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Zircon dating of the basalt and felsic dyke in Haveri, SW Finland

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Abstract

The E-MORB type Haveri basalt differs from the volcanic arc type rocks in the Tampere belt showing no subduction signature. It is considered to have formed in a marginal basin prior to the volcanic arc. We present here zircon U-Pb dating on two samples. The basalt and the felsic dyke yielded $^{207}Pb/^{206}Pb$ ages of 1902 ± 5 Ma and 1891 ± 2 Ma, respectively, interpreted as crystallisation ages. The basalt also contains older 1.98 Ga grains while the felsic dyke contains older 1.92 Ga, 1.94 Ga, 1.98 Ga and 2.0 Ga grains, which are inferred as inherited. The age dating of E-MORB type basalts can be used to identify the extensional episodes of the accretionary Svecofennian orogeny.

Keywords: Svecofennian orogen, age determination, Haveri formation, tectonic switching, extension

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1. Introduction

Absolute age determinations are essential in understanding and modelling stratigraphic relations and evolution of large-scale events in deformed Precambrian terrains. One such example is the Paleoproterozoic Svecofennian orogen (e.g., Lahtinen et al. 2005) in the Central Fennoscandian shield where tectonostratigraphic relations are difficult to construct without the information given by the single grain age dating. One intriguing unit in the Svecofennian in Finland is the Haveri formation (e.g., Mäkelä 1980; Kähkönen 1987), located in the northern part of the Tampere belt in SW Finland (Fig. 1b), as the basalts in Haveri show E-MORB type geochemical affinities clearly distinct from the surrounding arc-type volcanic rocks (e.g., Kähkönen et al. 1989; Kähkönen & Nironen 1994; Kähkönen 1999; Strauss 2003). The stratigraphic relations suggest that the basalts in Haveri are the lowermost and thus the oldest known unit in the Tampere belt (e.g., Kähkönen

1999). However, the Haveri basalt is only dated with a whole rock Pb/Pb method, which yielded quite an imprecise age between 2.0 and 1.9 Ga (Vaasjoki & Huhma 1999). This has led to variable interpretations and speculations on the origin of the Haveri formation such as (i) an initial stage island arc (Mäkelä 1980), (ii) a pre-1.91 Ga rift-related marginal basin preceding the volcanism in the Tampere belt (Kähkönen & Nironen 1994; Lahtinen & Huhma 1997, Lahtinen et al. 2017), (iii) possible correlation to ~ 1.95 Ga Outokumpu and Jormua ophiolite complexes (Vaasjoki & Huhma 1999), (iv) or a back-arc basin (Strauss 2003). Kähkönen & Nironen (1994) also speculated that Haveri might be an exotic fragment of an oceanic floor, i.e. an ophiolite. These interpretations have influenced the discussion of the evolution of the whole Svecofennian orogen. In this contribution we present the first U-Pb zircon ages of the Haveri basalt and crosscutting felsic quartz feldspar porphyric dyke and discuss the tectonic implications on the Svecofennian orogeny.

2. Geological setting

The Paleoproterozoic Svecofennian orogen formed approximately between 1.95 and 1.75 Ga ago as a consequence of accretionary processes. In Finland, the Svecofennian Province is further divided into the Southern Svecofennia and the Northern Svecofennia Subprovinces, which represent the main arc complexes (Kohonen et al. 2021; Fig 1a). For more detailed description, see e.g., Gaal & Gorbatschev (1987), Nironen (1997) and Lahtinen et al. (2005).

The Tampere belt (TB) in the central part of the Svecofennian Province comprises mature volcanic rocks ranging from basalts to rhyolites and associated sedimentary rocks. The TB probably formed in a mature arc or continental margin setting (Kähkönen 1987). The ages of the subduction-related igneous rocks are dated between 1897 ± 3 Ma (Sammatinjärvi dacite; Kähkönen & Huhma 2012) and 1880 ± 7 Ma (Lempiäniemi plagioclase porphyry; Kähkönen et al. 1989). Structurally, the TB is an upright syncline where primary sedimentary structures are well-preserved (Nironen 1989), but the internal structure of the syncline might be more complicated (Kalliomäki et al. 2014). The TB is metamorphosed in medium grade condition without melting and andalusite and occasional staurolite are the key porhyroblasts (Kähkönen & Nironen 1994; Hölttä & Heilimo 2017). Detailed description of the TB is presented, e.g., in Kähkönen (1987, 1999) and Nironen (1989b). The high metamorphic grade Pirkanmaa belt is situated south of the TB (Lahtinen et al. 2009). It has been suggested that the Pirkanmaa belt is a mid-crustal expression of a forearc region and an accretionary prism/subduction complex to the TB volcanic arc complex (Kähkönen 2005; Lahtinen et al. 2009; Kara et al. 2021).

The Haveri formation is located in the northern part of the TB. The Haveri formation has raised repeated economic interests as it contains a copper-gold mineralisation mined in the past (Mäkelä 1980; Strauss 2003; Eilu 2012). The Haveri formation is overlain by the sedimentary Osara and volcanic Harhala formations (Fig. 2; Kähkönen & Nironen 1994) and it is considered that the Haveri formation represents submarine conditions in an extension-related setting, based on the tholeiitic pillow lavas and depositional structures in sedimentary rocks (Kähkönen 1999, 2005).

3. Samples and research methods

Two samples were collected for the dating (Fig. 2). The sampling site of the basalt was guided by Kähkönen & Nironen (1994) where one basalt analysis (sample 28-maf-MN/88) shows Zr content of 235 ppm. Our sample was collected in the vicinity of this site. The quartz-feldspar porphyric felsic dyke sample was collected 900 m east of the mafic sampling site from the same outcrop as sampled by Kähkönen & Nironen (1994; sample 28-fels-MN/88).

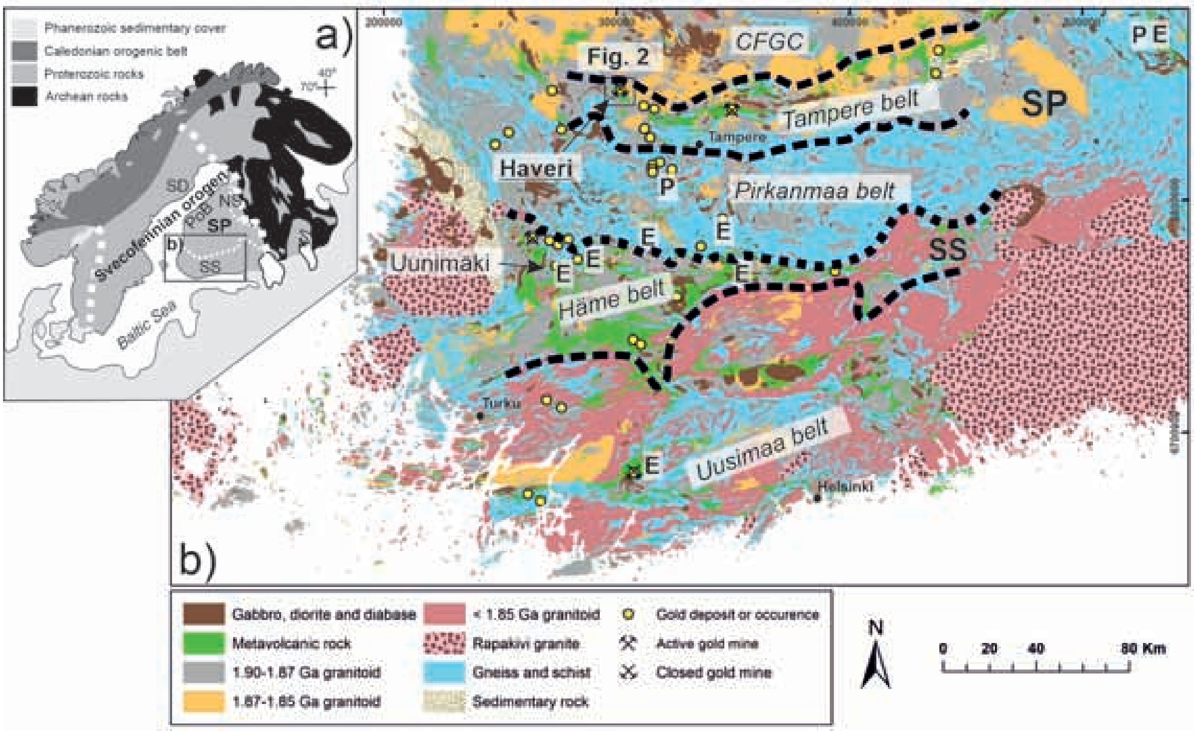

Figure 1. a) Geological overview of the Fennoscandian Shield, modified after Koistinen et al. (2001). SP = Svecofennian Province, SS = Southern Svecofennia Subprovince, NS = Northern Svecofennia Subprovince, PoB = Pohjanmaa belt; SD = Skellefte district. b) Lithological map of southern Finland, modified after Bedrock of Finland – DigiKP. The study area is indicated by a black rectangle. CFGC = Central Finland Granitoid Complex, E = E-MORBs, P = Picritic basalts.

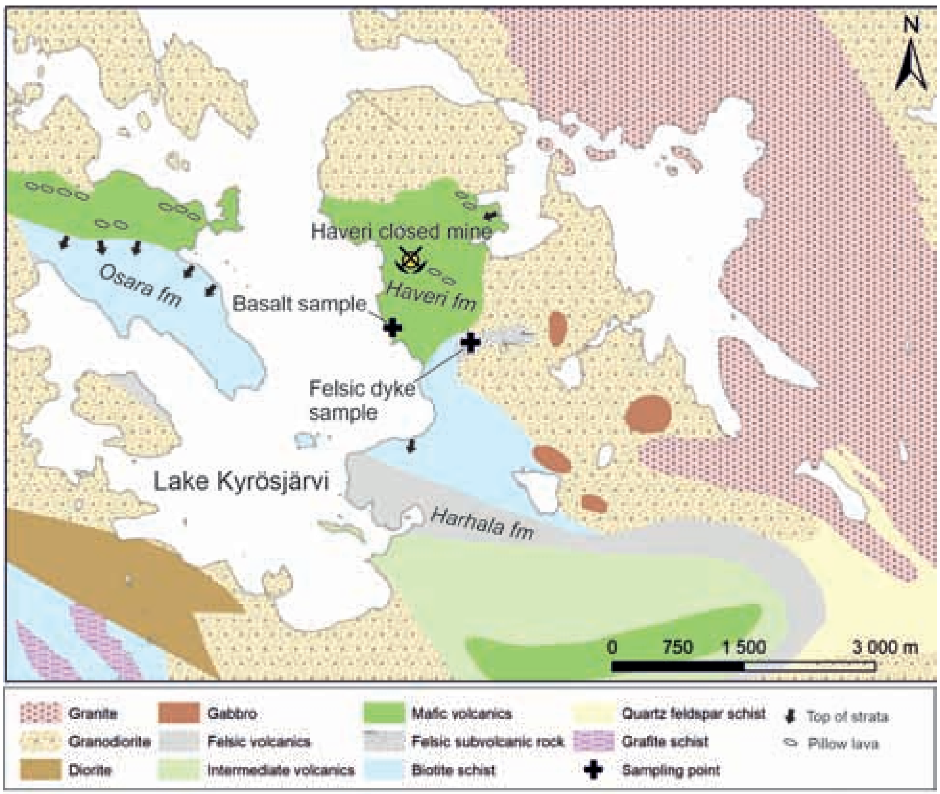

Figure 2. Lithological map of the Haveri region. Modified after Bedrock of Finland – DigiKP and Kähkönen & Nironen (1994).

Approximately five kilograms of the basalt was processed, and eleven zircons were recovered. The felsic rock sample was approximately 1.5 kg in weight, and it yielded abundant zircons. Standard procedure for zircon separation was used including crushing, grinding, panning, heavy liquid separation, magnetic separation, pyrite removal using nitric acid (basalt only) and hand picking. The grains were mounted in epoxy resin and sectioned approximately in half and polished. Back-scattered electron (BSE) images and mineral identification with energy-dispersive X-ray Spectroscopy (EDS) were done using a Thermo Scientific Apreo 2 SEM at the University of Turku.

Zircon U-Pb analyses were performed using a Nu Plasma AttoM single collector ICPMS in the Finnish Geosciences Research Laboratory (SGL) at the Geological Survey of Finland. The ICPMS was connected to a Photon Machine Excite laser ablation system. Samples were ablated in He gas (gas flows = 0.4 and 0.1 l/min) within a HelEx ablation cell (Müller et al. 2009). He aerosol was mixed with Ar (gas flow= 0.8 l/min) prior to entry into the plasma. The gas mixture was optimized daily for maximum sensitivity. Typical ablation conditions were: beam diameter 25 µm, pulse frequency 5 Hz and beam energy density 2.54 J/cm2. A single U-Pb measurement included a short pre-ablation with 35 µm beam, 10 s of on-mass background measurement, followed by 30 s of ablation with a stationary 25 µm beam. ^{235}U was calculated from the signal at mass 238 using a natural $^{238}U/^{235}U=137.88$. Mass number 204 was used as a monitor for common ^{204}Pb. In an ICPMS analysis, ^{204}Hg mainly originates from the He supply. The observed background counting-rate on mass 204 was 150–200 cps and has been stable at that level over the last five years. The contribution of ^{204}Hg from the plasma was eliminated by on-mass background measurement prior to each analysis. Age related common lead (Stacey and Kramers 1975) correction was used when the analysis showed common lead contents significantly above the detection limit (i.e., >50 cps). Signal strengths on mass 206 were typically 100000 cps, depending on the uranium content and age of the zircon.

The calibration standard zircon GJ-1 (609 ± 1 Ma; Belousova et al. 2006) and in-house standard A382 (1877 ± 2 Ma, Huhma et al. 2012) were run at the beginning and at the end of analytical session and at regular intervals during the session. The $^{207}Pb/^{206}Pb$ age offset from concordant ID-TIMS ages for the in-house reference zircon does not exceed 0.5%. Raw data were corrected for the background, laser induced elemental fractionation, mass discrimination and drift in ion counter gains and reduced to U–Pb isotope ratios by calibration to concordant reference zircons using the program Glitter (Van Achterbergh et al. 2001). Further data reduction including common Pb correction and error propagation was performed on site using excel spreadsheet written by Y. Lahaye and H. O'Brien. Errors are propagated by quadratic addition of within-run errors (2 SE), the reproducibility of standard during the run (2 SD) and the overall error on the certification of the selected standard. To minimize the effects of laser-induced elemental fractionation, the depth-to-diameter ratio of the ablation pit was kept low, and isotopically homogeneous segments of the time-resolved traces were calibrated against the corresponding time interval for each mass in the reference zircons.

Plotting of the U-Pb isotopic data and age calculations were performed using the Isoplot/Ex 4.15 program (Ludwig 2012). All the ages were calculated with 2σ errors and without decay constants errors. Data-point error ellipses in the figures are at the 2σ level.

4. Results

The Haveri basalt zircons are mostly elongated prismatic grains with length varying between 70 and 200 µm and width between 40 and 150 µm (Fig. 3a). The colour ranges from transparent to light brown. In the BSE images the grains are quite homogeneous: only faint oscillatory zoning, a few

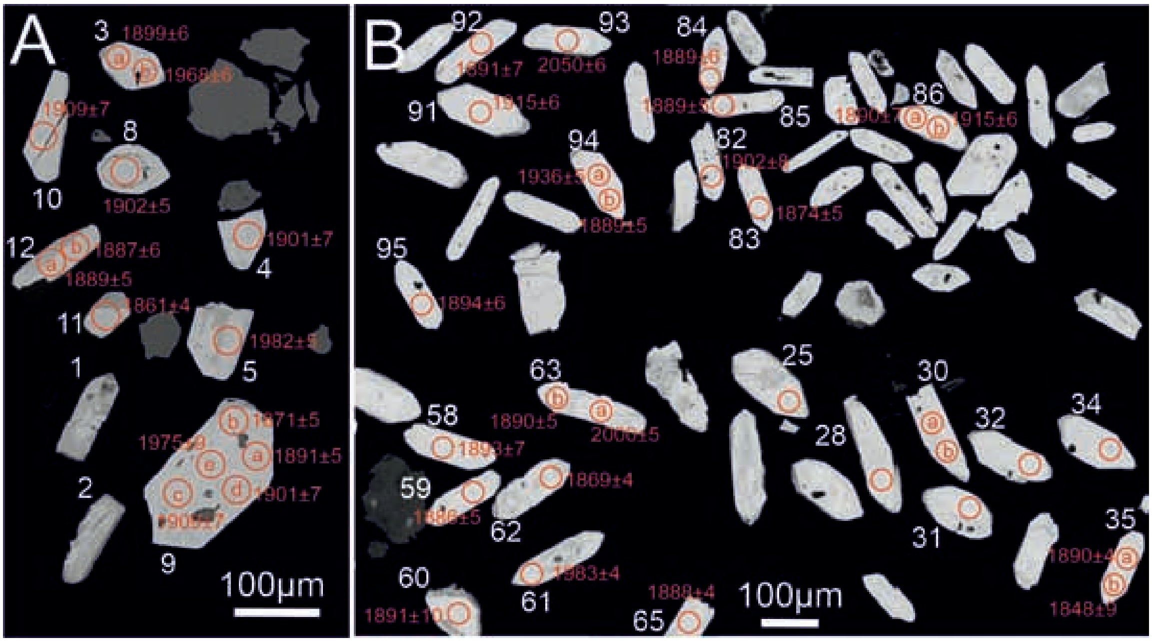

Figure 3. Zircon BSE-images of the basalt (A) and of the felsic dyke (B).

cracks and metamictic domains can be observed. One obvious core-rim pair is present. In total, 16 analyses were performed on 11 grains and three age populations were found from the analyses. The largest population contains seven analyses and yielded almost identical $^{207}Pb/^{206}Pb$ and upper intercept ages of 1902 ± 5 Ma (Fig. 4a and b). Four of the analyses yielded a concordia age of 1902 ± 6 Ma. An older population including three analyses, two concordant and one discordant, yielded a weighted average $^{207}Pb/^{206}Pb$ age of ~ 1.98 Ga. The youngest population consists of six analyses and shows $^{207}Pb/^{206}Pb$ ages between 1891 and 1861 Ma (Fig. 4 a and b).

The Haveri felsic dyke zircons are euhedral and elongated grains with length varying between 50 and 300 µm and width between 30 and 150 µm (Fig. 3b). All the zircons are quite homogeneous, but especially the larger ones show metamictic domains and abundant cracks. Cores or core-like structures, overgrowths and rims of varying thicknesses can be found, and many grains show faint oscillatory zoning. A total of 54 spots were analysed on 45 grains and several age populations were found. The largest population consists of 36 analyses, which yielded a concordia age of 1892 ± 3 Ma (26 analyses of 36), almost identical upper intercept age and similar weighted average $^{207}Pb/^{206}Pb$ age of 1891 ± 2 Ma (Fig. 4c and d). An older age population consists of 12 analyses and includes several subgroups showing varying $^{207}Pb/^{206}Pb$ age between 2050 and 1915 Ma. The six youngest analyses yielded $^{207}Pb/^{206}Pb$ ages between 1874 and 1848 Ma. A full list of the U-Pb analyses is given in Electronic Appendix A.

5. Discussion and conclusions

5.1. Age constraints

We suggest that 1902 ± 5 Ma represents the crystallisation age of the Haveri basalt. This is the main age population found from the zircons and concordia, upper intercept and weighted average $^{207}Pb/^{206}Pb$ age of the analyses yielded the same age. The elongated and angular grain shapes (Corfu et al. 2003) and Th/U ratios between

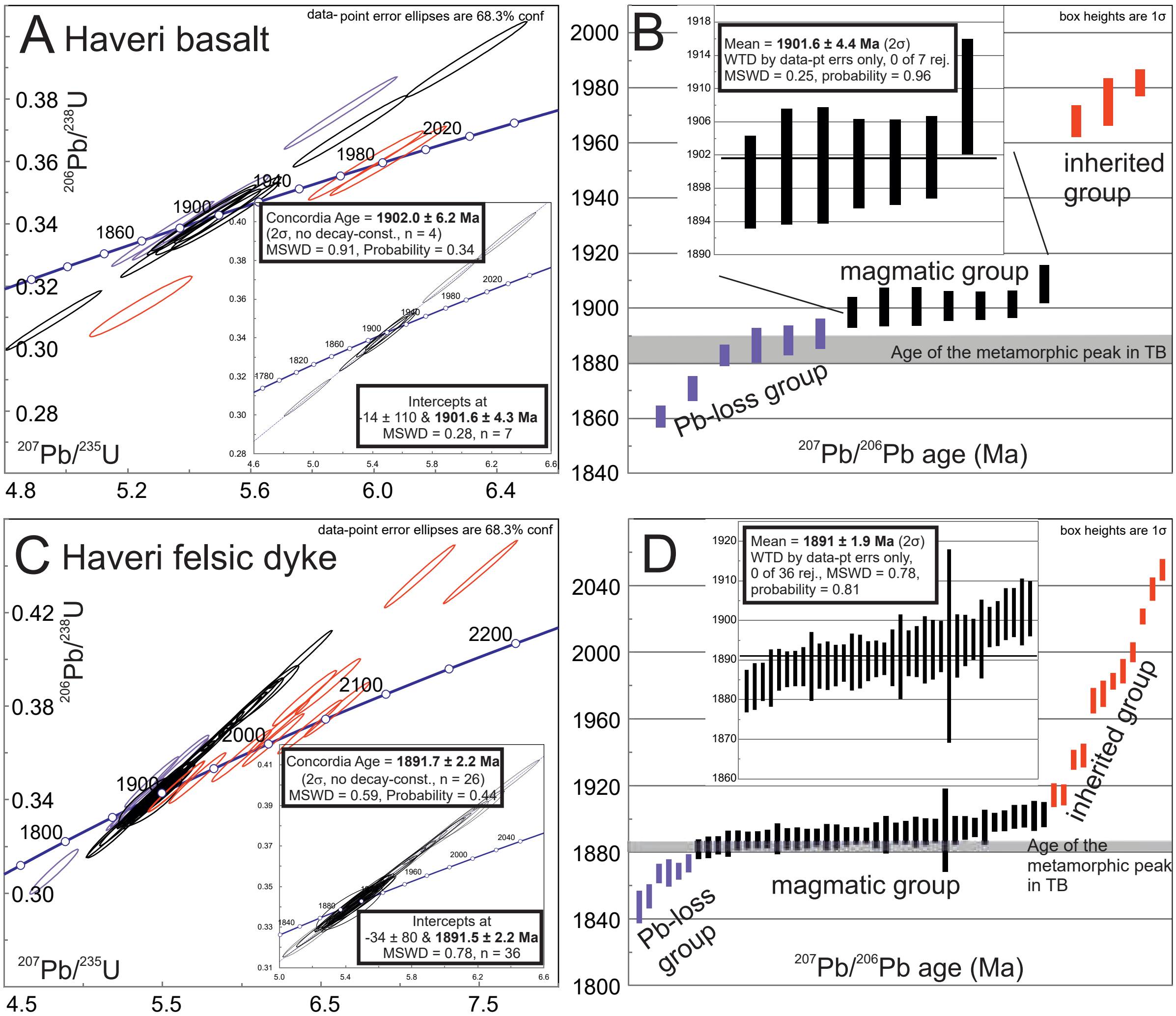

Figure 4. Zircon U-Pb results. A and B are for the basalt and C and D for the felsic dyke.

0.26 and 0.56 (Kirkland et al. 2015), with two outliers below 0.1, also support a magmatic origin. The crystallisation age of 1891 ± 2 Ma for the felsic dyke is unambiguous. The zircon record is extensive and zircon morphology suggests magmatic origin (Corfu et al. 2003). For both samples we prefer the weighted average $^{207}Pb/^{206}Pb$ age due to higher number of analyses used for calculation, lack of the U-Pb fractionation effect and less Pb-loss effect compared to the concordia age.

Although the basalt is the oldest dated rock in the TB and the oldest dated igneous rock in SW Finland, the age is younger than previously considered (c.f., Kähkönen & Nironen 1994; Vaasjoki & Huhma 1999; Lahtinen et al. 2017). Instead, with errors considered, the ages of the basalt and the felsic dyke are close to the ages of volcanic rocks in the TB (Kähkönen 2005; Kähkönen & Huhma 2012). The basalt is only marginally older than the oldest subduction-related volcanic rock in the TB (1897 ± 3 Ma; Kähkönen & Huhma 2012) and the age of the felsic dyke is similar to the ages of volcanic rocks in the TB (e.g., Kähkönen 2005). This indicates that the Haveri formation is related to the buildup of the volcanic TB (c.f., Kähkönen & Nironen 1994).

The basalt contains inherited ~1.98 Ga zircons whereas several inherited populations were found from the felsic dyke: ~1.92 Ga, ~1.94 Ga, ~1.98 Ga and ~2.0 Ga. Similar inherited zircon ages have been recognized from other magmatic and sedimentary rocks in SW Finland (Lahtinen et al. 2009). Although the number of grains is small, we suggest that the lack of other inherited populations in the basalt is simply due to the older age and more primitive origin compared to the felsic dyke. Therefore, the basalt intruded less sediments and the source have not been contaminated by subduction related sediments as probably is the case for the felsic dyke. Alternatively, the 1.98 Ga grains stem directly from the source of the E-MORB magma.

It is likely that the young ages are due to Pb-loss rather than metamorphic recrystallisation or new metamorphic growth. Some of the zircon domains might have recrystallised or grown due to metamorphism, which peaked in the Pirkanmaa belt to the south and apparently in the TB approximately at 1.88 Ga (Kilpeläinen 1998; Mouri et al. 1999, Lahtinen et al. 2009). The temperature remained around 500 °C and never exceeded 600 °C in the TB (Kilpeläinen et al. 1994), which does not support new zircon growth. The basalt sample was taken a few kilometres away from the mine site, affected by the most intense fluid flow, and the sample contains quartz veins and disseminated sulphides indicating moderate postcrystallization alteration and fluid influx conditions. As the younger grains and domains lack distinct metamorphic morphologies, we suggest that Pb-loss due to strong fluid influx is the cause for the young ages. Moreover, if a Pb-loss event is close to the time of crystallisation, the measured ages can travel down the concordia curve resulting in younger concordant ages (Corfu 2013). The age and duration of the fluid activity is uncertain, but it likely took place after the metamorphic peak and lasted a few tens of millions of years as indicated by the youngest $^{207}Pb/^{206}Pb$ ages, 1861 Ma in the basalt and 1848 Ma in the felsic dyke.

5.2. Tectonic implications

Extensional episodes and related magmatism in subduction-related volcanic arcs are common (e.g., Collins 2002; Xia & Li 2019). In the Fennoscandian shield, Hermansson et al. (2008), Saalmann et al. (2009) and Kara et al. (2021) modelled the accretionary Svecofennian orogeny using the tectonic switching model where subduction polarity stays constant at N-NE beneath a single continental margin. The subduction hinge switches between short-lived advancing episodes (contraction) and prolonged retreating episodes (extension; Hermansson et al. 2008; Kara et al. 2021). During extension the MORB/E-MORB type mafic magmatism occurred (Väisänen & Mänttäri 2002; Lahtinen et al 2017; Nironen 2017; Kara et al. 2021). With dating of the mafic magmatism, we can infer the age of the extensional stages.

Several E-MORB-type igneous rocks occur in southern and western Finland and in Skellefte district, central Sweden (Fig. 1a and b; Berge 2013). Those in the Uusimaa belt are regarded younger than the ~1.89-1.88 Ga arc rocks (Väisänen & Mänttäri 2002; Nironen et al. 2016), but those in the Tampere, Pirkanmaa and Häme belts are regarded coeval or slightly older than the oldest arc volcanic rocks. The E-MORB type Uunimäki gabbro in the northern part of the Häme belt (Fig. 1b) has a concordia age of 1891 ± 5 Ma and $^{207}Pb/^{206}Pb$ age of 1895 ± 6 Ma (Kara et al. 2021). Similar E-MORB type rocks occur within the boundary zone of the Pirkanmaa and Häme belts whereas those in the central Pirkanmaa belt show more depleted and primitive characteristics (Fig. 1b; Peltonen 1995; Lahtinen et al. 2017; Kara et al. 2021). Moreover, the 1889 ± 5 Ma Takamaa formation in the central TB shows evidence of incipient rifting (Kähkönen et al. 1989). These occurrences suggest that at least two extensional episodes with rifting occurred approximately at ~1.9 Ga and ~1.89 Ga, or alternatively one prolonged extension between 1.9 Ga and 1.89 Ga. Kilpeläinen (1998) suggested three separate deformation episodes related to contraction in the Tampere and Pirkanmaa belts: D_1 stage took place approximately between ~1910 and 1889 Ma but the maximum age is poorly constrained, followed by D_2 stage at 1889-1880 Ma and D_3 stage approximately at 1870 Ma. This suggests that two separate extensional episodes occurred. In addition, younger extension, approximately at 1.86 Ga, has been identified in the Pirkanmaa, Häme and Uusimaa belts (Lahtinen & Nironen 2010; Kara et al. 2020).

Most of the E-MORB occurrences and associated picrites remain undated but show similar geochemical characteristics to the dated units. The E-MORBs and picrites in the Rantasalmi region (Makkonen & Huhma 2007; Kousa et al. 2018) are similar to the Pirkanmaa belt E-MORBs and picrites, where the geochemical characteristics were suggested to be related to the extent of rifting (Kara et al. 2021). The age of the Rantasalmi E-MORBs were estimated to be 1.91-1.90 Ga (Kousa et al. 2018). The Pohjanmaa belt, which is the possible continuation of the Pirkanmaa belt (e.g., Lahtinen et al. 2017), host several E-MORB/WPL mafic lavas (Vaarma & Kähkönen 1994). According to Lahtinen et al. (2017) these are all related.

As the Haveri basalt was formed in an extensional setting, the question is what was rifted? Kähkönen & Nironen (1994) proposed that a slightly older continental crust was rifted to form a marginal basin filled with the Haveri lavas. These in turn were covered by turbidites and younger volcanic rocks. Kähkönen (2005) inferred that the older continent might be the 2.1-2.0 Ga hidden crust, based on the abundant detrital zircons of that age in many places (Kähkönen 2005 and references therein). The Haveri basalt only contains 1.98 Ga inherited grains, which might indicate the age of the rifted crust.

Supplementary data

Electronic Appendices are available via Bulletin of Geological Society of Finland web page.
Electronic Appendix A: Zircon U-Pb data of the Haveri basalt and felsic dyke

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