The Archaean Karelia and Belomorian Provinces, Fennoscandian Shield

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Abstract

The Archaean bedrock of the Karelia and Belomorian Provinces is mostly composed of granitoids and volcanic rocks of greenstone belts whose ages vary from c. 3.50 to 2.66 Ga. Neoarchaean rocks are dominant, since Paleoarchaean and Mesoarchaean granitoids (>2.9 Ga) are only locally present. The granitoid rocks can be classified, based on their major and trace element compositions and age, into four main groups: TTG (tonalite-trondhjemite-granodiorite), sanukitoid, QQ (quartz diorite-quartz monzodiorite) and GGM (granodiorite-granite-monzogranite) groups. Most ages obtained from TTGs are between 2.83–2.72 Ga, and they seem to define two age groups separated by a c. 20 m.y. time gap. TTGs are 2.83–2.78 Ga in the older group and 2.76–2.72 Ga in the younger group. Sanukitoids have been dated at 2.74–2.72 Ga, QQs at c. 2.70 Ga and GGMs

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at 2.73–2.66 Ga. Based on REE, the TTGs fall into two major groups: low-HREE (heavy rare earth elements) and high-HREE TTGs, which originated at various crustal depths. Sanukitoids likely formed from partial melting of subcontinental metasomatized mantle, whereas the GGM group from partial melting of pre-existing TTG crust.

The Karelia and Belomorian Provinces include a large number of generally NNW-trending greenstone belts, whose tectonic settings of origin may include an oceanic plateau, island arc and/or continental rift. The ages of volcanic rocks in these greenstone belts vary from 3.05 to 2.70 Ga.

Migmatized amphibolites occur as layers and inclusions in TTGs and fall into two main groups on the basis of their trace element contents. Rocks of the first group have flat or LREE-depleted trace element patterns, resembling the modern mid-ocean ridge basalts. Rocks of the second group are enriched in LILE and LREE may in part represent metamorphosed dykes with assimilated and/or diffused crustal signatures from their TTG country rocks.

Metamorphism of the TTG complexes occurred under upper amphibolite and granulite facies conditions at c. 2.70–2.60 Ga. The pressures of the regional metamorphism were mostly c. 6.5–7.5 kbar as constrained by geobarometry, and the corresponding temperatures were c. 650–740 °C. The granulites near the western boundary of the Karelia Province were equilibrated at c. 9–11 kbar and 800–850 °C. Subduction-related eclogites in the Belomorian Province were metamorphosed at pressures up to 20 kbars in two stages around 2.88-2.81 Ga and c. 2.72 Ga. In other greenstone belts the observed metamorphic conditions show significant variations. In the central parts of the Ilomantsi greenstone belt the observed metamorphic P and T values are c. 3–4 kbars and 550–590ºC, and in the Kuhmo greenstone belt 16–17 kbar and 650–690 °C, respectively.

Neoarchaean accretion of exotic terranes at c. 2.83–2.75 Ga and the subsequent collisional stacking at ~2.73–2.68 Ga were instrumental in the construction of the current crustal architecture of the Karelia Province. The Svecofennian orogeny strongly modified, however, this Neoarchaean crustal structure during the early Proterozoic

3.1 Introduction

There is no a widely accepted geodynamic framework for the Archaean eon, as exists for the modern Earth via plate tectonics and the Wilson cycle (Benn et al. 2006). Nevertheless, it is commonly thought that plate tectonics has been operating in some form at least since the Neoarchaean (e.g. deWit 1998; Condie and Benn 2006). However, there is considerable controversy concerning many of the fundamental aspects of the Archaean tectonics, including serious doubts about the applicability of the concept at all (Hamilton 1998, 2011). Some views emphasize the likelihood of a much faster convection and ocean floor spreading rate in the Archaean than in the Phanerozoic, whereas some researchers suggest that plate velocities have been fairly similar throughout geological time (Blichert-Toft and Albarede 1994; Kröner and Layer 1994; Blake et al. 2004; van Hunen et al. 2004; Strik et al. 2003; Korenaga 2006). Some recent simulations indicate that only after the development of the post-perovskite layer above the core-mantle boundary and after sufficient cooling of the early Earth, could the core have been able to release enough heat to the upper mantle to enable the rapid motion of plates. The earliest this event could probably have occurred is inferred to be in the early Palaeoproterozoic (Tateno et al. 2009; Hirose 2010).

It is generally thought that the Archaean mantle was hotter than today, and the critical petrological evidence for this is komatiites and their much higher abundance in the Archaean than in the younger geological environments. Komatiites have been related to mantle plumes and reported to record very high mantle temperatures and melting pressures. On the other hand, it has also been argued that komatiites were, similarly to modern boninites, produced by hydrous melting at relatively shallow mantle depths in a subduction environment. This alternative interpretation predicts that the Archaean mantle was only slightly hotter than the present one (Grove and Parman 2004; Condie and Benn 2006).

Complete ophiolite sequences that would attest to the existence of Phanerozoic-style oceanic crust are rare in the Archaean rock record. A Neoarchaean, 2.51 Ga ophiolite complex has been described from the North China craton (Kusky et al. 2004), and greenstone occurrences that share many—although not all—characteristics of Phanerozoic ophiolites have also been found in other Archaean cratons, including the Belomorian province of the Fennoscandian shield (Kusky and Polat 1999; Corcoran et al. 2004; Puchtel 2004; Shchipansky et al. 2004). Archaean oceanic crust could have been thicker than the Proterozoic and Phanerozoic oceanic crust, resembling that of modern oceanic plateaux. This would explain the scarcity of MORB-type ophiolites, as in this case only the upper basaltic, pillow lava-dominated sections of the oceanic crust were likely to be accreted or obducted and thus preserved in the rock record (Kusky and Polat 1999). Şengör and Natal'in (2004) pointed out that in many Phanerozoic accretionary orogens, as in the western Altaids, evidence of closed oceans also only occurs in the form of separated fragments of variably complete ophiolitic sequences, without a single example of well-preserved major ophiolite nappe such as those in Oman or Newfoundland. Moreover, as the Altaids cover about as much of the Earth's land area as the Archaean crust, the rarity of indisputable Archaean ophiolite sequences already appears a less forceful argument against modern-style plate tectonism. Dilek and Polat (2008) and Dilek and Furnes (2011) also argued that the structure of the Archaean oceanic crust differed from that of the modern one; for example the Archaean sheeted dyke systems could be rare because spreading rates and magma supply budgets were necessarily not in such a balance that is needed for their generation.

Tonalite-trondhjemite-granodiorite (TTG) gneisses are the major constituent of the Archaean crust, and more widespread in the Archaean than in the younger rock record (Condie and Benn 2006). The petrogenesis of HREE-depleted Archaean TTGs have been interpreted in terms of a process that begins with tectonic thickening and subduction of oceanic crust, followed by progressive metamorphism and partial melting of the basaltic rocks, producing TTG melts that for the most part were in equilibrium with anhydrous, eclogitic residues (Rapp et al. 2003). Foley et al. (2002) argued that even the earliest continental crust might have formed by melting of amphibolites in subduction zone environments rather than by the melting of eclogites or magnesium-rich amphibolites in the lower parts of thick oceanic crust. Nair and Chacko (2008) proposed a model where oceanic plateaux served as the nuclei for Archaean cratons, and that TTGs originated in intraoceanic subduction systems where thinner oceanic lithosphere subducted beneath the thick oceanic plateaux.

On the basis of isotopic age distribution and εNd data, the most important Archaean crust-forming event was at c. 2.70 Ga, when large volumes of continental crust were produced worldwide. In some theories, this flare-up in crust formation was related to a large-scale mantle overturn event, which gave rise to a large number of plumes (Condie 1998, 2000; Condie and Benn 2006). The period of 2.75–2.65 Ga was also a time period during which the pre-existing continental crust was intensively reworked in various Archaean provinces, as in the Kola and Karelia

Provinces in the Fennoscandian and Superior Province in the Canadian Shield.

Archaean rocks cover roughly one-third of the Precambrian region in the Fennoscandian shield. Most of these rocks are Neoarchaean in age (2.8–2.5 Ga). Mesoarchaean (3.2–2.8 Ga) juvenile or reworked units are essential constituents of the bedrock only locally, and Palaeoarchaean rocks (3.6–3.2 Ga) are, overall, rare. The Neoarchaean (2.75–2.65 Ga) crust-forming event was very strong, including juvenile TTG magmatism, sedimentation, metamorphism and crustal melting producing granites. In this chapter we provide a review of the extant literature on the geology of the Fennoscandian Shield and present new data on the geochemistry, age determinations, metamorphism and palaeomagnetism of the Archaean Karelia Province and the adjacent Belomorian Province in Finland and in Russia. We then discuss the magmatic and tectonic processes that strongly controlled the evolution of these provinces and their present constitution and structure, with an emphasis on the Neoarchaean evolution. All ages referred to are U-Pb zircon ages, unless otherwise stated.

3.2 Geological Setting

The Archaean of Fennoscandia is traditionally divided into Norrbotten, Murmansk, Kola, Belomorian and Karelia Provinces (Fig. 3.1; Slabunov et al. 2006a, b; Hölttä et al. 2008). Lobach-Zhuchenko et al. (2000b, 2005) and Hölttä et al. 2008 subdivided the Karelia Province into three terranes: the Vodlozero, the Central Karelia and the Western Karelia (Fig. 3.1), which they observed with several singular lithological, structural and age characters. We have adopted this provincial division, but use instead the term subprovince to reflect the uncertainty about the nature of the unit contacts. By definition, a tectonostratigraphic terrane should be a fault-bounded crustal block whose geological history differs from that of the surrounding areas (Jones et al. 1983; Jones 1990). The Karelia subprovinces clearly differ in their geological histories, but there is as yet little evidence that they are all bounded by major accretionary faults.

The Karelia subprovinces include Mesoarchaean 3.2–2.8 Ga volcanic rocks and granitoids that are also common in the Vodlozero subprovince, and Neoarchaean (≤2.8 Ga) granitoids and greenstones with locally occurring, recycled Mesoarchaean crustal material (Vaasjoki et al. 1993). The Belomorian province east of the Karelia Province largely consists of 2.93–2.72 Ga TTG gneisses, greenstones and paragneisses, and local occurrences of ophiolite-like rocks and eclogites, which have not been discovered elsewhere in the Archaean parts of the Fennoscandian Shield (Shchipansky et al. 2004; Volodichev et al. 2004; Slabunov 2008; Slabunov et al. 2006a, b). Reflection seismic studies suggest that the Belomorian province comprises a stack of eastward-dipping subhorizontal nappes and thrusts that is separated from the underlying Karelia Province by a major detachment zone (Mints et al. 2004). These nappes probably developed during the Palaeoproterozoic (Mints et al. 2004; Sharov et al. 2010).

3.3 Geochemistry of Granitoids and Migmatitic Amphibolites

Figure 3.2 presents a lithological map of the Finnish part of the Karelia Province, showing that most of the area (c. 80 %) consists of TTGs. The rest is mainly comprised of supracrustal rocks in greenstone belts and sedimentary gneisses, and migmatitic amphibolites in enclaves in the gneissic granitoids. The gneissic granitoids include both true orthogneisses and TTG migmatites with amphibolite and paragneiss mesosomes, as observed in the field.

3.3.1 Granitoids

Archaean granitoids in the Karelia Province can be divided into four main groups on the basis of their field and petrographic characters, major and trace element compositions and age. These are the TTG (tonalite-trondhjemite-granodiorite), sanukitoid, QQ (quartz diorite-quartz monzodiorite) and the GGM (granodiorite-granite-monzo-

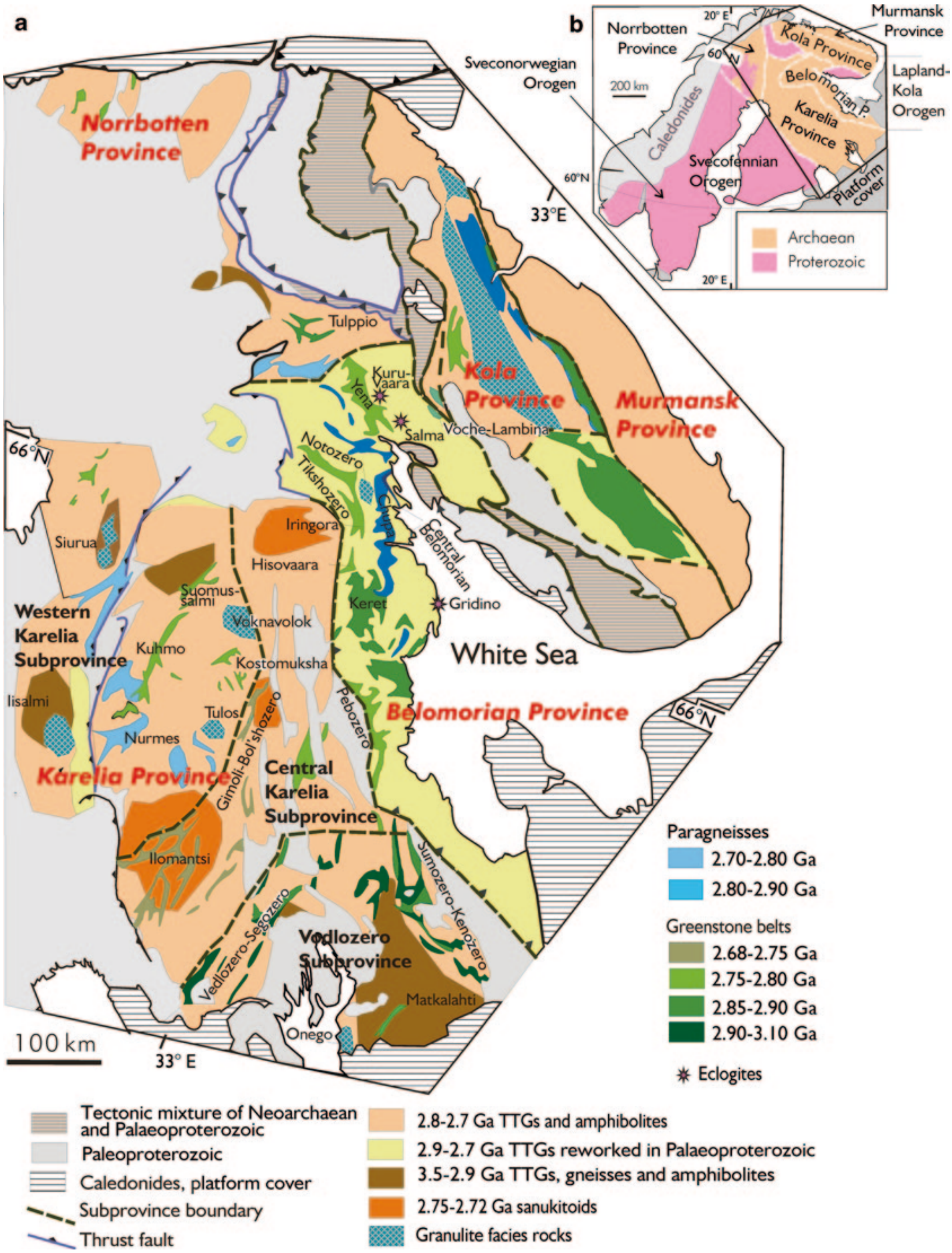

Fig. 3.1 A generalised geological map of the Archaean **a** of the Fennoscandian shield (**b**, inset), modified after Slabunov et al. (2006a) and Hölttä et al. (2008)

Fig. 3.2 A generalised geological map of the Finnish part of the Karelia Province. The inset shows the location of the map area in Finland

granite) groups (Käpyaho et al. 2006; Mikkola et al. 2011a).

3.3.1.1 TTGs

TTGs represent the majority of the Archaean bedrock in the Karelia, and understanding of their origin is thus crucial. It is well recognized that Archaean TTGs share many geochemical features with adakites, silica-rich volcanic and plutonic rocks in volcanic arcs, which are strongly depleted in Y and heavy rare earth elements, low in high-field-strength elements (HFSE) and high in their Sr/Y and La/Yb ratios (Defant and Drummond 1990). The chemical criteria for adakites are listed in the original paper by Defant and Drummond (1990), and later, for example, in Richards and Kerrich (2007), who use the following critical composition to define an adakite: $SiO_2 \geq 56$ wt%, $Al_2O_3 \geq 15$ wt%, $MgO < 3$ wt%, Mg number ~50, $Na_2O \geq$ wt 3.5 %, $K_2O \leq 3$ wt%, K_2O/Na_2O ~0.42, Rb≤65 ppm, Sr≥400 ppm, Y≤18 ppm, Yb≤1.9 ppm, Ni≥20 ppm, Cr≥30 ppm, Sr/Y≥20 and $La_N/Yb_N \geq 20$. Martin and Moyen (2003) divided the adakites into the low silica (<60 wt% SiO_2) and high silica (HSA, >60 wt% SiO_2) groups, relating the HSA group to slab melting with some interaction with mantle-wedge peridotite and the LSA group to melting of wedge peridotites, which were previously metasomatised by slab-derived melts.

Most TTGs at least in the Western Karelia Province fulfil the adakite criteria, apart from the Cr and Ni contents, which are normally below the detection limits of the XRF analysis used, 30 and 20 ppm, respectively. Mg contents are also normally low, in most of the TTGs in the range of Mg# 0.35–0.45, suggesting that they gained minor components from mantle peridotites (Halla 2005; Lobach-Zhuchenko et al. 2005, 2008). Halla et al. (2009) divided the TTGs into two major groups, low-HREE (heavy rare earth elements) and high-HREE TTGs, which likely originated at different lithospheric depths. Here, we follow the proposal of Halla et al. (2009), but instead of the HREE contents—which depend on the source composition and post-melting fractionation, besides melting pressure—we use the $(La/Yb)_N$ ratio, as low-HREE rocks generally have high $(La/Yb)_N$ ratios. We further divide the low-HREE group into Eu-positive and Eu-negative subgroups. Our dataset demonstrates that the two TTG groups are not separated in compositional space, but rather represent the end-members of a full compositional continuity (Fig. 3.3).

Moyen (2011) studied a large set of analyses from rocks generally regarded as TTGs, ultimately dividing these rocks into four groups: the potassic group and the high-, medium- and low-pressure groups of 'proper' juvenile TTGs. The high-pressure group, equivalent to the low-HREE group, has TTGs with high Al_2O_3, Na_2O, Sr and low Y, Yb, Nb and Ta. The TTGs of the low-pressure group, equivalent to the high-HREE group, show opposite values in these respects. The potassic group rocks that show enrichment in K and LIL elements are thought Moyen (2011) to have formed by melting of pre-existing crustal rocks (Moyen 2011). If the criteria for true TTGs include $K_2O/Na_2O < 0.5$ (Martin et al. 2005), approximately 20 % of our samples collected in the field as TTG suite rocks would actually belong to the potassic or transitional TTGs, as originally described by Champion and Smithies (2007). On the other hand, this is a definitional problem, because in granodiorites the K_2O/Na_2O ratio is typically 0.5–1, and those rocks whose K_2O/Na_2O is <0.5 are tonalites and trondhjemites, i.e. TTs rather than TTGs.

The rocks in the low-HREE group are trondhjemitic and tonalitic when classified using normative compositions and the QAPF diagram in Streckeisen and Le Maitre (1979) or the Ab-An-Or diagram in O'Connor (1965).In the Eu-negative subgroup, SiO_2 is generally 62–74 wt%, mg# 30–55 and Al_2O_3 15–17 wt%. Most of the samples have fractionated REE patterns with low HREE and high Sr/Y (median 81). They also have high La_N/Yb_N ratios of 20–120 with a median of 49, showing negative Eu anomalies (Fig. 3.5) and low abundances of compatible elements.

The SiO_2 values in the Eu-positive, low-HREE TTGs typically vary from 68–76 wt%, and they commonly have low abundances of FeO and MgO, as reflected in their near-white to light grey colour in outcrops and samples. Mg# is generally 28–55 and Al_2O_3 14.5–17.5 wt%. These TTGs

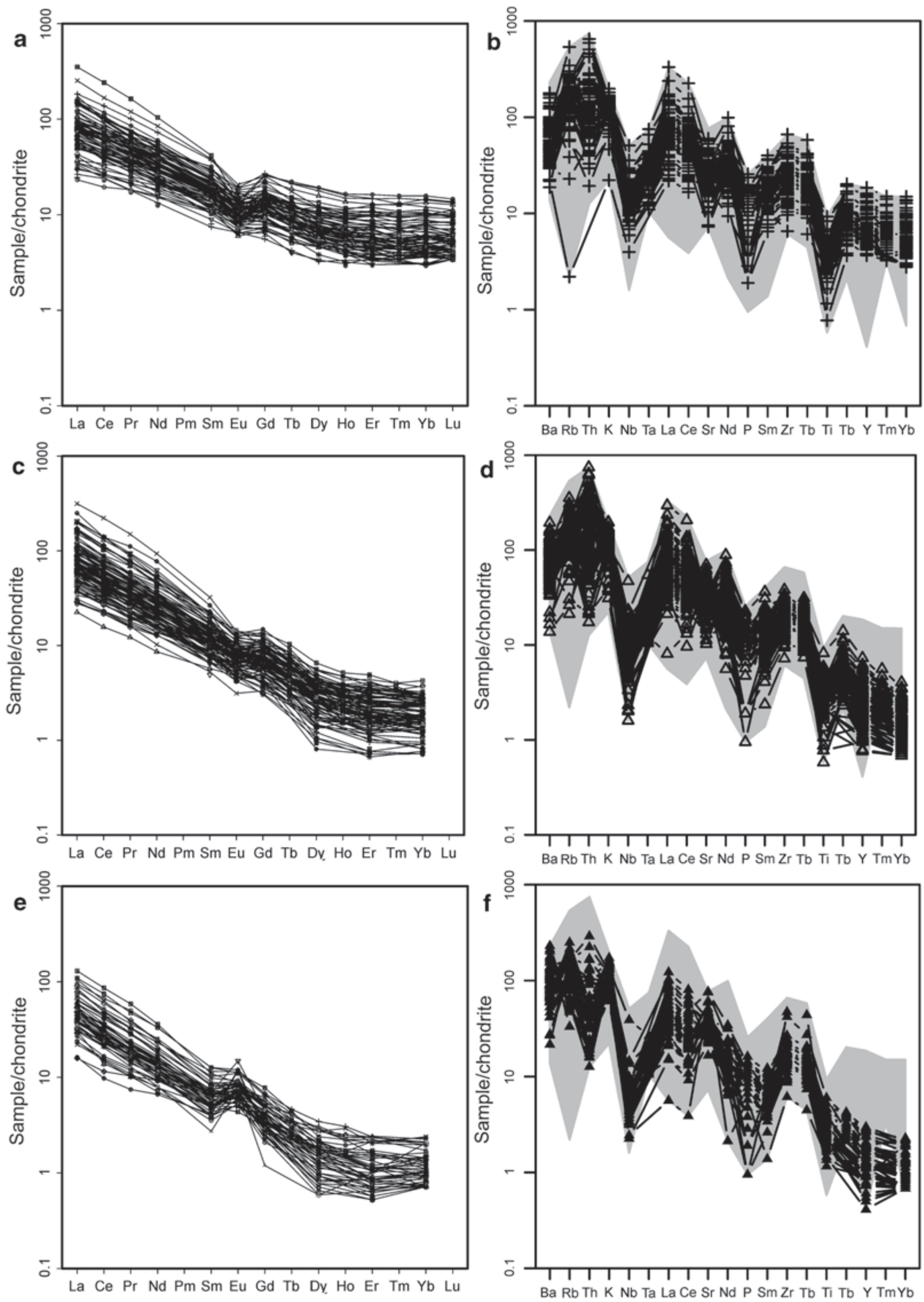

Fig. 3.3 Trace element patterns of TTGs: **a**, **b** = high HREE group, **c**, **d** = low HREE group, **e**, **f** = Eu-positive group. Diagrams represent 255 analyses where the K_2O/Na_2O ratio is <0.5, taken from the Rock Geochemical Database of Finland (Rasilainen et al. 2007). Normalizing values in the REE diagrams (a, c, e) are from Boynton (1984) and in the trace element diagrams (b, d, f) from Thompson (1982). The *shaded* area is for all data. Normalising factors are as in Fig. 3.3

are associated with negative magnetic anomalies on airborne magnetic maps, and especially in the northern part of the Karelia Province in Finland they cover large areas (Fig. 3.2). Most samples show, apart from positive Eu anomalies, strongly fractionated REE patterns with HREE mostly below the detection limits, low Y, Sc and Nb, high Sr/Y, La_N/Yb_N and Zr/Sm ratios and low abundances of compatible elements. In outcrops, part of these rocks show diatexitic migmatite structures and seem to represent almost complete fusion of amphibolites, but there are also many occurrences of this group that appear to be homogeneous orthogneisses.

The high-HREE group is normatively granodioritic and tonalitic. Its SiO_2 content is c. 60–72 wt%, mg# generally 30–55 and Al_2O_3 15–17 wt%. The Al_2O_3/SiO_2 ratio is generally lower in samples of this than the other TTG groups. Rocks in this group have low Sr/Y ratios (median 22) and higher abundances of compatible elements and HREE than the other TTGs. The La_N/Yb_N is <20 with a median of 10. On average, the high-HREE group has higher Nb contents (c. 5–15 ppm) than the low HREE group (c. 1–8 ppm).

3.3.1.2 Sanukitoids

The late tectonic, Neoarchaean sanukitoid intrusions occur in the Western and Central Karelia subprovinces, commonly associated with major shear zones (e.g. Lobach-Zhuchenko et al. 2005; Heilimo et al. 2010, Fig. 3.1). To this date ~20 sanukitoid intrusions have been described from the Karelia Province. The size of the intrusions is generally small, exceptions being the Koitere, Kuusamo and Njuk sanukitoids. The rocks are variably even-grained or K-feldspar porphyritic, and form a series from diorites to tonalites and granodiorites. U–Pb age determinations give an age of 2.74–2.72 Ga for the sanukitoid magmatism, which falls between the last peak of TTG (tonalite-trondhjemite-granodiorite) magmatism, ~2.75 Ga (Käpyaho et al. 2006; Mikkola et al. 2011a) and the age of the GGMs, ~2.70 Ga (granodiorite-granite-monzogranite) and QQ (quartz diorite- quartz monzonites) in the Karelia Province (Käpyaho et al. 2006; Lauri et al. 2011; Mikkola et al. 2011a, 2012).

The geochemistry of the sanukitoids is contradictory: they are enriched in the compatible elements Mg, Ni, and Cr and are also enriched in the incompatible elements LILE (Ba, Sr, and K), and LREE. Stern et al. (1989) gave a strict geochemical definition of sanukitoids: SiO_2 = 55–60 wt.%, MgO = 6 wt.%, Mg# = 60, Sr = 600–1800 ppm, Ba = 600–1800 ppm, Cr = 100 ppm, and Ni = 100 ppm. More recent studies (Lobach-Zhuchenko et al. 2005; Halla 2005) consider sanukitoids as a series of granitoids with high contents of compatible and incompatible elements, at a given SiO_2 content, that separates sanukitoids from the TTG series. The sanukitoid intrusions in the western Karelian Subprovince in Finland are alkali-calcic to calc-alkalic, magnesian, mostly metaluminous, and show geochemical features typical for the sanukitoid series: high content of LILE (K, Ba, and Sr), and high content of mantle-compatible elements (Mg, Cr, and Ni) and high Mg#.

Bibikova et al. (2005) and Lobach-Zhuchenko et al. (2005) have argued that sanukitoid intrusions in the Russian side of the Karelia Province occur as two temporally, spatially, and geochemically different groups, termed as the eastern and western sanukitoid zones. Compiled single-grain zircon U–Pb age data of the Finnish side of the Karelian Province confirm the occurrence of two temporally differing sanukitoid zones throughout the Province (Heilimo et al. 2011). The western zone shows a younger average age of ~2718 Ma, and the eastern zone an older average age of ~2740 Ma with an apparent age difference of ~20 Ma between the zones. Most felsic sanukitoids contain inherited zircon cores with ages up to ~3.2 Ga reflecting the importance of inheritance processes in the petrogenesis of sanukitoids (Bibikova et al. 2005; Heilimo et al. 2011).

The available isotope data from the Neoarchaean mantle-derived sanukitoids of the Karelia Province of the Fennoscandian Shield indicate well-mixed sources and reflect recycling of sediments from crustal reservoirs of different age in subduction processes. Oxygen isotope data from zircon show variable $\delta^{18}O$ values indicating a mantle input and also contributions from sources with high $\delta^{18}O$ values, such as sediments and altered upper oceanic crust (Mikkola et al. 2011a;

Heilimo et al. 2012, 2013). Slab breakoff at the end-stage of subduction has been proposed as a plausible trigger mechanism for sanukitoid magmatism in Karelia Province because it allows contributions from different $\delta^{18}O$ sources as well as explaining the contradictory elemental compositions (Heilimo et al., 2013)

3.3.1.3 QQs

Another important group of non-TTG granitoids comprises rocks whose normative composition is mostly quartz dioritic and quartz monzonitic (QQ), but locally also dioritic and monzodioritic. Rocks belonging to this group are especially found in the Western Karelia subprovince. Most QQ intrusions are located west of the sanukitoids described above, but whether there is a genetic link between the sanukitoid and QQ groups is not yet clear. The QQs include the enderbites in the Iisalmi complex (Hölttä 1997), the Ranua diorite (Mutanen and Huhma 2003) and some smaller plutons (Mikkola et al. 2011a). The QQs have zircon U-Pb ages close to 2.70 Ga. The ε_{Nd} values at 2.7 Ga are positive, for example +1 in the two analyzed enderbites of the Iisalmi complex (Hölttä et al. 2000a) and +0.8 in the Ranua diorite (Mutanen and Huhma 2003). Many compositional features in these rocks also fulfil the definition of adakites, as the SiO_2 content in most analysed samples is 52–62 wt%, Al_2O_3 a high 15.4–20.1 wt%, MgO 2.5–4.0 wt%, mg# 44–54, Na_2O 3.3–5.4 wt%, K_2O 0.8–1.8 wt%, Rb 13–52 ppm, Sr 630–1170 ppm, Sr/Y 21–80, Ba 200–800 ppm, Y < 18 ppm and Yb < 1.9 ppm. Furthermore, Cr in most samples is > 30 ppm and Ni > 20 ppm, except for the values of 10–28 ppm and 6–13 ppm, respectively, in the Ranua diorite. However, the QQs have only moderately fractionated REE patterns with La_N/Yb_N ratios of 6–15, and the enderbites in the Iisalmi area have a typically flat HREE (Fig. 3.4). This type of REE pattern is a distinctive feature of the enderbites and was not observed in other Archaean igneous rocks, apart from 2.74 Ga alkaline rocks in Suomussalmi, which have a similar REE distribution but generally higher REE abundances (Mikkola et al. 2011b). The enderbites have lower Rb, U and Th contents than the other quartz diorites, probably because of the loss in these elements during the granulite facies metamorphism that they underwent at 2.70–2.60 Ga (Mänttäri and Hölttä 2002).

3.3.1.4 GGMs

Granodiorite-granite-monzogranite (GGM) suite rocks dated to 2.73–2.66 Ga are the youngest Neoarchaean rocks occurring in large volumes in the Western Karelia subprovince (Käpyaho et al. 2006). They are relatively weakly deformed, reddish and medium- to coarse-grained rocks. Many of the granites have highly fractionated, HREE-depleted patterns with negative or no Eu anomalies, but some are less fractionated with relatively high HREE (Mikkola et al. 2012). At least part of the granites could represent the melting products of sedimentary gneisses. On the other hand, according to experimental studies, dehydration melting of sodic TTG gneisses can also produce granitic to granodioritic melts (Patiño Douce 2004; Watkins et al. 2007). Skjerlie et al. (1993) have shown experimentally that when interlayered, the melt productivity of tonalite and pelite increases via the interchange of components that lower the required melting temperatures. Anatectic granites, therefore, tend to contain material from two or more different source rocks, which is reflected in their isotopic and chemical compositions (Skjerlie et al. 1993). This could also be the case with the Archaean GGMs, which could represent the melting products of TTGs and paragneisses, and mixing of the derived melts with contemporaneous mafic magmas. The range in the ε_{Nd} (2700 Ma) values of the GGMs is c. −1.5 to +1.0 (Hölttä et al. 2012), indicating that some of them might represent the melting of juvenile 2.72–2.78 TTGs, whereas others could originate from older material. However, in the Suomussalmi area, the average $\delta^{18}O$ values of zircon from GGMs are normally only slightly higher (6.42 ± 0.10) than that of the TTGs, 6.10 ± 0.19. This means that at least in Suomussalmi the GGMs do not necessarily have a significant sedimentary input in their source but rather represent melting products of TTGs (Mikkola et al. 2012). However, in the Suomussalmi area the abundance of exposed

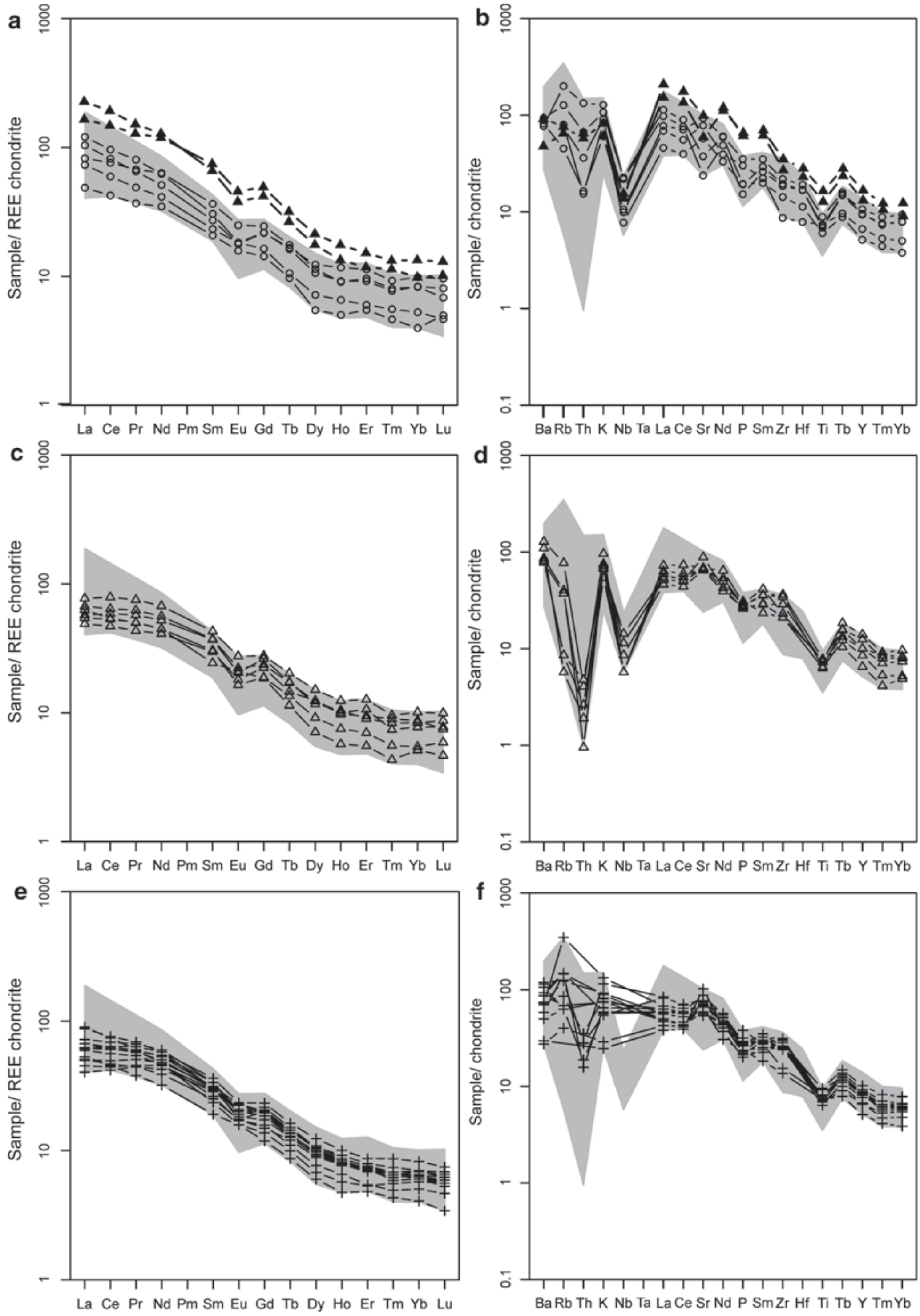

Fig. 3.4 Trace element patterns of QQs. **a**, **b** = Ranua, *black triangles* denote alkali gabbroic inclusions in quartz diorite; **c**, **d** = enderbites of the Iisalmi complex; **e**, **f** = opx-free QQs in Iisalmi north of the granulites. The *shaded* area is for all data

sedimentary gneisses is also low, whereas in other areas they may have had a more significant contribution to GGM genesis.

3.3.2 Amphibolites in gneissic complexes

In the Western Karelia subprovince amphibolites are a ubiquitous component of TTGs in which they can typically be found as layers and inclusions whose widths vary from a few tens of centimetres to tens of metres. Normally, the gneiss-associated amphibolites are all migmatized, the volume of felsic neosome ranging in the exposures from < 10 % up to c. 90 %. Extensively melted and deformed amphibolites closely resemble strongly deformed, plutonic TTGs with amphibolite rafts, making their distinction difficult. In metatexitic amphibolites, palaeosomes have mostly amphibolite facies mineral assemblages, typically hornblende-plagioclase-quartz, commonly with retrograde epidote. In granulite facies areas, amphibolites commonly have coexisting orthopyroxene and clinopyroxene. Garnet-bearing two-pyroxene mafic and intermediate granulites, which are common in the Iisalmi complex, represent high-temperature/medium-pressure equivalents of the common hornblende amphibolites.

The ages of the amphibolites are problematic to resolve, because they commonly appear to contain predominantly metamorphic zircon grains (e.g. Mutanen and Huhma 2003). Intermediate granulites and amphibolites in the Iisalmi complex have been dated for their protoliths at c. 3.2 Ga (Paavola 1986; Mänttäri and Hölttä 2002; Lauri et al. 2011), but this complex is older overall than most other areas in the western part of the Karelia Province.

On the basis of major element composition and using the TAS classification, the amphibolites are classified mostly as basalts and andesitic-basalts. Some amphibolites are andesitic and some have a slightly alkaline character. In the Jensen cation plot, most compositions are in the field of high-Mg tholeiites, and some even in the komatiitic basalt field, the latter probably because of the existence of cumulus olivine.

Basaltic amphibolites (SiO_2 < 52 wt%) fall into two main groups on the basis of their trace element contents. Samples of the first group have flat or LREE-depleted trace element patterns, resembling those of the modern mid-ocean ridge basalts. Further characteristics of these samples are high $(Nb/La)_N$ ratios, low Zr/Y ratios and high Ni and Cr contents (Hölttä 1997; Nehring et al. 2009; Hölttä et al. 2012). Samples of the second group are enriched in LILE and LREE and have lower Nb/La_N ratios and higher Zr/Y ratios than those in the first group (Fig. 3.5). Compatible elements, especially Ni but also Cr, are lower in the LREE-enriched than in the group of LREE-depleted samples.

Condie (2005) used the high field strength element (HFSE) ratios of Archaean basalts to demonstrate their possible mantle source domains, assuming that their magmatectonic framework was broadly similar to that of young oceanic basalts. He noted that on the basis of the Nb/Th, Zr/Nb, Nb/Y and Zr/Y ratios, most non-arc-type Archaean basalts from greenstone belts resemble oceanic plateau basalts, which are thought to originate from plumes variably comprising deep primitive and shallow depleted mantle. Figure 3.6 shows a plot of Nb/Y vs. Zr/Y for the basaltic amphibolites in the Finnish part of the Karelia Province. Given the depletion of granulite facies rocks in U and Th, thorium-based ratios are probably useless for the high-grade migmatitic amphibolites. As Zr/Nb ratios in all samples of the amphibolites are c. 10–30, they probably do not contain recycled oceanic lithospheric material. Most of the analysed amphibolites have LREE-depleted or flat REE patterns, relatively low Zr/Y ratios of c. 1–2, and tendency to plot between the primitive and deep depleted mantle compositions in the diagram in Fig. 3.6. A smaller number of the amphibolites are LREE enriched, have Zr/Y ratios of c. 3–5 and define on the Nb/Y vs. Zr/Y diagram a trend towards enriched sources that could be the lithospheric mantle or the continental crust. In the field, some of the amphibolites show dyke-like relationships with the host TTGs, suggesting that amphibolites in the second group could represent metamorphosed and deformed dykes, whose magmas were chemically reacted with the TTG crust.

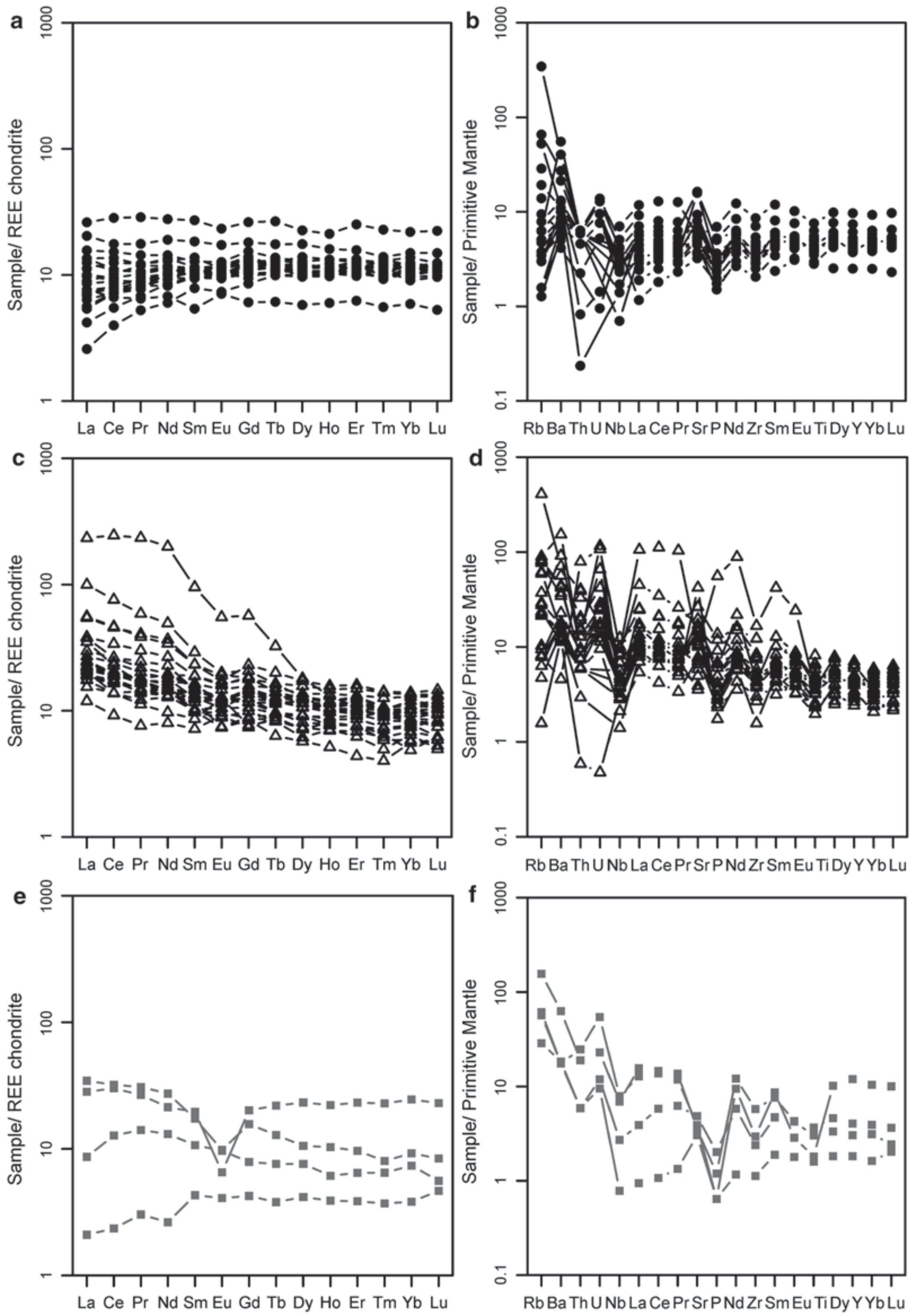

Fig. 3.5 Trace element patterns of basaltic amphibolites. **a**, **b** = LREE-depleted group; **c**, **d** = LREE-enriched group; **e**, **f** = komatiitic basalts. Normalising factors are as in Fig. 3.3

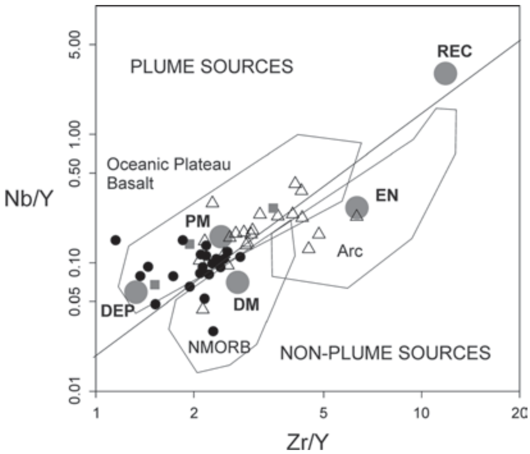

Fig. 3.6 Nb/Y vs. Zr/Y of basaltic amphibolites. *Black circles* = LREE depleted basalts, *triangles* = LREE enriched basalts, *grey squares* = komatiitic basalts. Abbreviations: *PM* primitive mantle; *DM* shallow depleted mantle; *ARC* arc-related basalts; *NMORB* normal ocean ridge basalt; *DEP* deep depleted mantle; *EN* enriched component; *REC* recycled component

This interpretation is supported by the higher LILE contents in these amphibolites compared with the first group of amphibolites, which could be restites of the plateau basalts after melting that produced the TTGs.

3.4 Greenstone Belts

The Karelia and Belomorian Provinces include numerous, generally NNW-trending greenstone belts (Slabunov et al. 2006a). Case studies have proposed distinct formative settings for the individual belts, i.e. an oceanic plateau setting for the Kostomuksha belt (Puchtel et al. 1998), an island arc setting for the Sumozero-Kenozero belt (Puchtel et al. 1999) and a continental rift setting for the Suomussalmi-Kuhmo and Matkalahti belts (Luukkonen 1992; Papunen et al. 2009; Kozhevnikov et al. 2006). A brief description of the best-known belts is given here.

3.4.1 Vedlozero–Segozero Greenstone Belt

The Vedlozero–Segozero greenstone belt is located at the western margin of the Vodlozero subprovince and consists of three volcanic complexes whose ages are 3.05–2.95 Ga, 2.90-2.85 Ga and 2.76-2.74 Ga (Svetov and Svetova 2011). The 3.05–2.95 Ga complex comprises two suites (Svetov 2009). The basalt–andesite–dacite–rhyolite (BADR) suite contains pillow and amygdaloid lavas, fragmental intermediate and felsic lavas, tuffs and dikes. The age of subvolcanic dacitic andesite is 2995 ± 20 Ma (Sergeev 1982); andesitic lavas are dated at 2945 ± 19 Ma (Ovchinnikova et al. 1994) and 2971 ± 59 Ma (Svetov 2010). The rocks belong to the calc-alkaline series with adakitic geochemical characteristics (Svetov 2009). This association is a relict of the oldest volcanic island arc system known in the Fennoscandian Shield (Svetov 2005, 2009).

Another suite consists of komatiites and basalts. The Sm–Nd isochron age of this association is 2921 ± 55 Ma; ε Nd (t) = +1.5 (Svetov et al. 2001). The sequence is composed of diverse lavas, including pillow, variolitic, and spinifex varieties. Pyroclastic interlayers occupy less than 5 % of the sequence's volume (Svetova 1988; Svetov et al. 2001). The Al-undepleted pyroxenites and basaltic komatiites and tholeiitic basalts are predominant, their intrusive equivalents consisting of magnesian gabbro and serpentinized ultramafic rocks. This association is interpreted to have evolved in a backarc basin of a suprasubduction zone environment (Svetov 2005).

The 2.90–2.85 Ga complex consists of andesites, dacitic andesites, dacites and rhyolites—i.e. ADR-type volcanic rocks and various kind of metasediments. The Janis palaeovolcano, for example, consists of lava breccias, lavas, and block agglomerate tuffs; a feeder is filled with subvolcanic dacite. The chemogenic silicites were deposited in the crater lake. Tuffs, tuffites, tuffstones, tuffaceous conglomerates, and silicites occur at the periphery of the paleovolcano. The subvolcanic intrusions are composed of dacite and rhyolite. The dacitic andesites erupted in a subaerial environment (Svetova 1988). Lavas were inferior in abundance to the products of volcanic explosions. Existing dates for the felsic volcanic rocks give ages of 2860 ± 15 Ma (Samsonov 2004) and 2866 ± 11 Ma (Svetov 2010); dykes are dated at 2862 ± 45 Ma (Ovchinnikova et al. 1994). The ADR-type volcanic rocks belong to the calc-alkaline series, the same rocks with adakitic and Nb-enriched andesite geochemical characteristics (Svetov 2005, 2010. This complex was evidently formed in a suprasubduction setting of the Andean type active continental margin (Svetov 2005).

The ADR-type, 2.76–2.74 Ga volcanic complex is found in the Malasel'ga area. The existing dates for these felsic volcanic rocks reveal zircon ages of 2765 ± 13 and 2743 ± 12 Ma (Svetov 2010).

3.4.2 Sumozero–Kenozero Greenstone Belt

The Sumozero– Kenozero greenstone belt is located at the eastern margin of the Vodlozero subprovince. Two volcanic complexes of different ages are recognized in Sumozero–Kenozero greenstone belt, the c. 3.00–2.90 Ga komatiite–basalt complex and the 2.88 Ga complex consisting of BADR and adakitic volcanic rocks (Rybakov et al. 1981).

The Sm–Nd whole rock isochrons for komatiite- basalt associations yield ages of 2.91–2.96 Ga (Sochevanov et al. 1991; Lobach-Zhuchenko et al. 1999; Puchtel et al. 1999). The komatiite–basalt complex consists mainly of metakomatiitic lavas, locally with spinifex texture, and pillowed tholeiitic basalts. Interlayers of felsic volcanic and tuffaceous rocks, and graphite-bearing schists, as well as crosscutting mafic and felsic dykes and plagiogranite bodies are found. Komatiites (~30 wt % MgO in the spinifex zone) belong to the Al-undepleted group. The komatiites and tholeiitic basalts have positive Nb anomalies and their REE patterns are similar to those of the MORB-type rocks. In terms of isotopic and geochemical signatures, the komatiite–basalt association is comparable with oceanic plateau complexes, genetically related to the mantle plumes (Puchtel et al. 1999).

The younger complex of Sumozero–Kenozero greenstone belt consists of metavolcanic BADR rocks with interlayers of carbonaceous and carbonate schists and quartzite. There also exist subvolcanic rhyolites and adakites (Rybakov et al. 1981; Puchtel et al. 1999). The ages of the BADR and adakitic rocks are 2875 ± 2 Ma and 2876 ± 5 Ma, respectively. The BADR and adakitic rocks are thought to have formed in a subductional setting as the products of melting of a mantle wedge and a subducting slab, respectively (Puchtel et al. 1999).

3.4.3 Matkalahti Greenstone Belt

The Matkalahti greenstone belt is situated in the central, poorly exposed part of the Vodlozero subprovince. The belt is composed of alternating metasedimentary rocks (quartz arenite, graywacke, carbonaceous schist) and mylonitized basaltic–komatiitic metavolcanic rocks with rare interlayers of foliated felsic metavolcanic rocks (Kozhevnikov et al. 2006). The age of detrital zircon grains from quartz arenite and graywacke varies from 3.33 to 2.82 Ga indicating that this complex is younger than 2.82 Ga. The complex may represent an intracratonic rift assemblage (Kozhevnikov et al. 2006).

3.4.4 Kuhmo Greenstone Belt

The Kuhmo greenstone belt forms the central part of the c. 220-km-long Suomussalmi-Kuhmo-Tipasjärvi greenstone belt in the Western Karelia subprovince (Fig. 3.2). The supracrustal succession in the Kuhmo belt starts with rhyolitic-dacitic lavas and pyroclastic rocks, whose depositional basement and original thickness are unknown. Felsic volcanic rocks occur in two age groups, 2.84–2.82 Ga and c. 2.80 Ga (Huhma et al. 2012a). The felsic volcanics are overlain by an up to one-kilometre-thick sequence of tholeiitic pillow lavas and hyaloclastites, with sporadic layers of Algoma-type BIFs. The tholeitic strata are overlain by a sequence of komatiites (total thickness ~500 m), komatiitic basalts (~300 m), interlayered high-Cr basalts (~250 m) and komatiites, high-Cr basalts (~250 m) and finally pyroclastic intermediate-mafic volcanics (Papunen et al. 2009).

The Kuhmo greenstone belt is bounded by TTGs, sanukitoids, QQ and GGM suite plutons. Several studies have suggested that the Kuhmo greenstone succession was deposited on older continental crust represented by the TTGs (Martin et al. 1984; Luukkonen 1992; Papunen et al. 2009). However, no unconformity or superposition relationship between the TTGs and the greenstone belt supracrustal units has ever been demonstrated. Nor is the concept of an old sialic basement supported by the more recent isotope geochemical and age data (e.g. Käpyaho et al. 2006; Huhma et al. 2012a). With the exception of one sample from the palaeosome of a migmatite c. 30 km east of the Kuhmo greenstone belt (2942 ± 6 Ma; Käpyaho et al. 2007), all samples from the surrounding TTGs yield crystallisation ages younger than the volcanic rocks of the greenstone belt. Furthermore, Käpyaho et al. (2006) concluded, based on an extensive Sm-Nd isotope study, that the plutonic rocks in the Kuhmo area represent relatively juvenile material without a major input from significantly older crust. Neither do the εNd from + 1.2 to + 2.4 of intermediate and felsic volcanic rocks in Kuhmo (Huhma et al. 2012b) indicate that they would contain significant amounts of older crustal material, as one could expect if they originated in a narrow continental rift setting as suggested by Papunen et al. (2009).However, in the northern part, the Suomusalmi section of the belt both volcanic rocks and granitoids display negative initial εNd values and oldest volcanic rocks yield ages close to 2.95 Ga (Mikkola et al. 2011a; Huhma et al. 2012a, b).

The Kuhmo komatiites and komatiitic basalts have generally flat primitive mantle normalised patterns showing only moderate depletion in LREE (Fig. 3.7), as is common for Al-undepleted komatiites. Owing to multiple episodes of post-eruption alteration, the Kuhmo komatiites are disturbed to various degrees in their isotope systems. For example, a Sm-Nd isotope study by Gruau et al. (1992) has shown that the Kuhmo komatiites produce a Sm-Nd isochron yielding an age of c. 1.9 Ga, indicative of resetting of the Sm-Nd isotope system of these Archaean rocks during a Proterozoic metamorphic event.

In terms of Nb/Y and Zr/Y ratios, Kuhmo komatiites plot close to the demarcation line between plume and non-plume source compositions (Fig. 3.8c). Most samples cluster between primitive mantle (PM) and shallow depleted mantle (DM) compositions, and there are many plottings towards the deep depleted plume component (DEP). In the Zr/Nb vs. Nb/Th plot (Fig. 3.8d), the Kuhmo komatiites also plot close to the oceanic plateau basalts, but at somewhat higher Zr/Y values, probably indicating minor pre-eruption fractionation of chromite. Overall, the Kuhmo komatiites show an affinity to oceanic plateau basalts derived from a slightly depleted primitive mantle (PM)-type source, with a minor deep plume signature. Importantly, these komatiites clearly have not been derived from a depleted MORB-type source, and there is also little indication of enriched and recycled mantle components in their source, or input from continental crust during their ascent and eruption.

3.4.5 Kostomuksha Greenstone Belt

The Kostomuksha greenstone belt consists of basalt–komatiite, rhyolite–dacite and banded iron formation (BIF) suites (Kozhevnikov 2000; Lobach-Zhuchenko et al. 2000a; Rybakov et al. 1981; Rayevskaya et al. 1992). The belt hosts an economic BIF that is under active mining.

Existing Sm–Nd isochron ages of basalt and komatiite are 2.84 (Puchtel et al. 1998) and 2.81 Ga, respectively, and the zircon age (SHRIMP-II) of the tuff coeval with the this basalts is 2792 ± 6 Ma (Kozhevnikov et al. 2006). In their geochemical signatures tholeiitic basalts and komatiites are comparable with oceanic plateau basalts.They are depleted in Th and LREE ((La/Sm)N = 0.66), characterized by a flat HREE pattern ((Gd/Yb)N = 1), a chondritic Ti/Zr ratio, a slightly positive Nb anomaly, and an average εNd(*t*) of + 2.8 (Puchtel et al. 1998). The evolved and unfractionated komatiitic lavas host rare tuffaceous units, and peridotitic sills are widespread. The Al-depleted komatiites are depleted in LREE ((La/Sm)N = 0.48; (Gd/Yb)N = 1.2, Puchtel et al. 1998). The Re–Os isotopic data (γ^{187}Os = + 3.6 ± 1.0) suggest that they were derived from a plume that arose at the core-mantle boundary (Puchtel et al. 2001).

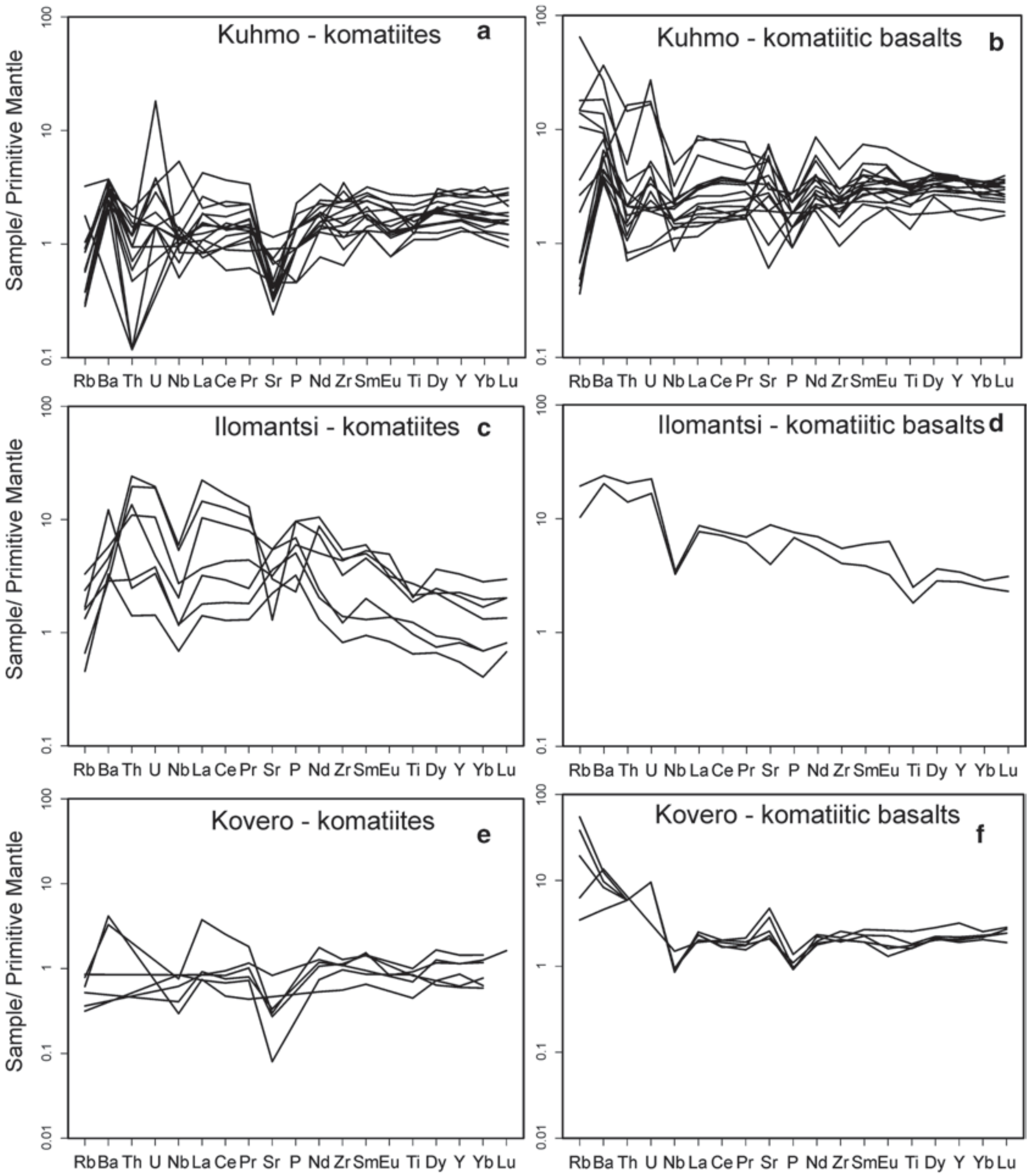

Fig. 3.7 Primitive mantle-normalized trace element patterns of komatiites and komatiitic basalts in Kuhmo, Ilomantsi and Kovero. Normalizing values are from Sun and McDonough (1989)

The rhyolite–dacite suite consists of various lavas, tuffs, and tuffites with interlayers of carbonaceous schist and jaspilite; dikes and subvolcanic intrusions also exist. Age estimates of 2790±21 Ma (Bibikova et al. 2005) and 2795±10 Ma (Lobach-Zhuchenko et al. 2000a) have been obtained for this suite. Rayevskaya et al. (1992) classified the volcanic rocks as calc-alkaline rhyolites and dacites (sporadic andesites) with distinct negative Nb anomalies. Two geochemically different groups of rocks are distinguished: (1) high-K rocks with highly fractionated REE patterns ((La/Sm)N=4.9–6.2, (Gd/Yb)N=2.5–3.7) and low Al_2O_3, Sr, and Y and

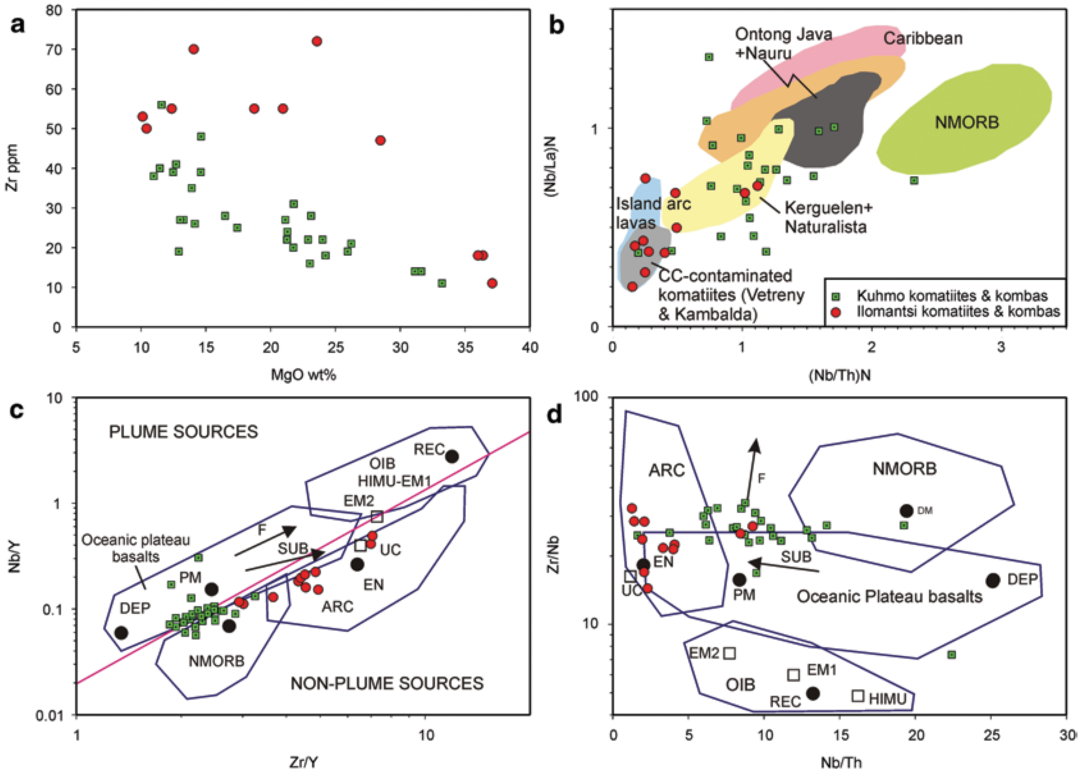

Fig. 3.8 Zr-Nb-Y-Th-La ratios in the komatiites and komatiitic basalts of Kuhmo and Ilomantsi. Abbreviations (Condie 2005): *UC* upper continental crust; *PM* primitive mantle; *DM* shallow depleted mantle; *HIMU* high mu (U/Pb) source; *EM1* and *EM2* enriched mantle sources; *ARC* arc-related basalts; *NMORB* normal ocean ridge basalt; *OIB* oceanic island basalt; *DEP* deep depleted mantle; *EN* enriched component; *REC* recycled component. *Arrows* indicate the effects of batch melting (F) and subduction (SUB)

(2) an adakitic group with fractionated REE patterns ((La/Sm)N = 2.9–5.3, (Gd/Yb)N = 1.4–2.1)) and high Y and Sr (Samsonov et al. 2005). These rocks have plutonic analogues in the area. The εNd(t) values are from −6.21 to +1.59 (Bibikova et al. 2005). These geochemical signatures indicate that the volcanic rocks were related to at least two mantle and ancient crustal sources, and that they likely formed in an arc setting (Samsonov et al. 2005; Kozhevnikov 2000).

The sedimentary suite consists of conglomerates, jaspilites including the BIFs and greywackes with thin interlayers of andesite–dacite–rhyolite tuffs and komatiite bodies. The terrigenous sedimentary rocks are poorly differentiated, immature, and comparable with graywackes from the Neoarchean greenstone belts. The provenance was composed of approximately equal abundances of basic and felsic rocks (Mil'kevich and Myskova 1998). The U-Pb age on zircon from dacitic tuffite interlayered with the sedimentary suite is 2787 ± 8 Ma (Bibikova et al. 2005), indicating the deposition time of the sedimentary suite.

All three rock suites in the Kostomuksha greenstone belt were formed at c. 2.80–2.79 Ga but evidently in different geodynamic settings. It is widely thought that the greenstone belt is a tectonic collage that arose as a result of accretion of almost coeval mantle- plume volcanic plateau and subduction-related igneous and sedimentary rocks (Kozhevnikov et al. 2006; Kozhevnikov 2000; Puchtel et al. 1998; Samsonov 2004).

3.4.6 Ilomantsi and Gemoli-Bol'shozero Greenstone Belts

In the Ilomantsi greenstone belt and in its NNE continuation, the Gemoli-Bol'shozero greenstone belt are located in the southwestern part of the

Central Karelia subprovince. The supracrustal sequence is dominated by sedimentary rocks intercalated with less abundant komatiites, tholeiites, low-Ti tholeiites, andesites, dacites and banded iron formations. The lowermost unit in Ilomantsi starts with mafic pillow lavas, but predominantly consists of felsic pyroclastic and epiclastic sedimentary rocks. The depositional environment was dominated by two distinct but overlapping felsic volcanic complexes, probably locally subaerial, but nevertheless developed within relatively deep, turbidite-dominated basins (Sorjonen-Ward 1993). Thin tholeiitic intercalations and komatiitic sheet flows occur in the upper part of the succession, typically associated with banded iron formations. The komatiites are generally massive recrystallized in structure, and only locally preserve such relict features as cumulus textures or flow top breccias (O'Brien et al. 1993). No primary silicate minerals are preserved in the komatiites, as they have been pervasively altered and recrystallised to tremolite-chlorite-serpentine rocks or chlorite-talc-rich schists. Locally, the komatiites are demonstrably intercalated with felsic volcanic rocks.

Komatiitic rocks from Ilomantsi have the characteristics of the aluminium-undepleted, Munro-type komatiites, having average Al_2O_3/TiO_2 ratios of c. 17.4, but there is considerable scatter. Komatiites and komatiitic basalts of the Kovero greenstone belt, which flanks the Ilomantsi belt in the SW, have high Al_2O_3/TiO_2 ratios, reflecting their low TiO_2 content at a given Al_2O_3 and MgO level (Hölttä et al. 2012). Ni, Co and Cr increase as a function of MgO and Mg# in a way that is indicative of low-pressure fractionation of olivine and chromite being the main factor controlling the compositional variation. This is consistent with the relatively differentiated nature of the lavas, as chromite is generally undersaturated in komatiites with >c. 25 % MgO (Barnes and Roeder 2001). Kovero komatiites seem to differ from the Ilomantsi and Kuhmo komatiites, having overall higher Cr/MgO ratios.

Ilomantsi komatiites have highly fractionated patterns with high LREE/HREE ratios and distinct negative Nb-Ta and Ti-anomalies (Fig. 3.7). The close similarity in trace element patterns between the komatiites and associated rhyolites and dacites (O'Brien et al. 1993) is a clear indication of extensive interaction of the komatiites with the felsic volcanics. Interestingly, the komatiitic basalts from Ilomantsi show less fractionated incompatible trace element patterns than the komatiites, inconsistent with derivation of the basaltic magmas from the komatiites by crystal fractionation. A likely explanation is that the komatiitic basalts evolved from the komatiitic magma already prior to the eruption, within transient storage chamber(s) at depth. During the subsequent eruptions, the komatiites assimilated more felsic magmas than the komatiitic basalts, possibly due to their higher eruption temperatures.

Vaasjoki et al. (1993) reported a TIMS U-Pb age of 2754 ± 6 Ma on zircon from a plagioclase-phyric andesite that represents the stratigraphically lowermost units of the Ilomantsi greenstone belt. The majority of detrital zircon grains in the nearby metasedimentary rocks are of the same age (Huhma et al. 2012a). Previous studies on the Ilomantsi belt have documented a close relationship between volcanism, sedimentation, deformation and pluton emplacement (Sorjonen-Ward 1993), implying rapid c. 2.75 Ga crustal growth in the region. All exposed contacts between the supracrustal rocks and granitoids have been interpreted as intrusive (Sorjonen-Ward and Luukkonen 2005), and the granitoids cannot, therefore, represent the basement to the Ilomantsi greenstone belt, nor a dominant source of the material in the supracrustal sequences.

3.4.7 Keret Greenstone Belt

The Keret greenstone belt is located in the western part of Belomorian Province and consist of three complexes whose ages are 2.88–2.82 Ga, 2.80–2.78 Ga and c. 2.70 Ga. The Keret'ozero greenstone complex is the oldest. It consists of three metavolcanic suites which are komatiite-tholeiite suite, intermediate-felsic suite and andesite basalt-basalt suite (Slabunov 1993, 2008).

Komatiite-tholeiite metabasalts are classified as Na-tholeiite series. High-Mg rocks are classified as Al-undepleted LREE-enriched komatiites and komatiitic basalts having 10–37 wt.% MgO, 0.19–0.9 wt.% TiO_2, $Al_2O_3/TiO_2 \approx 20$,

$CaO/Al_2O_3 = 0.64–0.9$ and $Zr/Y = 2–3$. This suite represents an oceanic plateau setting (Slabunov 2008).

The intermediate and felsic suite consists of metatuffs, metalava and subvolcanic bodies. The rocks vary in composition from andesite-basalt to rhyolite, with andesites and dacites being predominant. Geochemically the rocks in this suite are classified as calc-alkaline island-arc volcanic rocks, which were produced in a subduction zone. U-Pb datings on zircon give ages of 2877 ± 45 and 2829 ± 30 Ma (Bibikova et al. 1999).

The andesite basalt-basalt suite differs from the underlying unit in the predominance of basalts. Essential constituents are Cr- and Ni-rich metasedimentary interbeds, which could have been produced by breakdown of intermediate-felsic volcanic rocks, basaltic lavas and komatiites. These rocks are geochemically comparable with mature island-arc volcanic rocks. Greywackes of such composition are typically formed in ensimatic island-arc settings (Slabunov 2008, 2010).

Hisovaara greenstone complex is located in the northern part of the Keret belt and is dated at c. 2.80–2.78 Ga. It is subdivided into six suites (Thurston and Kozhevnikov 2000; Bibikova et al. 2003; Kozhevnikov et al. 2005; Kozhevnikov and Shchipansky 2008). The ultrabasic-boninitic-basaltic suite is composed of ultrabasic rocks (peridotitic cumulates), tholeiitic metabasalts, high-Mg basalts and komatiites with a 0.5–1 m thick metaboninite interbed and high-Ti ferrobasalts (Shchipansky et al. 2004). This suite is interpreted as a fragment of a suprasubduction ophiolite, the best-preserved portion of which has been described from the adjacent Irinogora area of the Tikshozero belt.

The andesitic suite consists of biotite-epidote-amphibole schists, in which amygdaloidal, massive and glomeroporphyric textures are locally preserved. The age of these andesites is 2777 ± 5 Ma (Bibikova et al. 2003). These rocks are extremely rich in Na_2O (Kozhevnikov 2000). In geochemical characteristics they are similar to tholeiitic andesites from initial intraoceanic island arcs (Bibikova et al. 2003).

The age of volcanic-sedimentary suite is estimated at 2778 ± 21 Ma (Bibikova et al. 2003). Thin intercalations of carbonaceous varieties and iron-cherty and aluminosilicate rocks and lenses of homogeneous metasomatic quartz-kyanite rocks are encountered. The latter are assumed to have been produced by hydrothermal processes that accompanied volcanism (Kozhevnikov 2000; Bibikova et al. 2001b). The U-Pb dating on zircon grains from quartz-kyanite rocks shows give an age of c. 2.77 Ga (Bibikova et al. 2001c). Abundant subvolcanic andesite, dacite and rhyolite bodies or dykes were identified in volcanic-sedimentary suite. These felsic rocks are compositionally comparable to adakites (Bibikova et al. 2003).

The suite of coarse clastic volcanics consists of agglomerate tuffs, oligomictic conglomerates, volcanic conglomerates with tuffaceous matrix and lava breccias that vary in composition from andesite to rhyodacite (Kozhevnikov 2000; Kozhevnikov and Shchipansky 2008). This suite is interpreted to represent a pull-apart basin (Kozhevnikov 2000).

An upper mafic suite consists of metabasalts with a well-preserved pillow texture. Sheet-like bodies of ultramafic composition occur only locally. Metabasalts rest on the underlying rocks along an angular unconformity (Kozhevnikov 2000).

A sedimentary-volcanic suite of c. 2.70 Ga consisting mainly of quartz arenites was identified in the Hisovaara area (Thurston and Kozhevnikov 2000). The protolithis of the sandstones were deposited between 2.71–2.69 Ga, as shown by U-Pb ages of detrital zircon grains (Kozhevnikov et al. 2006).

3.4.8 Tikshozero Greenstone Belt

The Tikshozero greenstone belt is located in the western part of the Belomorian province north of the Keret belt, and it consists of three complexes whose ages are 2.80–2.78 Ga, c. 2.75 Ga and 2.74–2.72 Ga.

The 2.80–2.78 Ga greenstone complex consists of intermediate-felsic volcanic and subvolcanic suite, suprasubduction ophiolite suite, metabasalt suite, metagreywacke suite, coarse clastic suite and sedimentary-volcanic suite. Well-preserved ophiolitic fragments constitute an important part of this complex (Shchipansky et al. 2004) and make up a significant component of gently NNE-dipping thrust sheets. These sheets consist of mafic metavolcanics which are analogues to the upper tholeiites and boninites of Hisovaara. Ophiolitic rocks are thrust over on island-arc type intermediate and felsic calc-alkaline metavolcanic rocks and related volcano-sedimentary rocks, which may be regarded as a paraautochthonous basement for the ophiolitic thrust sheets. Supracrustal rocks of the Iringora areal unit were subjected to intense deformation and metamorphism, but their primary volcanic and sedimentary structures are locally preserved.

A c. 2.75 Ga unit within the Hisovaara greenstone complex consists predominantly of intermediate volcanic rocks. Amygdaloidal and pillow lavas, sedimentary-volcanic rocks, and minor metabasalts and high-Mg basic rocks make up this unit. Porphyritic basalts from this units are dated at 2.75 Ga (Alekseev et al. 2004).

The 2.74–2.72 Ga units in the Hisovaara greenstone complex, Tikshozero belt consists of three suites represented by the komatiite-tholeiite suite, intermediate-felsic volcanic suite and sedimentary-volcanic suite (Stepanov and Slabunov 1989; Levchenkov et al. 2003). On the basis of their composition, the tholeiitic basalts of the komatiite-tholeiite suite are similar to modern N-MORB basalts, with their REE content 8–15 times chondritic with almost flat REE patterns ($(La/Yb)_N = 0.9$–1.0). Komatiites in this suite represent Al-undepleted type (Al_2O_3/TiO_2 18–25, $CaO/Al_2O_3 = 0.7$–1.6; $Zr/Y = 2$–2.8), and they are depleted in LREE $(La/Sm)_N = 0.7$–0.8).

The intermediate-felsic volcanic suite consists of amphibole-biotite schists. Well-preserved agglomerate-tuff and thinly-laminated felsic tuffaceous rocks are common. On the basis of their chemical compositions, intermediate-felsic rocks are calc-alkaline andesite basalts, andesites and rhyolites. They show a low Sr/Y ratio (about 20–25), Y concentration varying from 20 to 30 ppm. They exhibit a differentiated REE spectrum ($(La/Yb)_N$ c. 11–17) and a negative Nb anomaly relative to Th and La, typical of island-arc volcanic rocks. The intermediate-felsic volcanic rocks are dated at 2.74 Ga (Levchenkov et al. 2003). The volcanic rocks were slightly contaminated with older crustal material, as indicated by their Sm-Nd systematics (ε_{Nd} (2.73) ranging from +0.9 to −1.1; t_{DM} = ca. 3.0 Ga) and by the presence of c. 2.80 Ga zircon cores (Mil'kevich et al. 2007).

The sedimentary-volcanic suite consists of leucocratic (garnet-kyanite)-biotite-muscovite schist—greywacke that is locally tourmaline-bearing, with interbeds of biotite-amphibole schist—intermediate-felsic tuffs or lavas. The age of the lavas is estimated at 2.72 Ga Ma (Mil'kevich et al. 2007).

3.4.9 Central Belomorian Greenstone Belt

The Central Belomorian greenstone belt consists of amphibolites with intercalated serpentinite, orthopyroxene and hornblendite bodies. They contain no relics of primary textures, and are therefore classified on the basis of petrographic and geochemical characteristics.

Metabasalts (amphibolites) are similar in chemical composition to tholeiites, and are comparable in many respects to MORB, although some varieties of metabasalts are comparable with oceanic-island basalts (Bibikova et al. 1999; Slabunov 2008). The Sm-Nd systematics of these metabasalts (ε_{Nd} (2.85 Ga) = +2.3) shows that their protolith was not contaminated with older crustal material. This inference is consistent with their formation in an oceanic setting (Slabunov 2008).

Relict olivine (86–81 % *Fo*), orthopyroxene (89–85 % *En*) and spinel (iron-rich ferrichromite with 21 % Cr_2O_3) grains are preserved in serpentinites. The olivines are comparable in composition with those from the cumulate peridotites of a gabbroid ophiolite complex and deep cumulate peridotites. The serpentinites seem to have been formed after dunites and harzburgites (Stepanov et al. 2003). They are depleted in LREE ($(La/Yb)_N = 0.52$), but some varieties show a U-shaped REE distribution pattern. The Sm-Nd isotopic characteristics of the serpentinites (ε_{Nd} (2.85 Ga) = +1.9) rule out any crustal contamination, and suggest that they were derived from a depleted mantle, which is also consistent with their origin as part of an oceanic lithosphere (Slabunov 2008).

The age of the oceanic lithosphere of the Central Belomorian greenstone belt is estimated at 2.88–2.84 Ga on the basis of the U-Pb age on zircon from oceanic trondhjemites (2878 ± 13 Ma, Bibikova et al. 1999) and of the age of early metamorphism in metagabbroic amphibolites from the Lake Seryak area (2836 ± 49 Ma, Slabunov et al. 2009). This age agrees within analytical error with the age of the diorite massif (2.85 ± 0.01 Ga), which crosscuts the unit (Borisova et al. 1997). Based on geological, isotopic and other geochemical data from mafic-ultramafic rocks the Central Belomorian greenstone belt is interpreted as a fragment of a Mesoarchaean oceanic association (Lobach-Zhuchenko et al. 1998; Bibikova et al. 1999; Slabunov 2008).

3.4.10 Chupa Paragneiss Belt

The Chupa paragneiss belt of the Belomorian province consists of migmatized kyanite-garnet-biotite and biotite gneisses. Fine-grained garnet-biotite gneisses, relics of the least altered primary rocks, occur as small lenses of metasediments (Bibikova et al. 1999, 2001a, 2004; Myskova et al. 2003). Based on geochemical characteristics (enrichment in Ni, V, Co and Cr), paragneisses are reconstructed as metagraywackes produced by breakdown of felsic volcanics and mafic–ultramafic rocks in a fore-arc basin environment. Small intermediate and felsic volcanic (dacite-dominated) calc-alkaline interbeds, comparable with island-arc volcanics, and scarce tholeiite bodies are encountered in the graywackes. These observations collectively support the fore-arc basin deposition interpretation (Bibikova et al. 1999; Myskova et al. 2003).

Greywackes in the Chupa paragneiss belt likely formed in the interval 2.87–2.85 Ga, because the age of the cores of detrital zircons is 3.2–2.9 Ga, and the age of the earliest metamorphogenic zircons is 2.86–2.78 Ga. The U-Pb zircon age of the metadacites in the belt is 2870 ± 20 Ma (Bibikova et al. 2004).

3.5 Radiometric Age Determinations from the Karelia Province in Finland

3.5.1 U-Pb

During the past decades, a large number of thermal ionisation mass spectrometry (TIMS) U-Pb age determinations on zircon from Archaean rocks have been carried out at the Isotope Laboratory of the Geological Survey of Finland (GTK). Recently, the secondary ion mass spectrometer (SIMS) of the Nordsim laboratory and multiple-collector inductively coupled plasma mass spectrometer (LA-MC-ICPMS) of GTK have also been used in the age determination of Archaean rocks. Figure 3.9 presents most of the zircon age data available from plutonic and volcanic rocks in the Finnish part of the Karelia Province. It is evident from the diagram that the ages of the TTGs range mostly between 2.83–2.72 Ga, and within this range cluster are two groups separated by a c. 20 Ma time gap; in the older group, TTGs are 2.83–2.78 Ga, and in the younger group 2.76–2.72 Ga. The > 2.78 Ga TTGs occur almost exclusively outside the Ilomantsi complex (Fig. 3.9). In the Ilomantsi greenstone belt, volcanic rocks and related dykes are 2.76–2.72 Ga. From the Ilomantsi complex there have thus far been only two observations of Mesoarchaean rocks (Huhma et al. 2012a), consistent with observations from the Central Karelia subprovince in Russia (Lobach-Zhuchenko et al. 2005; Bibikova et al. 2005; Slabunov et al. 2006a), suggesting that most of the Central Karelian crust is relatively young, c. 2.76–2.72 Ga. However, some porphyritic dykes that intruded into mafic volcanic rocks have older, c. 3.0 Ga zircon populations (Vaasjoki et al. 1993). Two datings for felsic volcanic rocks from the Kovero greenstone belt SW of Ilomantsi give ages of c. 2.88 Ga (Huhma et al. 2012a). These results indicate that the Ilomantsi greenstone belt is not a completely juvenile Neoarchaean formation, but includes at least some reworked Mesoarchaean material. Volcanic rocks in the other greenstone belts are dated mostly at 2.84–2.80 Ga (Huhma et al. 2012a). Mesoarchaean c. 2.95 Ga ages are

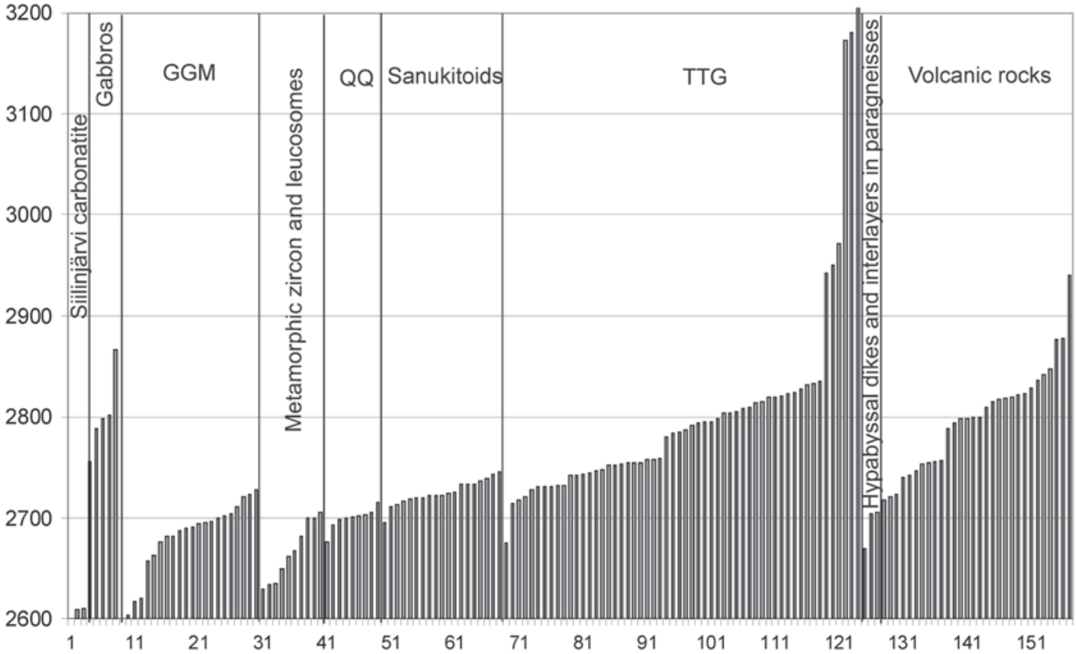

Fig. 3.9 Histogram showing the distribution of the U-Pb ages on zircon of various Archaean lithologies in the Finnish part of the Karelia Province

yielded by some volcanic rocks and TTGs in the northwestern part of the Western Karelia subprovince. In the Western Karelia subprovince, rocks whose zircon ages are > 3.0 Ga to exist only in its western part (Hölttä et al. 2000a; Mutanen and Huhma 2003; Lauri et al. 2011). The GGM suite rocks, mostly of c. 2.73–2.66 Ga, occur all over in the subprovince. In the Western Karelia subprovince the youngest Neoarchaean zircon ages are from granulites and leucosomes of migmatites giving ages of c. 2.65–2.63 Ga and c. 2.71–2.65 Ga, respectively. These ages have been interpreted to date the high-grade metamorphism of lower and mid-crust (Mutanen and Huhma 2003; Mänttäri and Hölttä 2002; Käpyaho et al. 2007; Lauri et al. 2011).

3.5.2 Sm-Nd

The TIMS/SIMS/LA-MC-ICPMS U-Pb zircon age determination localities do not yet evenly cover the Archaean area of Finland, and some relatively large areas remain untouched. Hence, some surprises may arise as the future studies fill up the existing gaps. Sm-Nd analyses have been carried out from most of the samples used for the U-Pb zircon analyses. For this work, many additional TTG samples were analysed to improve the regional cover of the Sm-Nd data. These samples are partly from our own sample sets and partly from the Rock Geochemical Database of Finland (Rasilainen et al. 2007). Whole rock chemistry was used to select samples with the least obvious metamorphic alteration. The analytical data are presented by Huhma et al. (2012b). Figure 3.10 shows the distribution of the T_{DM} model ages on a geological map of Finland.

The various geochemical groups of TTGs observed in this work do not correlate with Sm-Nd model ages or ε_{Nd} values in any simple way. The diagrams in Fig. 3.11 show the REE patterns and the Sm-Nd (T_{DM}) model ages of a large number of the analysed TTGs. Even in the youngest age group, the model ages are < 2.8 Ga and ε_{Nd} values range from + 0.9 to + 2.4, and the samples by and large represent juvenile Neoarchaean materials. Hence, all geochemical types are represented in our study, and the same pattern also holds true in the oldest model age group.

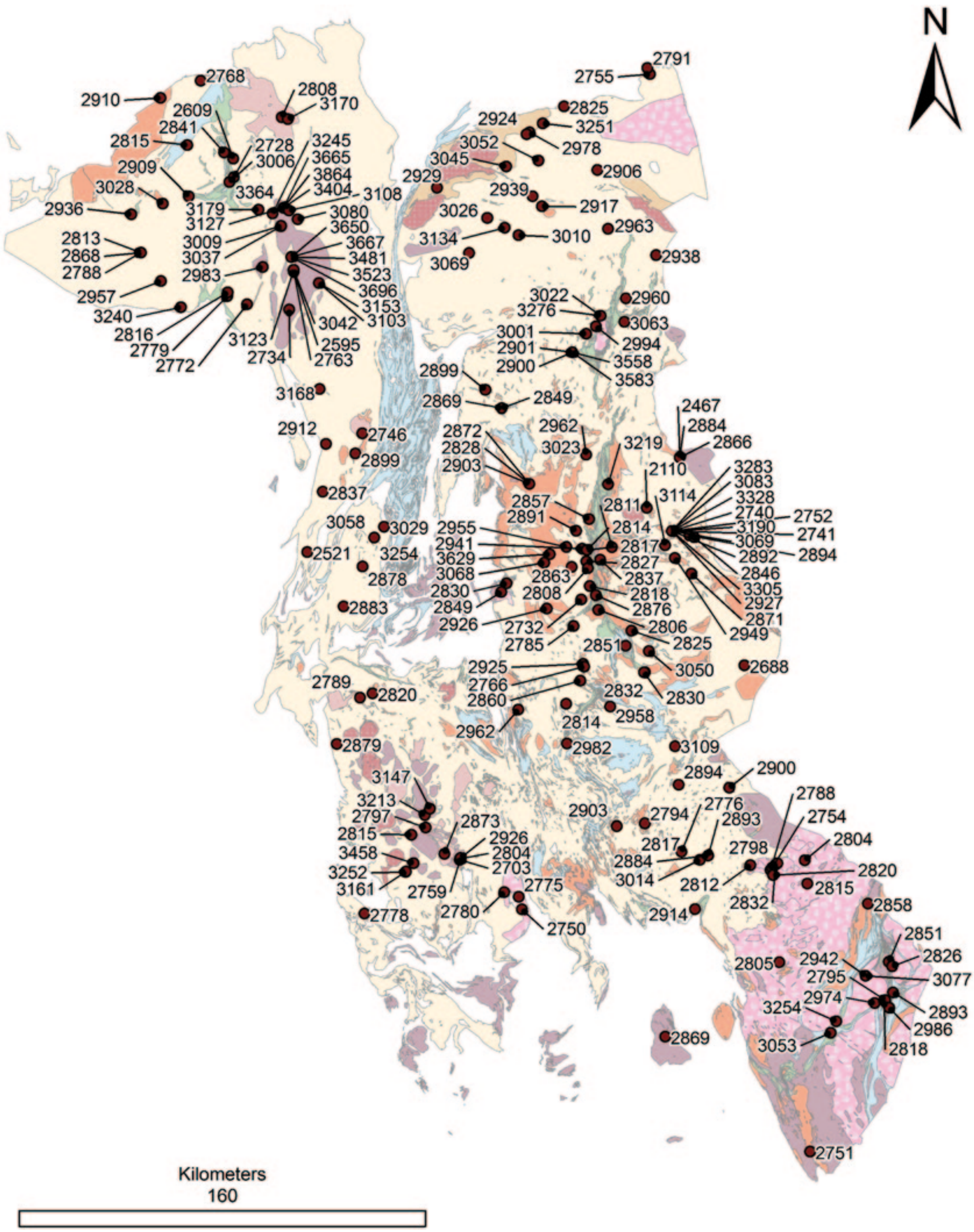

Fig. 3.10 Distribution of the T_{DM} model ages. The lithological base map is from Fig. 3.2

3.5.3 SIMS Ages on Detrital Zircon in Paragneisses

Kontinen et al. (2007) concluded in their study on the Nurmes paragneisses in the Western Karelia subprovince that the deposition of the protolith wackes took place at c. 2.70 Ga, which is in the range observed in U-Pb ages of the youngest dated detrital zircon grains. Nearly 50 % of the grains were dated at 2.75–2.70 Ga. The whole

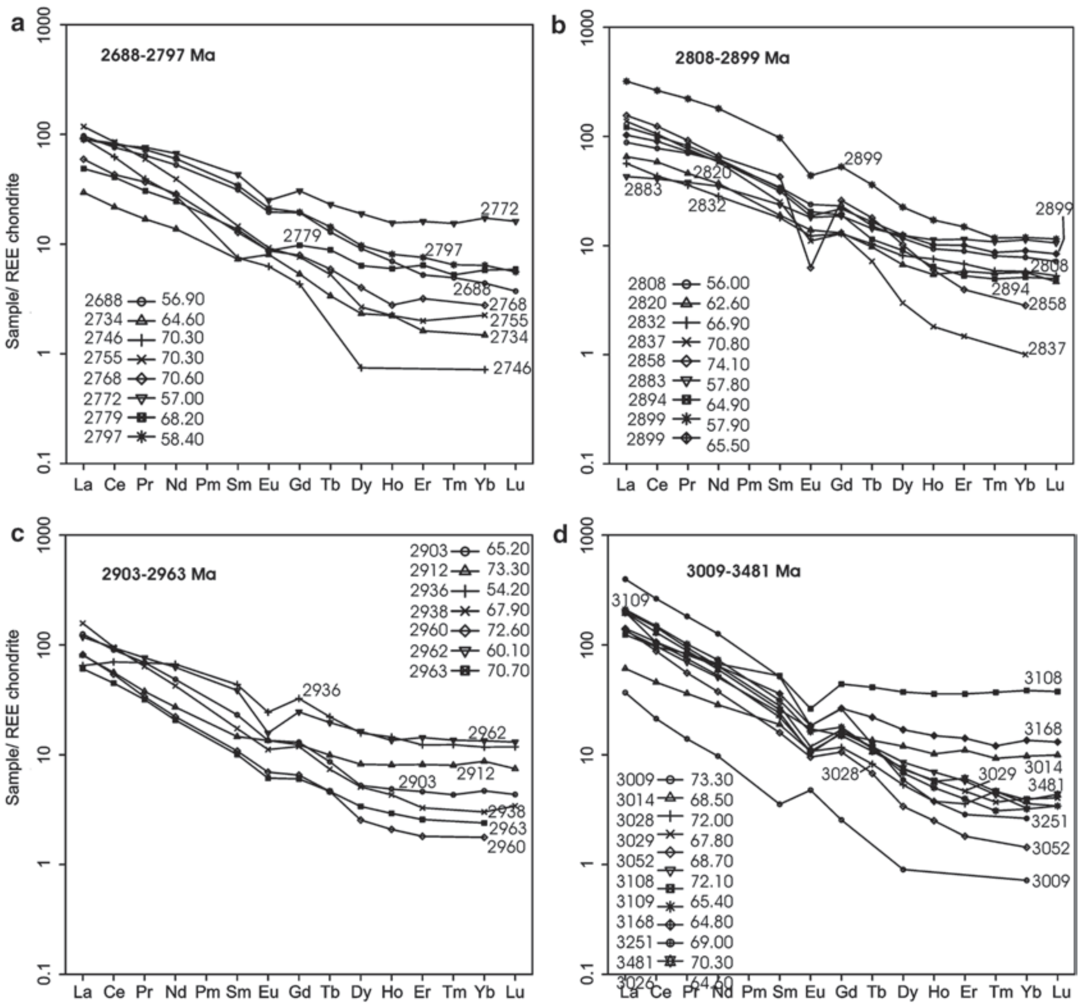

Fig. 3.11 REE patterns of the samples used for Sm-Nd analyses. The numbers next to the *curves* are for the model ages and the insets in the *upper right* corner of the *boxes* show the model age and the SiO_2 content of the sample

rock compositions of the Nurmes paragneisses suggest that the source terrains mainly comprised TTGs and sanukitoid-type plutonic and mafic volcanic rocks. Huhma et al. (2012a) analysed detrital zircon in metasediments from five other localities in the Karelia Province, and the results were similar to those from the previous studies. Most of these samples predominantly contain c. 2.73–2.75 Ga zircon grains, which suggests that the Neoarchaean intrusions of this age produced most of the sedimentary detritus. Mesoarchaean 3.2–2.8 Ga zircon grains were rare in all paragneiss and also other metasedimentary samples.

According to Bibikova et al. (2005), zircon from sanukitoids has higher Th/U ratios (>0.5) than zircon in TTGs (<0.5). Given that most 2.75–2.72 Ga zircon grains from paragneisses have Th/U ratios >0.5, sanukitoids indeed may have been one of their main sources, although it has to be noted that high Th/U in zircon is not restricted to sanukitoids, but is also found in samples of the other main rock groups.

3.6 Lower Crustal Xenoliths

Mantle and lower crustal xenoliths recovered from c. 500- to 600-Ma-old kimberlites near the southwestern boundary of the Western Karelia subprovince provide pertinent information on the petrology and physical properties of the lower crust (Hölttä et al. 2000b; Peltonen et al. 2006). The lower crustal xenoliths are almost exclusively mafic granulites. Mineral thermobarometry, together with isotopic, petrological and seismic velocity constraints, imply that the lower crustal xenoliths are derived from the weakly reflective, high-V_p layer at the base of the crust (40–58 km depth; Hölttä et al. 2000b; Peltonen et al. 2006). Single grain zircon U–Pb dates and Nd model ages (T_{DM}) from the xenoliths imply that the bottom high velocity layer is a hybrid layer consisting of both Archaean and Proterozoic mafic granulites. Many of the studied xenoliths record only Proterozoic zircon ages (2.5–1.7 Ga) and Nd model ages (2.3–1.9 Ga), implying that the lower crust contains a significant juvenile Palaeoproterozoic component (Hölttä et al. 2000b; Peltonen et al. 2006). Only a small fraction of zircon grains separated from the xenoliths give Archaean ages typically in the range of 2.7–2.6 Ga. Mesoarchaean zircon grains are almost absent. The oldest zircon ages, up to c. 3.5 Ga, and Nd T_{DM} model ages of c. 3.7 Ga of the xenoliths are similar to those from the oldest gneisses in the Siurua complex. Based on these data, the lowermost crust probably originated as Archaean mafic gneisses, but was repeatedly intruded by Proterozoic mafic magmas 2.5–1.80 Ga ago.

3.7 Metamorphism

3.7.1 Amphibolites and Paragneisses in the Western Karelia Subprovince

Ubiquitous evidence of melting and migmatisation of felsic but also mafic rocks within the TTGs of the Western Karelia Province implies that they were mostly metamorphosed in upper amphibolite and granulite facies conditions. Partial melting was commonly extensive, leading to the intense migmatisation characterizing so many localities. Lower grade rocks are found in the inner parts of the Kuhmo-Suomussalmi and Ilomantsi greenstone belts that commonly show mid- or low amphibolite facies mineral assemblages, well preserved primary structures and only a little or no migmatisation. Because of the dominance of the mineralogically monotonous gneissic TTG rocks, suitable mineral assemblages—especially garnet-bearing paragenesis—for the study of pressure-temperature evolution are not common. However, amphibolites and paragneisses locally have a garnet-hornblende-plagioclase-quartz or garnet-biotite-plagioclase-quartz mineral assemblages that can be used for geothermometry and geobarometry. Pressures and temperatures have been obtained using Thermocalc (Powell et al. 1998) and TWQ (Berman 1988, 1991) average PT calculations and grt-bt-pl-qtz thermobarometry (Wu et al. 2004). Metamorphic pressures obtained for the paragneisses and for the amphibolites around are mostly c. 4.7–7.5 kbar, and corresponding temperatures c. 650–740 °C (Hölttä et al. 2012). Many samples give lower temperatures of c. 600 °C, but they probably record post-peak cooling or Proterozoic metamorphism of the Archaean bedrock, because these rocks are normally migmatised, indicating high metamorphic temperatures.

Low P/T orthopyroxene-bearing, normally garnet-free granulite facies rocks occur slocally in the Western Karelia Province, but medium-pressure granulites, metamorphosed at c. 9–11 kbar and 800–850 °C, are only found in the Iisalmi complex near the western border of the Western Karelia subprovince (Hölttä and Paavola 2000). Sanukitoid suite granodiorites north of the Ilomantsi greenstone complex locally contain orthopyroxene, but it is not clear whether the mineral assemblages in these rocks were metamorphic or magmatic (Halla and Heilimo 2009). Amphibolites and paragneisses near these charno-enderbites were metamorphosed in upper amphibolite and granulite facies at c. 6.5–7.5 kbar and 670–750 °C (Hölttä et al. 2012).

3.7.2 Greenstone Belts

Garnet-bearing samples form supracrustal rocks in the Ilomantsi belt that typically have the assemblage grt-bt-pl-qtz ± ms, locally with andalusite and staurolite or more commonly their muscovite-filled pseudomorphs. Grt-bt thermometry for these samples indicates in most cases crystallization at c. 550–590 °C (Hölttä et al. 2012), similarly to the results of O'Brien et al. (1993) and Männikkö (1988), and these temperatures are in accordance with the observed mineral associations. In the NE part of the Ilomantsi greenstone belt, sillimanite is also present in pelitic rocks, and temperatures from grt-bt thermometry are also higher than in the SE, being c. 600–625 °C. Pressures indicated by the grt-bt-pl-qtz barometer are c. 3.5–5.5 kbar in the central parts of the greenstone belt but > 6 kbar in the NW in the sillimanite-bearing metasediments (Hölttä et al. 2012). The lower pressures are of the same order as those obtained by Männikkö (1988) using sphalerite barometry for samples from the Kovero greenstone belt SW of Ilomantsi. Garnet grains in the Ilomantsi belt samples are commonly zoned, with cores richer in Mg than rims, indicating decreasing PT conditions during garnet growth.

Previous studies on the Kuhmo-Suomussalmi greenstone belt have resulted in evidence of a decrease in metamorphic grade from outer to inner parts of the belt. According to Tuisku (1988), geothermometry suggests metamorphic temperatures as low as 500 °C for the inner and up to 660 °C for the outer parts of the belt. Pressures obtained using the sphalerite barometer applied to sphalerite inclusions in pyrite are mostly between 6–7 kbar but range in some cases as high as c. 13 kbar (Tuisku 1988).

An interesting observation was made for a patch of garnet-bearing amphibolites east of the Kuhmo greenstone belt. Noting the standard tholeiite basaltic whole-rock composition of these amphibolites, it is surprising that they do not comprise any matrix plagioclase, but only minor albite and oligoclase inclusions in garnet. The observed ranges of the anorthite content in the plagioclase inclusions in two microanalysed samples are An_{10}–An_{30} and An_1–An_{20}, indicating that some of the inclusions are almost pure albite. The garnet hosts are rich in grossular (X_{grs} 0.25–0.35, $X_{grs} = Ca/(Fe + Mn + Mg + Ca)$) and spessartine ($X_{sps}$ 0.10–0.12) but Mg-poor (X_{prp} 0.05–0.09), which indicates that the metamorphic temperatures were not very high during garnet crystallization. These rocks generally contain epidote, commonly only as inclusions in garnet but locally also in the matrix. If an albitic composition An_1 of plagioclase is used in the average P calculation, the Thermocalc gives average pressures of c. 16–17 kbar at 600–700 °C.

3.7.3 Age of Archaean High-Grade Metamorphism

Because the bulk of TTGs and volcanic rocks in the greenstone belts are from juvenile Neoarchaean additions to the crust, it is not surprising that signs of Mesoarchaean metamorphic events are difficult to distinguish in the preserved small enclaves of Mesoarchaean rocks. Mänttäri and Hölttä (2002) interpreted a c. 3.1 Ga zircon population to be metamorphic in the 3.2 Ga rocks of the Iisalmi complex. Käpyaho et al. (2007) found an obviously metamorphic zircon population of 2.84–2.81 Ga in a palaeosome of a 2.94 Ga migmatite west of the Kuhmo greenstone belt. However, most of the observed high-grade metamorphism and deformation appears Neoarchaean in age.

In the Western Karelia subprovince the ion probe U-Pb data on zircon and monazite from granulites and from leucosomes of migmatites appear to indicate partial melting over a broad time interval from 2.72–2.61 Ga (Hölttä et al. 2000a; Mänttäri and Hölttä 2002; Mutanen and Huhma 2003; Käpyaho et al. 2007; Lauri et al. 2011). Titanite from amphibolite facies rocks also yields broadly similar U-Pb ages. Zircon grains from leucosomes in the Iisalmi granulites give ages from 2.71–2.65 Ga. These age data indicate that migmatisation was a long-lasting event, and mostly coeval with GGM magmatism. The youngest Sm-Nd garnet-whole rock ages in the Iisalmi complex granulites are < 2.5 Ga, which indicates that cooling to the closure temperatures of the Sm-Nd system lasted until the Proterozoic era (Hölttä et al. 2000a).

3.7.4 Eclogites of the Belomorian Province

Archaean eclogites are found in the Belomorian Province at Gridino and Salma (Fig. 3.1).

Eclogites fall into three age groups. Mesoarchaean 2.88–2.86 Ga and 2.82–2.81 Ga eclogites are found only from the Salma area in the northern Belomorian province but Neoarchaean 2.72 Ga eclogites are known from both Gridino and Salma (Volodichev et al. 2004; Shchipansky and Konilov 2005; Mints et al. 2010a, b; Shchipansky et al. 2012a, b). Eclogites form up to tens of metres sized fragments and layers in TTG gneisses.

In the northern Belomorian province Mesoarchaean eclogites are found in fault bounded mélange zones whose length is several tens of kilometres. Some of the eclogites are rich in Fe and Ti and some are high-Mg picritic rocks. Garnetites and kyanite-bearing trondhjemites are associated with eclogites. Eclogites Garnetites are assumed to have been produced by the metasomatic alteration of eclogites (Shchipansky et al. 2012a, b; Mints et al. 2010b). Based on their composition (FeO*/MgO—0.5–2.5) eclogites are classified as tholeiitic basic rocks. Their REE concentration is 2–12 times that of chondrite, and their REE distribution pattern is either flat or weakly fractionated (La_N/Sm_N—0.99–1.8; Ga_N/Yb_N—0.77–1.17). Some of them are slightly enriched in Nb relative to Th and La (Shichipansky et al. 2012b). Compositionally, eclogites represent MORB and oceanic plateau basalts or gabbros. Kyanite-bearing trondhjemites have adakitic compositions, and they form small veins in eclogites. They are interpreted as melting products of eclogites in a subductional setting (Shchipansky and Konilov 2005).

In the northern Belomorian province metamorphic conditions for the eclogite-facies metamorphism are estimated at P ~ 13–14 kbar and T ~ 700–750 °C (Konilov et al. 2010; Mints et al. 2010b). However, signs of "pre-eclogite-facies" metamorphism in the form of inclusions of chlorite, zoisite and even pumpellyite-actinolite-albite intergrowth in garnet have been observed. Commonly the primary eclogite mineral assemblages (omphacite-garnet-amphibole-quartz-rutile) are replaced by diopside-plagioclase symplectites during decompression and later eclogites have partly altered into amphibolites. Younger granulite facies overprints have also been identified in eclogitic rocks (Shchipansky and Konilov 2005; Konilov et al. 2010).

Geochronological studies of eclogites suggest that Mesoarchaean oceanic rocks were metamorphosed at eclogite facies at 2.88–2.87 Ga and 2.82–2.81 Ga. The early retrograde alterations of eclogites were close in time to eclogitization. Symplectitic eclogites were metamorphosed at 2.72–2.70 Ga, 2.53–2.08 Ga and 1.91–1.89 Ga (Konilov et al. 2010; Kaulina 2010; Shchipansky et al. 2012b; Mints et al. 2010a, b).

Neoarchaean eclogites have been described from the Gridino and Salma areas. Neoarchaean eclogites in the Gridino area, together with amphibolites, ultrabasic rocks, TTG gneisses and zoisitites, form a part of a metamorphosed melange which constitutes a fault-bounded zone (Slabunov 2008). This zone is cross-cut by a swarm of near-N-S-trending Palaeoproterozoic mafic dykes whiche were also eclogitized to a varying degree.

In the Gridino area eclogites have the mineral assemblage omphacite-garnet-quartz-rutile ± kyanite, thermobarometric studies show that they were metamorphosed at c. 14–17.5 kbars and 740–865 °C (Volodichev et al. 2004). Pre-eclogite amphibolite-facies mineral assemblages (epidote, amphibole, carbonate, quartz and albite inclusions in cores of garnet grains) have been identified in the eclogites (Perchuk and Morgunova 2011; Volodichev and Slabunov 2011). Retrograde plagioclase-diopside symplectites after omphacite and amphibole-plagioclase-biotite assemblages after garnet are common in the Gridino eclogites. They were produced by decompression during the exhumation of the rocks (Volodichev et al. 2004), probably simultaneously with intrusion of small enderbite stocks and veins.

An ion microprobe study of zircon from Gridino eclogite gave an age of 2720 ± 8 Ma Ma, which probably is the age of eclogite facies metamorphism (Volodichev et al. 2004). This age is in good agreement with the geological data: the mélange is cross-cut by massive 2701 Ma trondhjemite veins (Volodichev et al. 2004). Geochronological data are difficult to interpret because eclogite-facies metamorphism was not the only metamorphic event. The area was also metamorphosed during the Palaeoproterozoic, as shown by c. 1.9 Ga zircon generations (Bibikova et al.

2003; Skublov et al. 2011) and by re-equilibration of Sm-Nd isotopic systems in minerals at that time (Kaulina and Apanasevich 2005).

Zircon grains of 2.72 Ga have been described also from eclogites in Salma (Konilov et al. 2010; Shchipansky et al. 2012a, b; Mints et al. 2010b). These Neoarchaean eclogites seem to be separated by a fault from Mesoarchaean eclogites. A contact between these two complexes is visible in the Kuru-Vaara quarry (Shchipansky et al. 2012a). Neoarchaean eclogites in the Kuru-Vaara area are classified as amphibole eclogites (omphacite + garnet + amphibole + rutile), metamorphosed at peak temperatures of 680–720 °C and pressures of up to 20 kbar (Shchipansky et al. 2012a). Based on their chemical composition, these eclogites are classified as low-K tholeiitic basalts whose REE concentration is similar with N-MORB. Some of them show an U-shaped or boninite-like REE distribution pattern. The formation of Neoarchaean eclogites is attributed to subduction where the oceanic crust submerged to a depth of at least 50 km.

3.7.5 Proterozoic Metamorphism

The Archaean bedrock in the western part of the Karelia Province underwent strong reheating during the Palaeoproterozoic Svecofennian orogeny. Evidence of this includes, for example, that biotite and hornblende sampled from the Archaean rocks have K-Ar ages typically in the range 1.8–1.9 Ga. Archaean K-Ar mineral ages are only present in samples from the Iisalmi and Taivalkoski granulites and the Ilomantsi complex (Kontinen et al. 1992; O'Brien et al. 1993). The heating of the Archaean crust has been explained by its burial under a massive overthrust nappe complex ca. 1.9 Ga ago (Kontinen et al. 1992).

Numerous ductile shear zones were developed in the Archaean bedrock during the Svecofennian orogeny. The widths of these shear zones vary from tens of metres to several kilometres (Kohonen et al. 1991). In many places the Proterozoic shearing was associated with high hydrous fluid flows and related chemical alteration, which is reflected, for instance, in alkali-deficient compositions and kyanite- and cordierite-bearing assemblages in originally TTG rocks (Pajunen and Poutiainen 1999). In some zones almost all Archaean rocks were ductilely deformed during the Svecofennian orogeny. West of the Iisalmi granulite complex, Archaean granulite facies mineral assemblages were decomposed in the Proterozoic metamorphism that took place at c. 550–650 °C and 5–6 kbar (Mänttäri and Hölttä 2002). Similar temperatures and pressures for the Proterozoic metamorphic overprint have also been reported near the Kuhmo greenstone belt (Pajunen and Poutiainen 1999). Deformation microstructures in the sanukitoid granodiorites near the Ilomantsi complex indicate temperatures of 400–500 °C during Proterozoic metamorphism (Halla and Heilimo 2009).

3.8 Palaeomagnetism

Several palaeomagnetic studies on Archaean rocks have been carried out in the Karelia Province, but only in a few cases has stable Archaean remanent magnetization unaffected by Proterozoic overprinting been revealed. The main use of palaeomagnetic data in Archaean geology has been in reconstructing the past positions and movement of the craton at different times and comparing its position with other similar-aged cratons. Continental reconstructions have been made particularly with the Superior Province because of the considerable geological similarity between these two cratonic masses. This section reviews the palaeomagnetic data from Archaean rocks in Finland and NW Russia, and present models of the Archaean plate configurations of the Karelia and the Superior Provinces.

The 2913 ± 30 Ma Shilos metabasaltic rocks located NW of Lake Onega in NW Russia, with preserved remanent magnetization estimated at 2800 Ma (Arestova et al. 2000 and references therein), are the oldest rocks from the Karelia Province for which palaeomagnetic data are presently available. Another case of old remanence is from the NW of Lake Onega, where the 2890 Ma Semch River gabbro-diorite is interpreted to preserve its primary magnetic remanence (Gooskova and Krasnova 1985). However, in both cases the age of remanence is defined based on comparison to the APWP (Elming et al. 1993). Therefore, due to poor age constraints, the data are not used in either case for continental reconstructions.

The most reliable Neoarchaean palaeomagnetic data from the Western Karelia Province are obtained from the granulite facies enderbites in the Iisalmi area (Neuvonen et al. 1981, 1997; Mertanen et al. 2006a), from the orthopyroxene-bearing sanukitoids NW of the Ilomantsi area (Mertanen and Korhonen 2008, 2011) and from the Panozero sanukitoid in the Vodlozero subprovince (Lubnina and Slabunov 2009). Younger, well-defined Neoarchaean data come from the Shalskiy gabbronorite dyke and granulite-grade gneisses in the Vodlozero subprovince in NW Russia (Krasnova and Gooskova 1990; Mertanen et al. 2006b). Common to all these cases is that they are generally well-preserved from the 1.9–1.8 Ga Svecofennian overprinting, they show high magnetic anomalies compared to surrounding TTG gneisses and their remanence has high stability, the directions of remanence clearly differing from the known Proterozoic remanences.

The sanukitoids NW of the Ilomantsi complex and the Iisalmi enderbites show a steep characteristic remanence component, but in the former inclination is negative and in the latter positive (Mertanen et al. 2006a; Mertanen and Korhonen 2011). It is interpreted that the steep remanence directions record the long-lasting Neoarchaean metamorphic event at different times, during which the polarity of the Earth's magnetic field has reversed at least once. The remanence of the sanukitoids NW of Ilomantsi is regarded as ca. 2.7 Ga ($^{207}Pb/^{206}Pb$ monazite age of 2685 Ma, Halla 2002) and the Iisalmi granulite complex as ca. 2.6 Ga (Sm-Nd garnet-whole rock ages 2590–2480, Hölttä et al. 2000a). Based on these data the Karelia Province moved from a high polar palaeolatitude of 83° to the palaeolatitude of c. 68°, respectively. The overall data from Ilomantsi and Iisalmi thus imply that at c. 2.7–2.6 Ga the Western Karelia subprovince was located at high palaeolatitudes.

The palaeomagnetic pole of Panozero differs significantly from those NW of Ilomantsi and in Iisalmi (Lubnina and Slabunov 2009). The difference between the poles can be explained by the different locations of the Vodlozero and Western Karelia subprovinces during the early Neoarchaean. It is possible that in the time interval of 2.77–2.74 Ga the Karelian subprovinces were separated, and the final joining of separate terranes took place at 2.70–2.65 Ga (Slabunov et al. 2011). Alternatively, it is also possible that the two terranes were joined already at ca. 2.77–2.74 Ga, and the amalgamated craton drifted from low palaeolatitudes to higher palaeolatitudes between 2.74 and 2.68 Ga.

3.9 Discussion

3.9.1 Adakitic Features of TTGs

The mutual compositional similarity of TTGs and modern adakites has been the basis of a suggestion of petrogenetic kinship between the two rock suites (Martin 1999; Martin et al. 2005). Adakites are spatially related to subduction, and the most likely source of their parental magmas has been the basaltic part of a subducted oceanic slab. They seem to be related to an environment where the subduction zone is abnormally hot, allowing the subducting slab to melt (Moyen 2009). Numerical and petrological models suggest that partial melting of a subducting slab is possible at 60–80 km depth, but only when the subducting oceanic crust is very young (<5 Ma), and therefore hot, or as a consequence of heating under abnormally high stresses in the subduction shear zone (Defant and Drummond 1990, 1993; Peacock et al. 1994). However, according to Gutscher et al. (2000), most of the known Pliocene-Quaternary adakite occurrences are related to the subduction of 10–45 Ma lithosphere, which, according to numerical models, should not produce melt under normal subduction zone thermal gradients. Gutscher et al. (2000) addressed this by flat subduction that can produce the temperature and pressure conditions necessary for the fusion of moderately old oceanic crust. Variation in the subduction angle has been proposed as a critical factor also controlling the variation observed in geochemical features of the Archaean TTGs. Smithies et al. (2003) proposed that Archaean subduction was predominantly flat and that the subduction regimes thus lacked well-developed mantle wedges that would produce melts or interact with possible slab-derived melts, in the latter case increasing the compatible element content and Mg contents of the slab melts.

Recently, the usage of the terms adakite, and especially adakite-like, has been expanded to encompass a wide range of rocks that exhibit the high Sr/Y and La/Yb ratios but not necessarily the other criteria of the original adakite definition. This loose usage has led Moyen (2009) to recommend that separate, more precise terms should be used to describe these "adakitic" rocks, and the term adakite should be reserved only for the high-silica adakites that closely correspond to the rocks originally described as adakites by Defant and Drummond (1990). Halla et al. (2009) argued that the term adakite should only be used for unmistakable slab melts, and therefore not for such rocks as the TTGs in the Karelia Province. Smithies (2000) also argued that despite the many compositional similarities, most Archaean TTGs actually differ from Cenozoic adakites, especially in that they have on average lower Mg# and higher SiO_2 contents, suggesting that, unlike adakitic melts, the TTG melts did not interact with mantle peridotite.

High Sr/Y and La/Yb ratios can have several causes, such as high Sr/Y sources, garnet-present melting and interactions with the mantle (Moyen 2009). Melts with an adakitic geochemical signature can also be generated by normal crystal fractionation processes from andesitic parental melts, and slab melting is not compulsory for adakite petrogenesis; but, adakites or adakite-like rocks can instead originate in various geodynamic settings (Castillo et al. 1999; Castillo 2006; Richards and Kerrich 2007; Petrone and Ferrari 2008).

Modelling by van Hunen et al. (2004) suggests that if mantle temperatures were indeed higher during the Archaean than presently, even flat subduction was an unlikely process. If this was the case, the obvious lack of interaction with mantle in many TTGs must be explained in some other way. Halla et al. (2009) attributed the low-HREE TTG group to high-pressure partial melting (>20 kbar) of a garnet-bearing basaltic source with little evidence of subsequent mantle contamination. The high-HREE group was generated by significantly lower pressure melting (c. 10 kbar) of a garnet-poor basaltic crust and shows interaction with the mantle by its higher Mg#, Cr and Ni contents. Halla et al. (2009) proposed that the high-HREE TTGs were produced in an incipient, hot subduction zone underneath a thick oceanic plateau/protocrust. For the low-HREE TTGs, they saw a non-subduction setting as probable, proposing that these rocks were generated by deep melting in the lower parts of thick domains of basaltic oceanic crust.

3.9.2 TTG Melts and PTX Relations of their Protoliths

There is a general agreement that the Archaean TTG suite rocks were formed by partial melting from a compositionally basaltic-gabbroic source. The process was evidently fluid-absent partial melting of amphibolites at temperatures of 900–1100 °C and over a large pressure range from 10 to 25 kbar. The composition of products from partial melting is controlled by pressure and temperature, the composition of the source, water availability during the process and the degree of melting. The composition of the partial melts is further modified, for instance, by magma mixing, fractional crystallization and wall rock contamination on their way from the loci of melting to the crystallization sites (Rapp et al. 1991; Martin 1995; Martin and Moyen 2002; Foley et al. 2002; Rapp et al. 2003; Moyen and Stevens 2006; Moyen 2011).

During the fusion process, pressure and temperature control the assortment and abundance of the residual minerals, such as garnet and plagioclase, and consequently the major and trace element content of the TTG melts. For example, the heavy REE is controlled by residual garnet, which is stable in mafic rocks at pressures above c. 9–12 kbar and increases in abundance with increasing pressure. Sr is controlled by plagioclase, which is stable below c. 15–20 kbar. Nb and Ta depend on the presence of residual rutile, which is stable above c. 16–18 kbar (Foley et al. 2002; Moyen and Stevens 2006; Moyen 2011).

In the low-HREE TTG group with positive Eu anomalies, Sr and Sr/Y ratios are high and Nb low. Positive Eu could be explained by plagioclase accumulation, but as TTGs in this group are no more enriched in Al_2O_3, CaO or Na_2O than the other low-HREE TTGs, it is more probable that they represent melting with residual rutile and garnet. In dacitic and rhyolitic melts, the garnet-melt partition coefficients are higher for Sm (2.66) and Gd (10.5) than for Eu (1.5) (Rollinson 1993). Thus, high pressure melting with abundant garnet in the residue would obviously lead to melts that are HREE poor, and show a positive Eu/Eu* ratio, which is also predicted by experimental work (e.g. Springer and Seck 1997).

The melting temperature strongly depends on the starting composition, and the solidus temperatures for arc basalts; tholeiitic basalts and komatiitic basalts are thus highly different (Moyen and Stevens 2006). The trace element composition of melts principally depends on their protolith concentrations, but through control of the mineral composition of the restite, the major element content of the protolith also affects the trace element composition of the coexisting melt. According to Nair and Chacko (2008), the increase in residual garnet from 5 to 15 wt% changes the La/Yb ratio in the melt fraction from c. 12 to c. 24. Variation in the La/Yb ratio of the melts may be considerable, even under constant P and T, if there is enough compositional variation in the source. Figure 3.12 shows simplified pseudosections with melting curves and breakdown curves of amphibole, orthopyroxene, garnet and rutile for two compositionally deviating samples analysed from amphibolite intercalations in TTGs, representing tholeiitic and komatiitic basalts with 1 wt.% H_2O. The pseudosections were constructed using the Perple_X 6.6.6 software (Connolly 1990, 2005; Connolly and Petrini 2002, http://www.perplex.ethz.ch/). The composition has a strong influence on the melting temperature but also on the mineral stability fields. In Na_2O and Al_2O_3-poor komatiitic basalt, plagioclase decomposes at pressures that are c. 7–10 kbar lower than in the case of basaltic composition (Fig. 3.12). This means that in >800ºC temperatures komatiitic basalt can produce melts with elevated Sr at pressures that are far below the c. 19–21 kbar range that is broadly the upper stability limit of plagioclase in tholeiitic basalts in these temperatures. Also, in komatiitic basalt rutile is present at several kbar lower pressures than in tholeiitic basalt, and consequently komatiitic basalt can produce Nb and Ta depleted melts at lower pressures than tholeiitic basalt.

According to the model calculations, also the abundance of garnet is strongly dependent on the composition so that at above c. 10 kbar the komatiitic basalt produces roughly twice as much garnet as the tholeiitic basalt (Fig. 3.12). Consequently in TTG melts the La/Yb ratio may significantly differ only on the basis of the composition of the protolith.

Sanukitoids and quartz diorites are partly too Mg-rich to represent melts from crustal sources. The high Sr content and low La_N/Yb_N ratio indicate that the quartz diorites originated from a hot and shallow mantle environment, where the temperature was so high that plagioclase was not stable, and the pressure so low that little garnet was present in the residue. Figure 3.13 shows the proportion of residual garnet during the dehydration melting of amphibolite, and the corresponding La/Yb ratios in the derived melts, according to the experimental work of Nair and Chacko (2008). The garnet mode in the residue in the case of QQs and high-HREE TTGs would have been 5–15 wt%, and much higher in low-HREE TTGs. This could reflect higher pressures of melting in the case of the low-HREE TTGs, but could also indicate a compositionally different source, which would promote a higher garnet mode in the residue.

In the Archaean Earth, low pressure TTGs could have formed, for example, at the base of thick oceanic plateaus (deWit and Hart 1993; Moyen and Stevens 2006). In the greenstone belts, most sequences of mafic volcanic rocks of plateau basalt signatures are not compositionally homogeneous, but typically comprise a variety of compositionally different tholeiitic and komatiitic basalts and komatiites. In addition,

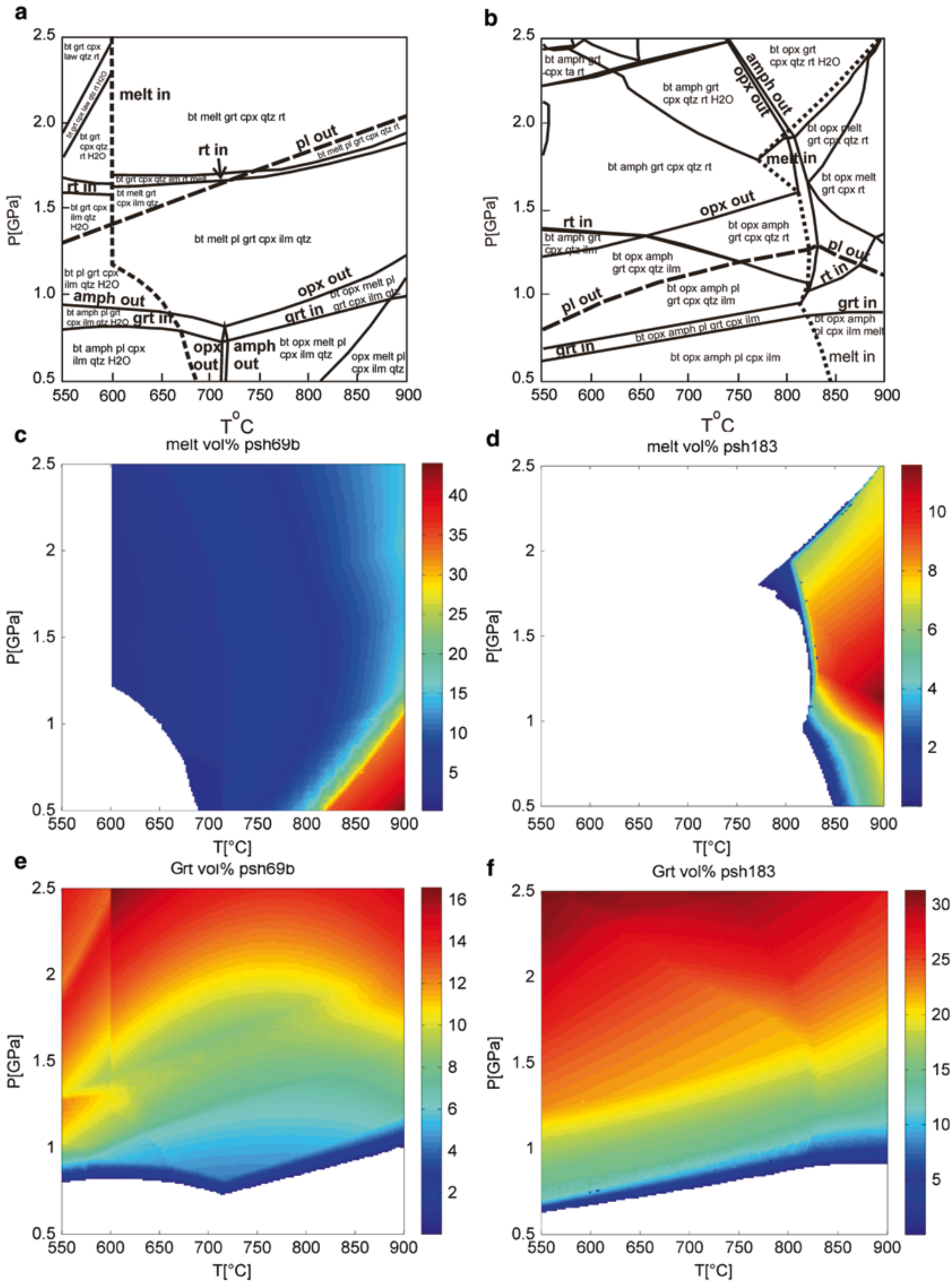

Fig. 3.12 NCKFMASHTi pseudosections for Fe-rich basaltic granulite (psh69b) and komatiitic basalt (psh183) and lower and upper stability limits of various minerals in these two whole-rock compositions. The solution models used in calculations were: biotite (TCC), amphibole (GlTrTsPg), garnet (HP), orthopyroxene (HP), clinopyro-

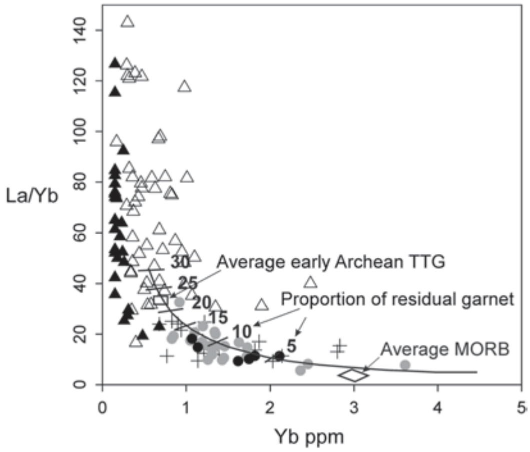

Fig. 3.13 La/Yb ratios of Karelian TTGs plotted on the La/Yb diagram after Nair and Chacko (2008), showing the effect of the abundance of residual garnet on the La/Yb ratio of the melt in 20 % melt fraction during dehydration melting of amphibolite. *Black triangles* = low-HREE TTGs, *crosses* = high-HREE TTGs, *grey dots* = QQs, *black dots* = orthopyroxene-bearing QQs (enderbites)

many greenstone belts also contain calc-alkaline basalts and andesites-dacites. Modern oceanic plateaux, such as the Kerguelen Plateau, consist of intermediate, felsic and alkaline volcanic rocks, as well as sediments (Frey et al. 2000). It is evident from the above discussion that melting of such heterogeneous packages would produce melts that would also be compositionally heterogeneous and show variable trace element patterns, even in cases where the melting depth was not very high or highly variable.

3.9.3 Greenstone Belts

A picture emerging from previous research suggests that the Karelia Province was a collage of TTG and greenstone complexes that originated in various tectonic settings related to subduction, collision, continental rifting and mantle plumes (Bibikova et al. 2003; Samsonov et al. 2005; Slabunov et al. 2006a, b; Papunen et al. 2009). The greenstone belts have also been interpreted as composite terranes comprising magmatic products from various tectonic settings involving plumes and arc magmatism (Puchtel et al. 1998, 1999). Archaean greenstone belts can be divided into autochthonous to parautochthonous and allochthonous based on their relationship with the underlying basement rocks (Polat and Kerrich 2000, 2006). According to Thurston (2002), evidence from the Superior Province suggests that many, if not all, greenstone sequences were in autochthonous or parautochthonous units fed from mantle plumes either in continental rift or continental platform settings. Allochthonous models favour the assembly of greenstone belts by horizontal tectonic transport and accretion of various types of oceanic crust in a plate-tectonic geodynamic regime (e.g. Puchtel et al. 1998, 1999; Percival et al. 2004, 2006; Polat and Kerrich 2006). The presence of such features as fold and thrust complexes, orogen-parallel strike-slip faults and tectonically juxtaposed terranes from different tectonic settings, as well as subduction zone geochemical signatures in part of the plutonics and volcanics in the greenstone belts, supports the concept that the accretion of alloch-

xene (HP), plagioclase (h), ilmenite (WPH), melt (HP). For the original references see http://www.perplex.ethz.ch/PerpleX_solution_model_glossary.html. Compositions used (in wt.%) are:

	psh69b	psh183
SiO_2	59.63	47.3
TiO_2	1.00	0.347
Al_2O_3	13.53	11.5
FeO	9.68	8.54
MgO	2.65	14.2
CaO	5.49	11.7
Na_2O	3.45	1.16
K_2O	1.97	0.918
H_2O	1.0	1.0

thonous terranes was elemental in the growth of many greenstone belts (Polat and Kerrich 2006).

Many komatiite-bearing sequences in Archaean greenstone belts have been interpreted as pieces of dismembered Archaean oceanic plateaux (Kusky and Kidd 1992; Abbott and Mooney 1995; Puchtel et al. 1998). Many greenstone belts are characterized by assemblages that suggest roughly coeval plume-type komatiite-tholeiitic basaltic and arc-type calc-alkaline volcanism. This situation has been explained in the Karelia Province in terms of a subduction setting where the arc-type plutonic volcanic rocks formed at the margins of plume-generated thick basalt plateaux that were not able to subduct because of their buoyant nature (Puchtel et al. 1998, 1999). Interlayering of komatiites with subduction-related volcanic rocks has been explained by the interaction of plume and subduction related magmas in such a subduction regime (Grove and Parman 2004).

Grove and Parman (2004) proposed that Archaean komatiites could have been formed by hydrous melting in a subduction environment, which would easily explain the close spatial and temporal association of many komatiites and island arc type volcanic rocks. This idea, which is supported by the experimental work of Barr et al. (2009), is still disputable, however. For example, based on their work on komatiites in Ontario and Barberton, Arndt et al. (2004) and Stiegler et al. (2010) saw little evidence for the hypothesis of hydrous melting.

3.9.3.1 The Kuhmo Greenstone Belt—an Oceanic Plateau?

The basic-ultrabasic volcanic assemblages in the Kuhmo greenstone belt consist of komatiites and their evolved counterparts, i.e. komatiitic basalts. The komatiite-basalt sequence is completely devoid of epiclastic and chemical interflow sediments, and it lacks geochemical evidence of contamination by any significantly older continental material. These data suggest an eruptive setting far from continental land masses and hydrothermal vents at oceanic ridges, and argues against the origin of the Kuhmo greenstone belt within a continental rift zone, as has been proposed, for instance, by Papunen et al. (2009). Immobile trace element (Zr, Y, Nb, Th) systematics are also inconsistent with the formation in a back-arc setting, but rather suggest an oceanic plateau setting and magma derivation from a mantle plume. The Al-undepleted nature and the trace element characters of the komatiites indicate that they were derived from a source more similar to primitive upper mantle rather than that of depleted MORBs. Furthermore, there is negligible geochemical evidence for the involvement of crust or enriched or recycled mantle sources (EM1, EM2, HIMU). Condie (2005) stressed the clustering of non-arc-related Archaean basalts, in terms of HFSE ratios, close to the primitive mantle values, and suggested on this basis that Archaean mantle plumes had their main source in the “primitive mantle”. Our findings support this conclusion, indicating that the late Archaean mantle was less fractionated, or better stirred, than either the early Archaean and post-Archaean mantle.

The U-Pb zircon data (Huhma et al. 2012a) and the Sm-Nd data (Huhma et al. 2012b) on the Kuhmo volcanic rocks provide no evidence for their deposition on a significantly older basement in the Kuhmo region. Felsic volcanic rocks in the central part of the Kuhmo belt formed 2.80 Ga ago, giving the minimum age for the mafic-ultramafic magmatism (Huhma et al. 2012a). We conclude that the Kuhmo komatiites represent fissure-controlled eruptions onto a pre-existing rhyolitic-dacitic-tholeiitic oceanic plateau.

3.9.3.2 The Ilomantsi Greenstone Belt—a Volcanic Arc within an Attenuated Continental Margin?

Komatiites within the Ilomantsi greenstone belt, which is located c. 150 km SSE of the Kuhmo belt, were emplaced within a volcano-sedimentary basin. Magmatism was dominated by felsic volcanism in two major pulses and coeval granitoid plutons. The komatiites were emplaced as thin but extensive sheet flows, and probably also as sills beneath the felsic volcanic edifices. The komatiites and dacites-rhyolites occur intercalated, suggesting that ultrabasic and felsic volcanism was coeval. Both the komatiites and the felsic volcanic rocks have distinctly low

Nb/Th, suggesting that the komatiites contain a significant component assimilated from the felsic volcanic rocks. Alternatively, the komatiites could have inherited their arc-signature from a subduction-enriched mantle wedge. However, we consider this model less likely, because the samples enriched in lithophile-incompatible elements are depleted in PGE, which is best explained by the segregation of a sulphide melt from the magma in response to contamination by sulphidic metasediments during emplacement. There is evidence of a significantly older cryptic granitoid basement in the region in the form of inherited c. 3.0 Ga zircon grains in plutonic and subvolcanic felsic rocks and old T_{DM} ages of metasedimentary units in the Ilomantsi belt (Vaasjoki et al. 1993; Huhma et al. 2012a, b). We suggest that the Ilomantsi volcanic rocks represent arc magmatism within an attenuated continental margin where older basement rocks were assimilated by younger arc magmatism.

3.9.4 The Greenstone Belts of the Belomorian Province—an Archaean Subduction System?

The Keret belt arc-type volcanic rocks, the Chupa belt metagraywackes, interpreted as fore-arc basin sediments, an ophiolite-like Central Belomorian greenstone belt and the Salma eclogite-bearing mélange form the 2.88–2.82 Ga lateral sequence. This sequence likely represents an arc-subduction-accretion system, in which arc-type volcanic rocks and metagreywackes belong to the suprasubduction zone whereas the ophiolitic and eclogites are part of a subducting slab. This system vanished at 2.82 Ga, and the accretion of the arc-ophiolite complexes gave rise to the first fragment of a continental crust.

The 2.81–2.78 Ga sequence consists of basalt-komatiite, suprasubduction ophiolite and island-arc complexes that include a large fraction of adakite-series rocks, TTGs, fore-arc basin metagraywackes, a medium pressure granulite-enderbite-charnockite complex and eclogites. These complexes mark an oceanic plateau, a back-arc spreading basin, a volcanic island-arc zone, a fore-arc basin, a deep section of a suprasubduction zone and fragments (2.82–2.80 Ga eclogites) of a subducting slab (Slabunov 2013). The fact that adakites are abundant among the island-arc volcanics formed in this period is consistent with the subduction concept.

After a short, c. 25 Ma period of no tectonic activity, subduction continued and a new island-arc complex was formed. A new continental crust was not generated over the next 20 Ma. The last growth event of the sialic crust began at 2.73 Ga and continued for 25–30 Ma. During this period the Gridino and Shirokaya Salma-Kuru-Vaara eclogites, calc-alkaline island-arc volcanics, molybdenum-bearing gabbro-diorite-granodiorite intrusions and a granulite-enderbite complex were formed, probably at a convergent boundary during subduction in an active continental margin setting (Slabunov 2013).

3.9.5 Supercontinent Reconstruction

Several similarities in their geological evolutions can be cited in support of the concept that the Archaean Superior, Hearne and Karelia Provinces were parts of the Neoarchaean-Palaeoproterozoic supercontinent Superia (Bleeker and Ernst 2006).

Comparison of palaeomagnetic data from the Karelia Province with similar-aged poles from the Superior Province (Mertanen and Korhonen 2011, and references therein) shows that Superior was also located at relatively high palaeolatitudes at 2.68–2.60 Ma (Figs. 3.14a, b). However, the Karelia and Superior Provinces were significantly separated, as there is a c. 30° difference between the latitudes both at 2.68 Ga and 2.60 Ga. Taking into account the maximum errors of the poles from both cratons, the latitudinal distances become shorter, and it may be possible that they were even joined at that time. The relative palaeopositions of the two cratons cannot be resolved unequivocally, especially at 2.68 Ga, due to the nearly polar position of the Karelia Province, which allows its rotation in several directions with respect to Superior.

The steeply inclined remanence of the Iisalmi and Ilomantsi rocks differs distinctly from the

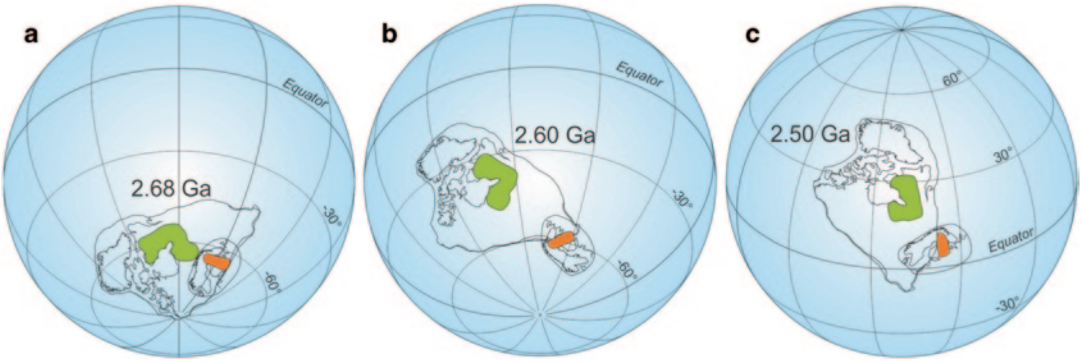

Fig. 3.14 Continental reconstructions between the Karelia and Superior cratons at 2.68 Ga (**a**), 2.60 Ga (**b**) and 2.50 Ga (**c**)

2.50 Ga remanence of the Shalskiy gabbronorite dyke dated at 2510 ± 1.6 Ma in the Vodlozero subprovince (Bleeker et al. 2008). The Shalskiy dyke, as well as the Shalskiy basement gneisses and Vodla River gneisses ca. 50 km east of the Shalskiy dyke, has a shallow southwards pointing remanence direction (Krasnova and Gooskova 1990; Mertanen et al. 2006b) which is interpreted to be 2.50 Ga. This remanence direction positions the Karelia Province at a low equatorial palaeolatitude. The data thus suggest substantial movement of the Karelia Province between the time of cooling at c. 2.60 Ga after the high grade metamorphism and subsequent rifting and minor reworking of the craton at c. 2.50 Ga.

Palaeomagnetic reconstruction between Karelia and Superior Provinces at 2.50 Ga is presented in Fig. 3.14c. Compared with the 2.6 Ga configuration, the palaeopositions between Karelia and Superior are now completely different. The Superior Province had drifted across the equator to the latitude of approximately 20° and rotated clockwise by about 45°, so that at 2.50 Ga both provinces were located near the equator, and the Karelia Province along the southern margin of the Superior Province. This (Slabunov 2008, 2010) reconstruction is in close agreement with the Superia model of Bleeker and Ernst (2006), who suggested that the provinces were together at 2.50 Ga.

3.9.6 Tectonic Evolution of the Karelia Province

If the Karelia province was indeed once a part of the supercontinent Superia, then many of the observations and models presented for the evolution of the Superior and Hearne Provinces must also be applicable to the Karelia Province. Several major characteristics in the crustal architecture and composition of the Superior province have long been considered to provide strong support for the operation of modern-style plate tectonics during the Neoarchaean (e.g. Goodwin 1968; Langford and Morin 1976; Card 1990). Tectonic build-up of the province, mostly between 2.72–2.68 Ga, is interpreted in terms of accretionary growth process that involved collisions of many microcontinental blocks and juvenile volcanic arcs (Percival et al. 2006).

Accretionary-type orogeny is commonly considered one of the main processes behind growth of continental crust with time (Şengör and Natal'in 1996; Cawood et al. 2003; Brown 2009). It may also be applicable to Archaean granite–greenstone complexes, as these are commonly characterized by linear belt structures in which early accretion of formations from various oceanic tectonic environments and microcontinents is followed by arc-type magmatism (e.g. Kusky and Polat 1999; Bibikova et al. 1999, 2003; Samsonov et al. 2005). Neoarchaean accretion of exotic terranes at c. 2.83–2.75 Ga, culminating in a subsequent major collisional event/orogeny at around 2.73–2.67 Ga, may have been the mechanism that generated the basic structure of the Karelia Province, and which was then strongly reworked during the Svecofennian orogeny. The subduction events enriched the lithospheric mantle in LIL elements. At c. 2.76 Ga, a new volcanic arc began to form above a subduction zone at the margin of the Western Karelia Province, represented now by the Ilomantsi plutonic-volcanic complex. A subsequent slab breakoff or some other subduction-related process at c. 2.72 Ga led to melting of the enriched wedge mantle, producing voluminous sanukitoids and also small amounts of TTGs (Lobach-Zhuchenko et al. 2008; Halla et al. 2009; Heilimo et al. 2010, 2011).

Kontinen et al. (2007) interpreted SHRIMP and TIMS U-Pb age determinations on zircon grains from the paragneiss mesosomes and cross-cutting granitoid plutons to constrain the deposition of protolith wackes to c. 2.70 Ga. Some caution should be taken with this interpretation, as the newly obtained precise age of 2715 ± 2 Ma for the Loso sanukitoid intrusion (Huhma et al. 2012a), which crosscuts the local Nurmes type paragneisses, suggests that metamorphic effects may have influenced the youngest ages obtained for detrital zircon grains in the paragneisses. Deposition more likely took place in a short (10 Ma or less) period just before 2715 Ma, and it is possible that deposition of the Nurmes sediments and sanukitoid plutonism were partly overlapping events. Trace element and U–Pb data suggest that the source comprised mainly 2.75–2.70 Ga TTG and/or sanukitoid-type plutonic and mafic volcanic rocks. The close similarity of the paragneiss and sanukitoid compositions is an important clue to the timing and tectonic setting of the deposition. It is clear that TTG-dominated crust, presently characterizing the Western Karelia subprovince, was not the dominant source of the Nurmes sediments. The presence of MORB-type volcanic intercalations in Nurmes wackes suggests that they were deposited in a back arc or intra-arc setting (Kontinen et al. 2007). The exotic nature of the Nurmes sediments as overthrust must be considered as a serious option.

After the intrusion of the last juvenile granitoids, the QQ quartz diorites at 2.70 Ga, the crust was deformed and metamorphosed. The related process could have been collisional stacking after closure of ocean basins, which thickened the crust between 2.71–2.64 Ga. Vibroseismic images of the crust, as well as tectonic observations from the exposed bedrock, indicate ductile thrusting and related crustal stacking with tectonic transportation from southeast to northwest (Kontinen and Paavola 2006; Korja et al. 2006; Sorjonen-Ward 2006). A similar seismic structure characterised by gently dipping, commonly listric reflections, is also common in other Neoarchaean cratons, and it is interpreted to result from horizontal compression (van der Velden et al. 2006). Fragments of Mesoarchaean (micro) continents, such as the Siurua and Iisalmi complexes, are present as slices in the thickened Neoarchaean crust. Reflecting heat production by radioactive decay in the thickened, predominantly felsic-granitoid crust, a Barrovian-type medium P/T metamorphic framework was developed. The middle and lower parts of the crust were partially melted, producing migmatites and the GGM suite intrusions (Mänttäri and Hölttä 2002; Käpyaho et al. 2007; Lauri et al. 2011).

The thickness of the Neoarchaean crust in the Iisalmi complex was at least c. 40 km on the basis of c. 10–11 kbar pressures from the granulites that represent lower crust and bear evidence of long-term residence at high-temperatures (Mänttäri and Hölttä 2002). The significance of the amphibolite facies high-pressure rocks from the other areas is more problematic to interpret as these rocks record a geothermal gradient that is lower than the normal continental gradient. Apart from Kuhmo garnet-amphibolites, other examples of Archaean high P/T rocks in the Fennoscandian Shield are the eclogites in the Belomorian Province which were metamorphosed at c. 700–800 °C and c. 14–17 kbar, possibly even at pressures exceeding 20 kbar (Volodichev et al. 2004; Mints et al. 2010a; Shchipansky et al. 2012b), i.e. in an eclogite–high-pressure granulite (E-HPG)-type environment, which is consistent with subduction of crustal rocks into the mantle depths (Brown 2007). In Kuhmo the rocks metamorphosed under high pressure only seem to occur in a restricted area surrounded by amphibolite facies rocks that were metamorphosed at c. 6–7 kbar. Therefore, their exhumation might be explained by similar subduction-related tectonic processes that exhume high-pressure rocks at present convergent margins (Beaumont et al. 1996, 1999; Agard et al. 2009).

The Ilomantsi greenstone belts shows low pressure metamorphism at 3.5–5.5 kbar and 550–600 °C, which indicates a gradient that is warmer than the normal continental geotherm. These rocks are juxtaposed with migmatites that normally show pressures of c. 6–8 kbar, and in the Kuhmo amphibolite even 15–16 kbar, which represents the amphibole-epidote eclogite facies in the classification of Brown (2009). The duality of thermal environments is typical for modern plate tectonics, where the belts representing different gradients are juxtaposed by plate tectonic processes (Brown 2009). Although the metamorphic structure in the Karelia Province is not quite similar to that in modern subduction-related orogenic belts, the significant differences in the metamorphic gradient between adjacent domains is easy to explain by subductional/collisional processes that assembled rocks representing various geodynamic settings.

One major problem is that we do not currently have a clear conception of the extent to which the present crustal structure is due to Archaean accretion/thickening or to Palaeoproterozoic orogenic events. Nevertheless, at least the western and eastern parts of the Karelia Province were strongly reworked in the Palaeoproterozoic Svecofennian and Lapland–Kola orogenies, when it was compressionally thickened and subjected to medium P/T type amphibolite facies metamorphism at 1.9–1.8 Ga (Kontinen et al. 1992; Daly et al. 2001, 2006; Bibikova et al. 2001b).

3.10 Conclusions

1. The Western Karelia Province mostly consists of Neoarchaean gneissic granitoids, whereas Palaeoarchaean and Mesoarchaean granitoids (>2.9 Ga) are only locally present. The granitoid rocks are classified into four main groups, which are the TTG (tonalite-trondhjemite-granodiorite), sanukitoid, QQ (quartz diorite-quartz monzodiorite) and GGM (granodiorite-granite-monzogranite) groups. Most ages obtained from TTGs are between 2.83–2.72 Ga, and they define two age groups separated by a c. 20 Ma time gap. TTGs are 2.83–2.78 Ga in the older group and 2.76–2.72 Ga in the younger group. Sanukitoids have been dated at 2.74–2.72 Ga, QQs at c. 2.70 Ga and GGMs at 2.73–2.66 Ga. Based on REE, the TTGs fall into two major compositional groups, low-HREE TTGs and high-HREE TTGs, which originated at different crustal depths; but, the compositions of their protoliths had a significant effect on the REE patterns. Sanukitoids are interpreted as products of melting of subcontinental metasomatized mantle. The GGM group represents partial melting of pre-existing TTG crust that also caused high-grade metamorphism and migmatisation.
2. Existing isotope data on volcanic rocks of the Kuhmo greenstone belt do not provide much evidence for their deposition on significantly older basement in intracratonic environment. The composition of the komatiites in Kuhmo indicates that they were derived from primitive upper mantle, representing fissure-controlled eruptions onto a pre-existing oceanic plateau. The volcanic rocks in the Ilomantsi greenstone belt represent arc magmatism within an attenuated continental margin where older basement rocks were assimilated by younger arc magmatism. Metamorphic evolution of the greenstone belts differs from that of the surrounding migmatites, indicating late-tectonic juxtaposition of greenstone belts and TTG migmatites.
3. At least four discrete subductional systems of 2.88−2.82 Ga, 2.81−2.78 Ga, c. 2.75 Ga and 2.74–2.72 Ga were identified in the greenstone belts of the Belomorian Province. Greywacke units were formed in fore-arc basins at 2.88−2.82 Ga and at c. 2.78 Ga. Volcanic rocks were generated at c. 2.70 Ga in extensional settings which arose fom the orogenic collapse. In the Belomorian Province there are three age groups (2.88−2.87 Ga, 2.82−2.80 Ga and 2.72 Ga) of crustal eclogites which were derived from oceanic basalts and high-Mg rocks in subductional processes.

4. Neoarchaean accretion of exotic terranes at c. 2.83–2.75 Ga and subsequent crustal stacking at around 2.73–2.68 Ga is a possible mechanism that largely generated the present structure of the Karelia Province, although it was again strongly reworked during the Svecofennian orogeny.

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