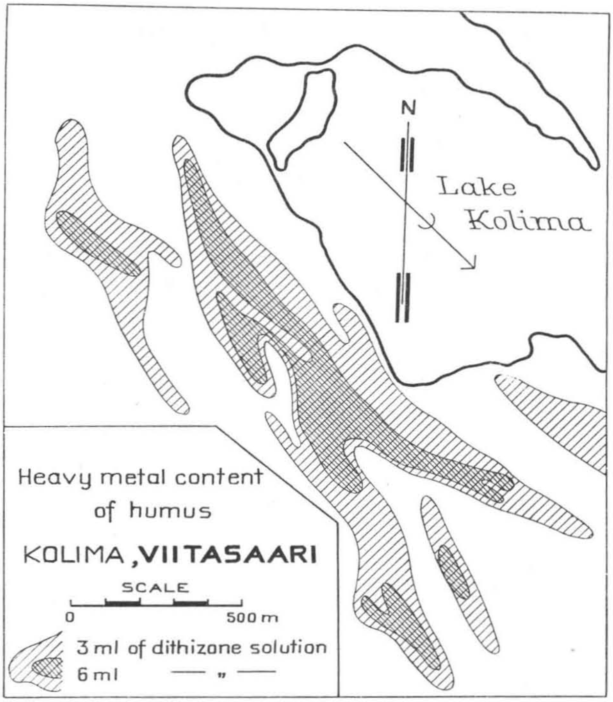

Fig. 3. Heavy metal concentrations in humus at Kolima Viitasaari, Central Finland.

not comparable with those of forested areas. In the area represented in Fig. 2 only this one ore deposit is known. The terrain slopes slightly northwards, which is also the direction of water flow in dikes. Transport by glacial ice appears to be the only cause on the elongated anomaly form; the anomaly in humus thus reflects the anomalous heavy metal contents of the underlying glacial till.

The highest values found are 4—6 times the background but most of the anomaly is only 2—3 times the background, the anomaly being consequently low but marked in shape. The apex of the fan-shaped anomaly points towards the outcrop of the ore body. The result of this testing may be considered to be positive.

KOLIMA

In the commune of Viitasaari, Central Finland many minor mineralizations are known. Some of them have been studied also geochemically. The exploration at Kolima was begun after the finding of some boulders with a

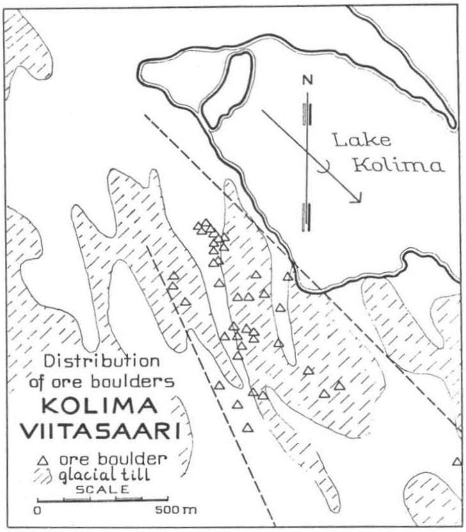

Fig. 4. Distribution of ore boulders at Kolima.

highly diverse mineralization: galena, sphalerite, arsenopyrite, chalcopyrite, pyrrhotite and pyrite. Later on, a garnet- and zinc spinel-bearing mica gneiss zone showing the same type of mineralization was discovered by diamond drilling (description according to Aatto Laitakari, pers. comm.). The region is sparsely settled forest land, the surficial deposits consisting of till, but there are also extensive areas of sorted sediments and bog peat.

The samples were taken at 20 meter intervals along lines perpendicular to the movement of ice situated 100 meters apart. Figure 3 presents the results of humus analysis in milliliters of dithizone solution. Only the central part of the area investigated is shown in the picture. Here likewise the anomaly is elongated along the direction of glacial striae (S 50°E). In the fan-shaped anomaly there occur also here some gaps caused by cultivated fields. Figure 4 shows the distribution of ore boulders and the outcrops of till. The strike and dip of the gneisses are approximately N 40°W 60°S. In Fig. 5 the zinc and lead concentrations of soils are presented.

The anomaly in humus is somewhat complex in form because of the disturbance produced by cultivated fields, but it clearly reveals the direction of the glacial ice transport and points to the known outcrop of the mineralized horizon. The result of this test may also be considered to be positive in a prospecting sense — the result together with results got from the analysis of till, C-layer, boulder tracing and geophysical investigations was used in planning the diamons drillings.

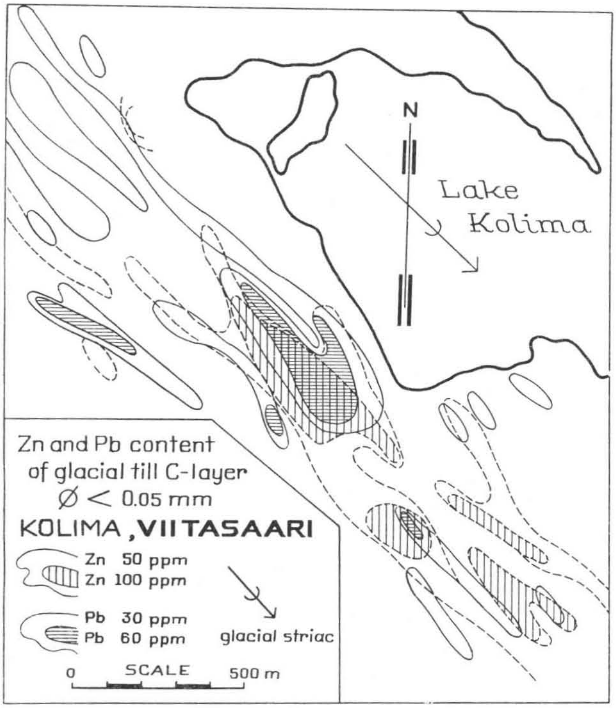

Fig. 5. Zinc and lead concentrations in soil, C layer, Ø < 0.05 mm at Kolima.

KOTANEN

The study of the Kotanen field in the commune of Viitasaari gave results differing from those described in the foregoing. In this area some chalcopyrite- and arsenopyrite-bearing boulders were met with, but diamond drilling did not bring to light ore bodies. Only scattered disseminations of chalcopyrite were found and the concentrations of precious metals in the arsenopyrite boulders were too low to justify continued exploration (description according to V. Pääkkönen, pers. comm.). The geochemical investigation was carried out however, together with a closer study of ice movements by means of till fabric analyses. This area is uninhabited and far from roads there are no possibilities for contamination.

The soil and humus samples were taken at 20-meter intervals along lines, perpendicular to the youngest ice transport, 100 meter apart. The copper content of the till (C-layer, —250 mesh = < 0.05 mm fraction) is shown in Figure 6 and, as may be seen, the anomalies are strong and narrow and elongated along the youngest ice movement S 30°E. In till at Kotanen two

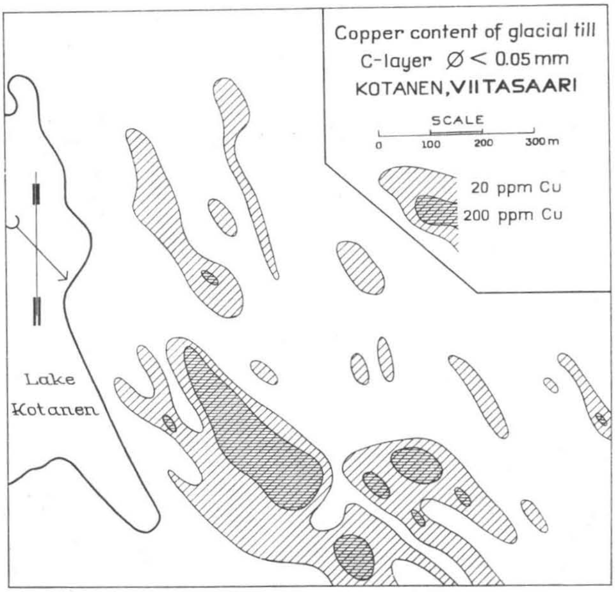

Fig. 6. Cu concentrations in till (C-layer, $\lesssim$ 0.05 mm fraction) at Kotanen.

different beds were found, the lower one was transported S 80°E according to preferred orientation of stones.

The areas of anomalous copper concentrations in till are elongated in the direction of the youngest ice movement, and because the strike of the mineralized gneisses almost parallels the direction of the youngest ice movement, the anomalies are also narrow and comparatively strong. The anomalous concentrations of heavy metals in humus (Fig. 7) do not, however follow anomalies in the till. Anomalies in humus are low and mostly situated on hillsides, whereas the anomalies in till do not reflect the rather marked topography of the area. The valley bottoms here are covered by bogs — it was observed that the humus of bogs gives quite erratic, mostly very low values, which do not correlate at all with those in the humus of the forested area nearby (cp. Ermengen, 1957), and therefore there are barren areas in anomalies in bogs.

The higher concentrations of heavy metals in humus in the Kotanen field do not correspond to the copper concentrations in till or the distribution of ore boulders. The anomalies in humus are slightly elongated in the direction of the latest ice movement but the real cause of the situation of these anomalies is not known. The result obtained from this experiment may be considered to be negative for prospecting purposes.

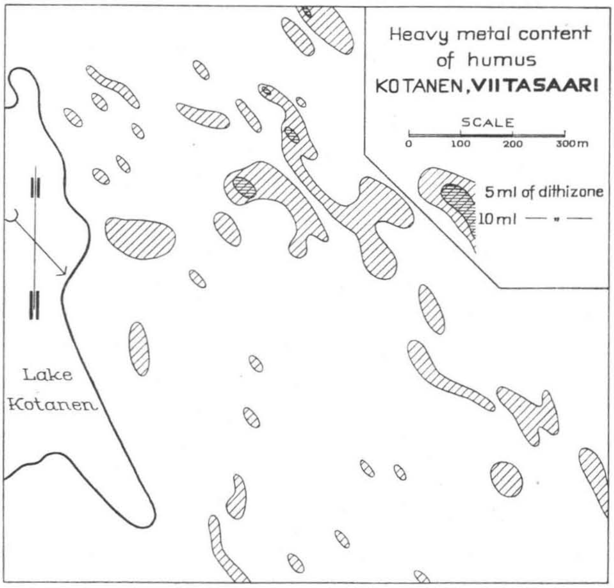

Fig. 7. Heavy metal concentrations in humus at Kotanen.

EXAMPLES FROM OTHER AREAS

In the Ritovuori area, commune of Pihtipudas, Central Finland, several vein-like galena, arsenopyrite mineralizations have been found and also a large number of ore boulders with the same mineral paragenesis. It was thought that here the geochemical investigation, especially the humus analysis should give good results. No anomalies of heavy metals in humus were found and only As was found in some samples of till in anomalous amounts, but these did not form any coherent anomalies either. Whether this is due to some peculiar reason in the nature of till or to certain other factors is not known. The result must be considered negative.

In the Ritomäki area, commune of Kärsämäki, Central Finland, some sphalerite-bearing pyrite boulders were met with. Later on a pyrite-sphalerite mineralization in the bedrock was encountered by diamond drilling. The ore boulders are concentrated in a small area some 1.8 km SE from the mineralized horizon, but the Zn contents in the till form a long and wide, indistinct, fan-shaped anomaly, which is mostly situated under a large bog. The samples from the till under the bog were taken with a motor-driven auger in winter time. Only the western flank of the anomaly in the till is situated on dry land and beside (not upon) this, close to the apex of the fan and close to the mineralized horizon, a 200-m-long weak heavy metal anomaly in humus was found. Humus anomaly like the Zn anomaly in the till run along ice transport S 50°E. Because the investigations in this area are still in progress no definite conclusions about the value of humus analysis can be made, but it seems that in an area like this, with large bogs, the possibilities to use humus analysis in prospecting are rather limited.

In the Pyrrönperä area, commune of Kärsämäki, Central Finland, a boulder very rich in chalcopyrite and a weak sphalerite mineralization in the bedrock were found and consequently a large area was studied also geochemically. Only some scattered anomalous concentrations of copper but quite an abnormally high zinc background were found in the till (C-layer, < 0.05 mm fraction). In the humus several uneven areas of high heavy metal concentrations were found, none of which, however, when investigated more closely by soil analysis or excavations led to the finding of ore mineralizations. The cause for the high background value for Zn and the many anomalies in the humus, except those found along roads, is not known. In places the bedrock exposures contained traces of a preglacial weathered mantle and the till presumably also contained weathered material. Perhaps this accounts for the high background value of zinc in the till and this in turn for many of the heavy metal anomalies in the humus. Along the roads there were always higher concentrations of heavy metals in the humus than in nearby forests, being evidently due to contamination by motor traffic (rubber, gasoline or traces of metals). Roads must therefore be avoided in humus investigations. In the light of the facts described, the result of this investigation may be considered to be negative. Neither humus, nor soil analysis did help us in prospecting for the source of the ore boulder mentioned.

It should be mentioned that with the exception of the Petolahti, Kolima and Ritomäki areas, in the exploration areas described here where no ore grade mineralizations have been found and where prospecting with humus analysis has failed, the other exploration methods did not yield any positive results either.

ACCURACY OF THE METHOD

As may be seen from the foregoing examples, the results of humus analyses are not always easy to interpret. Sometimes high anomalies are found in an area where no ore is known or found later on and sometimes no definite anomalies are met in connection with known ore bodies. Because the method has also given good results, the possibilities and limitations of its use should, however, be studied, especially in view of the fact that a method for geochemical reconnaissance work is urgently needed.

Table 1. The effect of the inhomogeneity of humus on the analytical reproducibility.

Weighing mg	100	100	200	500	1000	average for 100 mg
No.	ml of dithizone-xylene solution consumed in titration					
18261	1	2	2	9	18	1.7
18262	0	0	0	0	0	0
18263	0	0	0	0	0	0
18264	0	0	2	5	12	1.0
18265	2	5	2	9	15	1.7
18266	3	3	3	9	17	1.8
18267	5	5	8	12	32	3.3
18268	0	5	0	9	40	2.8
18269	6	8	5	18	53	4.7
18270	6	16	8	83	123	12.4

INHOMOGENEITY OF HUMUS

Humus may be defined as partly decayed organic litter mixed with mineral material above layers of mineral soil. Some samples of several types of humus were studied microscopically and found to contain organic tissue (leaves, roots, *etc.)*, mineral grains and remains of seeds and insects. The amount of organic material was studied by burning the samples and ashing them at 550°C. The amount of organic material in humus samples taken from forest soil varied from 50 percent from coniferous areas to 90 percent from areas rich in deciduous trees. In the case of bog humus very little mineral material was observed under the microscope and the ashing proof showed that more than 99 percent of the samples always consisted of organic material. Less than 20 percent of the humus taken from cultivated fields consisted of organic material. All types of humus appeared to be quite heterogeneous in composition and structure. An experiment to study the effect of the inhomogeneity of humus on the results of mass heavy metal analysis was carried out by taking several weighings from the same series of samples, which was known to contain both low and high heavy metal concentrations.

It is evident that a larger sample weight should give better results, but, as is shown in the foregoing table, the correlation between the 100 mg weighing and the average seems to be sufficiently good for prospecting purposes. When a series of 40 samples was studied in this manner the correlations between the average and a 100-mg weighing and the correlation between the average and a 1 000-mg weighing were calculated with the Pearson correlation coefficient formula:

$$r = \frac{\Sigma xy - \dfrac{\Sigma x \cdot \Sigma y}{N}}{\sqrt{\left(\Sigma x^2 - \dfrac{(\Sigma x)^2}{N}\right)\left(\Sigma y^2 - \dfrac{(\Sigma y)^2}{N}\right)}}$$

x and y = values of different samples in two series

N = number of samples in a series

The results obtained by calculation were r (100 mg/average) = 0.72 and r (1 000 mg/average) = 0.69. This time the lesser weighing thus happened to have a stronger correlation with the average than the greater — the difference being insignificant.

LEACHING POWER OF COLD BUFFER SOLUTION

Another experiment gives us an idea of the leaching power of the cold ammonium citrate buffer solution. A series of humus samples was analysed by total cold titration, afterwards ashed and the ash analysed spectrographically. The following table shows the relationship between the amount of exchangeable and total heavy metal contents of humus.

From a series of 21 samples from Kolima, analysed in the manner shown in Table 2, correlations were calculated, the results being as follows:

r (total Cu in dry humus/d) = 0.30 (slight correlation)
r (total Pb in dry humus/d) = 0.04 (no correlation)
r (total Zn in dry humus/d) = 0.70 (good correlation)
r (j/d) = 0.73 (good correlation)

Table 2. The relation between exchangeable and total heavy metal contents of humus.

No.	a	b	c	d	e	f	g	h	i	j	k
52	0	0	2	0	40	45	—	85	64	54	0
56	2	1	18	1.5	90	50	400	540	37	205	10
60	13	12	25	12.5	90	50	700	840	53	447	100
72	20	18	30	19	45	40	1 600	1 685	59	1 050	170
76	9	16	25	12.5	160	80	2 000	2 240	23	504	100
80	3	8	11	5.5	110	180	900	1 190	28	328	50
84	6	5	12	5.5	110	200	1 400	1 710	51	865	50
88	12	6	12	9	130	130	2 200	2 460	36	880	75
92	0	6	2	3	55	—	500	555	65	362	20
96	0	1	16	0.5	80	—	400	480	1	4	5

a, b and c = different titrations from the same humus samples from 100 mg weighing, given in ml (of them c was made with old dithizone solution and is shown here only to indicate that such a solution should not be used in analysis).
d = average of a and b, (a + b)/2
e, f and g = spectrographic determinations of ashed humus ppm Cu (e), Pb (f) and Zn (g)
h = e + f + g
i = ash content % of humus, after burning at 550°C
j = total heavy metal content of dry humus (i × h/100) ppm
k = approximate exchangeable heavy metal content (ppm) of humus in Zn equivalents, derived with values of d from the diagram Fig. 8.

From Table 2 it may be seen that only 5—20 percent of the total heavy metal content of humus is leachable by a cold citrate buffer (pH 8.5). It is

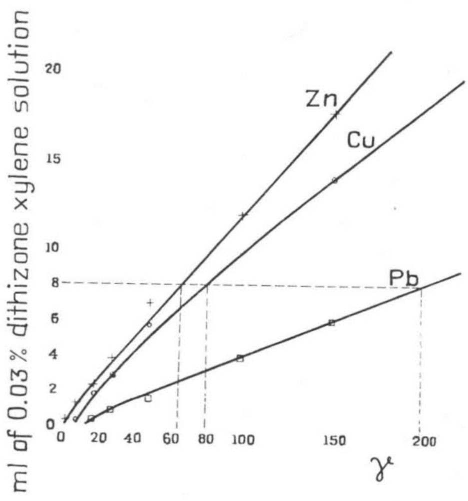

Fig. 8. The titration of standard solutions of Cu, Pb and Zn with 0.03 percent dithizone-xylene solution.

also seen that in all cases in this field (Kolima) zinc is much more abundant in humus than are the other two metals and therefore it must have a much greater influence on the results of humus analysis.

EFFECT OF DIFFERENT METALS ON THE TOTAL TITRATION

In the analysis of natural samples there are many interfering metals that have their influence on titration but it is also seen that all the three heavy metals (Cu, Pb and Zn) under discussion here have their own power of giving colour to the dithizone solution. The intensity of the colour of metal dithizonates was studied by standard solutions in the laboratory and the results are presented in the following diagram, Figure 8.

As is seen, much larger concentrations of lead are needed to give the same colour to a given amount of dithizone solution than, *e. g.*, zinc. For example, to give a bluish-gray colour to 8 ml of 0.03 percent dithizone xylene solution, 60 γ Zn, 80 γ Cu or 200 γ Pb are needed.

TYPE OF SOIL

The heavy metal content of humus is derived mainly from vegetation, which in turn extracts heavy metals from ground water solution via roots.

Table 3. Relation between mineral soil at a depth of 30 cm and exchangeable heavy metal contents of overlying humus, Kolima Viitasaari.

Soil type	Average heavy metal content of humus, ml	Number of samples
till	1.5	196
clay	1.3	143
silt	2.2	41
sand	2.1	70

It is therefore evident that the permeability and mineral composition of the sediment under the humus layer must have a considerable effect on the metal contents of humus. The effect of the mineral composition is seen, *e. g.*, in the Kolima area, where the heavy metal concentrations of the humus clearly reveal that of the underlying till (Figs. 3, 4 and 5). Flowing water may be responsible for the concentration of heavy metals in the humus on hillsides at the Kotanen field.

In the central area of Kolima field a study was made to determine the relations between heavy metal concentrations in humus and the type of mineral soil underneath. The relations are shown in Table 3.

The smallest concentrations in humus seem to overlie clay, which is a rather compact and impermeable sediment. Above more coarse sorted sediments the heavy metal concentrations are higher.

TYPE OF VEGETATION

When analysing humus in the sampling spot, Lyytikäinen observed that the heavy metal content under birches or stubs of them was somewhat higher than in the near surroundings. A statistical study of the influence of vegetation on the heavy metal concentrations in humus was therefore planned and made in the Kotanen area. The terrain was divided with index plants into the following types:

— cultivated field,

— bog *(Betula pubescens, Picea abies, Betula nana, Ledum palustre, etc.)*,

— rocky ground (lichens and moss + the trees of surrounding areas),

— thick damp forest *(Picea abies, Betula verrugosa, Populus tremula, Vaccinium myrtillus, Oxalis acetosella, Convallaria majalis, etc.)*,

— dry sparse forest *(Pinus silvestris, Juniperus communis, Calluna vulgaris, Vaccinium vitis idaeus,* lichens, *etc.)*.

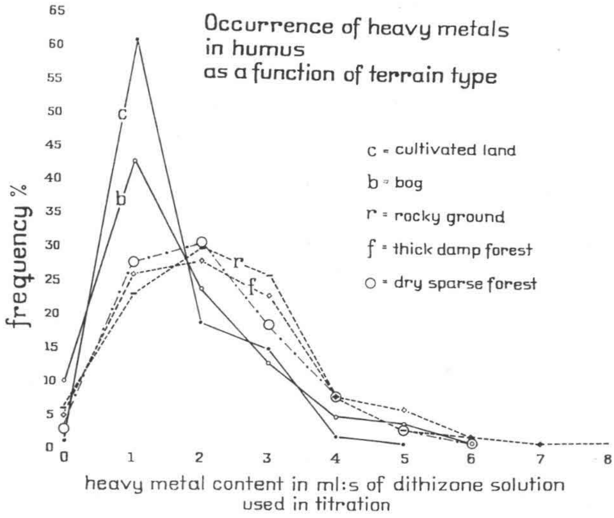

Fig. 9. The distribution of heavy metal concentrations in humus of different terrain types.

The boundaries of these types in nature are gradual, of course, and the determination of types somewhat subjective. The type of vegetation itself naturally depends on topography and the type of sediment beneath it (water support). The results of the present investigation, based on 3 100 samples, are presented in the following table and in Figure 9.

As may be seen, the average concentrations of heavy metals (Table 4) and the distribution of concentrations (Fig. 9) in the humus of bog and cultivated field differ markedly from those of the three other types. Humus of bogs and cultivated fields cannot, therefore, be compared with that of the others. Average concentrations in the humus of rocky ground and thick damp forest are rather similar and slightly higher than in the case of dry sparse forest, but do not differ to the extent that simultaneous use of all these 3 types in prospecting would be invalid.

Table 4. The relationship of vegetation to heavy metal concentration in humus, Kotanen, Viitasaari.

Terrain type	Average heavy metal content of humus, ml.
cultivated field	1.6
bog	1.7
rocky ground	2.4
thick damp forest	2.4
dry sparse forest	2.1

CONTAMINATION

In several connections in the foregoing, mention has been made of the fact that contamination along roads is strong. Roads have therefore been avoided as much as possible and samples closer than 20 meters from a road are not taken. Marked contamination has also been noted under electric wires and wire fenches as well as in inhabited areas, especially along dikes or in gardens and old house sites. When sampling humus, a manual is always held and observations of vegetation, topography and possible sources of contamination are made.

SUMMARY

Humus analysis is a cheap and rapid method for provisory geochemical prospecting of large areas. It gives the sum of exchangeable copper, lead and zinc of humus in milliliters of dithizone solution consumed in titration. The method of analysis is rough but its reproducibility seems to be fairly good. The heavy metal concentrations in humus depend on underlying soil, overlying vegetation, the topography of the region, the transport of glacial ice and on the bedrock of the area. The results of the analysis still depend on the relations of the metals in humus and the kind on leaching solution and the bond between metals and humus acids.

The form of anomalies in humus in investigated fields seems to be caused mostly by transport of glacial ice. The diffusion vertically in till is small (Kauranne, 1959), therefore the heavy metal concentrations seldom are high directly above the ore exposure. It might be possible to arouse an anomaly directly above the exposure if the mineral soil were sorted glacial sediments like silt and sand (Bishoff, 1954), but upon till the anomalies usually seem to be found farther away. The metals in humus are thus derived from the mineral content of transported till 1) where ore minerals are finely ground and the total surface area exposed to leaching is larger than that of the ore outcrop, 2) where glacial ice has pushed the metal-bearing material close to the surface (Hyvärinen, 1958) so that slow diffusion in tight glacial sediments is capable of transferring the heavy metals into the humus and is more easily attained by roots and 3) where near the surface, conditions are suitable for leaching the sulphides because deeper down the environment often has reducing effect. The background level, *e. g.*, for zinc is higher in areas with preglacial weathered material in the till, which is reflected in the humus by anomalies without any apparent source.

When prospecting by means of humus analysis all the facts mentioned in the foregoing must be observed, the background level determined for every case separately, and special care taken to avoid contamination.

Care must be taken in interpreting the results obtained also — results from different areas may not be directly comparable.

In spite of all its limitations and possible errors the analysis of humus can be successfully used in prospecting. The Geological Survey of Finland has made over 50 000 humus analyses up to the present (1959) and will continue the work.

Acknowledgements — We are grateful to the geologinen tutkimuslaitos (The Geological Survey of Finland) for extending permission to publish the results of the present study. We wish to express our special appreciation to the director, Vladi Marmo, Ph. D. for reading the manuscript and giving much valuable advice and also to our colleagues: P. Ervamaa, Lic. Ph., V. Pääkkönen, Lic. Ph., A. J. Laitakari, M. S. and P. Oivanen, M. S., who have assisted us by all possible means in their fields of investigation. Thanks are due also to our assistants: Mrs. Raili Niittykoski, Misses Lea Harlamo, Liisa Pajari and Marja Uotila and Messrs A. Heinonen, O. Niittykoski, K. Paavilainen, L. Ristimaa and J. Willman. The pectrographic determinations were made by A. Löfgren, M. S., the figures drawn by Miss Karen Dahl and the English text corrected by Paul Sjöblom M. A.

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28

MAGNETISCHE CASSITERITE [1]

VON

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EINLEITUNG

Lagerstätten mässiger oder relativ niedriger Bildungstemperatur führen keinen magnetischen Cassiterit. Sogar die manchmal tiefschwarzen Kristalle aus Pegmatiten, teils homogen und dann im Dünnschliff enorm pleochroitisch, teils — und zwar oft im selben Kristall, ja in derselben Zone — unter Bildung feinster Nädelchen von Tapiolit ($FeTa_2O_6$) entmischt, sind praktisch unmagnetisch. Demgegenüber fällt auf, dass neuerdings in technisch störendem Umfang Zinnsteine mit starkem Magnetismus bekannt wurden. Das ist deswegen so unangenehm, weil Zinnstein aus den Schwermineralkonzentraten von Seifen oder auch von tiefgründig zersetztem anstehendem Gestein durch das Fehlen von Magnetismus gegenüber Magnetit, Wolframit, Columbit, Ilmenit, u. s. w. getrennt wird, und dieses Verfahren dann eben nur teilweise — und zwar in unübersehbarer Weise — wirksam ist.

Da im Dünnschliff diese Cassiterite oft opake Einschlüsse führen, wurde kurzerhand angenommen, dass Verwachsungen mit Magnetit vorlägen, ein Eindruck, der durch oberflächliche Untersuchung im Auflicht bestätigt erschien. Dazu ist allerdings zu bemerken, dass der Versuch, von Konzentraten mit viel oder überwiegend Cassiterit gute Anschliffproben zu erhalten, bisher nur zu einem höchst mangelhaften Resultat führte.

In den letzten Jahren konnte Verfasser nun einige Untersuchungen durchführen, die zeigten, dass wenigstens in den behandelten Fällen die Annahme einer Magnetitverwachsung nicht zutrifft, vielmehr sehr komplizierte, schwer übersehbare Beziehungen vorliegen. Selbstverständlich soll damit keineswegs gesagt werden, dass Zinnstein-Magnetitverwachsungen unmöglich sind.

[1] Eingegangen den 14. April, 1961.

Sie sind sogar in kontaktmetasomatischen Bildungen durchaus wahrscheinlich, wenn auch kaum in wirtschaftlich bedeutsamer Menge vorhanden. Unser Problem bilden die armen Zinnvererzungen alkalireicher Granite, die in tropischen Regengebieten z. T. Anlass zur Bildung umfangreicher Zinnseifen gegeben haben, z. T. auch durch tiefgründige Verwitterung so aufgelockert sind, dass sie als Gestein ein armes Zinnerz bilden und im Grossbetrieb durchgewaschen werden können. Das trifft z. B. für die durch Williams (1955, 1956) beschriebenen Granite von Harwell und Rayfield-Gona bei Jos in Nigeria zu und ebenso für Kedah in Malaya, die von Knorring (1960) neuerdings kurz beschrieb. Sicher sind primär ähnliche Vorkommen auch anderswo entwickelt, aber bisher übersehen, da die zur Gewinnung als Gestein wie als Seifen notwendige extreme Zersetzung nicht vorliegt. Die abgebauten Vorkommen liefern neben Zinnstein Columbit (in Nigeria in wirtschaftlich sehr erheblichem Umfang), Monazit, Uranothorit, Zirkon, Orangit u. s. w. Die Aufbereitungsprobleme wurden von Williams *(op. cit.)* meisterhaft dargestellt und meist auch elegant gelöst.

Der Zinnstein fällt dabei in zwei Fraktionen an, einer unmagnetischen und einer magnetischen, die in allen Stadien in columbitreiche Fraktionen übergehen kann, was unreinen Zinnstein einerseits, besonders aber Columbitverlust andererseits bedingt.

Den Herren Williams und von Knorring verdankt Verf. sein Untersuchungsmaterial, er dankt ihrer Hilfe herzlich, ebenso den Herren Hurlbut und Frondel für pegmatische Zinnsteine aus den Black Hills zum Vergleich. Anderes Material aus Pegmatiten stand aus Südwestafrika, Malaya, Finnland, Bolivia reichlich zur Verfügung. Verfasser ist dem Geophysical Laboratory of the Carnegie Institution of Washington dankbar für die Arbeitsmöglichkeit, Herrn Dr. Freund der Fa. Leitz/Wetzlar und der Tochtergesellschaft E. Leitz Inc. in New York für die Leihgabe eines leistungsfähigen Mikroskops.

MIKROSKOPISCHE UNTERSUCHUNG

A. Die »unmagnetischen Konzentrate» führen »klaren», relativ wenig gefärbten Zinnstein, der wohl kleinen jüngeren Gängchen in den Graniten entstammen mag und in viel grösserer Menge solchen, der intensiv gefärbt und oft stark zonar ist. Daneben finden sich die spärlichen, in jedem Konzentrat vorhandenen mitgerissenen Fremdmineralien: vereinzelt Rutil, Uranothorit, Zirkon, Columbit, Tapiolith, sogar Quarz und Sulfide — insgesamt aber nur vielleicht 2—3 %. Ilmenit, Magnetit, Hämatit fehlen praktisch.

Für uns von Interesse sind nur die »dunklen, zonaren Zinnsteine». Diese enthalten nur gelegentlich und sehr untergeordnet Columbit als Entmischungskörper (im folgenden oft E-Körper), als aus solchen durch Sammelkristallisation »zusammengelaufene» Körner und in winzigen Mengen als primäre Verwachsung. Praktisch alle »dunklen» Zinnsteine führen aber in erheblicher Menge und auffallend zonenweise angereichert Entmischungen (oder ebenfalls »zusammengelaufene» Körner) von T a p i o l i t h. Dieser ist, wenn beide nebeneinander vorliegen, von Columbit leicht durch geringes Reflexionsvermögen und viel häufigere Innenreflexe (braun bis blutrot, aber auch lichtgelb in dünnsten Lamellen) unterscheidbar. Ist nur eines beider Mineralien vorhanden, so ist sorgfältigste Beobachtung nötig.

Man würde vielleicht erwarten, dass, bei der ausserordentlichen Gitterähnlichkeit von Zinnstein und Tapiolith, die Entmischung lappige oder myrmekitische Formen annähme oder auch in einer, vielleicht die Verdreifachung der c-Dimensionen bei Tapiolith irgendwie betonenden besonderen Form aufträte. Tatsächlich trifft das letztere, wie Verf. mit Frau Schachner-Korn (Ramdohr, 1960) und Noll (1949) zeigten, bei pegmatitischen Zinnsteinen in Nadeln || (902) oder (301) auf. H i e r i s t d i e S a c h e a b e r ü b e r r a s c h e n d a n d e r s!

Nicht weniger als 5 verschiedene Verwachsungsgesetze — alle aber anscheinend völlig achsenparallel — mit verschiedenen Trachtvarietäten der E-Körper sind gesichert, noch weitere wahrscheinlich. Es wäre fast hoffnungslos, bei den äusserst feinen Verwachsungen die Verhältnisse entziffern zu wollen, wenn, wie z. B. Tafel I Abb. 1 zeigt, viele dieser Gesetze nebeneinander auftreten. Das ist nun überraschenderweise im Gegensatz zu diesem Bild meist nicht der Fall: Im selben Kristall bevorzugen verschiedene Stellen, besonders Zonen, aber auch innerhalb derselben Zone verschiedene Flecken, jeweils eine oder nur wenige dieser Möglichkeiten, oft in ausgezeichnet scharfen Formen. Die Bestimmung der Möglichkeiten ist bei den verschiedenen Tracht- und Habitusmöglichkeiten von Zinnstein in Schnitten allgemeiner Lage sehr schwierig, wird aber erleichtert, sobald Schnitte || (001) vorliegen. Diese sind geometrisch wie optisch, manchmal auch durch die Formen der Columbitentmischung (s. u.) leicht erkennbar, und es sind auch die Prismen (110) — grösser entwickelt — und (100) leicht zu unterscheiden. So wurden die mit 1)—5) im Folgenden bezeichneten Fällen beobachtet:

1) Am eigenartigsten sind äusserst dünne ($\ll 1\,\mu$), bis 10 μ grosse scharfe quadratische Tafeln || (001), die ihre Quadratseite || (100) des Zinnsteins orientieren (Tafel I, Abb. 2) Dass im Photo z. T. die Täfelchen Interferenzstreifen zeigen, beruht darauf, dass der Schnitt nicht ganz genau || (001) liegt, die Täfelchen also nicht exakt horizontal liegen, sondern ganz flach in den Zinnstein einstechen. Zinnstein zeigt also hier »Farben dünner Plättchen». Die Täfelchen ohne solche liegen, völlig im Zinnstein eingebettet, unter der Schliffoberfläche.

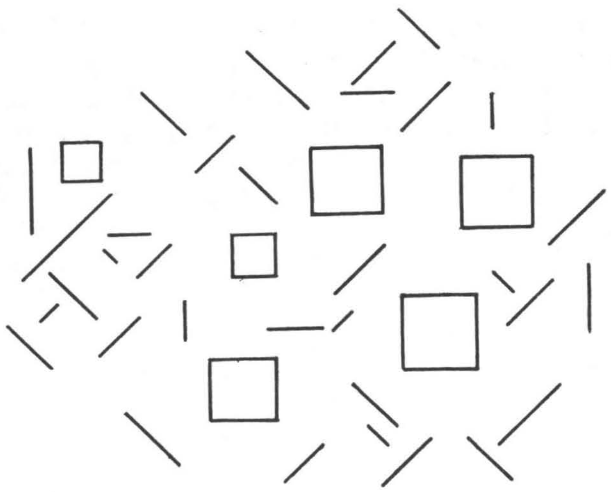


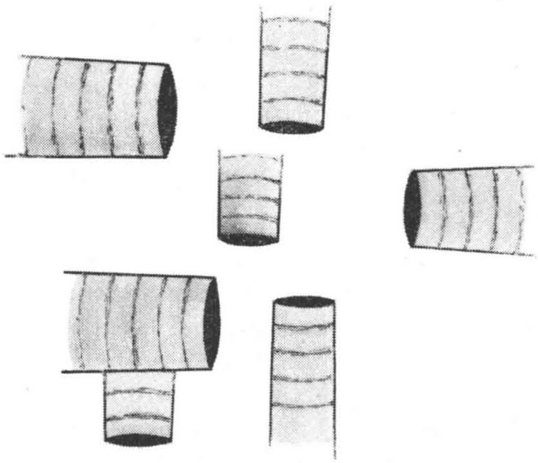
Abb. 1.

2) Mit den Basistäfelchen gern vergesellschaftet sind Scharen feinster, in Basisschnitten völlig scharfer, senkrecht einstechender dünn strichförmiger Lamellen || (110), mit den tetragonalen Formen von 1) also einen Winkel von 45° bildend (Tafel I, Abb. 2). In Schnitten anderer Lage bleiben sie fast immer strichförmig, sind also sehr dünn und erscheinen erst in Schnitten sehr nahe || (110) breit linealförmig bis rechteckig. Ihre Randbegrenzung ist dann nicht so scharf wie die der Basistäfelchen unter 1).

3) Andere Basisschnitte, die Tafeln nach 1) ebenfalls gut zeigen, führen, wieder strichförmig wie 2), aber jetzt || den Quadratseiten, also || [010] (Tafel I, Abb. 3) den Ausbiss von Täfelchen, die in Schnitten || [100] denen unter 2) ähneln. In seltenen Fällen können 1), 2) und 3) Lamellen nebeneinander auftreten, also für einen Basisschnitt schematisch (Abb. 1).

4) Schnitte beliebiger Orientierung zeigen oft strichförmige Lamellen oder, wie man am Verlauf der »Lamellen» im Inneren beim Heben und Senken des Tubus bei Ölimmersion erkennen kann, lineal- oder nadelförmige Einlagerungen, die schräg zur Auslöschung, d. h. [001] des Zinnsteins liegen. Sie müssen einer relativ steilen Pyramide des Zinnsteins parallel laufen. Wieder gibt der Basisschnitt die Deutung:

Basisschnitte zeigen manchmal äusserst feine spindelige Querschnitte von Einschlüssen (α). Beim Senken des Tubus kann hier leicht verfolgt werden, dass die spindeligen Querschnitte in vier verschiedenen, aber gleichwertigen Richtungen steil einstechenden Linealen entsprechen, die entweder den Kanten von (922) : ($9\overline{2}2$) oder (311) : ($3\overline{1}1$) — beide Möglichkeiten sind im Schrifttum für nadelförmige Einschlüsse schon erwähnt — parallellaufen. Im Bild



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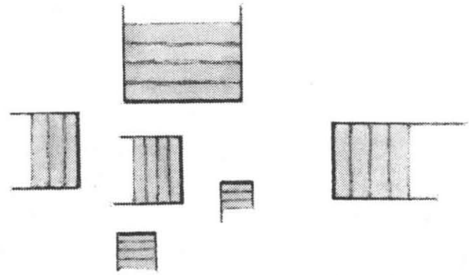

Abb. 3.

sieht das, durch die schnell folgenden Interferenzfarben in der Richtung leicht verfolgbar (etwa wie Abb. 2).

Manchmal sind diese Körper keine Lineale, sondern feinste Nadeln (β) von etwa rundlichem Querschnitt — also etwa wie die schon lange bekannten Beispiele —, manchmal aber auch rechteckige Tafeln (γ), von den unter 3) und 2) genannten leicht durch das schiefe Einstechen unterscheidbar (Abb. 3). Lägen nur Nädelchen vor, so wäre die Beziehung auf eine Kantenrichtung berechtigt, da es sich aber meist um Tafeln oder Lineale handelt, ist es einfacher, die parallele Fläche anzugeben, d. h. (902) oder (301).

5) Als letzte Form treten in Basisschnitten als allerfeinste quadratische Schnitte, in allen anderen Schnitten aber als Nädelchen || zu [001] erkennbar E-Körper auf. Da [001] ja eine singuläre Richtung darstellt, sind diese Nadeln von den unter 1)—4) beschriebenen Fällen leicht unterscheidbar.

Warum alle diese fünf Möglichkeiten, von denen 4) noch dazu in drei Habitusvarietäten zu unterteilen wäre, wahllos nebeneinander, dabei aber jeweils in den verschiedenen Kristallen oder in Zonen oder Flecken desselben Kristalls gehäuft auftreten, ist unverständlich. Denkbar wären winzige, mit der Abkühlungskontraktion zusammenhängende Druck- oder Zugspannungen. Die kristallographische Orientierung ist, soweit das bei den winzigen Körpern nach Auslöschung und Orientierung der Bireflexionen möglich ist, in allen Körpern die gleiche und in O und E mit Zinnstein übereinstimmend. Nur im Falle 4) lässt sich Zwillingsstellung zum Zinnsteinkristall nicht ausschliessen (etwa nach (301) entsprechend den herzförmigen Rutilzwillingen).

Der Entstehungszeit nach ist zwischen 1)—5) kein Unterschied bemerkbar. Über die Erscheinungen der Sammelkristallisation vergl. bei B.

B. Die »magnetischen Konzentrate» ähneln zunächst den unmagnetischen so sehr, dass bei schwacher Vergrösserung kaum ein Unterschied zu sehen ist, allerdings fehlen die »klaren», wenig gefärbten Körner. Ganz vereinzelt sind mitgerissene Columbite, grobmechanische Columbit-Zinnstein- und Zinnstein-Ilmenit-Verwachsungen.

Die dunklen Cassiterite zeigen Entmischung in praktisch jedem Korn, doch kommen im Einzelkorn grosse »klare» Zonen, zum Teil überwiegend, vor. Die Formen der Entmischung sind äusserst verschieden und kompliziert, doch sind auch hier nach Zonen oder fleckenförmig gewisse Bilder jeweils vorherrschend. Wieder sind kleine scharf begrenzte, orientierte Körper neben grösseren unregelmässig lappig rundlichen, durch Sammelkristallisation »zusammengelaufenen» vorhanden. Eine gewisse, aber oft nicht zutreffende Relation zwischen absoluten Anteil und der Form der Körper ist deutlich: Wo sehr viel Entmischungsprodukt vorhanden ist, ist die Neigung zum Zusammenwandern viel grösser, oft sind dann überhaupt keine kristallographisch abgegrenzten Formen mehr vorhanden. Auch nahe den Zwillingsgrenzen im Zinnstein sind die E-Körper unregelmässig lappig und grobkörniger geworden. Beides ist ja mechanisch ohne weiteres verständlich. Während beim unmagnetischen Zinnstein als E-Körper praktisch allein Tapiolith vorlag, tritt hier stets in wechselndem, manchmal überwiegenden Anteil auch Columbit auf. Gerade das stete Nebeneinander beider erleichtert hier die Erkennung sehr.

Die Entmischung des Columbits ist stets älter als die von Tapiolith und auch meist gröber. Es sind kurze, gedrungene Täfelchen, meist mit gerundeter Endbegrenzung, die ersichtlich nur nach zwei Richtungen eingelagert sind. Vergleich mit der Lage der Tapiolithkörper zeigt, dass das nur das Prisma [100] sein kann. Dem entspricht, dass a_0 des Zinnsteins genau mit $1/3$ a_0 des Columbits (Aufstellung nach Dana) übereinstimmt (4.73 bezw. 4.75). Die weitere Orientierung ist, da die Täfelchen in der Zone [100] keine scharfe Begrenzung zeigen, nicht anzugeben; doch ist Übereinstimmung der O-Packung anzunehmen.

Die Columbitkörper werden viel weniger von Sammelkristallisation erfasst als die von Tapiolith; oft sind sie, vielleicht höchstens etwas kantengerundet, in grösseren rundlich-lappigen Tapiolith-Körnern noch in alter Form und Orientierung erhalten (Tafel II, Abb. 1). Immerhin gibt es Fälle genug, wo auch sie zu völlig unregelmässigen Massen wechselnder Grösse »zusammengelaufen» sind. Das sind dann die Körnchen, die bei mangelnder Aufmerksamkeit, schlechten Schliffen und Nichtanwendung von Ölimmersion leicht für Magnetit gehalten werden.

Die Sammelkristallisation des Tapioliths lässt sich hier ausgezeichnet verfolgen, wobei die Erhaltung der Columbittafeln gut mithilft. Jede der im Abschnitt A) beschriebenen Formen wird erfasst, ohne sichere Bevorzugung der einen oder anderen. Hat die Sammelkristallisation einmal eingesetzt, so pflegt sie kräftig durchzugreifen, so dass Übergänge (wie etwa in Tafel I, Abb. 1) nicht allzu häufig sind. Da auch submikroskopische, ja vielleicht noch nicht entmischte Restgehalte mitergriffen werden, sind, wo ursprüngliche und »zusammengelaufene» Entmischungen nebeneinander vorliegen, die rundlichen Formen von den kristallographisch begrenzten durch charakteristische Entfärbungspartien getrennt. Dass Nachbarschaft von Zwillings- und Zonengrenzen im Zinnstein die Körper oft grösser werden lässt, ist schon oben gesagt.

Im Falle der feinkristallinen Entmischung lässt sich wegen der Winzigkeit der oft nur mit stärksten Ölimmersionen erkennbaren Lamellen u. s. w. und wegen des Durchleuchtens von unterhalb der Schnittebene liegenden Körnchen der Mengenanteil der Entmischungsprodukte schwer abschätzen. In den Fällen der grob »zusammengelaufenen» Entmischung ist das kein Problem und kann quantitativ durch Planimetrierung im Photo (z. B. Tafel II, Abb. 2) — durchgeführt werden. Sicher werden 12, vielleicht 15 % des Tapiolithmoleküls erreicht! Das ist erheblich mehr, als die bisherigen Analysen nachweisen, aber diese Differenz ist nicht überraschend, da die Analyse ja das ganze Zinnsteinkorn erfasst, die höchstmöglichen Gehalte aber nur in einzelnen Zonen auftreten. — Nur nebenbei sei erwähnt, dass in manchen der groben Tapiolithkörner eine »Unterentmischung» einer Phase mit einem noch niedrigeren R. V. als Zinnstein zu beobachten ist.

C. Überlegungen über den in A) und B) geschilderten Befund führen zu einigen wichtigen Schlüssen:

1) Es ist bemerkenswert, dass nicht der gesamte Anteil an Nb + Ta zusammen mit dem aequivalenten Fe + Mn in der gitterdynamisch für eine Entmischung sicher bequemeren tetragonalen Form der $(Fe, Mn)(Nb,Ta)_2O_6$-Mischkristalle sich ausscheidet, sondern dass in vielen Fällen zuerst Columbit erscheint. Das scheint die vor kurzem gemachte Feststellung (oder Behauptung), dass diese Mischkristallserie auf der Nb-Seite (»Mossit») überhaupt nicht besteht, sondern irgendwo, vielleicht erst bei Nb : Ta $\backsim$ 1 : 1 beginnt, zu bestätigen. Höhere Nb-Anteile müssen sich dann zuerst in Entmischung eines Nb-reichen Columbits äussern. Eine solche Entmischung lässt dann den relativen Ta-Anteil im Zinnstein schnell anwachsen und die Entmischung des Tapioliths kann beginnen. Warum die einmal begonnene Columbit-Entmischung sich nicht fortsetzt bis zur Bildung von Ta-reicheren Gliedern, ist schwer verständlich, da die o'rhombische Mischkristallreihe aus Hunderten von Analysen wohl belegt ist. Immerhin scheint, das ist aber weitgehend hypothetisch, es so zu sein, dass bei sinkender Tem-

peratur wenigstens für die Ta-reichen Glieder die tetragonale Form stabiler wird; das soweit, dass in gitterenergetisch festgelegter Zinnsteinumgebung die tetragonale Form zum Zuge kommt.

2) Der M a g n e t i s m u s des Zinnsteins ist nach dem hier vorliegenden Befund eindeutig n i c h t mit einem Gehalt von Magnetit verknüpft, sondern mit dem Vorhandensein einer früheren Entmischung von Columbit. Die spätere Entmischung eines Ta-reichen tetragonalen Mischkristalls, hier immer »Tapiolith» genannt, obwohl nach den Angabe von W. Noll manchmal keine typische Trirutil-, sondern statistische Kationenverteilung im Gitter vorliegt, hat k e i n e n Einfluss auf den Magnetismus, wenigstens nicht in den beim Aufbereitungsvorgang verwendeten Feldstärken. Es muss natürlich offenbleiben, ob nicht bei noch höheren, aber bisher nie beobachteten und wegen der relativen geochemischen Häufigkeit auch unwahrscheinlichen Gehalten an $FeTa_2O_6$ in tetragonaler Entmischungsform doch noch Magnetismus aufträte.

Anzuregen wäre das Experiment, durch Temperung die Columbitentmischung zum Verschwinden zu bringen und damit vielleicht Verschwinden des magnetischen Verhaltens zu erreichen.

Die Frage, warum im Konzentrat magnetische, d. h. Columbit-entmischende und unmagnetische Zinnsteine nebeneinander vorkommen, ist schwer zu beantworten, da die Körner im Konzentrat ja aus dem geologischen Verband gelöst sind. Es ist möglich, vielleicht wahrscheinlich, dass erstere etwas älteren, wärmeren Kristallisationsstadien entsprechen.

ZUSAMMENFASSUNG

Es werden Zinnsteine aus Alkaligraniten beschrieben, die sich teils magnetisch, teils unmagnetisch im Aufbereitungsvorgang verhalten. Beide enthalten massenhafte feinste Entmischungskörper, die oft durch Sammelkristallisation »zusammengelaufen» sind. Die unmagnetischen führen praktisch nur Tapiolith, der in Täfelchen || (001), || (110), || (100), in Linealen, Säulen, Tafeln || (902) oder (301), schliesslich in dünnsten Nadeln || [001] eingelagert ist. Die magnetischen führen in denselben Formen ebenfalls Tapiolith, daneben aber wechselnd reichlich gröbere, aber immer noch winzige Tafeln von Columbit, der || (100) dem Zinnstein eingelagert ist mit gemeinsamer a-Achse. Tapiolith unterliegt stärker der Sammelkristallisation als Columbit.

Magnetit enthalten die untersuchten Proben nicht; was dafür gehalten war, ist sammelkristallisierter Columbit.

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ERLÄUTERUNGEN ZU DEN TAFELN

TAFEL I

Abb. 1. Schnitt eines Zinnsteinkristalls mit Entmischungskörpern von Tapiolith nach einer Vielzahl von Gesetzen. Daneben (in der Mitte) Partie, in der die Körper durch Sammelkristallisation zu lappigen Aggregaten »zusammengelaufen» sind. Kedah, Malaya. Vergr. 500 mal, Immersion.

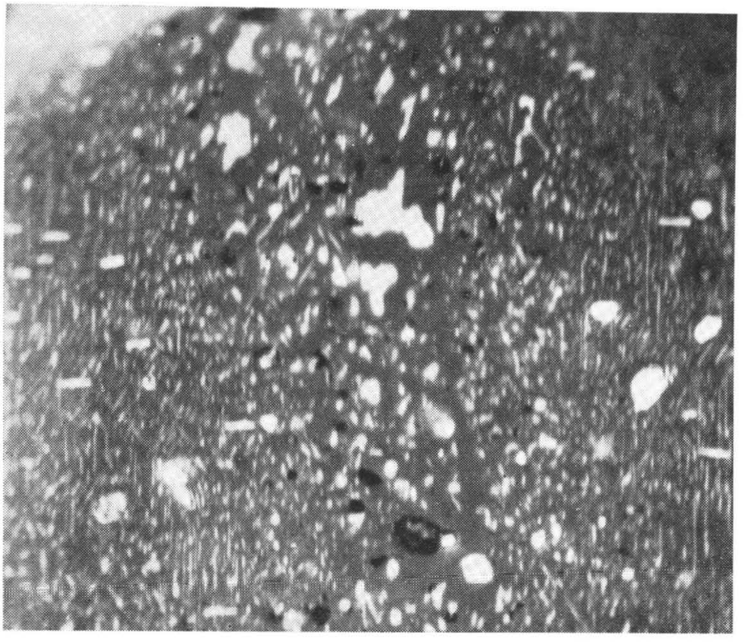
Abb. 2. Schnitt ∥ (001) in einem unmagnetischen Cassiteritkristall Entmischungskörper von Tapiolith in 1.) feinsten basalen quadratischen Täfelchen ∥ (001), 2.) in sehr dünnen Lamellen ∥ (110), 3.) sehr vereinzelt ∥ (100). Dazu feinster Staub unbekannter Orientierung. Kedah, Malaya. Vergr. 980 mal, Immersion.

Abb. 3. Täfelchen von Tapiolith ∥ (100) neben viel zahlreicheren solchen ∥ (110). Die Täfelchen mit den Interferenzstreifen liegen wieder ∥ (001). Schnitt hier nicht so gut ∥ (001) wie bei Tafel I Abb. 2. Vergr. 980 mal, Immersion.

TAFEL II

Abb. 1. Schnitt eines magnetischen Zinnsteinkristalls. Schnitt liegt roh ∥ (001), etwas um [010] gedreht. Entmischungskörper von Columbit nach (100), weiss, idiomorph, grossenteils eingebettet in »zusammengelaufene» Tapiolith-Körper. Die runden Gebilde sind Flüssigkeitseinschlüsse, die in Schnüren ∥ [001] im Zinnstein häufig sind. Kedah, Malaya. Vergr. 980 mal, Immersion.

Abb. 2. Zinnstein mit Entmischung von Columbit in Zonen. Man sieht, dass die ursprünglich praktisch submikroskopischen Entmischungskörper in einzelnen Zonen völlig, in anderen teilweise »zusammenlaufen» sind. Vergr. 250 mal, Immersion.



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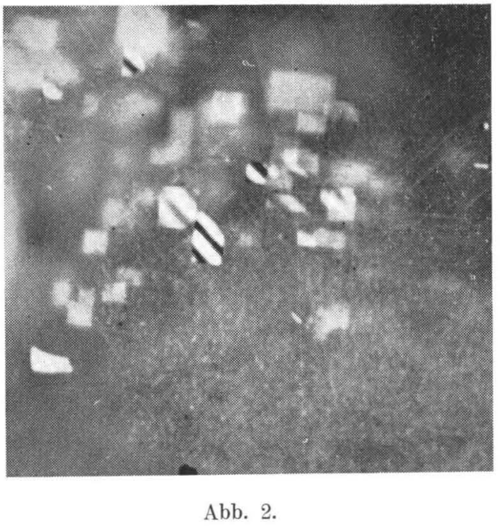

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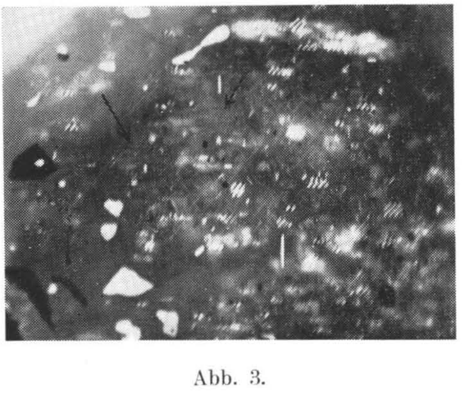

Abb. 3.

Paul Ramdohr: Magnetische cassiterite.

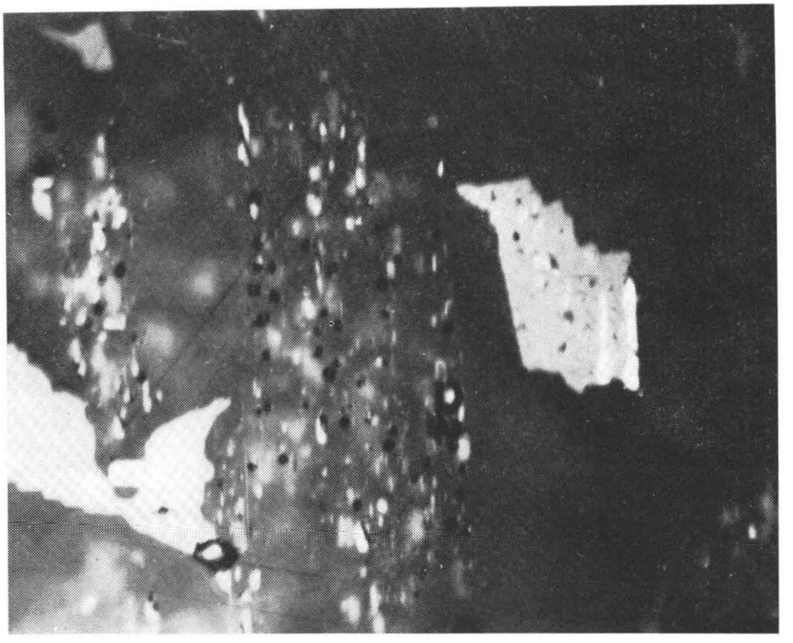

Abb. 1.

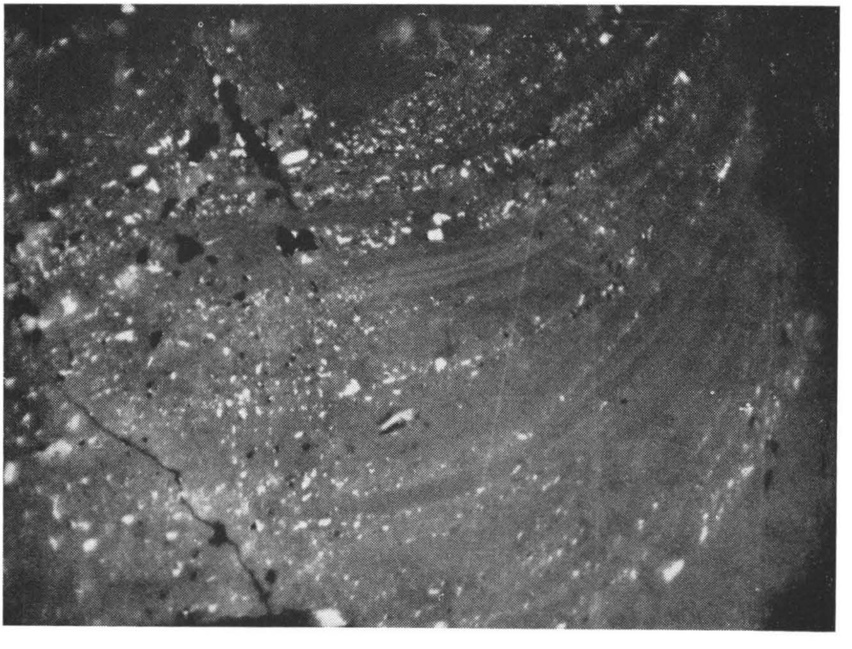

Abb. 2.

Paul Ramdohr: Magnetische cassiterite.

GRANITIZATION OF QUARTZOSE ROCKS [1]

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CONTENTS

QUARTZITE MIGMATITE

Hoping it will be permitted that one of the contributions to the Geological Society's 75'th anniversary volume is retrospective, I will give some glimpses from various times during the course of my activity as geologist. To begin with I quote a paragraph of my diary from my first summer in field work, as assistant of the Geological Survey in Pudasjärvi in Kainuu, Northern Finland. The work started on 27'th June, 1904, and the third outcrop observed was what now is called quartzite migmatite. The note (in translation) reads:

»Hirvaskoski 3. Korkiavaara. The steep top of the *vaara* (= highland) consists for the most part of quartzite. The same rock also occurs as bands and fragments in granite in the lower northern part of the *vaara*. The quartzite shows a very pronounced planar schistosity, is rich in mica and light grey in colour. It is in many places traversed by younger granite. Near the contact against granite the quartzite is more glassy and less schistose. The fragments are elongated and in places the granite was intruded as narrow veins, so a veined gneiss resulted. The fragments in this mixed rock are rich in mica, the quartz having been almost entirely refused by the granite magma.

[1] Received May 9, 1961.

The surrounding granite also contains small fragments of very glassy quartzite. Most of the fragments have been displaced from their original positions along the strike.»

In the sequel numerous occurrences of the same kind were described. Often the quartzite fragments in the mixed rock are said to be contorted. Veined gneisses occurring in wide areas west of the quartzite zone were suspected to have been derived from quartzite, although no remains of this rock were observed. The following note, of 23. 8. 1904 from northern Pudasjärvi, is one of many examples:

»Eastern slope of Nuukalanvaara. Mixed rock like veined gneiss in wich the schistose (older) portion is rich in mica. Probably, however, the older rock here was quartzite and not mica schist, as the fragments contain the more quartz the thicker they are.»

The notes quoted above were written three years before Sederholm (1907) published his first classic memoir on granite and gneiss in which he defined the conceptions granitization, migmatite, palingenesis, and anatexis. In this work, and long times thereafter, he interpreted the veined migmatites or veined gneisses as formed by the injection of granite magma in mica schist or any other schistose rock and called the product arterite. As is easy to see this was also how I understood the formation of the quartzite migmatites in Kainuu, more especially in the Pudasjärvi—Taivalkoski region, where this kind of migmatization is dominant.

Another well-known kind of migmatite is venite, formed by the segregation of granitic veins from the material of primary arkose, commonly promoted by stress and shearing. Anna Hietanen (1938) described from Olostunturi in Muonio, Western Lapland, arkosic quartzite in which sillimanite-bearing, *i. e.*, somewhat argillitic portions alternate with merely microcline bearing quartzite, while certain bands of quartzite contain plagioclase An_{20-23}, diopside, and sphene, suggesting admixture of marl in the original sediment. The metamorphic changes in these quartzites have hardly reached the stage of granitization, although the texture is not far from granitic, the psammitic fabrics having turned granoblastic, and quartz, that in blastopsammitic quartzites forms the biggest rounded grains, here fills the interspaces, while feldspar shows a tendency of developing porphyroblasts. Phenomena of metasomatism, important in all granitization, are also traceable in the granitization of quartzites.

QUARTZ-SILLIMANITE NODULES AND GRANITIZATION

In the summer of 1906, while engaged with field work in Enontekiö, I found in a gorge called Paatsikkakuru in the valley of Palojoki River a granitic-looking rock containing ellipsoidic nodules of quartz with needles of sillimanite. Fig. 1 shows two samples of nodules from the same area. For the photographs I am indebted to A. Matisto, who recently published the

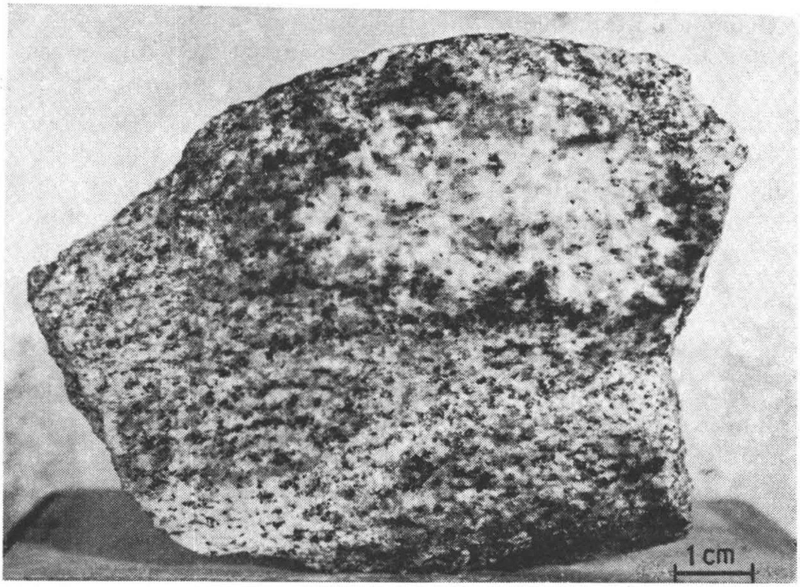

Fig. 1a. Sillimanite-quartz nodule in sillimanite gneiss. Appr. 6 km northeast of Palojärvi, Enontekiö. Photo A. Matisto.

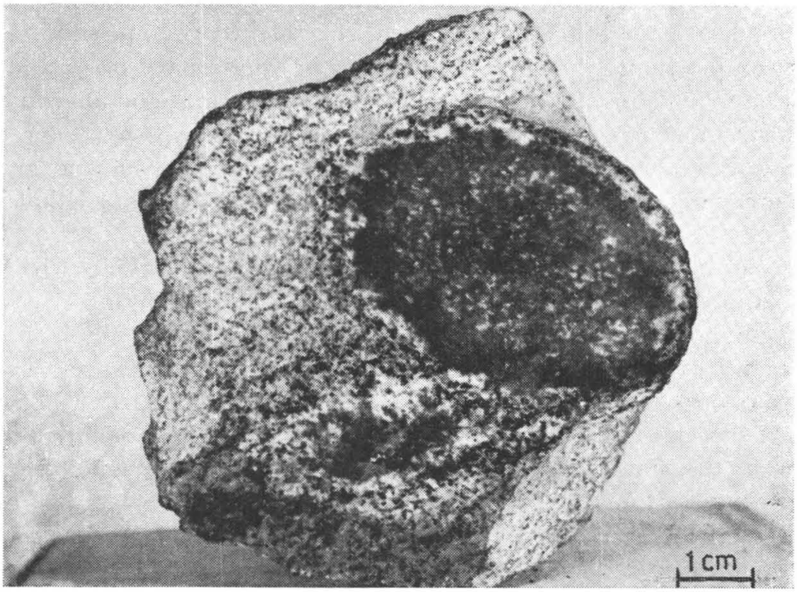

Fig. 1b. Quartz-sillimanite nodule, dark from feldspar pigment. Southern end of Luovavaara, Enontekiö. Photo A. Matisto.

General Geological Map, quadrangle B 8, Enontekiö (Matisto, 1959). Sederholm (1928) mentioned this occurrence in connection with »quite similar» nodular granites. He gave many examples from various countries and discussed a newly found occurrence on the solitary island Storklyndan about 25 km southwest of Uusikaupunki. I have visited this islet together with Sederholm, according to whom (1928, p. 80) the main rock of the island is granite of the Hanko—Uusikaupunki type enclosing fragments, up to several hundred meters in length, of migmatic rocks, composed of a schistose matrix and numerous veins of granite. Both the schist and the granitic veins contain quartz-sillimanite nodules, some of which occur on the border-line of both rocks. Sederholm's interpretation was, that »the nodules both in the schist and the granitic veins are, no doubt, due to the contact action of the surrounding granite».

I cannot see, how contact action of granite could produce such nodules and rather assume that the granite was formed by metasomatic granitization of the aluminous schist in which the nodules existed already before the granitization, or were formed at the early phases of the process, because they are common in certain highly metamorphic schists which are not at all granitized. I remember them, *e. g.*, from the island of Kimito, on the roadsides between the villages Vestlaks and Östermark. Sederholm compared the nodules with calcium-rich concretions in argillaceous schists and said that these »seem to be genetically related to the nodules in granites», adding: »There is little doubt that they are due to the secondary action of granitic magma». These concretions, however, have evidently originated in the same way as compositionally similar concretions in Quaternary clays and would seem to have little in common with the quartz-sillimanite nodules in quartz-rich schists in which they, in one way or other, perhaps by concretionary clustering, have originated under high-grade metamorphism and granitization (the concretion principle of metamorphic differentiation; Eskola, 1932, 1933).

Granites in the strict sense possess compositions varying within very narrow limits only. As has been established by constructing AbOrQ diagrams from great numbers of granite analyses, a dominating majority of the granites in their composition actually agree very closely with the lowest melting mixture in the four-component system $AbOrQH_2O$, or what may be called the eutectoid granite composition. The amount of anorthite and mafic minerals in the common granites is small enough to be negligible for purposes of classification. Only the potassium feldspar in a widely spread and important group of granites, the microcline granites or kaligranites, is present in a considerable excess over the eutectoid proportion. Here the eutectoid granites will be considered in the first place, because they are most common among the granitized quartzite.

The formation of arteritic and venitic migmatites is a preliminary stage of granitization, and the products, veined gneisses, are not homogeneous granites nor is their bulk composition, as a rule, the truly granitic composition. This is attained at the granitization of earlier rocks in a solid state by metasomatism, in most cases probably directly in a single act, the ichors, or the diffuse stream of ascending light ions of the granite elements soaking through elder rocks.

GRANITIZATION OF HURONIAN QUARTZITES IN CANADA

In 1922, while engaged in field work for the Geological Survey of Canada on the North Shore of Lake Huron, I had the opportunity of getting acquainted with an instructive instance of granitization of quartzite by metasomatism in the region where in the middle of the previous century W. E. Logan and A. Murray had accomplished a pioneer work in Precambrian stratigraphy and established the Huronian sequence of rocks. Assisted by R. C. Emmons, now Professor of Wisconsin University, I completed the mapping previously pursued in several summers by W. H. Collins and T. T. Quirke, of Lake Panache and Blind River areas, the last-named now famous for the large uranium deposits discovered in the 1950'es. The Huronian bedrock consists largely of moderately metamorphosed sediments regularly folded in open anticlines and synclines. Characteristic horizons can be followed throughout the areas and their stratigraphic succession has been determined with certainty. The lowermost member, Mississagi quartzite, from 300 m to 3 000 m in thickness, with basal conglomerates, rests upon the pre-Huronian basement analogous to the pre-Karelian basement gneiss in Finland, and including gneissose granites which traverse different, also banded sediments, volcanites and an »iron formation», or banded magnetite-quartzite. Upon the Mississagi quartzite follows a layer of polymict Bruce conglomerate enclosing the uranium ores, then, after limestones and graywackes of Espaniola formation, another thick bed of quartzite, the Serpent quartzite, overlain by the remarkable glacial Gowganda formation with thick beds of tillites and interbeds of glaciofluvial varved pelites and graywackes, in all about 1 000 meters thick. The wide spread of the morainic conglomerate in Canada points to the ancient existence of a truly continental ice sheet. In the uppermost part of the Huronian series there is again a quartzite horizon, the Lorrain quartzite, up to 2 000 meters thick, in its lower part arkosic, but upwards grading into very pure and shining white quartz rock. Due to its hardness this rock forms the highest ridges (Fig. 2), as do the quartzites at Koli and Tiirismaa in Finland.

All horizons of the sedimentary series have been invaded by intrusive rocks formed after the diagenesis of the sediments, probably during or after

Fig. 2. Ridge of white Lorrain quartzite, southern shore of Big Mountain lake south of Lake Panache, Township 82, Ontario. Photo P. Eskola 1922.

their folding. Mafic sills of quartz-bearing diabases, probably synchronous with the so-called Keweenawan diabases of Lake Superior region were intruded first, and later microcline granites occurring on the coast of Lake Huron near Cutler in the Blind River area and on the Killarney Peninsula and extending long way inland in Panache lake area. The granite traverses the different Huronian sediments and, as a rule, has caused in them notable but not very intense contact metamorphism, thus behaving like an ordinary magmatic rock. Only on the Killarney Peninsula and neighbouring islands the contact aureole is broader and materially different, *viz.*, metasomatically granitized.

On the shore of Threemile lake, about 10 km south of the eastern end of Panache lake, the Gowganda formation is near the Killarney granite boundary, and the glacial conglomerate is here a migmatite with a mica-schistose paleosome and somewhat diffuse lighter metatect. In this rock the glacial boulders are on the verge of disappearance while, in the neighbouring unaffected outcrops, the rock may be very much like Quaternary morainic drift in appearance. But its transition into migmatite can be followed step by step, and it is almost dramatic to see the rapid change of moraines to migmatite. Varved slate, the glaciofluvial associate of the till, is here like venite migmatite (Fig. 3).

The granite boundary on Threemile lake strikes southwest and the rock adjoining the granite from here southwestwards is Lorrain quartzite extending to Lake Huron at Killarney, some 25 km farther southwest. It is here the best example of metasomatic granitization is to be seen. Quirke and Collins (1930) described and discussed the phenomena observed on Killarney Peninsula and George Island (Fig. 4). Northwest of Killarney is the Lamorandière Bay that separates the granite area from the Lorrain northwest of the bay. Quartzite (Table 1, analysis 1) is here present only as fragmen-

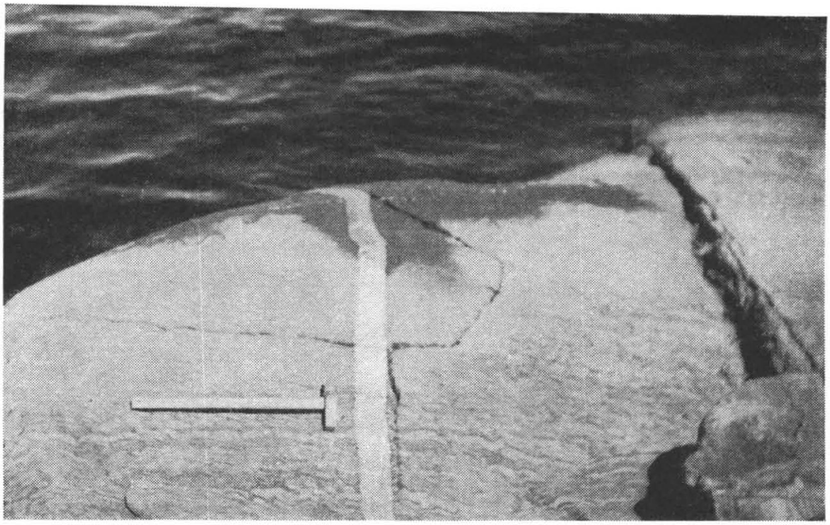

Fig. 3. Varved glaciofluvial Gowganda sediment interbedded with tillite, showing incipient granitization. Threemile lake southeast of Lake Panache, Township Goschen, Ontario. Photo P. Eskola 1922.

tary portions enclosed in porphyry which is an intermediate product of granitization. Unexpectedly the final product, the typical Killarney granite, here joins directly the unchanged Lorrain quartzite. The explanation to this circumstance seems to be that just here is the front of granitization on its way from the east, from the French River area (Quirke, 1929) where the bedrock is made up of gneisses representing intermediate products of granitization, in some places almost like the granite but in other localities showing recognizable relict properties of the Huronian sediments, which westwards are continuous and well preserved up to Sault S:te Marie at the eastern end of Lake Superior. It is at Killarney the »disappearance of the Huronian» takes place to which the title of the memoir of Quirke and Collins alludes.

Around the inclusions of quartzite the transition to the porphyry (Table 1, analysis 2) is described in detail (Quirke and Collins, 1930, pages 45—51). Referring to the original work and the drawings of thin sections in it, it may suffice here to mention the main stages only. The transition takes place within a distance varying from one to 15 meters. First phenocrysts of quartz begin to grow in the quartzite, then phenocrysts and groundmass grains of feldspar, largely potassium feldspars. Evidently the feldspar was formed in part at the expence of sericite, present in the quartzite, but much more of it has been added from outside sources, as appears from analyses 1—3, Table 1. In its alkali proportions and bulk composition the porphyry is

Table 1. Analyses.

	1	2	3	4	5
SiO_2	82.15	76.71	72.14	74.46	74.98
Al_2O_3	12.46	12.87	13.32	13.36	13.64
Fe_2O_3	.70	1.12	1.91	.83	.00
FeO	.16	.31	2.31	.99	2.46
MnO	—	—	.11	.02	.03
MgO	.54	.38	.24	.34	.81
CaO	.38	.40	1.11	1.63	.62
Na_2O	.22	2.05	3.41	4.42	2.80
K_2O	1.39	6.04	4.41	3.82	3.29
TiO_2	.11	.15	.35	.19	.31
P_2O_5	.11	.05	.06	.01	.01
H_2O+	2.23	.25	.32	.50	.91
H_2O-	—	—	.04	.02	.09
	100.45	100.33	99.73	100.59	99.95

1. Schistose, sericitic Lorrain quartzite. Inclusion in porphyry, eastern shore of Killarney Peninsula. Quirke and Collins, 1930, p. 51. Analyst R. J. C. Fabry.
2. Dark red porphyry. West of East Lighthouse, Killarney Peninsula. Quirke and Collins, 1930, p. 51. Analyst R. J. C. Fabry.
3. Killarney granite. Island in Killarney Bay about 1 km north of the northern end of George Island. T. T. Quirke and W. H. Collins, 1930, p. 51. Analyst F. F. Grout.
4. Granite. Paasinmäki, Simsiö Lapua. Hietanen, 1938, p. 15. Analyst Elsa Ståhlberg.
5. Granite vein in quartz vein. East of Vehkalahti, Orivesi. Not published before. Analyst Elsa Ståhlberg.

characteristically like the kaligranites common in the Precambrian. From the quartzite it differs very little, even in its aluminum content, except that much potassium and the major part of its sodium content has been added. At the next step biotite appears and grows more abundant towards the gneiss. The grain size in the matrix increases and a little magnetite is introduced. The foliation, inherited from the quartzite, becomes more accentuated as the rock grades into gneiss. Quirke presented several analyses of gneiss from Killarney and French River areas. Some of them show the kaligranitic composition, others a eutectoid composition like the Killarney granite (Table 1, analysis 3) to which the gneiss grades without a sharp boundary by increase of the grain size and diminuition of the matrix to a mere vestige, while the structure becomes wholly nonfoliated. According to Quirke and Collins the porphyry and gneiss have remained solid rocks during their transformation from quartzite. They continue (1930, p. 53):

»Some of the granite, on the other hand, must have been liquid, for along the southeast shore of Killarney bay it contains angular blocks from the neighbouring Huronian sedimentary formations. — But vestiges of an original quartzite texture are so persistent, even to the point where the gneiss becomes a coarse and practically massive rock, and there is so little difference in composition between the gneiss, which is apparently a metamorphosed quartzite, and the granite, that the writers are strongly inclined to regard the granite as an extreme metamorphic product of the quartzite».

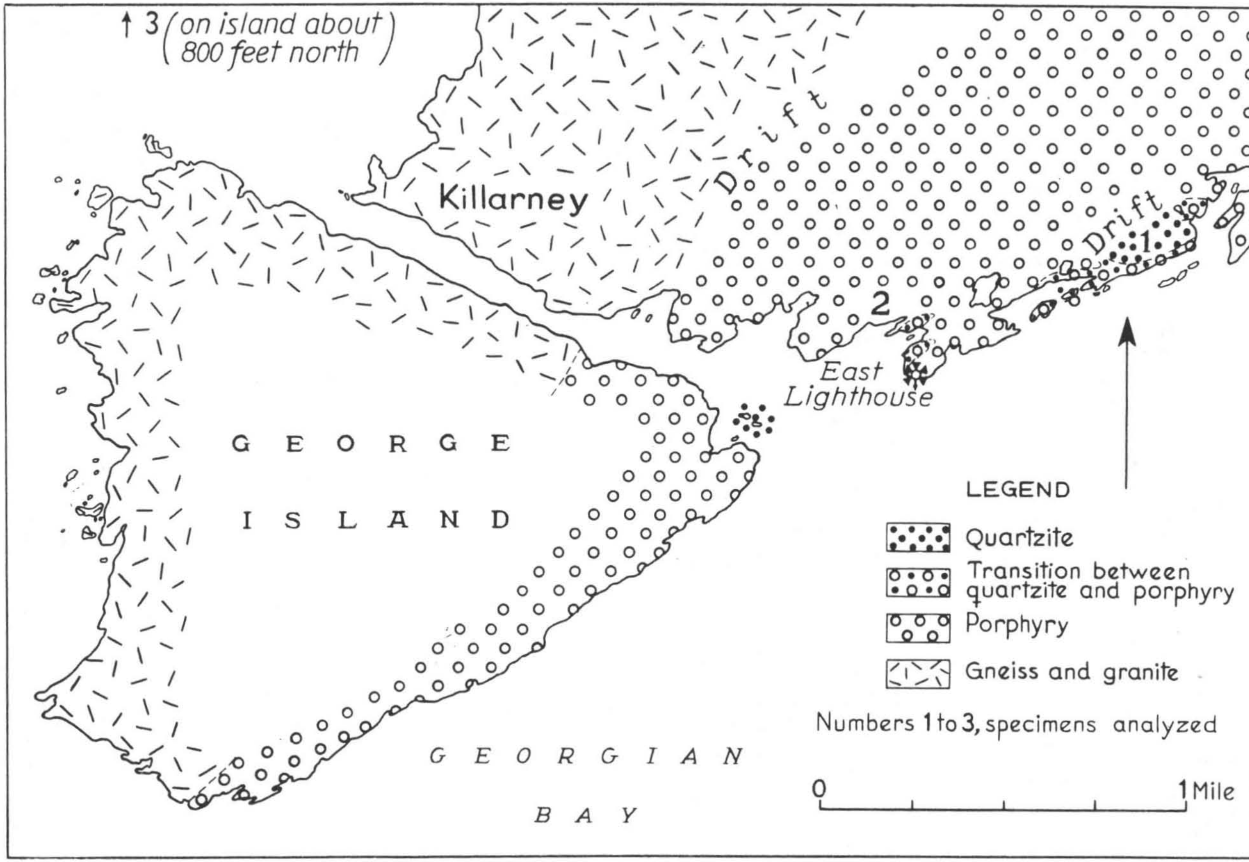

Fig. 4. Killarney and vicinity showing residual patches of Lorrain quartzite in relation to porphyry, Killarney gneiss and granite, and intermediate rocks. From Terence Quirke and W. H. Collins (1930).

The transformation of quartzite to granite implied introduction of alkalis, iron, calcium, and magnesium. The alkalis came first during the rapid stage in which the quartzite graded to porphyry, while iron and magnesium were not added until the transition from porphyry to gneiss in which they formed biotite and magnetite. The transition of gneiss into granite seems to have involved little if any compositional change, while the conclusion that a considerable part of material at that stage had been in a liquid state appears to me well founded.

As stated above, the granites farther west, at Cutler and elsewhere, also appear to have intruded as liquid magma. Quirke and Collins ascribe the absence of a broad aureole of intense contact metamorphism there to the existence of a great fault west of Killarney. The eastern block including the French River area has risen, and what there is now at the Earth's surface, was at the time of the granitization some six miles deep, while the westerly ranges never were buried to any great depth.

Considering the state of the granitizing substances it would seem probable to me that granite magma was present all the time. This would not preclude the introduction of the ions in a definite sequence. Of course an ichor, or a diffuse stream of migrating ions, would also contain all granite elements.

GRANITIZATION OF QUARTZITES IN FINLAND

The North Shore area near Lake Huron offers many features interesting for comparison with the Precambrian geology of Finland. In the relationships between the Huronian sedimentogenic series and the pre-Huronian basement there are close analogies to those between the Karelian series and the basement gneiss complex of eastern and northern Finland, to an investigation of which the efforts of the Finnish geologists henceforth must be seriously intensified. In Canada the old basement probably is more promising than in Finland, but here also much can be learned by systematic surveying which now can be aided by radioactive datings. The Huronian and Karelian formations are rather different in many respects but also exhibit certain analogies; *e. g.*, the lowermost sediments, in Huronian the Mississagi and in Karelian the Jatulian quartzite, are orthoquartzites deposited on deeply denuded and peneplained platforms.

Some years ago it seemed probable that the Huronian were younger than the Karelian. According to radioactive datings by Wetherill, Davis and Tilton (1960) the Cutler granite, which is intrusive in the assumedly pre-Huronian Sudbury formation of graywackes and quartzites, »is older than 1 350 m. y. and probably older than 1 750 m. y.» After my field work in 1922 I felt sure that the Cutler granite traverses the Lower Huronian Mississagi quartzite and even suspected that the Sudbury sediments of Cutler area might not lie below the great sub-Huronian unconformity (see Collins, 1925, p. 26). Whatever the truth in this question may be, the Cutler granite, anyway, in its petrography and tectonic position so closely resembles the late-Huronian Killarney granite, especially its intrusive satellites near Panache lake, that its belonging to the late-Huronian can hardly be questioned. Pre-Huronian granites do occur, *e. g.*, on the Serpent River not far north of Cutler, where I saw much of them and know that they are granite gneisses like the basement gneiss of Finland. They have indeed as little in common with Cutler granite or Killarney granite as the late-Karelidic granites have with the basement gneiss. It is desirable that the geochronology of the Killarney granite be established. At present, already, it appears probable that the Huronian and Svecofennian-Karelian granites as well as the Huronidic and Karelidic orogenies are synchronous. In Canada, as may appear from the brief description above, granitization of quartzites has taken place on a regional scale in the French River area east of Killarney. This »disappearance of the Huronian» is matched by the disappearance of the Karelian, as appears at a look to even a small-scale geologic map, at the eastern end of the Kemi—Rovaniemi zone in the migmatite and granite areas south of Kemijärvi, or in western Lapland east of Kolari. A detailed survey of the granitic migmatites might possibly reveal traces of quartzites.

The Svecofennian territory of the southwestern half of Finland, west of the Karelide zone where quartzites are abundant, is of old known for its poorness in quartzites. Instead it is a country of granites and migmatites, now known to have originated during the period of the Karelidic diastrophism about 1 800 million years ago. The paleosome of the migmatites, to a considerable part, can be recognized as derived from supracrustal rocks, volcanites and sediments. In many regions metamorphosed sediments have been preserved in large areas and well enough to allow the determination of their original sedimentary facies. The majority of them appears to belong to miogeosynclinal grauvacke-pelite associations. In such formations quartzites cannot be prevalent, but still one might expect to find more of them, at least arkosic and other impure quartzites, such as are quite common in the Huronian zone. Now, in the general regeneration of the Svecofennian, very much true granites have been formed by means of metasomatic granitization. Thus it seems safe to assume that very much quartzite has disappeared in the regeneration.

Among the sporadic occurrences of quartzite the greater part belongs to the type that Pettijohn (1957) called orthoquartzite. They are very pure quartz rocks formed on stable platforms of old peneplanated kratons on which chemical weathering has been complete. In Finland the best example is the Jatulian quartzite in the Karelian. Probably the whole basement gneiss area in eastern Finland, extending far east of the present-day frontier, was once covered by kaolinic products of weathering and later overlain by eolian sand. Väyrynen (1933) called it the Jatulian continent. The deposits were preserved in certain depressions of the peneplain and more generally on the margin of the Karelidic geosyncline stretching through the country from Lake Ladoga in a north-northwesterly direction. In Kainuu and southern Lapland the kaolin deposits are still preserved, while in some parts of Karelia the kaolinite has been dehydrated by metamorphism to kyanite and andalusite. Granitization and migmatization of quartzite on a large scale is known only from the northern Karelide zones, as mentioned earlier in this paper.

In the Svecofennian a well known red orthoquartzite is present on Tiirismaa, west of Lahti. The rock contains, besides quartz, only 11 per cent sillimanite, conceivably formed at metamorphism from original kaolinite, and a little iron oxides and sericite. The red colour, due to hematite pigment, suggests original sand formed in humid and warm climate. In the absence of vegetation the conditions, however, were probably desert-like. A remarkable feature of this and all other Svecofennian orthoquartzites is the absence of any traces of the original basement which, if found, would throw much light upon the oldest geological history of the country. In the case of Tiirismaa, for instance, a dating of the basement could decide the question whether or not it is of the same age as the basement gneiss in Karelia.

While the Tiirismaa quartzite has not been visibly granitized, the quartzite of Simsiö in southern Pohjanmaa (Hietanen, 1938) evidently represents a residue from a larger quartzite mass, in a way corroded by granitization (Fig. 5). The quartzite lens is enclosed in migmatite and surrounded by a continuous zone of granite from which an embayment intrudes in the quartzite which is a very pure quartz rock, but encloses bands and small accumulations of different kinds. First, manganese minerals, such as rhodonite, pyroxmangite, spessartite garnet, rhodochrosite, manganocalcite *etc.*, in another similar quartzite also tephroite and knebelite. These may be interpreted as metamorphosed remains of original sedimentary manganese ores that reacting with quartz formed silicates. Second, bands rich in diopside and tremolite, suggesting former presence of limestone layers intercalated with layers of quartz sand. Third, accumulations of magnetite and pyrite are enclosed in quartzite, naturally being remains of sedimentary, mainly limonitic iron ores. It is a common experience from the Precambrian that iron oxides do not readily react with quartz, but are preserved as oxides, magnetite and hematite. In southern Pohjanmaa there are, however, a few quartzite occurrences, *e. g.*, in Nurmo near Lapua containing grünerite, others ferroan hypersthene or Fe-rich hornblende and other ferrous silicates. The common occurrence of carbonates or their reaction products in the orthoquartzites, again suggests that the ferrous silicates were formed from original siderite. Calcium-aluminium silicates, such as anorthitic plagioclase, epidote, and

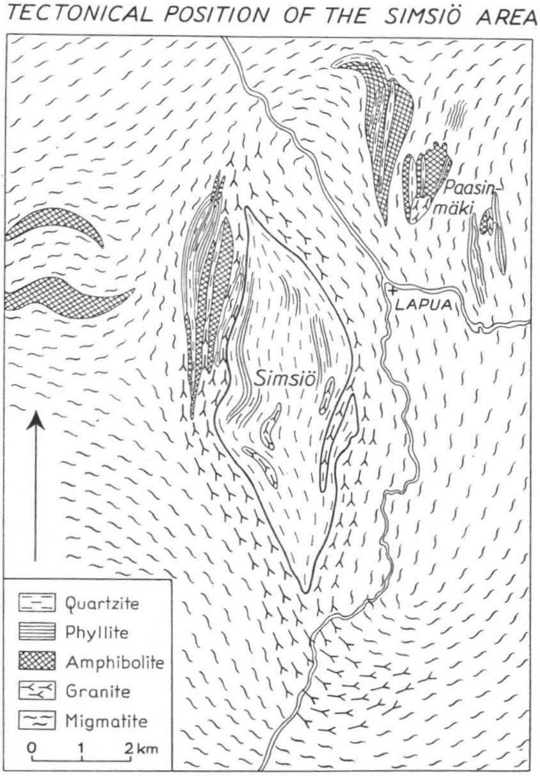

Fig. 5. Quartzite of Simsiö, Lapua. According to Anna Hietanen (1938).

clinozoisite, present in many orthoquartzites, were most probably formed of original marl.

A definite evidence of the granitization of orthoquartzite at Simsiö is the occurrence of the same peculiar silicate mineral accumulations and bands, especially of diopside, also in the granite around the quartzite.

Hietanen (1938) published an analysis of granite at the quartzite contact from Paasinmäki near Simsiö (Table 1, analysis 4). It shows a composition fairly near the eutectoid granite composition, as is also the case of the Killarney granite. This composition is commonly regarded as a proof of magmatic origin, an interpretation which is not precluded by the fact that this granite is a product of granitization of quartzite, *i. e.*, a product of metasomatism, the excessive silicon having been replaced by the other cations of granite. I have earlier repeatedly presented evidence for the idea that granite magma may be a metasomatizing agent *(e. g.*, Eskola, 1948). In general, metasomatic and magmatic crystallization are not controversial.

Pettijohn classed with the orthoquartzites also the calc-arenites, *e. g.*, the calcite-quartz sandstones. The first product of metamorphism of such a rock may be either wollastonite quartzite or wollastonite-bearing marble, depending on the proportion of the source minerals. Both assemblages are known from Finland. Wollastonite quartzite occurs in the quarry of Kuparsaari in Antrea (since 1944 no more belonging politically to Finland). The texture of this rock looks psammitic, rounded quartz grains being embedded in a matrix of wollastonite. (See Hietanen, 1938, p. 74.)

Another kind of wollastonite-quartz rock is encountered in the limestone quarry of Ihalainen near Lappeenranta, where a large mass of limestone is enclosed in rapakivi. There quartz veins are intruded in the limestone and are surrounded by zones of wollastonite, that may form thick bands, now used to the production of »stone wool», melting the wollastonite.

Wollastonite quartzite is no granite but, like the quartzite migmatites, it may be a first step towards granitization. Some migmatic calc-gneisses in process of granitization may well have passed through the earlier wollastonite stage, although experience from the Finnish Precambrian indicates that carbonate-bearing sediments are relatively resistant against granitization. From southwestern Finland I know numerous examples of limestone layers preserved in the middle of granite masses, although the country rock, usually leptite, has disappeared without a trace. Basalts and, still more, peridotites also are fairly resistant, while quartzites seem to be more readily granitizable than other nongranitic rocks, a feature that of course is connected with the porosity of the quartz lattice.

GRANITIZATION IN QUARTZ VEIN

During an excursion to the eastern part of the Tampere schist zone, about thirty years ago, I picked up at the road side east of Vehkalahti in Orivesi, a few samples from a thick quartz vein that had been blasted for quarrying quartz and contained veinlets of granite in the pure vein quartz. A few years later, when H. G. Backlund's radical theories on granitization were under eager discussion, I investigated the specimens microscopically and had the granite analyzed (Table 1, analysis 5) thinking it might throw some light upon the granitization problem.

The granite veins are in maximum about 1.5 cm thick. Their boundaries are curved, and in places minute apophyses from them have bored their way some distance in the quartz. Some veins taper out as a wedge, while some other veins have rounded ends. This manner of occurrence makes it evident that the granite veins have formed by means of replacement and do not fill up fissures.

The mineral constituents are quartz, feldspar and its alteration products, furthermore chlorite and iron ores. The feldspar is seemingly homogeneous and shows no twinning or perthite texture; its refringence is below that of Canada balsam. The observation, hampered by decomposition and dense pigment, does not justify the conclusion that it would be an homogeneous anorthoclase. The chlorite occurs as dirty green scales of the same size as the other minerals, in maximum 1 mm in diameter. It is pleochroic with extremely low anomalous blue interference colours. In composition the chlorite may possibly be close to the aluminous series (amesite-daphnite), but even so the high excess in aluminum shown by the analysis, which gives 4.3 weight per cent normative corundum, cannot be contained in the chlorite alone. The major part of it may well enter into the abundant decomposition products of the feldspar.

In texture the rock is fine-grained, eugranitic. Many feldspar grains have an idiomorphic thick table shape. Quartz mainly fills up the interspaces, but also forms rounded grains of its own. Part of quartz and feldspar are intergrown in a micropegmatitic manner.

The country rock of the quartz vein is a schistose rock composed chiefly of a fine sericite mass and thick plates (porphyroblasts) of oligoclase. The sericite mass seems to contain anhedral pseudomorphs, probably of some aluminous silicate. The rock is, anyway, an aluminous quartz-free schist, and not at all migmatic. No granitic or pegmatitic intrusions occur in the near environment, and the microgranitic veins do not continue outside their host rock, the quartz vein.

The geologic importance of this microgranite lies in the fact that it assuredly has not been derived from any magmatic intrusion, but must have obtained its material exclusively from the next surroundings of the big quartz vein by means of nonmagmatic metasomatism. Furthermore, it adds an instance of a nearly eutectoid composition of metasomatic granites of which the Killarney granite and the Simsiö granite were presented as examples previously in this paper.

As all Finnish geologists well know, the metasomatic Svecofennian granites, whose source rocks had been either sediments or intermediate and mafic plutonites, are invariably typical microcline granites (kaligranites). The three analyses of metasomatic granites quoted in this paper as examples of metasomatic granites, however, are all more like eutectoid granites in composition. In the cases of Killarney granite and Simsiö granite this could be accounted for by the fact that metasomatism was mediated by granitic magma. For the microgranite last described this interpretation is excluded, but actually it is not a genuine granite, for it has a considerable excess of silicon (from its quartzeous host!). Here must also be remembered that the origin of kaligranitic magmas, such as the rapakivi magma, is not yet fully understood. Petrogenesis still has many puzzles to be solved. As N. L. Bowen used to say: »We still know very little about rocks.»

63 2192—61

I am obliged to Mr. Richard Ojakangas for calling my attention to the recent investigations of P. D. Krynine and other American geologists who have proved that the colour of the »red beds», containing hematite pigment, is due to a warm climate during the weathering of the rock material and not, as was believed earlier, to an arid climate. It seems most probable that the same interpretation applies to the red Tiirismaa quartzite mentioned in this paper (p. 494).

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30

THE STRUCTURAL POSITION OF THE ORIJÄRVI GRANODIORITE AND THE PROBLEM OF SYNKINEMATIC GRANITES [1]

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ABSTRACT

The Orijärvi granodiorite is a typical representative of synkinematic Svecofennian granites. The rock forms a phacolith-shaped core in an open asymmetric anticline, but in detail shows discordant features relative to bedding planes.

Zones of different magnetite concentration, which coincide with certain strata of the surrounding rocks, continue uninterrupted through the granodiorite. In the granodiorite they form a relict bedding which indicates that the pluton was formed by replacement.

The relict beds become thicker toward the axial zone of the phacolith. The thickening is associated with a change of composition ranging from cordierite-anthophyllite rocks, situated at the lowest edge of the gentler limb, to adamellitic rocks of the axial zone.

This phenomenon is interpreted as the result of migration of elements from the limbs toward the axial zone during the folding, *i. e.*, down the gradient of shearing stress and pressure. Due to different mobility of the elements the migration has caused a chemical fractionation of the original complex. The contracting limbs were basified while granitization took place in the expanding axial zone. The pluton possibly first formed, or began to form, as an assemblage of more hydrous minerals.

In its present state the granodiorite is a metamorphic rock of the regional amphibolite facies. Unrotated and undeformed crystals in places of intensive synkinematic shear and plastic flow indicate that its present mineralogy is the result of late- or postkinematic metamorphism.

Conventional criteria used to prove that the synkinematic Svecofennian granites are allochtonous igneous rocks also fit the Orijärvi granodiorite. Hence, such criteria seem invalid as proof of magmatism.

[1] Received May 9, 1961.

CONTENTS

INTRODUCTION

»One of the difficulties in proving granitization is that nearly all criteria are subject to alternative interpretations.»

A. F. Buddington (1959, p. 733)

The rocks called synkinematic Svecofennian granites are slightly gneissic, quartzose granitoid rocks in which sodic plagioclase usually is the dominant feldspar. Since times antedating Sederholm, the »official» Finnish geology has considered them as purely magmatic (See: Simonen, 1960). Some other interpretations also have been advanced, but have not raised serious discussion (Backlund, 1936; Tuominen, 1951 *etc.;* Marmo, 1956 *etc;* Parras, 1958.

Yet some uneasiness has lately been observable around the subject. A glimpse of this appears from the following words of Simonen (1960, p. 93; *cf.* Simonen, 1948).

»Indisputable evidence bearing on the problem of the magmatic versus metasomatic origin of granitic rocks is, however, difficult to find, because considerable data previously regarded as indicating only a magmatic origin may be applied also to metasomatic and metamorphic rocks.»

Our motto thus seems to hold also *vice versa.*

In lack of indisputable evidence, Simonen (1960) piles up some one hundred pages of disputable »evidence» to support the magmatic views. Using numerous variation diagrams, based on purely petrographic data, he classifies the synkinematic plutons into four different petrographic provinces — the granodiorite, trondhjemite, charnockite and granite provinces — each forming a differentiation series ranging from basic to acidic members. The chemical differences between the different provinces are interpreted as resulting from »slight primary chemical differences between the parent magmas» *(op. cit.,* p. 94).

The plutons in question normally represent the regional mineral facies. The hornblende-biotite granodiorites, for instance, are rocks of amphibolite facies in regions of amphibolite facies, while the corresponding charnockitic rocks are rocks of granulite facies in regions of granulite facies. According to Simonen *(op. cit.*, p. 90) this »suggests that the grade of metamorphism is controlled by the type of plutonism». He further states that the regional metamorphism connected with deformation and synkinematic plutonism was roughly isochemical.

The present paper discusses one of the typical synkinematic plutons included in Simonen's data, the Orijärvi granodiorite. This is done in order to show that the conventional criteria of magmatism enunciated by Simonen and others are thoroughly inadequate.

In the present paper the term granite is defined according to Chayes (1957), and the subdivision (trondhjemite, granodiorite, adamellite) follows his classification. Terms like pluton, phacolith, dike, *etc.*, are used in a descriptive sense only. The terms metasomatism, replacement, granitization and basification are not restricted to constant volume, »dry», or »100 percent solid state» processes. With regard to the Orijärvi granodiorite, the name granodiorite means the average composition of the pluton.

GEOLOGIC ENVIRONMENT

The Orijärvi region, which represents the amphibolite facies (Eskola, 1915), forms the southwest part of a large gneiss block. This may be called the Karjaa—Karkkila block (Fig. 1). The northeast part of the same block, called the West-Uusimaa region, is a region of granulite facies (Parras, 1958). The block which roughly has the outline of a parallelogram is surrounded and partly infringed upon by »latekinematic» microcline granites. Several wider or smaller gneiss offshoots extend in different directions from the block. Only the widest two are shown on Fig. 1.

Numerous more or less isolated areas of synkinematic granites are enclosed in the block. Their mineral facies is the mineral facies of the surrounding rocks.

The layered rocks of the block are here, as usual, called the leptite formation (of south-western Finland). They are metamorphic derivatives of argillites, graywackes, conglomerates, marls, limestones and basic volcanic rocks. In t h i s respect there seems to be no essential difference between the Orijärvi and West-Uusimaa regions which explains their different mineralogy. The character of the border separating the two areas of different metamorphism is unknown. It may be a product of faulting as suggested by Parras (1958).

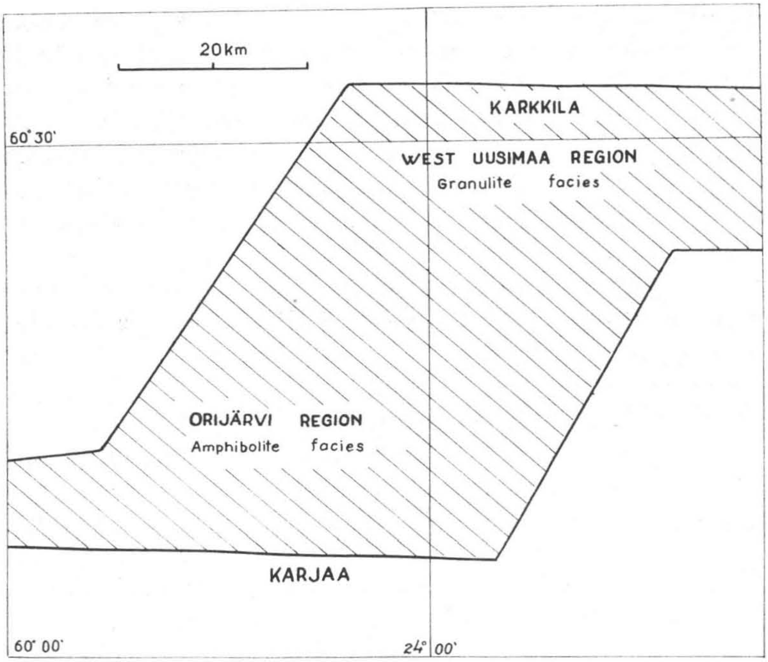

Fig. 1. The Karjaa—Karkkila gneiss block. Simplified from Härme (1960).

The layered rocks of the block reveal an open folding over axes generally plunging gently to the east. In the Orijärvi region, the folds are slightly asymmetric with steeper limbs dipping south.

The easterly plunge of the fold axes is associated with monoclinal cross folding and thrust faulting towards the west, which eliminate the long-distance effect of the easterly plunge. This and many other types of faulting split the complex into a mosaic in which the older geologic continuities are more or less broken. The largest faults form shear zones up to one kilometer or more wide (Tuominen, 1957).

Many of the faults are older than the present mineralogy of the rocks and became inactive before these minerals formed. Other faults have been repeatedly active or formed in later times.

GENERAL POSITION AND FORM

In his Orijärvi memoir, Eskola (1914) explains that the different areas of granodiorite (or »oligoclase granite») of the Orijärvi region represent concordant, widely differentiated igneous batholiths occupying anticlinal areas of large folds. In later papers he suggest that these »anticlinal batholiths»

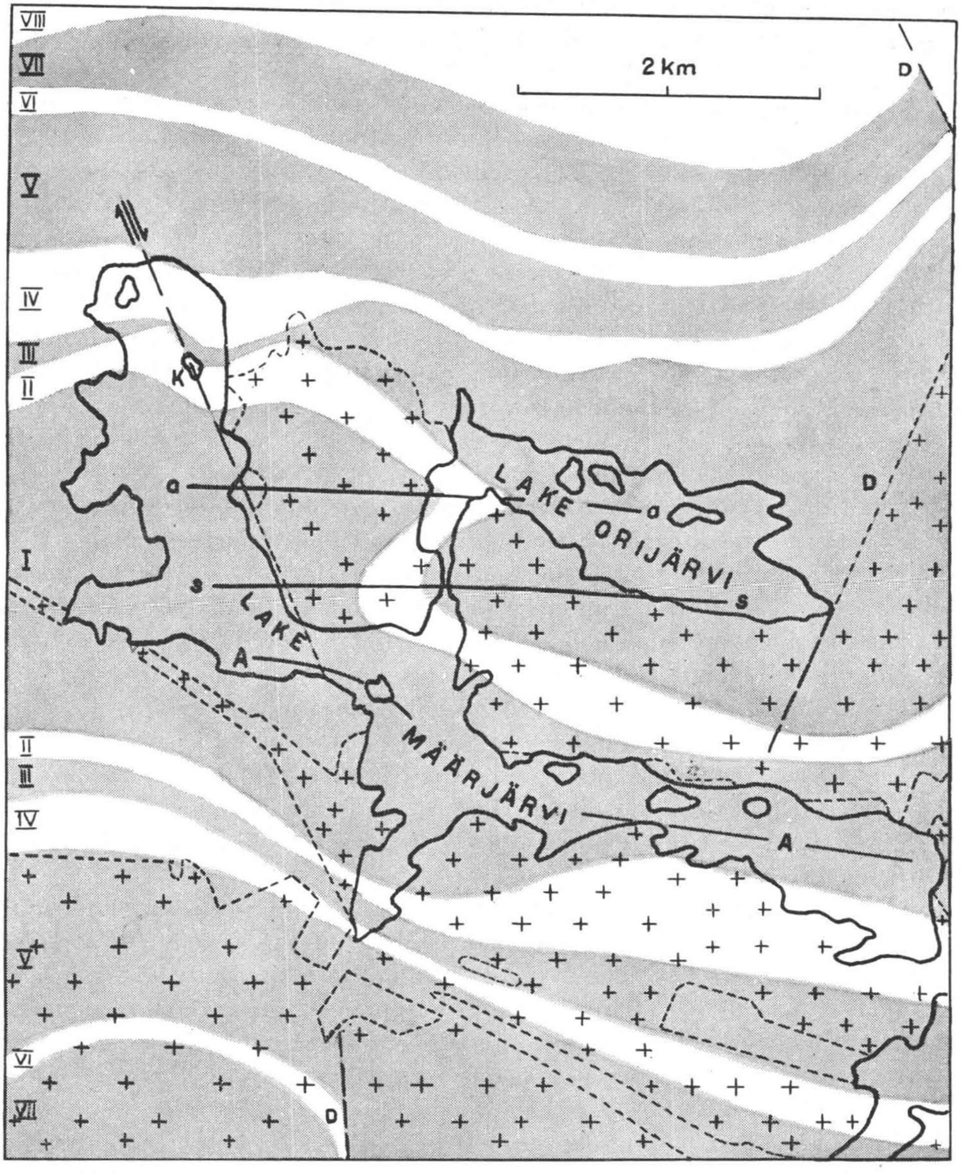

Fig. 2. Map of the Orijärvi granodiorite and surroundings. Crossed areas are granodiorite. A—A = Määrjärvi anticline. a—a = secondary anticline. s—s = secondary syncline. D = downthrown side of thrust fault. Arrows = strike-slip fault. K = Isle of Kurksaari. Shaded = zones of higher magnetic intensity. White = zones of lower magnetic intensity. Roman numbers I—VIII show the apparent sequence of strata. The magnetic zoning seen in the map is compiled from measurements at over 50 000 stations.

are steeply dipping domelike diapirs as a consequence of the diapiric upward movement of the granite magma during the folding. (Eskola, 1949, p. 468 and Fig. 7, II; 1950, p. 82).

The name »Orijärvi granite» or »Orijärvi batholith» was originally (Eskola, 1914) reserved to the mass of granodiorite situated between Lake Orijärvi and Lake Määrjärvi (Fig. 2). The granodiorite area south of Lake Määrjärvi was supposed to represent another anticlinal batholith, separated from the former by the narrow leptite zone underlying the eastern branch of Lake Määrjärvi. Consequently, this leptite zone was supposed to form a syncline.

Careful re-investigations in the Orijärvi region (Tuominen, 1957) showed that the leptite zone of Lake Määrjärvi represents in fact the crest zone of a large open anticline or anticlinorium, called the Määrjärvi anticline. This anticline is slightly asymmetric towards the south and plunges gently eastward. The axes of the neighboring synclines are 9 km north and 4 km south from the crest of the anticline.

The dragged layers of the crestal leptite zone dip 40—70 degrees south under the granodiorite of the south shore, and 0—40 degrees north under the granodiorite of the north shore of Lake Määrjärvi [1]. There is no question, therefore, that the granodiorites south and north of Lake Määrjärvi are parts of a single folded body which anticlinally overlies the leptites outcropping in the islands and shores of the lake.

The top of the north limb of the granodiorite also dips north and underlies the leptites which form the basin of Lake Orijärvi and the area north of this. The northwest edge of the surface section of this limb represents the lowest edge of the whole limb. Here the granodiorite fades away, changing gradually to cordierite (-anthophyllite) gneisses and other rocks allied to these.

The top contact of the southern limb cannot be defined with the same accuracy. However, certain beds characteristic of the area north of Lake Orijärvi are also met with in the narrow leptite strips 1—2 km south of Lake Määrjärvi. These strips obviously separate the granodiorite under consideration from the Simjärvi granodiorite occupying the neighboring synclinal through south of the Määrjärvi anticline.

The flanks of both limbs of the granodiorite are roughly concordant with the bedding of the enclosing leptites. In several places, however, faults and other planar structures deviating from bedding form the original border of the rock. For instance, the west border of the north limb partly follows the Kurksaari fault, Figs. 2 and 3. This fault as well as other faults of the area

[1] At the north shore, the gently dipping foot contact of the granodiorite is visible in closely spaced outcrops for a length of some 600 meters. Four diamond drill holes through the granodiorite (500—600 meters north of first »r» in the word Määrjärvi, Fig. 2) have also verified the gentle dip of the foot which was crossed at depths of 85 to 100 meters. The holes continued into the underlying leptites to depths of over 200 meters without meeting more granodiorite (Mikkonen, 1952).

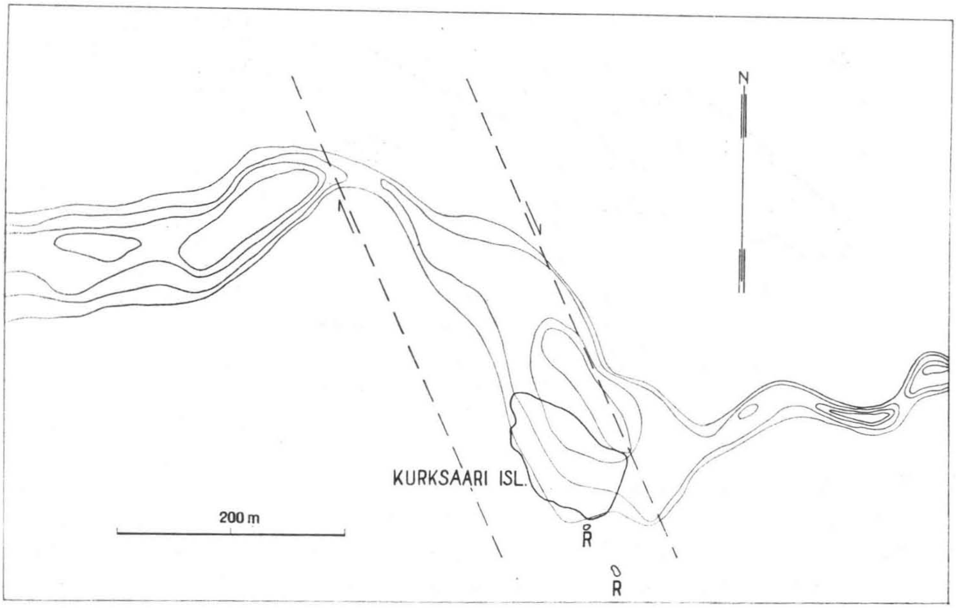

Fig. 3. Magnetic vertical intensity map revealing the Kurksaari strike-slip fault. Contours: 400, 800, 1 600 and 3 200 gammas. Dashed lines indicate the borders of the fault zone. R = small rocks.

are best revealed by the magnetic maps. The Kurksaari fault is primarily a strike-slip fault with an apparent slip of 300 meters. The sheared zone is over one hundred meters wide.

The fault zone is best exposed on the isle of Kurksaari inside the zone. This rocky islet consists mainly of cordierite-anthophyllite gneiss and partly of cummingtonite amphibolite. Embedded in the cordierite-anthophyllite gneiss are sharp-edged rotated fragments of the surrounding leptites. These fragments are partly replaced by porphyroblasts of cordierite up to 30 cm long (Eskola, 1914). The longest axes of the cordierite porphyroblasts show some tendency to be parallel to the faint foliation which is clearly related to the fault. The foliation, the brecciated structure and hence the whole fault, are, however, only relict features relative to the present mineralogy of the gneiss. No essential movement seems to have occurred along the fault during or after the growth of the cordierite crystals.

The granodiorite, too, at the border of the Kurksaari fault, seems to be without traces of crystal deformation related to the fault. It seems obvious, therefore, that the fault has not dissected the granodiorite nor been active after the final crystallization of this rock. This means that the fault is here the original border of the pluton.

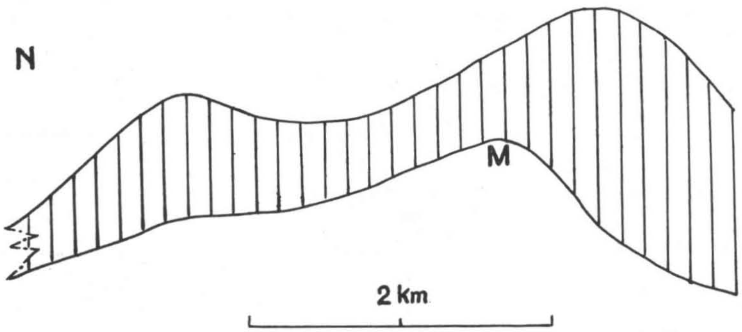

Fig. 4. Schematic cross section of the Orijärvi phacolith.
M = crest of the Määrjärvi anticline.

In spite of the fact that the original contacts of the pluton in many places follow faults and other cutting structures, its structural position seems to be mainly determined by folding. In this respect the simplest reasonable picture inferred from the features presented above is: The granodiorite is, by form and position, roughly a phacolithic core in the Määrjärvi anticline. This core is in the following referred to as the Orijärvi granodiorite or the Orijärvi phacolith (Fig. 4).

What is said above about the contacts of the pluton concerns merely their general trends. In detail there are marked deviations from these. As seen in single outcrops, even the flank contacts, if sharp enough for direct observation, in many cases intersect the bedding planes. In addition, a multitude of fragments from the surrounding rocks is enclosed in the granodiorite all over the mass. Thus, with regard to detailed features, the Orijärvi phacolith is a highly discordant body.

NATURE OF CONTACTS

The contact zones of the Orijärvi granodiorite are not widely enough exposed to provide a detailed picture. Plenty of information in this respect is, however, obtained from the other granodiorite plutons of the region. This information shows that the discordant contact relations described in the previous chapter are not the result of a »forceful magmatic intrusion».

The contacts with the leptitic rocks are generally gradual. Toward amphibolites the apparent change is sharper (Tuominen, 1957). Accordingly, the fragments of the surrounding rocks embedded in the granodiorites are mostly amphibolitic.

The rock of the border zones is in many places porphyritic (Eskola, 1914, p. 53; Tuominen, 1957): Crystals and nodules of quartz and sodic plagioclase,

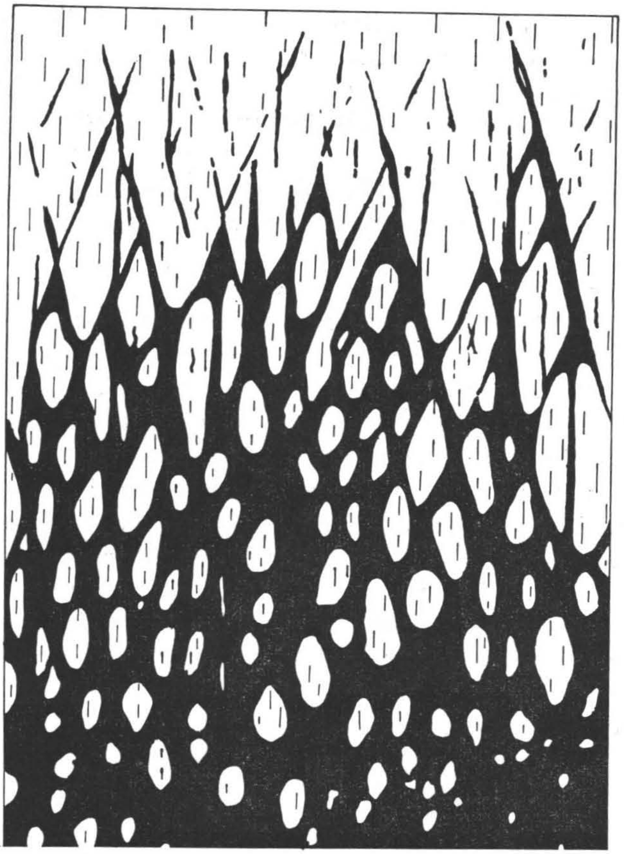

Fig. 5. Sketch illustrating typical contact zone between quartz-porphyritic border type of granodiorite (black) and rocks of leptite formation (white). Dashes indicate the strike of foliation. Scale ≃ 1:50.

up to several millimeters in diameter, are embedded in a microgranitoid matrix. Toward the inner parts of the plutons the number of the larger crystals increases. The result is a more or less even-grained granodiorite. In the opposite direction the number of the larger crystals decreases and the rock becomes, say, an even-grained leptite. At the northwest corner of the Orijärvi granodiorite, the porphyritic border rock grades over to cordierite gneiss.

At Kuovila, some ten kilometers southwest of Lake Määrjärvi, the porphyritic type of granodiorite forms a conjugated system of steeply dipping dikes. The dikes are up to six meters wide. They intersect the wrinkled bedding and steep eastwest striking foliation of the surrounding rocks, which are mainly fine-grained amphibolites. The walls of the dikes are not even, but indented parallel to the foliation and some fractures of the host rock. Numerous fragments of the host rock are embedded in the dikes.

Internally, the dike rock reveals a steep lineation and a slight foliation parallel to the dike. The fragments, however, are more or less flat and parallel to the indentation of the walls, and their foliation is closely parallel to the foliation of the wall rock. They are rounded and thus do not match each other or the dike walls. The borders of the fragments are generally sharp. They are »corroded», however, and inside the fragments there are scattered crystals of quartz and feldspar similar to those in the dike.

In many places the porphyritic contact zones of the larger granodiorite masses reveal features similar to those seen in the dikes of Kuovila. Relations like these, illustrated in Fig. 5, are characteristic of metasomatic phenomena and are thus strong indications of granitization. This has advanced along certain structural planes. The rock between has been left more or less unaffected and forms the present inclusions.

In the examples above the fragments have not been noticeably rotated relative to each other or the source rock. It is probable, however, that some laminar slip has taken place. This seems natural, as the contacts of this type follow zones of gentle dislocation. In some other cases there are slight rotational features. These may be expected in a synkinematic pluton whether magmatic or not.

RELATION TO SURROUNDING ROCKS

Apart from the contact zones, little direct evidence of granitization is seen in the Orijärvi granodiorite. Critical evidence in this respect is, however, obtained from the ground-magnetic survey (Tuominen, 1957, Plate III).

The magnetic map shows zones of different average intensity of the vertical component. These zones coincide with certain recognizable groups of strata of the leptite formation. Further, the zones continue, with lower intensity, into and through the granodiorite where no strata can be directly observed. The different zones are seen in Fig. 2. [1] They are marked with Roman numbers I—VIII, corresponding to the apparent sequence of strata. Zone I represents the lowest members of the sequence. They are exposed along the crest of the Määrjärvi anticline.

In the leptite formation the zonal variation of magnetic intensity corresponds to a zonal variation in the concentration of highly magnetic minerals such as magnetite and pyrrhotite.

[1] The zones have been tested on the original unpublished magnetic maps; courtesy of Finnish Ore Company.

Table 1. Average modal composition of magnetic zones in Orijärvi granodiorite.

Zone No. of specimens	I 39	II 34	III 21
Quartz	35.4	36.8	33.7
Potash feldspar	7.8	6.2	4.5
Plagioclase	43.8	42.5	46.0
Muscovite	.2	.9	—
Biotite	10.0	10.9	13.0
Chlorite	.1	.8	.1
Hornblende	1.6	.7	1.2
Epidote minerals	.7	.9	1.3
Accessories	.4	.2	.3

The low-intensity Zone II follows beds of cordierite-bearing and other leptites poor in magnetic minerals.

The high-intensity Zone III, specifically, is caused by a single bed of banded magnetite ore. The bed was discovered by the magnetic »ridge» seen in Fig. 3 and penetrated by several drill holes. North of Kurksaari the bed in question dips steeply north. Towards the east (Fig. 2), its general dip becomes gentle. In detail, however, it is here intimately folded and mixed together with the low-magnetic beds of the under- and overlying Zones II and IV. In the surroundings of Lake Orijärvi, magnetic Zone III thus reveals a tectonic mixture of banded iron ore and low-magnetic beds.

Zone I also indicates a similar mixture. Whether the banded iron ore of this zone represents a single bed or several is unknown.

The Orijärvi granodiorite, as defined above, is mainly situated within these three zones (I, II, III). No corresponding zoning is directly seen in the pluton. Fragments of banded iron ore are met with in the high-intensity zones. It is obvious, therefore, that a part of the magnetic intensity comes from them.

In order to study the compositional variation of the granodiorite, 94 specimens were collected systematically from both limbs of the pluton. To avoid »contamination» they were taken as far as possible from inclusions. A point counter analysis was made of every specimen. Results related to the magnetic zoning are given in Tables 1 and 2.

Table 1 gives the average modal composition of each zone. Compositions are very similar. The amount of lime minerals is slightly higher in Zones I and III than in Zone II. This difference corresponds to the differences appearing between the zones outside the granodiorite. It seems, however, to lack statistical significance.

Table 2 shows the concentration of opaque minerals in the three zones. The minerals were recorded in connection with the general modal analyses. Thus the results give merely the order of magnitude. Of four hundred grains

Table 2. Opaque minerals in magnetic zones of Orijärvi granodiorite.

	Zone	I	II	III
1	No. of specimens (Total 94)	39	34	21
2	Average percent of opaque minerals	.2	.04	.2
3	No. of specimens in which opaque minerals were recorded (Total 36)	20	5	11
4	Average percent of opaque minerals in specimens of line 3	.5	.3	.5
5	Highest percent of opaque minerals in single specimen	1.4	.7	1.6

of the opaque minerals studied in eight polished sections, only nineteen (5 percent) seemed to be other than magnetite. This means that the opaque minerals of Table 2 can practically be considered as magnetite. The specimens in which magnetite was recorded show an even distribution along the zones.

With regard to what is said above, the following relations appear from Table 2: The average concentration of magnetite in Zones I and III is five times that of Zone II. In Zones I and III magnetite was recorded in every second specimen versus every seventh in Zone II. From these specimens, Zones I and III show two times higher magnetite percents than Zone II. The same applies to the maximum values of line 5.

This means that in Zones I and III the concentration of magnetite is generally higher than in Zone II. Hence, in the granodiorite also, the zones of higher and lower magnetic intensity coincide with zones of higher and lower magnetite concentration. The magnetic zoning thus shows that the stratiform distribution of magnetite of the leptite formation continues into and through the granodiorite.

The only reasonable explanation to fit this and the other features described above is that the strata of the leptite formation continue, in the form of relict strata, into and through the granodiorite. This indicates that the Orijärvi granodiorite is an authoctonous product of granitization — or granodioritization if preferred.

RELATION TO DEFORMATION

The presence or absence of gneissous fabric is the traditional basis to distinguish between synkinematic and other plutons. Foliation parallel to the foliation of the surrounding rocks and »flow lines» parallel to the tectonic *b*-axes are specifically favored as evidence of synkinematic origin.

In the Orijärvi granodiorite there are several systems of foliation and lineation, all of which are present in the surrounding rocks as well (Tuominen, 1957). Most of the systems follow trends of old faulting and other structures deviating from the trend of the general folding. The most frequent foliation and lineation are, however, roughly parallel to the axial plane of the main folding. A very pronounced pencil lineation parallel to the tectonic *b*-axes appears along the main crest zone of the phacolith. This lineation results from a girdle pattern of biotite and a parallel orientation of elongated quartz grains. The optic axes of the quartz grains show a maximum in *b*.

In a granitized pluton, features like these may descend from the pre-existing rocks and thus do not show whether the granitization has been synchronous with the folding or not. Other structural relations must be studied in order to find reliable criteria in this respect.

The phacolithic position of the pluton no doubt suggests a synkinematic origin. Further evidence of at least equal importance is obtained from the internal construction of the pluton. This subject will be discussed in a coming paper and is only briefly reviewed below.

The relict beds become thicker toward the axial zone of the phacolith. This phenomenon is best seen in Zone II of Fig. 2. The steep south limb of the corresponding bed is several times thicker than the gentle north limb of the same bed. Ordinarily, the visible part of the south limb represents the axial zone of the asymmetric fold (Fig. 4).

By composition the thinner part of this bed dipping north is trondhjemitic while the thicker part dipping south is adamellitic.[1] As a whole, the composition of the phacolith varies gradually with the folded forms. The gradual series begins with cordierite-anthophyllite rocks and allied rocks situated at the lowest edge of the north limb. It continues through trondhjemitic and granodioritic compositions of the north limb, and ends with adamellites (Eskola, 1952, p. 128) occupying the axial zone of the phacolith. A clear peak toward adamellitic composition also appears along the axial zone of the secondary anticline situated at the north limb (Fig. 4).

The intimate relations between the form of the folds, the form of the phacolith, and the composition of the phacolith, all being similarly asymmetric, prove that the pluton was formed during the folding.

The relations described suggest that the original beds to a certain degree yielded to the deforming stress by means of a chemical flow (Mehnert, 1960). This possibly was a one-way migration of chemical elements from the limbs toward the axial zone of the fold or, generally, down the gradient of shearing stress and pressure (Ramberg, 1952; Bennington, 1956, 1959). Due to different »mobility» of the elements — increasing in the order: Mg-(Fe, Ca, Ti)-Al-(Si, Na)-K — the migration resulted in a gradual chemical fractionation of the original complex (Tuominen and Mikkola, 1950; Tuominen

[1] Disregarding the structural relationships, Simonen (1960), following Eskola (1952, p. 128), places the north-dipping part in his synkinematic granodiorite province and the south-dipping part among the late-kinematic microcline granites.

1951, 1958). The contracting limbs were basified while granitization took place in the axial zone which was expanding.

This means that the metamorphic differentiation which produced the pluton and its basic residuum, was a part of the folding mechanism.

RELATION TO REGIONAL METAMORPHISM

The synkinematic ancestry of the granodiorite still does not give a full picture of its position in the tectonic evolution. As a metamorphic rock of the regional amphibolite facies, it is undoubtedly contemporaneous with the surrounding metamorphic rocks (Eskola, 1915, p. 118).

The argillites of the leptite formation, as well as the restites derived from them, characteristically contain porphyroblasts of cordierite, andalusite (sillimanite), garnet and, in places, radiating »suns» of anthophyllite. Even in places of strong synkinematic shear and plastic flow these minerals show no traces of rotation or deformation (Tuominen and Mikkola, 1950; Eskola, 1950; Tuominen, 1951). Similar undisturbed features are seen in every rock type of the region and not least in the granodiorites. The present mineralogy of the rocks, that is the amphibolite facies [1], thus resulted from a recrystallization which took place after the folding, possibly during the development of the »late-kinematic» microcline granites.

No critical mineral relicts revealing the physico-chemical conditions of the synkinematic stages have been recognized. The nature of the restites may, however, be significant in this respect. The cordierite-anthophyllite gneiss of Kurksaari (p. 505) is a premetamorphic fault breccia and has the bulk composition of dehydrated gouge. It seems possible, therefore, that it is a metamorphozed gouge. The composition and the highly plastic past of the cordierite-anthophyllite-bearing and other restites situated at the flanks of the large folds suggest a similar origin (Tuominen and Mikkola, 1950; Mikkola, 1955).

In hydrous fault breccias, the original rocks commonly show differentiation into fractions rich in phyllosilicates (with high Mg, Fe and Al) and fractions rich in zeolites (Kerr, 1955). The latter generally occupy positions of relative expansion [2]. Similar differentiation can be expected in large sliding breccias and mylonites formed in layered systems during the early stages of folding.

It is thinkable, therefore, that the phacolith was originally formed, or began to form, as an assemblage of more hydrous minerals which after the

[1] The contact metamorphic hornblende-hornfels facies of Turner (Fyfe, Turner and Verhoogen, 1958) does not fit the conditions described.

[2] It is not difficult to imagine how such rocks are interpreted after a high-grade normal metamorphism.

folding were converted into the present assemblage of amphibolite facies. As a granodiorite the pluton would thus be late- or postkinematic.

OTHER GRANODIORITE MASSES OF THE ORIJÄRVI REGION

The above discussion on the Orijärvi granodiorite also applies to other granodiorite masses of the region. The larger plutons mainly follow the crest and trough zones of major folds. In all cases they are partly controlled by faults as well. Many of the smaller bodies are confined only in faults. In the faults, too, the granitization seems to have taken place in places of relative expansion while the strongly sheared and contracted parts have been basified.

The Simjärvi granodiorite, situated immediately south of the Orijärvi granodiorite, is a synclinal phacolith. It replaces the leptite formation above Zone IV (Fig. 2) and thus represents a higher stratigraphic level than the Orijärvi granodiorite. These two plutons together contain almost the entire known sequence of the leptite formation and thus replace widely different rock types.

The restites, too, come from different rocks. By composition they vary from cordierite-anthophyllite-bearing rocks to peridotitic (Mikkola, 1955) and other more or less basic rocks.

These relations indicate that the metamorphic fractionation under consideration was not restricted to rocks of specific bulk compositions but was rather omnivorous in this respect. Whether a rock was granitized or degranitized, or preserved its bulk composition more or less intact, was mainly determined by its position in the deforming system and its physical behavior toward the deformation.

CONCLUDING REMARKS

Strata of the surrounding rocks, continuing uninterruptedly through the pluton, seem to remain as the only reliable evidence proving large scale granitization (Buddington, 1959). This requirement seems established with regard to the Orijärvi granodiorite.

Taken liberally, the convential criteria of magmatism presented by Simonen (1960) also fit the pluton. This, however, only means that such criteria do not distinguish between granitization and magmatism.

There may thus be purely magmatic rocks among the synkinematic Svecofennian granites. The evidence, however, remains to be shown. Otherwise the »slight primary chemical differences» between some of Simonen's

»parent magmas» could be explained as primary differences in the pre-granitic rocks. This, of course, would increase the number of possible »petrographic provinces» almost unlimitedly.

Metasomatic compositions seen in high-metamorphic rocks are normally believed to result from high-temperature (magmatic-anatectic) processes. There is no doubt that large-scale granitization and, hence, basification too, takes place in those conditions.

On the other hand, considerable chemical migration probably also takes place during the low-temperature hydrous stages of folding. I would think, therefore, that the study of the lowest grades of deformational (open system) metamorphism, or »alteration», could explain some of the metasomatic mysteries of the high-metamorphic terrains.

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