Geology, lithogeochemistry and paleotectonic setting of the host sequence to the Kangasjärvi Zn-Cu deposit, central Finland: Implications for volcanogenic massive sulphide exploration in the Vihanti-Pyhäsalmi district



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Abstract

The Kangasjärvi Zn-Cu deposit is a highly deformed and metamorphosed Paleoproterozoic volcanogenic massive sulphide (VMS) deposit located in the Vihanti-Pyhäsalmi base metal mining district of central Finland. The host sequence to the deposit, referred to as the Inner Volcanic Sequence (IVS), is comprised of a bimodal suite of metavolcanic rocks and a regionally extensive tonalite-trondhjemite gneiss (sub-volcanic intrusions?). A separate and perhaps younger sequence of mafic volcanic rocks, with irregular intervals of undifferentiated intermediate to felsic schists and metalimestones, referred to as the OuterVolcanic Sequence (OVS), are separated from the IVS sequence by intervals of metagreywacke and U-P-bearing graphitic schists.

A stratigraphic scheme for rocks within the IVS is proposed based on outcrop observations, locally preserved volcanic textures, aspects of seafloor-related hydrothermal alteration and lithogeochemistry. In this scheme, rare andesites form the lowermost volcanic stratigraphy and are overlain by typical island-arc basalts that were erupted in a subaqueous setting. Tonalite-trondhjemite subvolcanic intrusions were locally emplaced within andesites and coeval rhyolites were extruded on the basaltic substrate. The extrusion of rhyolites, including high-silica rhyolites, was coeval with regional-scale, pre-metamorphic seafloor hydrothermal alteration and local sulphide mineralization. Extensively altered rhyolites envelope massive sulphides and are underlain by altered basalts. The latter rocks are now characterized by a variety of low-variance metamorphic mineral assemblages (e.g. orthoamphibole-cordierite rocks) and define a domain of intense pre-metamorphic chlorite \pm sericite alteration in the stratigraphic footwall of the deposit. The altered nature of these rocks is attributed to reaction with seawater-related hydrothermal fluids within a zone of upflow at or near the seafloor.

The fundamental controls on convective hydrothermal circulation and subsequent alteration and massive sulphide mineralization at Kangasjärvi, and possibly elsewhere in the district, share many characteristics with other well-described, ancient VMS deposits (e.g. massive sulphide deposits in the Flin Flon Belt, Manitoba, Canada). These characteristics include: 1) an association with bimodal volcanism developed in extensional settings; 2) a close spatial association with regionally extensive felsic subvolcanic intrusions; and 3) petrogenesis of ore-associated volcanic rocks (e.g. high-silica rhyolites, felsic subvolcanic intrusions) indicative of substantial heat transfer from the mantle to the upper crust and the development of anomalous thermal corridors. These features translate into geochemically distinctive rock types that, when combined with aspects of stratigraphy and pre-metamorphic alteration, may be used to develop regional exploration strategies in the Vihanti-Pyhäsalmi district.

Key words: massive sulfide deposits, copper ores, zinc ores, metavolcanic rocks, gneisses, lithogeochemistry, chemostratigraphy, genesis, tectonics, Paleoproterozoic, Kangasjärvi, Keitele, Finland

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I. Introduction

Paleoproterozoic rocks in central Finland are host to many recently active and past-producing Zn-Cu mines and prospects (Gaál, 1990) and comprise a geologically and economically important base metal metallogenic belt referred to as the Vihanti-Pyhäsalmi district. In this polydeformed and highly metamorphosed belt, massive Zn-Cu sulphide deposits are hosted by ca. 1.9 Ga bimodal volcanic sequences (Kousa et al., 1994) and are associated with distinctive metamorphosed alteration halos (e.g. orthoamphibole-cordierite rocks) indicative of seafloor-related hydrothermal alteration (Huhtala, 1979; Mäki, 1986; Ekdahl, 1993; Lahtinen, 1994; Roberts et al., 2003). Hence, the bulk of these deposits are considered to be syngenetic, volcanogenic massive sulphide (VMS) deposits.

Although recent workers have shown that syntectonic mobilization of ore has occurred on the mine scale (metres to hundreds of metres; Mäki & Luukas, 2001), the geological and geochemical signature of syngenetic ore formation remains the most significant exploration tool on a regional scale. However, due to a combination of polyphase deformation, upper amphibolite to granulite facies metamorphism and poor exposure, the fundamental controls on syngenetic mineralization in the district remain unclear. For example, several geologic relationships that are considered essential to VMS exploration, such as the depositional setting and stratigraphy of ore-associated host sequences (e.g. Allen et al., 1997) and the style and extent of systemscale alteration (e.g. Galley, 1993; Brauhart et al., 2001), have not been described in detail.

This paper presents textural descriptions and stratigraphic relationships of ore-associated rocks in the Kangasjärvi area, and these are used to elucidate the stratigraphy of the mineralized host sequence. Whole-rock geochemistry is used to classify least-altered volcanic and plutonic rock types, and immobile elements are employed to differentiate their altered equivalents (e.g. orthoamphibole-cordierite rocks). Immobile-element chemostratigraphy, in combination with the identification of pre-metamorphic, seafloor-related alteration facies proximal to Zn-Cu mineralization, is used to resolve the stratigraphic younging direction of the mine sequence. Within this stratigraphic framework, the petrogenesis of key volcanic and plutonic rocks is then examined to determine the paleotectonic setting in which VMS-style mineralization formed. These data provide insights into the evolution of ca. 1.9 Ga island arc systems in central Finland and highlight the fundamental controls on syngenetic mineralization in the Vihanti–Pyhäsalmi district.

2. Geological background

2.1. Geology of the Vihanti-Pyhäsalmi district

The Vihanti-Pyhäsalmi district is a moderate-size (~100 Mt) base metal mining camp in central Finland (Fig. 1) (Huhtala, 1979; Gaál, 1990). The volcanic host rocks to these deposits comprise remnants of primitive ca. 1.93-1.91 Ga island arc complexes within the Savo Belt of the Lower Svecofennian domain (Kousa et al., 1994; Lahtinen, 1994). The Savo Belt is a collage of arc-related volcano-sedimentary rocks accreted to the southwestern margin of the Karelian Archean craton and stitched by a variety of syn- to post-orogenic granitoids (Lahtinen, 1994). The eastern margin of the Savo Belt presently represents a major northwest- to southeast-trending Proterozoic suture zone (Raahe-Ladoga Zone) between the Karelian Archean craton to the northeast and other Paleoproterozoic terranes to the southwest (Koistinen, 1981; Lahtinen, 1994).

General models for the evolution of the Svecofennian crust and orogen (Lahtinen, 1994; Lahtinen & Huhma, 1997; Nironen, 1997) indicate that a system of several arcs, basins and subduction zones were located to the present southwest of the Archean craton prior to 1.9 Ga. Volcanic and plutonic rocks of the Savo Belt, and associated volcanogenic massive sul-



Fig. I. Geological map of the south-central portion of the Vihanti–Pyhäsalmi district (V-P; see inset) indicating the location of major massive sulphide deposits and the grade and boundaries of metamorphic blocks (modified after Lundqvist et al., 1996).

phide deposits, are thought to have formed in this realm at the active margin of an older (ca. ~2.0 Ga) Paleoproterozoic nucleus (Lahtinen, 1994; Lahtinen & Huhma, 1997; Nironen, 1997). The geology and tectono-metamorphic history of this complex suture zone is described elsewhere (Lahtinen, 1994 and references therein).

2.2. Geology of massive sulphide deposits in the Vihanti–Pyhäsalmi district

Massive Zn-Cu sulphide deposits are typically hosted by rhyolites in bimodal volcanic successions which also include tholeiitic basalts, basaltic andesites (Lahtinen, 1994) and lesser occurrences of andesites (Kangasjärvi; Rasilainen, 1991). These mineralized bimodal successions occur discontinuously within the Archean-Proterozoic suture zone (Fig. 1). Massive sulphide deposits are spatially associated with, or enveloped by, altered rocks comprised of a variety of unusual metamorphic mineral assemblages (e.g. cordierite-orthoamphibole rocks), ranging from the amphibolite to granulite metamorphic facies. Regardless of metamorphic grade, however, host rocks preserve the geochemical signature of pre-metamorphic, seafloor-related hydrothermal alteration (Huhtala 1979; Mäki, 1986; Ekdahl, 1993; Lahtinen, 1994, Roberts et al., 2003).

Due to complex deformation, detailed stratigraphic relationships between and within isolated volcanic successions (e.g. way-up directions, unit correlations) are difficult to determine with certainty. Furthermore, the basement to these bimodal successions remains enigmatic. Most volcanic rocks appear to be underlain by voluminous, coarse-grained gneissic tonalites. Textural, geochemical and geochronological data indicate that these rocks are likely the subvolcanic equivalent of ore-hosting rhyolites (Lahtinen, 1994; Lahtinen & Huhma, 1997). However, contacts between gneissic tonalites and country rocks are rarely exposed or are strongly deformed; hence, the setting of these intrusions (i.e. host rocks, character of contact aureole, depth of emplacement) is poorly understood.

2.3. Host sequence to the Kangasjärvi deposit

The Kangasjärvi Zn-Cu deposit is situated in the southern extension of the Vihanti-Pyhäsalmi district (Fig. 1) and is one of several small satellite deposits (≤ 3 Mt) broadly similar in character (i.e. host rocks, pre-metamorphic alteration styles) to the much larger Pyhäsalmi Zn-Cu deposit (68 Mt). Mafic volcanic and volcaniclastic rocks dominate the map pattern in the vicinity of the deposit, while the deposit itself is hosted within an apparently thin and extensively altered "corridor" of felsic volcanic rocks (400 x 3000 m) (Fig. 2). Pervasive, premetamorphic, seafloor-related hydrothermal alteration of rocks in this corridor is expressed by metamorphic assemblages (e.g. biotite + sillimanite ± garnet ± cordierite ± staurolite) that are the prograde equivalent of chlorite ± sericite alteration. An apparently thick (-1.3 km) succession or mafic volcanic rocks appears to form the core of a megascopic F₂-F₃ fold complex, such that felsic volcanic rocks to the north of deposit (Fig. 2) may be a fold or thrust repetition of the felsic volcanic rocks that host ore. Felsic volcanic rocks in this region underwent pervasive, pre-metamorphic, low-temperature seawater-related Na-K metasomatism (Roberts, 2002) and hence are thought to represent the regional stratigraphic hanging wall to the Kangasjärvi Zn-Cu deposit. To the southwest of the deposit, the volcanic sequence is truncated by a large (map pattern area ~30 km²), coarse-grained to locally mylonitic gneissic tonalite, locally referred to as quartz-plagioclase gneiss (QPG). Contacts between the QPG and country rocks are strongly deformed (see below), indicating that this unit was broadly coeval with the bimodal volcanic sequence (i.e. pre-collisional). The largely bimodal sequence of volcanic rocks that host the Kangasjärvi Zn-Cu deposit (Fig. 2), bound to the south by the QPG and to the west and north by an interval of graphitic schist



Figl 2J Geological map of the Kangasjärvi area showing the location of the Kangasjärvi Zn-Cu deposit (see inset), the Inner Volcanic Sequence (IVS) and the Outer Volcanic Sequence (OVS) (modified after Luukas et al., 2002).



Fig. 3. Schematic southwest- to northeast-trending cross-section of the Kangasjärvi Zn-Cu deposit showing the distribution of rock types and the halo of pre-metamorphic alteration.

(see below), is referred to as the Inner Volcanic Sequence (IVS).

A separate sequence of volcanic and sedimentary rocks, separated from the IVS by an interval of sulphide and U-P-bearing graphitic schist, occurs outboard of the F_2 - F_3 fold complex and is referred to as the Outer Volcanic Sequence (OVS; see Fig. 2). The OVS is comprised of mafic volcanic rocks with irregular intervals of intermediate to felsic schists of uncertain volcanic, volcaniclastic or clastic origin, migmatitic metasediments (metagreywacke) and locally dolomitic calc-silicate rocks. Whether the sulphide- and U-P-bearing graphitic schist represents a structural or stratigraphic contact is presently unknown due to poor exposure.

2.4. Tectonic and metamorphic setting

Rocks in the Kangasjärvi region underwent a protracted tectono-thermal history as evidenced by multiple phases of deformation (D_1-D_4) and the occurrence of low-pressure – high-temperature (Korsman et al., 1984; Hölttä, 1988; Roberts, 2002) metamorphic mineral assemblages. Structural relationships coupled with the results of conventional geothermobarometry suggest that the peak of metamorphism, which re-

sulted in partial melting of favourable lithologies, culminated in the upper-amphibolite facies (~700°C, ~5 kbar) broadly synchronous with D_2 - D_3 (Roberts, 2002).

2.5. Deposit geology

The Kangasjärvi Zn-Cu deposit was mined by Outokumpu Oy for a total production of 86 000 tonnes grading 5.12 % Zn, 0.06 % Cu, 5.0 g/t Ag, 0.3 g/t Au and the total geologic resource of the deposit has been estimated at 300 000 metric tonnes (http://www.gsf.fi/ explor/zinc/kangasjarvi.htm). The deposit occurs within the hinge region of a tight D₃ antiform (F₃) with an axial plane that dips steeply to the southwest and plunges forty-five degrees to the southeast (Fig. 2 and 3). The ore is predominantly massive and is comprised of pyrite + sphalerite ± pyrrhotite ± chalcopyrite ± galena ± native gold and silver. Sulphides are strongly recrystallised and annealed, with pyrite locally forming cubic porphyroblasts within a sphalerite ± pyrrhotite ± chalcopyrite matrix. Gangue minerals include quartz, cordierite, anthophyllite, garnet, sillimanite and sericite, and are commonly included in the ore as angular to milled silicate aggregates.

The ore is hosted by felsic volcanic rocks comprised of a variety of metamorphic mineral assemblages that include combinations of quartz, plagioclase, sillimanite, biotite, garnet, potassium feldspar, muscovite, staurolite and orthoamphibole (Roberts, 2002). These rocks are interpreted to be the metamorphic equivalent of hydrothermally altered rhyolites (Rasilainen, 1991). A distinctive suite of mafic to intermediate orthoamphibole-cordierite rocks lies adjacent to this package and is a typical metamorphic assemblage proximal to massive sulphide deposits elsewhere in the district (Huhtala, 1979).

3. Rock types and textures

The following is a brief description of textures and stratigraphic relationships for rock types in the Kangasjärvi area (see also Roberts, 2002), with an emphasis on rocks proximal to the massive sulphide deposit in the IVS sequence. Multiple phases of ductile deformation have obliterated textures relating to stratigraphic facing direction and complicate the interpretation of unit dimensions (i.e. layer thickness) and the apparent repetition of units (i.e. primary repetition or fold repetition). In the following descriptions, therefore, emphasis is placed on textures observed in outcrops and rock-type associations (i.e. adjacent rock types) determined from outcrops and drill hole profiles.

3.1. Mafic volcanic rocks (IVS and OVS)

A variety of variably deformed volcanic and volcaniclastic textures are observed for mafic rocks in the Kangasjärvi area (IVS and OVS): 1) pillow lavas (Fig. 4a); 2) porphyritic lava flows or sills (uralite porphyries); 3) mafic volcaniclastic rocks (agglomerates; Fig. 4b-c); and 4) deformed mafic dykes hosted by basalts (IVS and OVS) and rhyolites (IVS). In both the IVS and OVS, the presence of pillow lavas and mafic volcaniclastic rocks with epidote-altered scoria suggest that mafic volcanism occurred in a subaqueous setting. Mafic volcaniclastic rocks in the IVS commonly contain andesite clasts (Fig. 4b), suggesting that basalts were erupted on an andesitic substrate (see below).

3.2. Intermediate rocks (IVS)

A geochemically distinct suite of andesites ("footwall andesites") occurs adjacent to, and interlayered with, mafic volcanic and volcaniclastic rocks in the altered sequence of rocks that is thought to represent the stratigraphic footwall to the deposit (see below). Footwall andesites in outcrop are rare and therefore their mode of occurrence (lavas, sills?) and contact relationships are unknown. Altered footwall andesites adjacent to altered basalts are comprised of similar metamorphic mineral assemblages (e.g. orthoamphibole–cordierite), hence



Fig. 4. Outcrop photographs of mafic and felsic volcanic rocks from the IVS and OVS. a) Outcrop of deformed pillow lava with irregular calc-silicate-bearing pillow rims and clasts (OVS?; 3463795E, 7038743N). b) Mafic volcaniclastic rock with sub-rounded, matrix-supported andesitic clasts (IVS; 3456640E, 7032050N). c) Well-developed stretching lineation described by sub-angular, matrix-supported felsic clasts in a mafic matrix (OVS; 3463420E, 7038400). d) Outcrop of massive, coherent quartz ± plagioclase-phyric high-silica rhyolite HSR_a (3456780E, 7033100N, sample K37). e) Weakly deformed, blocky rhyolite breccia (in-situ hyaloclastite?). Darker coloured clasts locally display a jigsaw-like fit with adjacent clasts (3455650E, 7034030N, sample K120). f) Strongly deformed rhyolite breccia composed of plagioclase ± quartz-phyric clasts within a more micaceous matrix (3456020E, 7034130N, sample K25).

both rock types were similarly affected by premetamorphic, seafloor hydrothermal alteration (see below).

3.3. High-Ti and esites and dacites (IVS)

A distinct suite of high-Ti andesites and dacites occurs interlayered with felsic volcanic rocks along strike from the deposit ("altered corridor"). Similar to footwall andesites, these rocks are only observed in drill core and therefore their morphology and contact relationship are unknown.

3.4. Felsic volcanic rocks (IVS)

Outcrops of felsic volcanic rocks are rarely observed in the vicinity of the now-filled Kangasjärvi open-pit mine. Geochemically similar rhyolites in the regional stratigraphic hanging wall consist of several distinctive textural varieties: 1) coherent, quartz-phyric rhyolite lavas or sills (Fig. 4d); 2) weakly deformed, blocky rhyolite breccias (Fig. 4e); 3) deformed, plagioclase-± quartz-phyric rhyolite breccias (Fig. 4f); and 4) deformed rhyolite layers or lobes within a tuffaceous matrix. Similar to mafic volcanic rocks, strong deformation and poor exposure hinders detailed reconstruction of volcanic facies. In this region, however, a general transition from coherent, sill-like, quartz-phyric rhyolites to blocky rhyolite breccias (in-situ hyaloclastite?) and tuffaceous rhyolites may represent a textural transition from the inner core of a coherent felsic volcanic dome complex to a brecciated carapace.

3.5. Gneissic tonalites (IVS)

Gneissic tonalites (QPG) are moderately to strongly foliated, coarse-grained quartz + plagioclase-bearing rocks with minor biotite ± hornblende. Outcrops of QPG are typically massive and contacts with country rocks are rarely exposed. Two different QPG - country rock contacts have been observed: 1) strongly foliated to locally mylonitic QPG in sharp contact with an intermediate, fine-grained, banded hornblende + plagioclase + quartz ± biotite-bearing rock, similar in bulk composition to the footwall andesite suite; and 2) thin 0.1-1.0 m wide, isoclinally folded dykes or sills of moderately siliceous quartz-phyric rhyolite interfingered with the QPG. In the first occurrence, a sharp contact between andesite and the QPG is interpreted to be a primary intrusive contact. Some banding within the andesite may represent primary layering, and hence it is inferred that the QPG intruded into the layered andesite. Irregular and highly deformed plagioclase + quartz + epidote + Ca-amphibole + magnetite + pyrite veins (2-5 cm wide) occur within the andesite near the contact between these two units. This style of vein-like alteration is typical of magmatic - hydrothermal alteration associated with felsic intrusive complexes (e.g. Galley, 1996), and

therefore may support the above hypothesis. In the latter instance, contacts between the QPG and rhyolite dykes or sills are clearly intrusive and deformed by the earliest recognizable phase of deformation (i.e. isoclinal F, folds).

4. Lithogeochemistry

4.1. Sampling protocols, analytical methods and terminology

Surface samples were obtained using a sledgehammer or a portable diamond mini-drill. Profiles and geochemical samples in the vicinity of the mine were obtained from diamond drill holes archived at the Geological Survey of Finland Loppi Core Facility. Large (2–4 kg) samples were collected and all weathered rinds and hammer marks were removed using a diamondedged saw.

Whole-rock major and minor elements were determined using x-ray fluorescence spectrometry (XRF). Trace elements were analysed by inductively coupled mass spectrometry (ICP-MS) after dissolution of the sample by hydrofluoric acid – perchloric acid treatment followed by a lithium metaborate/sodium perborate fusion of the insoluble residue. The estimated uncertainty is 1–5% for major elements and 3–10% for trace elements.

The term "least-altered" is used to identify rocks that contain no visible sulphides (<0.5 wt.% S), shear fabrics or metamorphic phases (prograde and retrograde) unlikely to occur in typical, unaltered metavolcanic rocks. Analytical data for representative least-altered lithologies are presented in Table 1 and a summary of selected high field strength element (HFSE) and rare earth element (REE) contents and their ratios are given in Table 2. The entire geochemical database is available from the authors upon request.

In the following section, immobile trace-element ratios and patterns are presented to examine aspects of the paleotectonic setting and petrogenesis of mafic and felsic igneous suites. Only samples that are deemed "least-altered"

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Suite	OVS B/BA	OVS B/BA	IVS B/BA	IVS B/BA	IVS B/BA	IVS FWA	IVS FWA	IVS Ti-A	IVS Ti-A	IVS Ti-D	IVS Ti-D
Sample ID	K144- 3	K60- 6	317- 172.05	028- 128.8	K137- 2	304- 217.4	027- 85.55	045- 77.6	046- 114.4	045- 47.5	047- 133
SiO,	51.3	55.7	50.7	52.1	54.3	63.6	62.8	58.6	59.1	65.9	66.6
TiO,	0.57	0.58	0.46	0.46	0.65	0.83	0.83	1.53	1.36	0.65	0.64
Al2O	18.27	18.73	17.05	16.96	16.18	15.21	15.30	13.59	13.97	13.12	13.04
FeO	11.19	8.25	10.51	11.85	11.03	7.49	8.78	14.22	13.69	10.02	9.29
MnO	0.16	0.20	0.26	0.32	0.27	0.23	0.26	0.13	0.11	0.15	0.14
MgO	4.35	3.46	7.17	5.92	4.65	2.20	2.10	2.49	1.33	2.27	2.34
CaO	9.98	7.75	11.21	8.94	9.57	4.28	4.25	4.52	4.56	1.22	2.93
Na ₂ O	3.02	4.42	1.87	2.81	2.77	5.20	4.81	3.61	4.28	4.68	4.66
K,Ô	1.05	0.76	0.67	0.53	0.43	0.57	0.50	0.80	1.28	1.95	0.28
P ₂ O ₅	0.07	0.14	0.07	0.08	0.11	0.41	0.38	0.45	0.32	0.09	0.09
S	0.000	0.008	0.037	0.069	0.000	0.057	0.077	0.226	0.368	0.117	0.161
Cr	n.a.	28	n.a.	126	n.a.	7	<5	n.a.	n.a.	n.a.	n.a.
Ni	12.7	15.3	47.7	52.5	29.1	<5.0	<5.0	< 5.0	<5.0	10.1	10.6
Co	33.7	22.5	41.8	41.6	26.4	10.9	10.6	15.2	12.2	9.3	8.9
Sc	39.7	33.6	43.1	42.7	37.6	29.2	30.3	35.1	33.1	26.5	26.8
V	344	275	251	237	270	40	36	27	67	137	139
Cu	94	24	37	44	33	13	6	44	15	22	31
Zn	89	84	117	255	171	129	98	47	31	73	99
Rb	17	11	11	8	3	7	6	20	25	148	6
Ba	596	363	175	180	135	345	322	256	303	271	100
Sr	675	603	196	241	195	186	166	105	111	92	70
Ga	16.4	n.a.	13	n.a.	16.1	n.a.	n.a.	19.5	17.9	16.9	16.1
Та	< 0.20	0.24	< 0.20	< 0.20	< 0.20	< 0.20	< 0.20	0.56	0.42	0.45	0.4
Nb	2.71	3.68	1.08	1.33	2.24	2.83	3.28	7.40	6.31	6.16	5.80
Hf	0.90	1.38	0.80	0.68	1.15	1.65	1.74	4.41	3.99	3.87	3.87
Zr	31.0	59.0	26.1	30.1	42.2	61.9	71.0	148	135	134	142
Y	15.2	15.4	9.90	11.2	14.9	26.8	28.6	56.4	42.6	28.2	38.5
Th	1.15	1.61	< 0.50	< 0.50	0.65	0.98	1.13	1.78	1.52	1.43	1.56
U	0.29	0.73	0.28	0.32	0.63	2.88	1.25	2.06	0.77	1.16	1.46
La	7.98	7.76	3.30	3.92	5.29	7.79	8.35	17.1	12.2	16.4	12.6
Ce	17.8	18.0	6.90	8.30	12.0	17.0	19.1	36.2	28.5	31.6	26.0
Pr	2.45	2.53	0.98	1.09	1.59	2.35	2.69	4.82	3.70	3.77	3.38
Nd	10.6	11.7	4.16	4.84	7.16	10.8	11.4	21.8	17.4	15.6	15.3
Sm	2.26	2.61	1.10	1.23	1.92	2.58	2.93	5.87	4.69	3.59	3.63
Eu	0.66	0.67	0.41	0.51	0.65	0.91	1.11	1.41	1.25	0.51	1.03
Gd	2.45	2.58	1.50	1.36	2.32	3.59	3.91	7.66	6.71	4.41	5.41
Tb	0.38	0.39	0.24	0.26	0.36	0.56	0.62	1.29	1.00	0.74	0.88
Dy	2.34	2.44	1.55	1.73	2.33	4.09	4.31	8.31	6.52	4.52	5.77
Ho	0.51	0.49	0.32	0.35	0.50	0.84	0.86	1.91	1.46	0.94	1.36
Er	1.55	1.46	1.10	1.16	1.41	2.69	2.80	5.88	4.36	3.06	4.04
Tm	0.23	0.23	0.17	0.18	0.24	0.39	0.41	0.83	0.64	0.50	0.58
Yb	1.51	1.51	1.10	1.23	1.65	2.64	2.89	5.51	4.49	3.40	4.12
Lu	0.22	0.24	0.16	0.19	0.24	0.43	0.42	0.92	0.66	0.60	0.6

Table I. Whole-rock geochemistry of representative least-altered rocks from Kangasjärvi

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Suite	IVS QPG	IVS Ra	IVS Ra	IVS Rb	IVS Rb	IVS Rc	IVS Rc	IVS Rd	IVS HSRa	IVS HSRa	IVS HSRb
Sample ID	K11- 1	K7- 4	319– 248.3	K45- 1	028- 157.2	043- 52.3	K134- 1	K87- 1	K125- 1	045- 14.1	047- 164.4
SiO ₂	73.5	69.9	68.7	73.0	74.1	75.4	77.8	75.7	79.3	77.4	78.3
TiO,	0.34	0.51	0.39	0.34	0.35	0.31	0.27	0.26	0.14	0.16	0.14
Al ₂ O ₃	13.21	14.25	15.36	12.41	12.72	12.03	10.90	12.74	10.55	11.63	12.08
FeO	3.50	4.98	4.79	4.29	4.28	4.00	3.57	2.95	2.66	3.42	1.88
MnO	0.06	0.08	0.12	0.17	0.12	0.12	0.05	0.08	0.13	0.11	0.07
MgO	1.20	1.33	1.40	1.89	0.73	1.23	1.14	0.54	1.06	0.55	0.33
CaO	2.10	3.10	4.36	3.55	2.31	1.96	1.30	2.44	2.14	1.32	1.88
Na,O	4.46	4.09	3.50	2.41	4.18	3.58	3.79	4.37	2.60	4.45	4.09
K,Õ	1.51	1.60	1.27	1.89	1.12	1.30	1.17	0.87	1.41	0.95	1.26
P_0_5	0.09	0.14	0.12	0.10	0.08	0.05	0.04	0.05	0.01	0.02	0.01
S	0.064	0.078	0.100	0.068	0.079	0.020	0.000	0.008	0.000	0.039	0.016
Cr	<5	<5	n.a.	<5	<5	n.a.	n.a.	<5	n.a.	n.a.	n.a.
Ni	<5.0	< 5.0	8.8	7.9	< 5.0	<5.0	<5.0	<5.0	<5.0	<5.0	<5.0
Co	4.5	7.5	8.5	9.0	3.4	4.4	2.5	2.2	< 0.5	1.9	< 0.5
Sc	13.0	17.7	10.8	17.5	15.2	12.3	11.9	10.9	12.0	9.31	4.50
V	19	25	47	59	14	6	2	7	1	12	5
Cu	11	16	41	25	7	6	9	<5	<5	5	<5
Zn	29	59	100	95	34	50	69	34	83	37	18
Rb	25	21	17	24	15	17	13	15	29	15	12
Ba	447	464	554	237	361	547	174	316	499	388	466
Sr	199	145	479	168	147	90	73	109	36	53	46
Ga	n.a.	n.a.	16.2	n.a.	n.a.	12.8	9.0	n.a.	12.7	13.1	11.1
Ta	0.46	0.41	0.36	0.35	0.41	0.40	0.39	0.34	0.58	0.77	0.71
Nb	7.61	6.54	5.25	5.80	6.63	7.30	6.18	5.37	6.96	10.1	8.90
Hf	3.66	3.51	2.57	2.59	3.52	4.15	3.31	3.21	4.18	5.56	4.55
Zr	147	138	98	107	143	144	122	131	162	187	129
Υ	31.2	27.1	17.3	28.8	29.0	32.8	28.2	23.2	32.4	36.1	42.1
Th	2.69	2.19	1.44	2.00	2.24	2.59	2.08	2.72	2.90	4.20	5.22
U	0.99	0.91	0.73	0.73	1.16	1.36	0.95	0.95	1.18	1.83	1.64
La	17.2	13.5	11.7	17.1	12.9	19.5	10.8	11.2	14.5	31.8	24.5
Ce	35.9	30.3	24.6	35.6	28.5	40.9	22.1	25.9	31.5	61.7	48.7
Pr	4.82	4.20	3.18	4.79	3.66	5.16	2.81	3.34	4.25	7.15	5.76
Nd	20.2	18.0	13.0	19.2	16.0	22.4	11.6	13.6	17.5	28.1	22.2
Sm	4.53	3.98	3.06	4.28	3.98	5.02	2.90	3.25	4.09	5.67	4.55
Eu	0.92	1.04	0.84	1.07	0.95	1.17	0.65	0.74	0.71	0.68	0.65
Gd	5.02	4.28	3.02	4.96	4.36	5.48	3.76	3.49	4.40	5.98	5.24
Tb	0.77	0.67	0.47	0.78	0.75	0.86	0.66	0.58	0.79	0.93	0.96
Dy	4.66	4.25	3.07	4.67	4.52	5.18	4.27	3.48	4.94	5.51	6.05
Ho	0.99	0.92	0.59	0.93	0.94	1.08	0.94	0.77	1.10	1.24	1.34
Er	2.98	2.64	1.80	3.03	2.96	3.44	2.73	2.32	3.37	3.67	4.27
Tm	0.45	0.40	0.26	0.43	0.43	0.53	0.43	0.37	0.51	0.65	0.62
Yb	3.28	2.80	1.84	3.10	3.00	3.84	3.04	2.66	3.74	4.49	4.52
Lu	0.47	0.44	0.26	0.46	0.43	0.54	0.46	0.39	0.56	0.75	0.69

Table | (cont.)

Major element oxides recalculated on an anhydrous basis and reported in wt. %.

FeO reported as total iron.

Trace elements reported in ppm.

Abbreviations: OVS = Outer Volcanic Sequence; IVS = Inner Volcanic Sequence; B/Ba = basalt/basaltic andesite; MD = mafic dyke; FWA = footwall andesite; Ti-A = high-Ti andesite; Ti-D = high-Ti dacite; QPG= quartz-plagioclase gneiss; R = rhyolite (a-d denotes rhyolite types); HSR = high-silica rhyolite (a-b denotes high-silica rhyolite types); n.a. = not analysed.

	e in a initiation of	כוברובת	וו אר מווח	INEL CUITERILS AI		יחוום בובוום			ar ivalige						
	IVS Basalts.	/Basaltic	Andesites	IVS Ve	olcanicla	stic	IVS Foot	wall Anc	lesites	IVS Hig	h-Ti An	desites	IVS Hi	gh-Ti D	acites
		n = 33			n = 13		<i>c</i>	<i>η</i> = 16			n = 9			n = 8	
	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ
TiO_2	0.29-0.96	0.69	± 0.14	0.61 - 1.07	0.85	± 0.11	0.79 - 1.30	0.91	± 0.15	0.99 - 1.53	1.25	± 0.18	0.51-0.78	0.65	± 0.08
Та	< 0.2			0.2 - 0.4 (6)			0.2-0.3 (9)			0.3 - 0.6	0.4	± 0.1	0.4 - 0.6	0.4	± 0.1
Nb	0.7–2.7	1.6	± 0.5	1.4-5.9	3.6	± 1.2	2.8-3.8	3.2	± 0.3	4.9-8.3	6.8	± 1.0	5.8-8.5	6.7	± 0.9
Zr	13.3 - 48.0	32.0	± 7.8	31.5-92.4	62.2	± 15.7	48.7-83.0	67.6	± 8.29	113.0-166.0	141.8	± 16.7	134.0-165.0	148.9	± 12.1
Ηf	0.6 - 1.2 (31)			0.9–2.4	1.6	± 0.4	1.5-2.3	1.8	± 0.2	3.0 - 4.8	4.0	± 0.6	3.9-4.7	4.1	± 0.3
Y	5.3-17.5	13.0	± 2.8	12.8–28.2	22.0	± 5.3	26.2-36.8	29.2	± 3.1	32.7-56.5	46.3	± 8.9	28.2–38.5	33.5	± 3.9
Th	0.5-0.8 (13)			0.6-2.1 (11)			0.9-1.2 (15)			1.2-2.3	1.7	± 0.3	1.4–2.1	1.8	± 0.2
La	2.0-16.2	4.4	± 2.3	3.8-13.8	7.9	± 3.0	7.4–12.9	8.6	± 1.5	9.0 - 17.1	13.6	± 2.6	12.6–22.1	16.1	± 3.5
Yb	0.6 - 2.0	1.3	± 0.3	1.2 - 3.1	2.2	± 0.6	2.5-3.9	3.0	± 0.4	3.7-5.9	4.8	± 0.8	3.3-4.1	3.7	± 0.3
Zr/Nb	15.2–26.0	20.6	± 2.5	14.9–22.5	17.9	± 2.4	17.0-22.4	21.3	± 1.3	19.0 - 23.2	21.0	± 1.5	19.5–24.5	22.4	± 1.5
Zr/Y	1.4 - 3.3	2.5	± 0.4	2.3-4.8	2.9	± 0.7	1.9–2.5	2.3	± 0.2	2.5-4.1	3.1	± 0.5	3.7-5.7	4.5	± 0.6
Zr/Ti	0.002-0.007	0.005	± 0.001	0.005-0.012	0.007	± 0.002	0.004 - 0.009	0.008	± 0.001	0.010 - 0.014	0.011	± 0.002	0.021 - 0.029	0.023	± 0.002
Zr/Sc	0.3 - 1.1	0.7	± 0.2	0.7-4.5	2.0	± 0.9	1.2 - 2.4	2.1	± 0.3	3.0-5.8	4.3	± 0.9	5.1-9.5	6.9	± 1.8
Nb/Y	0.1 - 0.2	0.1	± 0.0	0.1 - 0.3	0.2	± 0.1	0.1 - 0.1	0.1	± 0.0	0.1 - 0.2	0.1	± 0.0	0.2 - 0.3	0.2	± 0.0
Nb/Ta				13.8 - 18.4 (6)			14.1 - 17.3 (9)			13.2–16.4	15.2	± 1.0	13.5–16.3	14.9	± 1.0
Ti/V	12-26	21.5	± 3.1	22-107	47.9	± 25.6	41-232	168.0	± 52.7	36-578	276.8	± 189.3	46 - 100	65.5	± 24.3
Ti/Sc	62-232	154	± 33	139–383	261	± 62	239-324	283	± 24	276-447	377	± 58	239-406	300	± 81
La/Ybn^2	1.3-5.8	2.3	± 0.8	1.8 - 7.0	2.7	± 1.4	1.7 - 3.4	2.1	± 0.4	1.6-3.2	2.0	± 0.5	2.2-4.1	3.1	± 0.6
Th/Nbn^3	2.0-3.5 (13)			1.6–3.8 (11)			2.1-3.2 (15)			1.8-2.7	2.1	± 0.3	1.9–2.5	2.2	± 0.2
Eu/Eu*3	0.69 - 1.30	1.06	± 0.14	0.79 - 1.28	0.96	± 0.15	0.84 - 1.41	1.03	± 0.14	0.64 - 1.00	0.80	± 0.12	0.39 -0.72	0.57	± 0.12
Ti/Ti*3	0.3-0.5	0.40	± 0.06	0.2 - 0.4	0.29	± 0.06	0.2-0.5	0.28	± 0.06	0.2 - 0.3	0.22	± 0.03	0.1 - 0.2	0.13	± 0.02

Table 2. Summary of selected HFSE and REE contents and immobile element ratios for lithologies at Kangasjärvi

Table 2 (cont.)

				IVS Rhyo	dacite/I	Shyolite									
	IVS Quartz	-Plagiocl.	ase Gneiss		(R)		IVS R	hyolite (]	R,)	IVS R	hyolite	(R _c)	IVS R	hyolite	(\mathbf{R}_{d})
		n = 21			n = 8			n = 34			n = 27			n = 11	
	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	lσ	range (n^1)	mean	Ισ
TiO_2	0.20 - 0.38	0.32	± 0.05	0.38-0.51	0.45	± 0.05	0.26-0.43	0.36	± 0.03	0.25-0.47	0.32	± 0.04	0.22-0.39	0.28	± 0.05
Та	0.4 - 0.6	0.4	± 0.1	0.3 - 0.4	0.3	± 0.1	0.3 - 1.7	0.4	± 0.2	0.3 - 0.5	0.4	± 0.1	0.3-0.5	0.4	± 0.1
Nb	6.1-9.0	7.3	± 0.6	3.9-6.5	5.3	± 0.9	4.7-8.0	6.4	± 0.6	4.8-9.0	6.7	± 0.8	5.4-9.2	7.2	± 1.2
Zr	68-309	149.4	± 48.1	90.7-138.0	114.1	± 20.1	102-171	132.1	± 13.4	106-229	143.7	± 22.7	130-214	161.5	± 26.8
Ηf	2.2-6.5	3.7	± 0.9	2.3-3.5	2.9	± 0.4	2.6-4.1	3.4	± 0.4	2.6-5.3	3.6	± 0.6	3.2-6.0	4.0	± 0.8
Y	22.1–36.2	30.6	± 3.7	17.3–39.7	28.4	± 7.5	16.9–36.3	28.9	± 3.6	19.1-44.3	33.4	± 5.8	23.2–59.3	35.1	± 11.3
Th	2.1 - 3.4	2.6	± 0.4	1.4-2.2	1.9	± 0.3	1.8–2.8	2.3	± 0.2	1.6–2.8	2.3	± 0.3	2.1 - 3.0	2.5	± 0.3
La	12.5-21.9	17.1	± 3.1	9.8–13.7	12.7	± 1.4	2.8–21.5	14.7	± 3.4	8.2-20.8	14.8	± 2.7	11.2-18.6	15.1	± 2.2
Yb	2.5-3.9	3.1	± 0.3	1.8-4.1	3.0	± 0.7	2.0-3.8	3.0	± 0.3	2.2-4.7	3.6	± 0.6	2.6-6.3	3.7	± 1.2
Zr/Nb	8.8-38.0	20.3	± 5.6	18.4–24.0	21.4	± 2.2	17.6–27.7	20.8	± 1.9	19.3-25.6	21.6	± 1.8	19.2–29.6	22.7	± 2.8
Z_{r}/Y	1.9-12.9	5.0	± 2.1	3.1-5.7	4.2	± 0.9	3.5-7.6	4.6	± 0.8	3.0-6.2	4.4	± 0.8	3.2-6.2	4.9	± 1.1
Zr/Ti	0.027-0.138	0.049	± 0.026	0.021-0.028	0.025	± 0.002	0.031 - 0.040	0.037	± 0.003	0.040-0.049	0.045	± 0.003	0.050-0.073	0.058	± 0.008
Zr/Sc	6.8–36.2	13.2	± 6.8	4.3–9.1	7.0	± 1.6	6.1–15.3	9.6	± 1.5	6.2-13.5	10.3	± 1.9	9.6–15.7	13.2	± 2.1
Nb/Y	0.2 - 0.3	0.2	± 0.0	0.1 - 0.3	0.2	± 0.1	0.2-0.4	0.2	± 0.0	0.2-0.3	0.2	± 0.0	0.1 - 0.3	0.2	± 0.1
Nb/Ta	14.0–18.5	16.4	± 1.1	14.6–18.5	16.1	± 1.3	3.9–21.3	16.1	± 2.6	13.8 - 18.6	16.3	± 1.2	15.1–18.5	16.4	± 1.1
Ti/V	105-576	189	± 97	31-422	213	± 157	58-1502	265	± 245	68–5028	1399	± 1509	78-4088	974	± 1286
Ti/Sc	234–308	271	± 20	179–362	277	± 54	195–378	263	± 38	147-290	230	± 38	169–280	227	± 30
La/Ybn^2	3.1-5.0	3.9	± 0.6	2.1-4.6	3.1	± 0.8	0.6-5.5	3.5	± 0.8	2.1-4.2	3.0	± 0.6	1.9-4.8	3.1	± 0.9
Th/Nbn^3	2.5–3.8	3.0	± 0.3	2.3–3.6	3.0	± 0.4	2.4 - 3.9	3.0	± 0.3	2.5-3.6	2.9	± 0.2	2.1-4.2	3.1	± 0.6
Eu/Eu*3	0.52-0.81	0.62	± 0.06	0.66–0.88	0.75	± 0.08	0.37 - 1.45	0.71	± 0.16	0.57-0.79	0.65	± 0.05	0.45-0.78	0.61	± 0.11
Ti/Ti*3	0.0 - 0.1	0.07	± 0.02	0.1 - 0.1	0.11	± 0.01	0.1 - 0.2	0.09	± 0.03	0.1 - 0.1	0.07	± 0.01	0.0 - 0.1	0.06	± 0.01

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	IVS High-Silic	a Rhyol	ite (HSR _a)	IVS High-Silica	Rhyoli	te (HSR_b)	OVS Basalts/Ba	saltic Aı	ndesites	OVS Interm	ediate 3	Schists	OVS Fe	lsic Sch	sts
	n	= 18		= <i>u</i>	: 11		<i>u</i> =	21		и	= 3		и	= 3	
	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	Ισ	range (n^1)	mean	lσ	range (n^1)	mean	Ισ
IiO_2	0.11 - 0.19	0.13	± 0.02	0.09-0.27	0.13	± 0.05	0.40 - 1.11	0.60	± 0.17	0.59-0.74	0.64	± 0.09	0.10 - 0.18	0.14	± 0.04
Га	0.4 - 0.6	0.5	± 0.0	0.5-0.8	0.7	± 0.1	0.2 - 0.3 (10)			0.4 - 0.7	0.6	± 0.2	0.4 - 0.7	0.6	± 0.2
Nb	5.8-8.6	7.6	± 0.9	5.1-9.5	8.3	± 1.3	0.6-6.2	3.1	± 1.3	6.1 - 10.1	8.3	± 2.0	5.0-8.3	6.6	± 1.7
Zr	135-201	174.5	± 18.5	86.5-129.0	115.1	± 13.0	13.7–93.1	45.3	± 21.2	92.7-133.0	117.2	± 21.5	99.9–177.0	134.6	± 39.1
Hf	3.5-5.8	4.8	± 0.6	3.0-4.6	4.0	± 0.4	0.6-2.3 (20)			2.3-3.3	2.9	± 0.6	2.6-4.8	3.6	± 1.1
Y	28.3-46.1	40.4	± 5.3	14.4 - 42.1	34.3	± 8.1	7.2–26.4	15.3	± 4.3	18.6–25.3	21.9	± 3.4	5.3-15.6	9.0	± 5.8
Th	3.0-4.3	3.6	± 0.4	3.3-5.8	4.9	± 0.7	0.6-1.7 (19)			4.1 - 11.1	7.3	± 3.5	14.3-16.3	15.2	± 1.0
La	14.4–21.5	18.8	± 2.4	7.1–29.4	21.3	± 6.6	2.7-11.1	7.7	± 2.2	16.6-33.2	24.2	± 8.4	20.3 - 34.1	26.4	± 7.0
Yb	3.3-5.0	4.4	± 0.5	1.9-4.8	3.9	± 0.9	0.8–2.7	1.5	± 0.5	1.8-2.4	2.2	± 0.3	0.4 - 1.4	0.8	± 0.6
Zr/Nb	21.0–25.7	22.9	± 1.3	10.5-17.1	14.1	± 1.9	8.8-23.2	15.2	± 3.8	10.8-21.8	15.0	± 5.9	15.2–26.8	20.7	± 5.8
Z_{r}/Y	3.9-4.8	4.3	± 0.3	2.6-6.0	3.6	± 1.0	1.6 - 4.6	2.9	± 0.9	4.2–7.2	5.5	± 1.5	8.1-33.2	19.4	± 12.7
Zr/Ti	0.087-0.179	0.142	± 0.029	0.045-0.114	0.095	± 0.019	0.003-0.013	0.007	± 0.003	0.015-0.021	0.018	± 0.003	0.085 - 0.102	0.095	± 0.009
Zr/Sc	15.1–18.9	17.7	± 1.1	16.1 - 42.9	31.0	± 8.1	0.3 - 3.1	1.3	± 0.8	4.0-6.9	5.7	± 1.5	37.2-56.0	46.0	± 9.4
Nb/Y	0.2-0.2	0.2	± 0.0	0.2 - 0.4	0.3	± 0.1	0.1 - 0.4	0.2	± 0.1	0.3 - 0.4	0.4	± 0.0	0.5-1.2	0.9	± 0.4
Nb/Ta	13.1 - 16.7	15.4	± 1.1	10.0 - 13.6	11.7	± 1.1	7.1–20.6 (10)			13.2–14.5	13.9	± 0.7	11.5-11.9	11.7	± 0.3
Ti/V	57-489	205	± 152	62-493	245.7	± 142.2	16.1–52.4	23.8	± 8.7	33.5-61.2	45.4	± 14.3	160–239	187.2	± 44.6
Ti/Sc	102-207	130	± 31	247-415	329	± 57	86-418	169	±74	260.7–381.7	312	± 63	437–565	481	± 73
La/ Ybn^2	2.5–3.7	3.1	± 0.3	1.6–5.8	3.9	± 1.1	1.4-5.2	3.7	± 0.9	6.6–10.0	7.9	± 1.8	12.4–56.9	34.4	± 22.2
$Th/$ Nb n^3	3.7-4.4	4.0	± 0.2	3.8–5.9	5.0	± 0.6	1.3-4.3 (19)			5.7-9.2	7.1	± 1.9	14.4-27.6	20.4	± 6.7
Eu/ Eu*³	0.36-0.52	0.43	± 0.05	0.20-0.74	0.34	± 0.15	0.76-1.14	0.88	± 0.10	0.6–0.8	0.70	± 0.14	0.37-0.42	0.39	± 0.03
Ti/ Ti*3	0.0-0.0	0.03	± 0.01	0.0-0.1	0.03	± 0.02	0.18 - 0.40	0.26	± 0.06	0.1 - 0.2	0.15	± 0.05	0.0 - 0.1	0.05	± 0.01

TiO₂ reported as wt. %.

Trace elements reported in ppm.

Trace element ranges and means are reported for *n* samples within each rock suite. ¹ (*n*) denotes number of samples used to define range whereby one or more trace element was below detection limits or not analysed; ² denotes trace element ratios normalized to CI chondrite (Sun and McDonough, 1989); ³ denotes trace element ratios normalized to primitive mantle (pm) (Sun and McDonough, 1989); Eu/Eu^{*} = Eu_n/ (Gd_m+Sm_{pm})^{0.5}; Ti/Ti^{*} = 0.5Ti_{pm}/(Gd_m+Sm_{pm}).



Fig. 5. Modified Zr/Ti-Nb/Y plot (Pearce, 1996) of Winchester and Floyd (1977) showing representative lithologies from the (a) IVS and (b) OVS. Rock type abbreviations are as follows: HSR,; high-silica rhyolite b; HSR : high-silica rhyolite a; IVS Rhy: IVS rhyolites (including R_a to R_d varieties); QPG: quartz-plagioclase gneiss (gneissic tonalite-trondhjemite); Ti-D: high-Ti dacites; Ti-A: high-Ti andesites; FWA: footwall andesites; IVS B/Ba: IVS basalts and basaltic andesites; Int. Dyke; intermediate dykes; Fel. Schist; felsic schists; Int. Schist: intermediate schists; OVS B/Ba: OVS basalts and basaltic andesites.

were employed when interpreting trace-element ratios or fractionation patterns unless otherwise stated. The major elements Si, Fe, Mn, Mg, Ca, Na, K and the metals Cu, Pb, Zn, Ag, Sn, As, Tl are generally considered mobile in seafloor hydrothermal systems owing to a variety of waterrock reactions involving the breakdown and replacement of primary igneous phases, infill or addition of new mineral phases and the precipitation of sulphides and native metals. The REE, except for Eu, are generally considered immobile, although they may become mobile, especially the light rare earth elements (LREE), during intense alteration (Pearce, 1996). Due to low ionic potential, the low field strength elements (LFSE: Ba, Rb, Cs, Sr) are considered mobile, whereas the high field strength elements (HFSE: Nb, Th, Zr, Hf, Y, Al, Ti) have high ionic potentials and are considered immobile (see review by Jenner, 1996).

4.2. Volcanic and intrusive rocks of the IVS and OVS sequence

Representative least-altered samples from the IVS sequence include subalkaline basalts, basaltic andesites, andesites, high-Ti andesites and dacites, rhyodacites – rhyolites, high-silica rhyolites and trondhjemitic to tonalitic gneisses (Fig. 5a). The largely mafic-dominated OVS is comprised of a suite of subalkaline basalts, basaltic andesites, intermediate schists and alkaline felsic schists (Fig. 5b) and lesser carbonaceous units (not shown). The geochemistry of intermediate and felsic schists from the OVS is not presented in this paper other than to note that their chemistry is distinct from IVS andesites and rhyolites (Table 2).



Fig. 6. Immobile-element binary plots for mafic to intermediate volcanic rocks from the IVS and OVS. See Fig. 5 for key to rock-type abbreviations.

4.2.1. IVS and OVS mafic volcanic rocks

Mafic volcanic rocks from the inner and outer volcanic sequence are characterized by low HFSE (Nb, Ta, Zr, Hf, Y) abundances (Table 2) and share similar incompatible-element ratios (e.g. avg. $Zr/Y_{VS} = 2.5$; $Zr/Y_{OVS} = 2.9$) (Fig. 6; Table 2). Both sequences are depleted in HFSE and HREE relative to MORB, are weakly LREE-enriched (avg. La/Ybn _{IVS} = 2.3; La/ Ybn $_{OVS}$ = 3.7) and have elevated LILE relative to HFSE (range of Th/Nb n_{IVS} = 2.0-3.5; Th/ Nb n_{OVS} = 1.3-4.3; note that less than ~50 % of IVS samples and ~90 % of OVS samples have Th abundances above the detection limit of 0.5 ppm) ("n" indicates normalized ratios; see Table 2 for references). These features translate into negatively sloping MORB-normalized incompatible-element profiles with troughs at Nb, Zr-Hf, \pm Ti (Fig. 7a–b). Although both the IVS and OVS suites are similar, mafic volcanic rocks of the OVS have consistently higher LILE (Sr, Ba, Rb, Th) and LREE abundances.



Fig. 7. MORB-normalized profiles for a) IVS basalts, b) OVS basalts, c) IVS footwall andesites, and d) high-Ti andesites and dacites. N-MORB values from Sun and McDonough (1989).



Fig. 8. Major element variation diagrams for mafic to intermediate volcanic rocks from the IVS and OVS. Fields in a) from Le Bas et al. (1986), b) Gill (1981) and c) Miyashiro (1974).

Least-altered mafic volcanic rocks of both suites have major element compositions typical of low- to medium-K, subalkaline island arc tholeiites (Table 1). Within the sample suite, least-altered IVS basalts range in composition from primitive basalts to basaltic andesites, whereas basalts from the OVS are generally more evolved (Fig. 8a). Least-altered IVS mafic volcanic rocks typically have $K_0 O < 0.5$ wt.%, whereas OVS samples have slightly higher K₂O values (~ 1 wt.% K₂O; Fig. 8b). Scatter of K₂O in the selected least-altered sample suite is indicative of the mobility of this element. Both the IVS and OVS suites plot within the tholeiitic field on the discriminate diagram of Miyashiro (1974) (Fig. 8c). Harker-type diagrams (Fig. 9) show fractionation trends typical of island arc tholeiites (e.g. Perfit et al., 1980). Lower FeO, TiO₂ and higher Al₂O₃, CaO and Sr (Rb, Ba) at a given MgO value, broadly distinguish the OVS from the IVS suite. High Al₂O₃ in OVS basalts does not appear to be a function of primary magma composition, but is restricted to fractionated end-members. For both suites, Ni and Cr values are low (Ni < 100 ppm; Cr <200 ppm) and, as shown for Ni, decrease with the fractionation of olivine (Fig. 9h). Several high Ni and Cr values may be partly due to olivine cumulation.

4.2.2. IVS Footwall andesites

In comparison to mafic volcanic rocks, the bulk composition of footwall andesites is relatively uniform and unfractionated (Fig. 9). Least-altered andesites have low HFSE (Table 2) and incompatible-element ratios similar to IVS (and OVS) mafic volcanic rocks (Fig. 6; Table 2). Footwall andesites have elevated LILE relative to HFSE (range of Th/Nb*n* = 2.1-3.2; Th < detection limit for one sample) and are weakly LREE enriched (avg. La/Yb*n* = 2.1). These rocks have MORB-normalized incompatible-element profiles similar to IVS basalts at higher relative HFSE and REE abundances (Fig. 7c).

Silica contents for least-altered footwall andesites range between 61-63 wt.% SiO₂, K₂O is generally < 0.5 wt.% and Na₂O contents are high (4.8–5.4 wt.% Na₂O). Andesites plot within the tholeiitic field in Figure 8c and are distinctive due to their relatively high P_2O_5 content (-0.4 wt.%). Similar incompatible-element ratios and low Sr (and Ba, Rb) contents suggest



Fig. 9. Major- and trace-element Harker-type diagrams for mafic to intermediate volcanic rocks from the IVS and OVS. Symbols as in Fig. 8.

that footwall andesites may have a similar parentage to IVS rather than OVS basalts. Harker diagrams (Fig. 9) indicate that footwall andesites are depleted in FeO, V and Sc relative to IVS basalts, whereas TiO_2 is moderately enriched.

4.2.3. IVS High-Ti andesites and dacites

The major- and trace-element chemistry of high-Ti andesites and dacites in our sample suite is quite variable (Table 1 and 2) and may partly vary as a function of pre-metamorphic alteration as many of these rocks contain variable proportions of biotite, garnet, cummingtonite and orthoamphibole. All high-Ti andesites (TiO₂ 1.0-1.5 wt.%) share similar incompatible HFSE ratios (avg. Zr/Nb = 21.0; avg. Zr/Y = 3.1) as IVS basalts and footwall andesites and have Zr/Ti ratios (~0.01–0.03) between those of mafic and felsic volcanic rocks (0.002-0.007 and 0.02-0.12, respectively). High-Ti andesites have major element abundances similar to footwall andesites; however, apparently least-altered varieties (hornblende + plagioclase + quartz + biotite + Fe-Ti oxide) are distinguished from these rocks by their higher FeO (11-14 wt.%) and TiO₂ (1.3-1.6 wt.%) contents. Least-altered high-Ti dacites (-65 wt.% SiO₂) have similar incompatible-element ratios as high-Ti andesites and are also characterized by relatively high FeO (9-10 wt.%,) and TiO, (~0.65 wt.%) abundances. MORB-normalized incompatible-element patterns for high-Ti andesites and dacites are similar in profile to those of IVS basalts and footwall andesites (i.e. arc signature) at higher relative REE and HFSE contents (Fig. 7d). However, profiles for dacites are characterized by prominent negative Eu and Ti anomalies.

4.2.4. IVS Rhyodacites-rhyolites

No systematic major- or trace-element distinction between rhyolites that preserve coherent, brecciated or "tuffaceous" textures is observed, possibly due to the ambiguity of these textures in outcrop and drill core. The only exception to this observation is a suite of high-silica, quartzphyric rhyolites (HSR_a, see below) that have coherent textures (sills or lavas?).

In Figure 10a, Zr/Ti ratios are used to highlight four rhyolite suites (R_a , R_b , R_c , R_d) and two high-silica rhyolite suites (HSR_a and HSR_b) from the IVS sequence. In this plot, Zr and Ti values are normalized (divided by Yb) to remove the effects of mass/volume changes. Common Zr/Ti trajectories for rhyolites may arise due to variable temperatures of fusion of



Fig. 10. Immobile-element diagrams for felsic volcanic rocks from the IVS. a) Yb-normalized Zr/Ti plot highlights Zr-Ti arrays subdivided into $R_{a,b,c,d}$ rhyolite types and HSR_a and HSR_b high-silica rhyolites (plot after Lentz, 1998). b) Zr/Ti-Y/Ti plot indicating the magmatic affinity of IVS rhyolites, high-silica rhyolites and tonalitic gneisses (QPG) (from Lentz, 1998, 1999). c) Nb-Y discriminate diagram of Pearce et al. (1984). See Fig. 5 for key to rock-type abbreviations.

a similar source (Lentz, 1999), whereas the observed range of Zr/Ti ratios may reflect mixing processes (magmatic or sedimentary?). The recognition of four (R_a to R_d) rhyolite types is supported by the following observations: 1) R_a



Fig. I I. Primitive mantle-normalized trace-element plots for a) least-altered IVS rhyolites (R_{a-d}), b) least-altered highsilica rhyolites HSR_a, c) least-altered high-silica rhyolites HSR_b and d) least altered quartz-plagioclase gneisses (QPG). Primitive mantle values from Sun and McDonough (1989).

rhyolites are spatially restricted to the altered footwall sequence; 2) R_b rhyolites are typical of, but not restricted to, the altered footwall – altered corridor and are commonly composed of metamorphosed chlorite ± sericite alteration assemblages; and 3) R_c rhyolites occur mainly within the regional stratigraphic hanging wall sequence and are commonly enriched in K, depleted in Na and variably silicified (Roberts, 2002). However, these suites have similar and overlapping bulk compositions and therefore R_a through to R_d (Zr/Ti = ~0.02-0.12) rhyolites are treated as a geochemically similar entity (IVS rhyolites).

IVS rhyolites have average Zr/Nb ratios (avg. Zr/Nb = 21.4) identical to IVS basalts. Immobile-element ratios (e.g. Zr/Y; Fig. 10b) indicate that rhyolites are tholeiitic to transitional calcalkaline in character (Zr/Y = 3.0-7.6; La/Yb*n* = 1.9-5.5; Th/Yb = 0.37-1.10; discriminating ranges from Barrett & MacLean, 1999). These rocks have I-type, island arc affinities (Fig. 10c), relatively low to moderate Ti/Sc ratios (147-378; crustal Ti/Sc values = 250-450; Wedepohl, 1995) and mantle-like Nb/Ta ratios (avg. Nb/ Ta = 16.2; mantle Nb/Ta values = 17.5, Taylor & McLennan, 1985).

Primitive mantle-normalized incompatibleelement profiles for IVS rhyolites (Fig. 11a) indicate that these rocks are moderately enriched in LILE over HFSE, weakly enriched in LREE relative to HREE, contain moderate HFSE abundances, flat HREE profiles, weak negative Nb (avg. Th/Nbn = 3.0) and Eu (avg. Eu/Eu* = 0.68) anomalies and moderate negative Ti (avg. Ti/Ti*= 0.08), Sc and V anomalies. For samples with higher Zr/Ti ratios (lower Ti), the topology of normalized profiles remains similar although negative Ti, Sc and V anomalies become more pronounced and may indicate the onset of magnetite fractionation. Several rhyolite samples, particular those in the R_suite, have anomalously low V abundances, which may be in response to this process.

Least-altered rhyolites have rhyodacitic (R_a) to rhyolitic silica contents (R_b , R_c , R_d), Al_2O_3 – 12-13 wt.%, $K_2O < 1.5$ wt.% and FeO/MgO – 2-4. Least-altered rhyolites are relatively sodic (Na_2O/K_2O ratios ~2.5-3.5), although these ratios are quite variable due to background albitic and sericitic alteration.

4.2.5. IVS High-silica rhyolites (HSR)

By definition, high-silica rhyolites in the sample suite have relatively high silica contents (~75-78 wt.% SiO₂, although some high values may be a result of pre-metamorphic silicification), and are distinctive due to relatively low TiO₂ (0.11-0.19), high HFSE contents (e.g. Zr 135-201 ppm) and hence higher incompatible to compatible element ratios (e.g. Zr/Ti 0.14-0.30) compared to IVS rhyolites. Two high-silica rhyolite suites are distinguished (HSR, and HSR_b). Similar to IVS rhyolites, high-silica rhyolites "HSR," have Zr/Nb ~23 and tholeiitic to transitional Zr/Y (3.9-4.8), La/Ybn (2.5-3.7) and Th/Yb (0.7–1.0) ratios. HSR have similar Nb/Ta values (avg. = 15.4) as IVS rhyolites although Ti/Sc ratios are distinctly lower (102-207). Primitive mantle normalized incompatible-element plots for HSR_a samples (Fig. 11b) are similar in profile to IVS rhyolites at slightly higher absolute HFSE and REE contents, whereas negative Nb (avg. Nb/Thn = 4.0), Eu (avg. Eu/Eu* = 0.49), Ti (avg. Ti/Ti* = 0.01), Sc and V anomalies are greater.

A second distinct high-silica, low TiO₂ (Zr/ Ti = 0.14–0.20) rhyolite "HSR_b" is distinguished from HSR_a by its lower Zr contents (87–129 ppm), and hence lower Zr/Nb (~14) and Zr/Th (~24) ratios. Similar to HSR_a, these rocks have a similar range of tholeiitic to transitional La/Ybn (1.6–5.8) and Th/Yb (1.1–1.7) ratios. HSR_b rhyolites have Ti/Sc ratios (247– 415) similar to IVS rhyolites, yet have distinctly lower crustal-like Nb/Ta ratios (avg. Nb/Ta = 11.7). Primitive mantle normalized incompatible-element profiles for HSR_{b} are similar in profile to those of HSR_{a} (and hence IVS rhyolites), although relative Zr contents are lower and negative Nb (avg. Nb/Thn = 5.0), Eu (avg. Eu/Eu* = 0.34), Ti (avg. Ti/Ti* = 0.01), Sc and V anomalies are on average greater than IVS rhyolites and HSR_a (Fig. 11c).

4.2.6. IVS Quartz-plagioclase gneiss (QPG)

QPG samples have similar immobile element systematics and least-altered major element abundances to IVS rhyolites with several evolved samples similar to HSR. For example, in comparison to IVS rhyolites, the QPG suite has similar Zr/Nb (avg. Zr/Nb = 20.3) ratios and overlapping Zr/Ti (0.03-0.11), Zr/Y (1.9-12.9), La/Ybn (3.1-5.0) and Th/Yb (0.7-1.1) ratios (Table 2). Figure 10c shows that these rocks have I-type affinities and plot within the volcanic arc granite field of Pearce et al. (1984). Primitive mantle-normalized incompatible-element plots are similar in profile to IVS rhyolites (avg. Th/Nbn = 3.0; avg. Eu/Eu* = 0.62; avg. $Ti/Ti^* = 0.07$) (Fig. 11d). This suite also has mantle-like Nb/Ta ratios (avg. = 16.4) and moderate Ti/Sc ratios (234-308). Least-altered QPG have whole-rock δ^{18} O values of ~8.0 ‰ (Roberts, 2002), further suggesting a mantledominated signature.

Quartz-plagioclase gneisses range in composition from tonalitic to trondhjemitic. Similar tonalites-trondhjemite gneisses have been reported from elsewhere in the district (Lahtinen, 1994), of which the Kangasjärvi QPG broadly resembles the Rastinpää high-Y suite of Lahtinen (1994). Lahtinen (1994) has shown that the composition of the Rastinpää high-Y tonalite varies in part due to fractional crystallization of plagioclase, a ferromagnesian phase (pyroxene) and magnetite. In Figure 12, Harker diagrams indicate a similar fractionating assemblage for the variety of QPG compositions at Kangasjärvi. FeO, MgO, TiO₂, V (Fig. 12a-d) and Sc (not shown) decrease with increasing degree of fractionation and indicate the involve-



Fig. 12. Major- and trace-element Harker-type diagrams for quartz-plagioclase gneisses (QPG). Arrows indicate igneous fractionation trends (FC) and pre-metamorphic alteration trends (Alt.).

ment of a ferromagnesian phase and a Ti-bearing phase (titanite, rutile, magnetite, ilmenite) during fractionation. Over the fractionation interval, the V content (36–4 ppm) decreases sharply compared to Sc (14–8 ppm). Due to the high partition coefficient of V relative to Sc for magnetite, these trends indicate the importance of magnetite fractionation over pyroxene (see Lahtinen, 1994). CaO and Sr, as well as Ba (not shown) also decrease systematically and indicate fractionation of plagioclase (Fig. 12e–f). K₂O does not show any trend with increasing fractionation (Fig. 12g) and indicates that biotite was not a significant fractionating phase. Note that several samples describe trends of FeO + MgO gain and CaO + Sr loss, which may indicate a slight degree of pre-metamorphic alteration (Fig. 12a–b and e–f, respectively).

With increasing degree of fractionation Y contents remain the same or increase slightly (Fig. 12h), indicating that hornblende fractionation was negligible. As magnetite and hornblende fractionation are characteristic of hydrous, lower-temperature, oxidized, calc-alkaline magmas (Lentz, 1998), the contrasting behaviour of V and Y in the QPG suite is indicative of a dry, moderate- to high-temperature and moderate fO_2 magma conditions transitional between tholeiitic and calc-alkaline systems.

5. Chemostratigraphy

In order to constrain relative stratigraphy in the local vicinity of the ore deposit, immobile-element ratios (Zr/Ti) are plotted on a chemostratigraphic profile through the ore horizon (Fig. 13). Sections through the host sequence are subdivided into three sub-sections (A, B, and C). Section A contains a complex association of variably altered basalts, rhyolites and footwall andesites that occur along the northeast margin of the relatively massive and more competent QPG. This package is strongly sheared and, from drill core, the nature of the QPG country rock contact is equivocal. Weak mineralization and the sporadic occurrence of altered rocks in A suggest that this section is a structural repetition of section B. The stratigraphic base of section B (i.e. furthest from ore) commences with a mixed package of mafic volcanic rocks (basalts - basaltic andesites), footwall andesites and variable mafic to intermediate volcaniclastic rocks. Unaltered varieties of these rocks are interlayered with mafic to intermediate orthoamphibole + cordierite rocks with pyrite ± pyrrhotite ± chalcopyrite disseminations, the later rocks becoming more prevalent towards the deposit. A variably to intensely altered package of felsic volcanic rocks (quartz + plagioclase + biotite + sillimanite + cordierite ± garnet assemblages) succeed these rocks and are the host to ore. The base of this package commences with a ~5-10 m interval of variably altered high-silica rhyolite and is typically succeeded by R₂ and R₄ rhyolites that host ore. Due to irregular disseminations of sulphides and competency contrasts between massive sulphides and host rocks, rocks in the vicinity of ore are commonly sheared and variably altered to retrograde biotite + muscovite assemblages. Intervals of mafic orthoamphibole-cordierite rocks occur within the felsic package, although these are interpreted to be fold repetitions of the same rocks lower in section B. Rock types and metamorphosed alteration assemblages described above occur symmetrically on either side of the massive sulphide layers (section C), and hence form the basis for the interpretation that the ore horizon sits within the hinge region of an F_3 antiform.

6. Orthoamphibole-cordierite rocks: origin and significance

From the above description of relative stratigraphy in the vicinity of massive sulphides, it is apparent that the origin and significance of mafic to intermediate orthoamphibole-cordierite rocks may help to constrain true stratigraphy in the context of "footwall" versus "hanging wall" alteration typically associated with VMS deposits (see review by Large et al., 2001). For the purpose of this discussion, orthoamphibole-cordierite rocks refers to altered rocks comprised of metamorphic assemblages that include orthoamphibole + cordierite ± plagioclase ± cummingtonite ± garnet ± biotite. Immobile and incompatible-element binary plots (e.g. Zr versus Nb; Fig 14a) for samples of least-altered IVS basalts, footwall andesites, volcaniclastic rocks and orthoamphibole-cordierite rocks describe a tight co-linear array. Samples trend along the linear array as a function of igneous fractionation and mass/volume changes incurred during alteration and metamorphism.



Fig. 13. Chemostratigraphic profile of the altered footwall sequence to the Kangasjärvi Zn-Cu deposit. Rock-type abbreviations as in Fig. 5.

Note that orthoamphibole-cordierite rocks cluster amongst basalts and andesites with some scatter for orthoamphibole-cordierite rocks associated with volcaniclastic signatures. Immobile incompatible-compatible element binary plots (Fig 14b–c) shows that orthoamphibole – cordierite rocks, especially those associated with IVS basalts, mimic the igneous fractionation pattern of the least-altered rocks. Therefore orthoamphibole-cordierite rocks appear to be the altered and subsequently metamorphosed equivalent of basalts, basaltic andesites and mafic volcaniclastic rocks.

In general, orthoamphibole-cordierite rocks are enriched in FeO + MgO (Fig. 14d) \pm K₂O and depleted in CaO (Fig. 14e), SiO₂ \pm Na₂O. Gains in FeO + MgO are associated with the addition of sulphides and the stability of metamorphic phases such as gedrite, garnet and cordierite. Gains in FeO +MgO \pm K₂O and the loss of CaO are typical of chlorite \pm sericite formation at the expense of primary igneous phases (e.g. Barrett & MacLean, 1994). Mafic orthoamphibole-cordierite rocks also have high Zn (up to ~900 ppm; Fig. 14f) and weak to moderate Cu (up to ~180 ppm).

These data clearly indicate that orthoamphibole-cordierite rocks are the metamorphosed equivalent of altered IVS basalts, footwall andesites and volcaniclastic rocks. Major element systematics show that orthoamphibole-cordierite rocks describe relative mass gains and losses similar to chlorite-altered rocks that are located in the proximal footwall zone of modern and ancient VMS deposits (e.g. Galley & Koski, 1999). Similar conclusions have been drawn for orthoamphibole-cordierite rocks found elsewhere (e.g. Treloar et al., 1981; Wolter & Seifert, 1984; Trägårdh, 1991; Araujo et al., 1996). Consequently, the domain of mafic orthoamphibole-cordierite rocks which precedes the orehosting felsic volcanic sequence at Kangasjärvi is considered to be part of the stratigraphic footwall and therefore likely represents a fossilized hydrothermal upflow zone related to mineralization. This interpretation has two implications: 1) deposition of footwall andesites and the bulk of mafic volcanism likely preceded mineralization; and 2) the bulk of felsic volcanism, as implied by the immediate host rocks to massive sulphides and the widespread K-metasomatism in the inferred "regional hanging wall", is like-



Fig. 14. Immobile (a-c) and mobile (d-f) major- and trace-element variation diagrams for least-altered rocks and orthoamphibole-cordierite rocks (Oam-Crd) from the altered footwall of the deposit.

ly to be broadly coeval with or post-dates mineralization.

7. Discussion

7.1. Petrogenesis and paleotectonic setting

7.1.1. Basalts and footwall andesites

Low Ni, Cr and HFSE abundances, coupled with low to moderate LILE contents and moderate LREE-enrichment of the IVS and OVS basalts are features typical of island arc tholeiite basalts (Perfit et al., 1980; Pearce & Peate, 1995). Basalts with low HFSE and REE contents, such as those from the IVS and OVS, have been reported from modern and ancient island arc settings (e.g. Ewart & Hawkesworth, 1987 and Stern et al., 1995, respectively) and indicate a strongly depleted arc mantle source likely due to the extraction of MORB or back-arc basin basalts (Stern et al., 1995). The relatively low-K and tholeiitic nature of the IVS basalts suggests an oceanic island arc setting, rather than a rifted continental margin (Pearce, 1996). Elevated LILE and LREE, however, preclude derivation in a juvenile oceanic setting (e.g. intraoceanic rift, forearc). Rather, these features indicate that basalts were erupted in an intra-arc setting underlain by thin juvenile crust. This conclusion is in accordance with Lahtinen (1994), who concluded that basaltic volcanism associated with VMS deposits in the Vihanti-Pyhäsalmi district occurred above a west-dipping subduction zone during rifting of a ca. 1.94-2.0 Ga primitive oceanic island arc underlain by depleted mantle. Higher LILE and LREE abundances (higher Th/Nb and La/Nb ratios, respectively) for the OVS suite relative to the IVS suite may indicate a greater contribution of these components from subduction-related processes (fluids or melts) or intracrustal contamination (see Stern et al., 1995). This suggests that OVS basalts may have been generated or modified in a thicker, mature arc relative to basalts in the IVS sequence. The chemistry of mafic volcanic rocks from the OVS is similar to mature, ca. -1.88 Ga mafic island arc basalts of the Western Volcanic Sequence in the Pyhäjärvi area (Kousa et al., 1994), and therefore OVS basalts are tentatively correlated with this younger suite.

The relatively uniform composition of footwall andesites and the observation of andesitic clasts within IVS mafic volcaniclastic rocks is interpreted to indicate that footwall andesites are not the fractionated equivalent of IVS basalts, but were likely erupted in a similar tectonic setting (similar mantle signature) and formed the substrate onto which IVS basalts were extruded. The transition from andesites to basalts likely indicates a transition from a compressional or neutral arc-building stage (>1.93 Ga; Lahtinen, 1994) to extensional rifting of the arc and the rapid extrusion of a bimodal volcanic suite (1.93 – 1.91 Ga; Kousa et al., 1994; Lahtinen, 1994).

7.1.2. IVS rhyolites, high-silica rhyolites and gneissic tonalites

The relatively flat REE profiles and low Zr/Y, La/Yb*n* ratios of IVS rhyolites generally indicate the lack of extensive hornblende fractionation, typical of tholeiitic to transitional-calc-alkaline sequences in extensional settings (see review by Lentz, 1998). High-silica rhyolites at Kangasjärvi have similar incompatible-element ratios to IVS rhyolites and gneissic tonalites and are therefore broadly cogenetic. Relative to IVS rhyolites, high-silica rhyolites at Kangasjärvi largely fit the definition of Barrie et al. (1993) for highsilica rhyolites (>73 wt.% SiO₂, high HFSE and REE contents, flat REE profiles, negative Eu anomalies). High silica contents indicate either fairly low degrees of partial melting or a high degree of fractional crystallization. Similarly, higher HFSE and REE contents indicate low degrees of partial melting and high temperatures of crustal fusion and emplacement (rapid ascent to the near surface). Pronounced Eu anomalies can result from either partial melting with a plagioclase residuum, subsequent feldspar fractionation or a combination of both processes (Lentz, 1998).

The Kangasjärvi gneissic tonalite is geochemically indistinguishable from the IVS rhyolites and hence it is considered to be a cogenetic subvolcanic intrusion. Both suites have incompatible-element ratios (e.g. Zr/Nb) that are identical to the IVS basalts and moderate LILE and LREE enrichment indicating derivation from a similar source region and a similar subduction component, respectively. Trace-element modelling by Lahtinen (1994) indicates that the Rastinpää gneissic tonalite may have been derived by small degrees of partial melting (10-15%) of basaltic, low-K island arc tholeiites at the base of the primitive island-arc crust. Due to the similar inferred geologic setting and geochemical composition as the Rastinpää gneissic tonalite, the model of Lahtinen (1994) is also considered viable for the Kangasjärvi area. This model is also supported by a high ε_{NJ} value (-3.2 ± 0.4) for one sample of QPG in the Kangasjärvi area and similar values (~3.0) for ca. 1.92 Ga rhyolites and tonalite gneisses associated with VMS deposits elsewhere in the district (Lahtinen and Huhma, 1997). These Nd isotope data and recent ion probe U-Pb measurements of zircons from felsic rocks of the 1.92 Ga suite (Vaasjoki et al., 2003) confirm a primitive, mantle-derived source for felsic magmas associated with bimodal volcanic suites in the Vihanti–Pyhäsalmi district.

7.1.3. High-Ti andesites and dacites

Whether high-Ti andesites and dacites have a sedimentary or igneous origin is presently unresolved due to a lack of outcrop-scale observations (e.g. textures, contact relationships). Although high-Ti andesites and dacites have Zr/ Ti ratios intermediate between mafic and felsic volcanic rocks, they are unlikely to be mechanical mixtures of these two end-members. For example, pure mechanical mixing of basalts or andesites with rhyolites would require these mixtures to have undergone a minimum mass or volume loss of -30 % in order to residually enrich the immobile elements Ti and Zr. A residual enrichment of this magnitude is not observed for other immobile elements (e.g. Al) or elements that would partition into heavy mineral fractions during sedimentation (e.g. Sc, V, P). The evolved nature of these rocks (higher FeO, TiO₂ and lower MgO, CaO and Al₂O₃ relative to IVS/OVS basalts) is similar to some Icelandic lavas (Icelandites; Wood, 1978), differentiated mid-oceanic ridge basalts (FeTi basalts, andesites and rhyolites; Juster et al., 1989) and differentiated rocks of the Skaergaard intrusion (Brooks & Nielson, 1978). In general, these evolved suites are thought to be the result of extensive fractionation of tholeiitic magmas at shallow levels in the crust under relatively oxidizing conditions (Juster et al, 1989).

Similar differentiated tholeiitic volcanic rocks overlie ancient VMS deposits in Canada (Bear Lake section, Flin Flon Belt: Syme et al., 1999; Kamiskotia camp, Abitibi Subprovince: Barrie et al., 1991). Given that the relatively oxidized conditions (Ni-NiO buffer) required to produce the observed fractionation trends necessitate open-system conditions (Juster et al., 1989), Barrie et al. (1991) suggested, in relation to evolved basalt-andesite-rhyolite suites at Kamiskotia, that potentially oxidizing hydrothermal fluids, contemporaneous with Cu-Zn deposition, may have influenced the redox state of shallow-level magma chambers. The similar stratigraphic relationship of high-Ti rocks at Kangasjärvi to those associated with VMS deposits elsewhere (i.e. within the upper levels of a seafloor hydrothermal system) may support the interpretation that high-Ti andesites and dacites are indeed the differentiated equivalent of underlying basalts. Without further geologic observations of high-Ti andesites and dacites at Kangasjärvi (i.e. textures, contact relationships), however, their origin remains equivocal. The occurrence of ferrobasalts in association with other VMS deposits in the Vihanti-Pyhäsalmi district (Ruostesuo; Roberts et al., 2003) and other world-class VMS districts (see above), suggests that the petrogenesis of these rocks warrants future research.

7.2. Depositional setting and stratigraphy

7.2.1. Inner volcanic sequence

A compilation of lithologies, textures, contact relationships and some aspects of pre-metamorphic alteration are shown in Figure 15 to illustrate the stratigraphy and geologic framework of the Kangasjärvi area. It is proposed that IVS basalts were extruded onto an andesitic substrate and that the bulk of felsic volcanism, in association with Zn-Cu mineralization, post-dated this event. The observation of QPG, the subvolcanic equivalent of ore-associated felsic volcanic rocks, in intrusive contact with footwall andesites further supports the above stratigraphic scheme. It is acknowledged, however, that these findings are based on relatively few and highly deformed outcrops and that a variety of intrusive relationships may be possible. In the Mullikkoräme region, for example, a felsic intrusion with textures indicative of a high level of emplacement (e.g. relict miarolitic cavities) is the local footwall to highly epidotized basalts (A. Galley, pers. comm., 2002). As epidosites are indicative of high-temperature reactions zones



Fig. 15. Compilation diagram of rock types, outcrop textures and alteration features for rocks in the Kangasjärvi area. Rock-type abbreviations as in Fig. 5. Solid lines between lithologies indicate contact relationships observed in outcrop or drill core that are considered primary. Dashed lines with question marks indicate inferred contact relationships. Qtz = quartz; PI = plagioclase; Bt = biotite; Oam = orthoamphibole; Sil = sillimanite; Grt = garnet; Crd = cordierite; Epi = epidote; Py = pyrite.

at the base of seafloor hydrothermal systems (Schiffman and Smith, 1988), the above-noted relationship at Mullikkoräme strongly suggests that shallow-level felsic intrusions provided the necessary heat to drive convective hydrothermal circulation. The lack of epidosites at Kangasjärvi, and the narrow distance between the QPG and the ore deposit (~200–250 m), may indicate that portions of the QPG represent a resurgent phase of felsic magmatism that post-dated mineralization (similar to the multi-component Sneath Lake intrusive complex, Snow Lake arc assemblage, Flin Flon Belt, Canada; Bailes & Galley, 1999) and possibly assimilated previously altered rocks (e.g. epidosites).

7.2.2. Outer volcanic sequence

The outer volcanic sequence is a complex association of mafic volcanic rocks, undifferentiated sedimentary rocks, locally occurring calc-silicates (metalimestones), metagreywackes and U-P-bearing graphitic schists. The relative stratigraphy and sedimentary provenance of this sequence is presently unresolved. In contrast to rifting and active volcanism described by rocks in the IVS, the occurrence of abundant sedimentary rocks in the OVS indicates that this sequence is characterized, at least in part, by a basin-style sedimentary setting. As noted previously, mafic volcanic rocks in the OVS broadly resemble the ca. 1.88 Ga mature island arc volcanic rocks and therefore the OVS sequence as a whole is considered to be younger than the IVS.

7.3. Metallogenic implications

We have shown that pre-metamorphic, seafloor hydrothermal alteration associated with VMS-style mineralization at Kangasjärvi occurred within a bimodal suite of volcanic rocks that comprise the stratigraphic hanging wall to the Kangasjärvi gneissic tonalite. This bimodal sequence was the result of intra-arc rifting of a fairly thin and juvenile oceanic island-arc (see also Lahtinen, 1994). Distinctive high-silica rhyolites were emplaced within the immediate footwall to massive sulphides prior to or during intense hydrothermal alteration associated with mineralization.

Bimodal volcanic sequences with cogenetic high-level tonalite-trondhjemite intrusions are typical of primitive submarine extensional settings such as oceanic spreading centres, oceanic proto-arcs, rifted-arcs and back-arcs (Galley, 2003). The generation of isotopically juvenile felsic magmas at Kangasjärvi, their rapid emplacement as high-level intrusions in extensional settings and their common association with high-temperature high-silica rhyolites is indicative of significant heat transfer from the upper mantle - lower crust to the upper crust (Bailes & Galley, 1999; Galley, 2003). While high-level intrusions have been postulated as the "heatengine" that drives long-lived convective hydrothermal systems (e.g. Campbell et al., 1981;

Cathles, 1983), Bailes & Galley (1999) noted that the intrinsic thermal properties required to generate well-endowed VMS districts may reflect an anomalous thermal corridor related to the broader geodynamic context (i.e. arc rifting) rather than individual high-level intrusions. In that regard, tonalite-trondhjemite intrusions and high-silica rhyolites at Kangasjärvi may be the near-surface expression of the high overall heat flux to the upper crust due to rifting of a juvenile arc, which subsequently led to longlived convective hydrothermal circulation and mineralization. High-silica rhyolites also occur in association with sulphide mineralization elsewhere in the district (Pyhäsalmi: Mäki, 1986; K. Rasilainen, pers.comm., 2002; Ruostesuo: Roberts et al., 2003). Therefore, by virtue of association with the altered footwall of the Kangasjärvi deposit (and a spatial association with other deposits) and by their intrinsic petrogenetic properties, delineation of high-silica rhyolites in the Vihanti-Pyhäsalmi district may be a powerful exploration tool for identifying favourable volcanic sequences.

8. Conclusions

Although many geologic relationships in the Kangasjärvi area are obscured by deformation and metamorphism, many attributes of stratigraphy and alteration can be resolved by a combination of map-scale patterns, outcrop observations and whole-rock geochemistry. Several of these observations appear to be fundamental to the localization of mineralization at Kangasjärvi, and possibly other deposits in the district, and are therefore applicable to districtscale exploration strategies. These relationships are summarized below:

- The bulk of basaltic volcanism occurred in a subaqueous setting on an arc-related andesitic substrate.
- Large (multi-phase?) trondhjemite-tonalite intrusions were emplaced as stocks or sills within the andesitic basement and represent

the subvolcanic equivalent of ore-associated felsic volcanic rocks.

- Sulphide mineralization occurred on or within a felsic volcanic – volcaniclastic package and was associated with moderate to intense chlorite + sericite alteration.
- 4) Orthoamphibole-cordierite rocks in the altered footwall of the deposit occur stratigraphically below the ore-hosting rhyolites and are the altered equivalent of mafic and intermediate volcanic and volcaniclastic rocks. The origin of these rocks is attributed to pre-metamorphic chlorite ± sericite alteration typical of hydrothermal upflow zones that underlie many ancient seafloor massive sulphide deposits.
- 5) High-silica rhyolites (HSR_a) are spatially associated with the immediate stratigraphic footwall of the deposit and were variably chloritized during the mineralizing process.
- 6) The occurrence of quartz-plagioclase gneisses and high-silica rhyolites represent the nearsurface expression of an anomalous thermal corridor that was likely conducive to longlived, convective hydrothermal circulation of modified seawater and the precipitation of base metal mineralization.

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