3.9. GEODYNAMIC CRUSTAL EVOLUTION AND LONG-LIVED SUPERCONTINENTS DURING THE PALAEOPROTEROZOIC: EVIDENCE FROM GRANULITE-GNEISS BELTS, COLLISIONAL AND ACCRETIONARY OROGENS

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Introduction

Proterozoic palaeomagnetic poles from the major shields point to a single apparent polar wander path (APWP) (Piper, 1983), which supports a possible single coherent continental lithospheric plate from c. 2.9 Ga to c. 1.1 Ga. However, the APWP method has intrinsic problems, such as large uncertainties in palaeopole ages and large gaps in the APWP record (e.g., Buchan et al., 1996). Consequently, geodynamic reconstructions of the history of early Precambrian supercontinents are based mostly on geological considerations; however, models reflect significantly different understandings of key geological structures, especially orogenic belts (e.g., Gaal, 1992; Rogers, 1996; Condie, 1998).

Geochronological data demonstrate episodicity in Palaeoproterozoic geological evolution, preceded by a prominent 2.7 Ga peak in the geochronological record, postulated to reflect creation of the first supercontinent (section 3.2) or a small number of composite continents. Palaeoproterozoic crustal evolution encompassed at least incomplete disruption of the supercontinent(s) (Khain and Bozhko, 1988; Mints, 1998; Condie, 2002a), commencing at c. 2.5 Ga. Reassembly at c. 1.75-1.65 Ga followed increased production of juvenile continental crust, which began at c. 1.9 Ga, followed by rapid accretion of arc systems at 1.88-1.84 Ga. Geochronological data also indicate a prolonged period of very low magmatic activity within continental areas between 2.45 and c. 2.1 Ga (Condie, 1998) (this is supported by an apparent lack of large igneous provinces at c. 2.4-2.2 Ga; section 3.3).

Palaeoproterozoic juvenile assemblages dominate within two types of mobile belt: (1) low-grade (greenschist to low-temperature amphibolite facies) volcano-sedimentary and volcano-plutonic belts; analogous Archaean belts are generally termed greenstones—e.g., sections 2.3, 2.4, 3.6, 4.3 and 4.4), and (2) high-grade (high-temperature amphibolite to ultra-high temperature granulite facies) "granulite-gneiss" belts (see also section 3.8). The former belts are interpreted as sutures (collisional orogens) or collapsed continental rifts. Extended volcano-plutonic assemblages at the margins of ancient continents are usually termed accretionary orogens (e.g., Hoffman, 1989c; Windley, 1992) (see also section 3.6).

However, ideas on the nature and tectonic and geodynamic significance of granulite-gneiss belts remain controversial. Structural constraints indicate that many large-scale Palaeoproterozoic granulite terranes evolved within a broadly collisional context (e.g.,

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section 3.8). Charnokite and enderbite intrusions geochemically resembling granitoids of the calc-alkaline and tonalite–trondhjemite series have been ascribed in many cases to arc environments (e.g., Van Kranendonk, 1996; Jackson and Berman, 2000). Consequently, granulite–gneiss belts have been interpreted as sutures/collisional orogens also (e.g., Hoffman, 1989c; Rosen et al., 1994; Daly et al., 2001). However, many features of granulite–gneiss belts are in conflict with this interpretation; reassessment of existing geodynamic models of the granulite–gneiss belts and their relationships with low-grade belts is thus one of the aims of this paper. A new evolutionary model is discussed here, emphasising the interaction of plume- and plate tectonic-related crustal-forming processes during the Palaeoproterozoic (see also section 3.6).

Magmatic and thermal activity during the early Palaeoproterozoic was largely concentrated within Laurentia (defined here as comprising North American and Fennoscandian cratons; Condie, 1990). The Fennoscandian crustal segment forms the northwestern part of the present-day East European craton, which also includes the Volgo-Uralian and Sarmatian segments (see Fig. 3.9-2b). The latter segments existed as separate cratons by c. 2.05–2.0 Ga; their amalgamation with Fennoscandia and Laurentia as a whole may have occurred at c. 1.7 Ga (Bogdanova et al., 2001a). In contrast, late Palaeoproterozoic assemblages are distributed within all continents. The Laurentian and Siberian cratons, where post-Palaeoproterozoic reworking was limited, are utilised here as type examples. The temporal and spatial distribution of the main Palaeoproterozoic assemblages within mobile belts of Laurentia and Siberia are shown in Fig. 3.9-1.

Geological Evolution of Laurentia

The supercontinent interior

2.51–2.44 Ga superplume event and initial rifting of the Archaean Supercontinent. The Palaeoproterozoic evolution of Laurentia (Figs. 3.9-1 and 3.9-2) at c. 2.5 Ga was marked by

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Fig. 3.9-1. Correlation between main events in the evolution of the Palaeoproterozoic mobile belts in Laurentia and Siberia. 1—poorly known continental crust of suggested arc-related affinity; 2—mafic dykes; 3—alkaline mafic–ultramafic intrusions; 4—layered mafic–ultramafic bodies; 5—gabbro-anorthosites; 6—sedimentary successions with volcanic intercalations, within platform cover and rift-related basins; 7—rift-related volcano-sedimentary successions; 8—granulite–gneiss complexes; 9—within-plate granites; 10—MORB-like volcanic rocks and ophiolite assemblages; 11—oceanic island mafic–ultramafic rocks; 12—island arc volcanic rocks; 13—arc-related granitoids; 14–15—time lines: first-order (14) and second-order (15). Shading of columns: blank = granulite–gneiss belts; light grey = accretionary orogens; grey = low-grade volcano-sedimentary belts; dark grey = passive margins. Abbreviations of orogens: T-T—Taltson–Thelon; CBcB—belts about the Cumberland Batholith; TH—Trans-Hudson; EA—eastern North American realm (K, Ketilidian; P, Penokean; Y, Yavapai–Mazatzal); Sv-ac—Svecofennian accretionary orogen; BSB—Belarus–South Baltica; S-p—Svecofennian passive margin; PV-CKB—Pechenga–Varzuga and Circum-Karelian belts; LGB—Lapland granulite belt; A—Akitkan belt; Stan—Stanovoy Province.

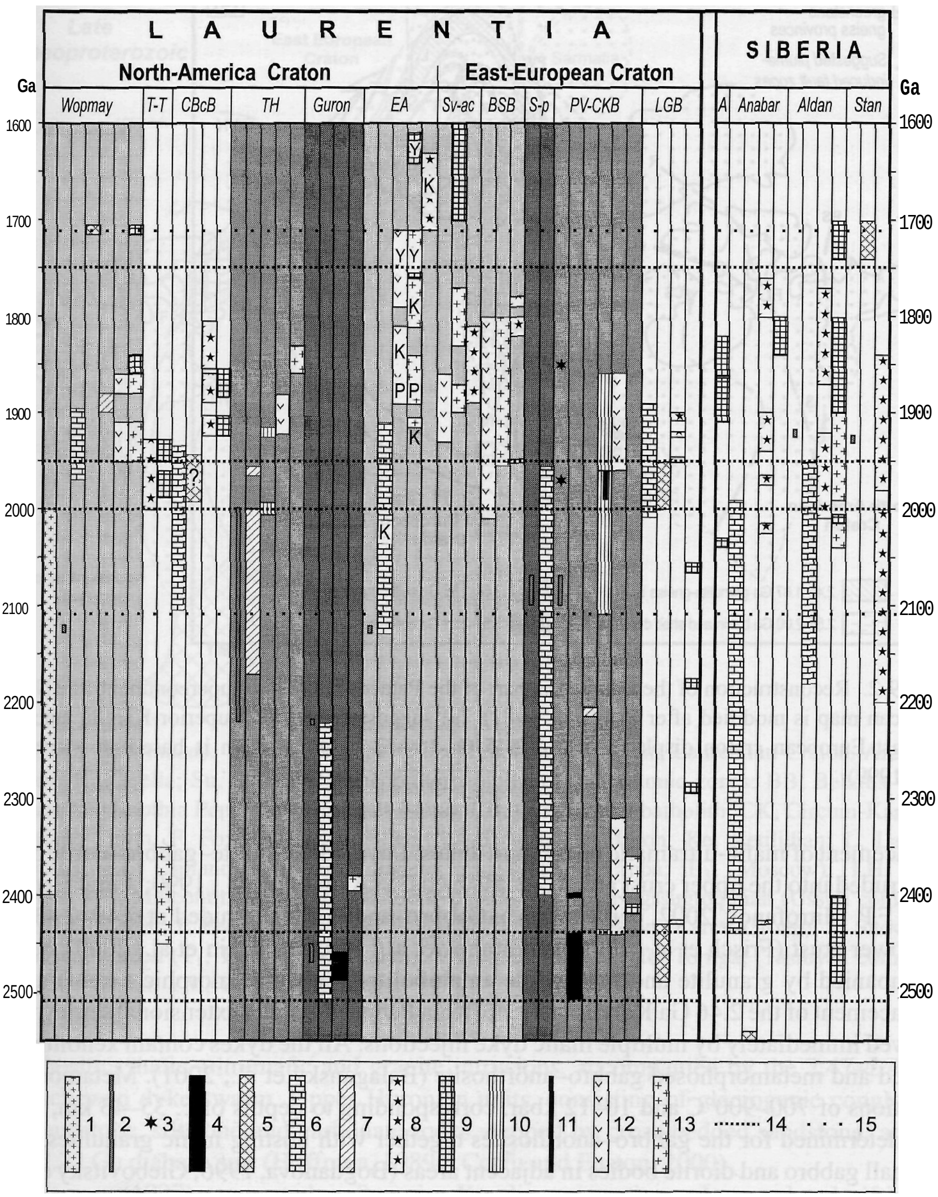
L A U R E N T I A
North-America Craton
East-European Craton
SIBERIA
Ga
Wopmay
T-T
CBcB
TH
Guron
EA
Sv-ac
BSB
S-p
PV-CKB
LGB
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Fig. 3.9-1.

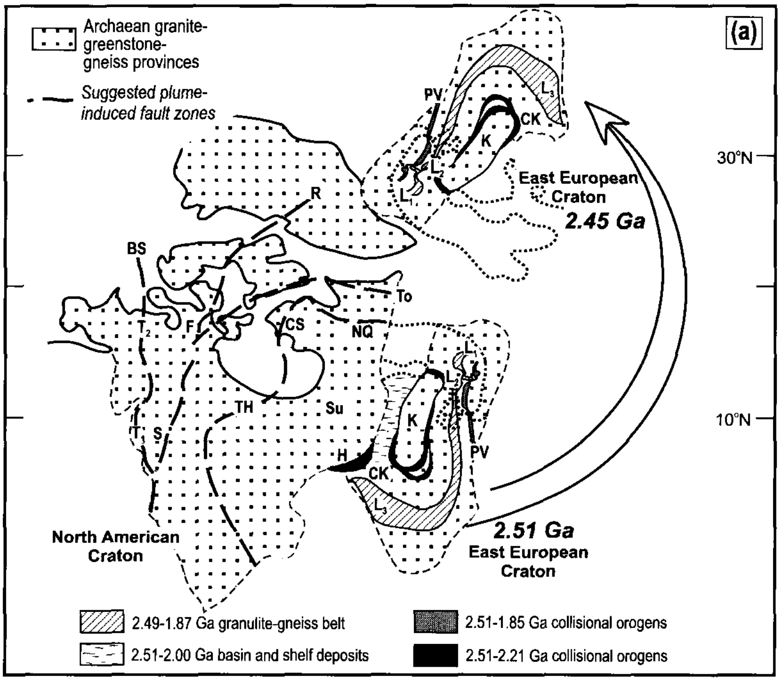


Fig. 3.9-2. Reconstruction of the Laurentian part of the Palaeoproterozoic supercontinent (the North American map is modified after Hoffman, 1989). (a) suggested 2.51 Ga Superior-Karelia junction and East European craton displacement to 2.45 Ga (the 2.45 Ga position is based on Mertanen et al., 1999).

emplacement of mafic-ultramafic bodies and dykes. Layered peridotite–gabbro–norite bodies intruded into the upper crust at 2.51–2.44–2.40 Ga (Alapiety et al., 1990; Amelin et al., 1995; F.P. Mitrofanov, 2002, pers. comm.). Gabbro–anothosites intruded at 2.49–2.43 Ga into lower crust (Frisch et al., 1995; Mitrofanov et al., 1995; Kislitsin et al., 2000), being accompanied by granulite and high-grade amphibolite facies metamorphic assemblages. Emplacement of the 2.46 Ga Kolvitsa massif (Kola Peninsula) in an extensional setting was followed immediately by multiple mafic dyke injections. All the dykes contain xenoliths of sheared and metamorphosed gabbro–anorthosite (Balagansky et al., 2001). Metamorphic conditions of 700–900°C and 10–12 kbar, corresponding to depths of c. 35–43 km, have been determined for the gabbro–anorthosites together with hosting mafic granulites, and for small gabbro and diorite bodies in adjacent areas (Bogdanova, 1996; Glebovitsky et al., 1997; Balagansky et al., 2001).

Initial rifting of the Archaean crust along the southeastern margin of the Superior craton was associated with unconformable deposition of the Huronian Supergroup (section 5.6). The lowest unit consists of flood basalts, felsic lavas and arkosic metasediments. The base of this volcano-sedimentary prism is cut by 2.49–2.46 Ga, within-plate, gabbro–

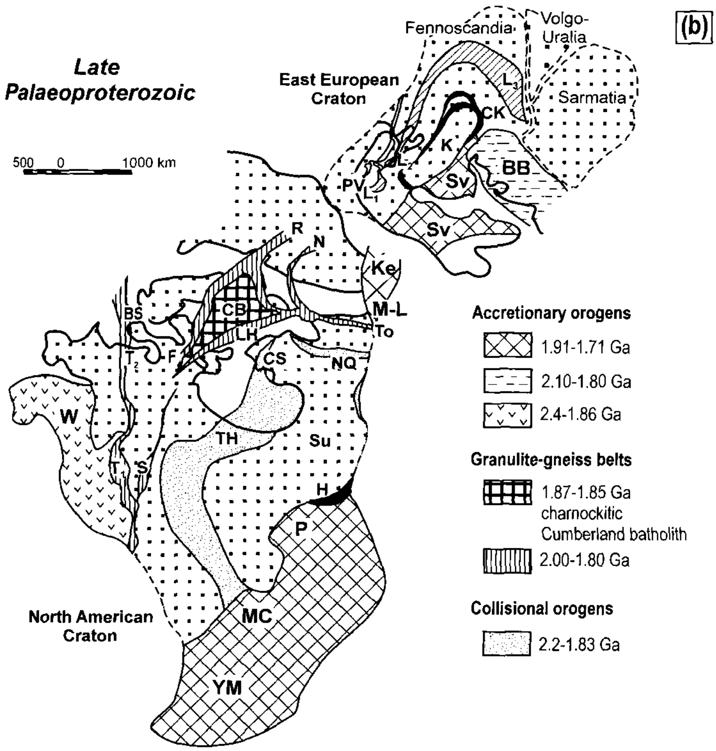


Fig. 3.9-2 *(continued)*. (b) the Palaeoproterozoic mobile belts in the Laurentian Craton. Archaean cratons: K, Karelia; Su, Superior. Palaeoproterozoic belts and tectonic zones: BB, Belarus–South Baltian; BS, Boothia Peninsula–Somerset Island; CB, Cumberland batholith; CK, Circum-Karelian; CS, Cape Smith; F, Foxe (Piling and Penrhyn Groups); H, Huron; Ke, Ketilidian; L, Lapland (branches: L_1, Lapland belt *sensu stricto*; L_2, Kolvitsa–Umba; L_3, Moscow); LH, Lake Harbour; MC, Midcontinent; M–L, Makkovik–Labradorian; N, Nagssugtoqidian (Ussuit complex); NQ, New Quebec; P, Penokean; PV, Pechenga–Varzuga; R, Rinkian (Karrat Group); S, Snowbird; T, Taltson–Thelon (T_1, Taltson; T_2, Thelon); TH, Trans-Hudson; To, Torngat (Tasiuyak complex); Sv, Svecofennian; YM, Yavapai–Mazatzal.

anorthositic, mafic–ultramafic and granite intrusions, accompanied by the 2.47–2.45 Ga Mattachewan dyke swarm. Upper Huronian units, consisting of glaciogenic conglomerate, mudstone, siltstone and carbonate rocks, capped by cross-bedded sandstone, are cut by a 2.22 Ga diabase suit (Hoffman, 1989c; Corfu and Easton, 2000).

Heaman (1997) suggested a Superior-Karelia connection, plume-related rifting at 2.45 Ga, further spreading of the Mattachewan ocean and corresponding breakup of the pre-2.5 Ga continent, with transformation of the Huronian sedimentary basin to a rifted passive margin. Although palaeomagnetic data do not support this model (Mertanen et al., 1999; Buchan et al., 2000), we note the following aspects: (1) a striking uniformity of the 2.51–2.44 Ga Laurentian mafic–ultramafic magmatism, concomitant with its restriction to the Superior and Karelia areas (Vogel et al., 1998); (2) analogy of the 2.49–2.22 Ga Huronian and 2.5–2.3 Ga Lapponian (Karelia) shelf-type sequences; (3) the structural style of the reconstructed palaeocontinent includes oval, concentric mobile belts that permit speculation on inherited initial plume geometry (Fig. 3.9-2); (4) evidence for a Superior-Karelia connection at 2.51–2.50 Ga, predates the palaeomagnetic data of Mertanen et al. (1999) by c. 50–70 My; (5) displacement and rotation permit superposition of Fennoscandia above the 2.44 Ga position based on palaeomagnetic data (Fig. 3.9-2).

2.44–2.0 (2.11) Ga quiescent within-plate development. The period from 2.44 to about 2.0 Ga was tectonically quiescent not only in Laurentia but also worldwide (Condie, 1998) (see also, sections 3.3 and 3.7 for indications of superplume quiescence). Sedimentary–volcanic successions with c. 2.44–2.4 Ga, predominantly mafic eruptive rocks occur in the eastern Fennoscandian Shield, within the Pechenga–Varzuga and Circum-Karelian belts (Melezhik and Sturt, 1994; Mints et al., 1996; Puchtel et al., 1996; Sharkov and Smolkin, 1997). Recent seismic data indicate that these belts are formed by trans-crustal overthrust–underthrust packages (Mints at al., 1996, 2001, 2002; Smithson et al., 2000). Initial volcanism in the eastern part of the Pechenga–Varzuga belt comprised low-K_2O tholeiites associated with shallow water sandstones, succeeded by lacustrine quartzites and predominant low-Ti andesite-basalts, less abundant low-Ti and high-Mg (komatiitic) basalts, and rare andesites and dacites. A suite of 2.44–2.42 Ga dacitic rhyolites and 2.32 Ga subaerial andesites terminates this part of the sequence (Mitrofanov et al., 1991; Amelin et al., 1995; Mitrofanov and Smolkin, 1995). A suite of 2.41 Ga komatiitic basalts and andesite-basalts predominates in the Vetreny belt (southeastern branch of the Circum-Karelian belt) (Puchtel et al., 1996). Geochemistry of these various mafic lavas suggests plume or T-MORB-related sources, although some features can be attributed to subduction-related processes or to crustal contamination.

The second part of this period commenced at c. 2.2 Ga with extrusion of alkaline basalts, trachybasalts with subordinate alkaline picrites, and rhyolites. They are associated with red epicontinental sediments, thick greywacke turbidites, ironstones, siltstones, black shales, dolomites, stromatolitic carbonate rocks, and phosphate-bearing and Mn-rich lithologies within the Pechenga structure (Mitrofanov and Smolkin, 1995), at the rifted passive margin of the Superior craton, in the New Quebec belt and in the Trans-Hudson orogen (Hoffman, 1989c). Between 2.3 and 2.1 Ga a significant part of the Karelian craton was covered by continental shelf sediments which were intruded by diabase sills at 2.2–1.97 Ga (Vuollo, 1994). Similarly, 2.22–1.99 Ga dyke swarms are common in North America (Ernst and Buchan, 2002a, and references therein).

(2.11) 2.0–1.95 Ga superplume event. The second Palaeoproterozoic superplume event (sections 3.2 and 3.3) was more widespread. Plume-related rift-like extension was followed by transition to oceanic spreading in areas of future collisional orogens. Indisputable evidence for ocean growth and closure comes from the 2.0 Ga Purtunq (Scott et al., 1991) and 1.95 Ga Jormua ophiolites (Kontinen, 1987) (northern margin of the Superior craton and western Karelia craton, respectively) and the 2.0 Ga Kittilä ophiolite-like com-

plex (see also section 3.7) in northern Finland (Hanski et al., 1998). The 2.11–1.96 Ga T-MORB-type pillowed tholeiites in the Pechenga structure are interlayered with picritic lavas and felsic ash-flow tuffs that are thought to have erupted from oceanic island volcanoes at 1.99–1.96 Ga. Subvolcanic, Cu- and Ni-bearing gabbro–wehrlite bodies of the same age cut the lower part of the MORB-type succession, but mostly are tectonically included in the accretionary prism known as the "Productive Layer" of the Pechenga ore field (Melezhik and Sturt, 1994; Mints et al., 1996; Sharkov and Smolkin, 1997). The submarine flood basalt province originated at the southeastern margin of the Karelia craton at 1.98 Ga (Puchtel et al., 1998b).

A younger generation of rift-related gabbro–anorthosite intrusions, known at present within the Lapland granulite belt, was emplaced at 2.0–1.95 Ga at the same crustal level as the above-mentioned older generation (Bernard-Griffiths et al., 1984; Kaulina, 1999; Nerovich, 1999; Mints et al., submitted).

1.95–1.75 (1.71) Ga combined plume- and plate tectonic-related evolution. The youngest, 1.87–1.86 Ga assemblages of the Pechenga–Varzuga belt include rhyolites, dacites, andesites, picrites, high-Mg basalts of suprasubduction affinity and subordinate N-MORB-like basalts associated with volcaniclastic sediments, black shales and sporadic cherts (Melezhik and Sturt, 1994; Sharkov and Smolkin, 1997). Circa 1.96 Ga felsic calc-alkaline gneisses have been discovered in the southern vicinity of the Pechenga–Varzuga belt (Daly et al., 2001). The Ticksheozero alkaline mafic–ultramafic–carbonatite complex intruded in northern Karelia at 1.85 Ga (Belyatsky et al., 2000). Abundant subduction-related assemblages are known within the c. 500 km-wide Trans-Hudson belt. Arc magmatism from c. 1.92 to 1.88 Ga was followed by the early outboard accretion of oceanic elements at c. 1.87 Ga, and extensive plutonism between c. 1.86 and 1.83 Ga (Hoffman, 1989c; Stern and Lucas, 1994) (section 3.5).

An extensive basin formed approximately at the same time in the area between the Pechenga–Varzuga and North-Karelian volcano-sedimentary belts. The metasedimentary and meta-igneous rocks deposited therein can be observed presently within the Lapland granulite belt *sensu lato* (LGB). Ages of detrital zircons from the metasediments range from 2.71 to 1.88 Ga; together with Sm–Nd data, this indicates that the youngest sediments were deposited at c. 1.9 Ga (Huhma and Meriläinen, 1991; Sorjonen-Ward et al., 1994; Balagansky et al., 1998; Daly et al., 2001; Bridgwater et al., 2001). Taking into account the 2.0–1.95 Ga age of a younger gabbro–anorthosite generation, sedimentation must have begun not earlier than 2.0 Ga. The relationship of this succession to subduction processes in the neighbouring volcano-sedimentary belts remains unclear.

High-grade metamorphism followed immediately after emplacement of the gabbro–anorthosites of each generation. The corresponding events ($M1_1$ at 2.46–2.43 Ga and an $M1_2$ at c. 1.95 Ga; Kaulina, 1996, 1999; Nerovich, 1999) at similar conditions of 860–960°C, 10.3–14.0 kbar (Fig. 3.9-3) and restricted to the lowest part of the belt, were practically coeval with the intrusions. The M2 event was characterised by lower p–T conditions: 800–860°C, 5.8–12.4 kbar. For the most pervasive M3 metamorphism, the p–T conditions were 640–770°C at 4.8–10.7 kbar. The M2 and M3 events, which affected both igneous and sedimentary rocks, date at 1925 Ma and 1917–1902 Ga, respectively

(Bibikova et al., 1993; Kaulina, 1996, 1999; Kislitsyn et al., 1999). The age of thrusting and related exhumation was c. 1.89–1.87 Ga (Bibikova et al., 1993; Alexeyev et al., 1999; Tuisku and Huhma, 1999; Daly et al., 2001). Rapid overthrusting of the hot crustal slices during the M4 stage (550–650°C at 4.5–8.4 kbar) at 1.87 Ga resulted in inverted metamorphic zoning in the "para-autochthonous" rocks. Deep-seated garnet granulite-hosted xenoliths of norite and gabbro–norite to gabbro–anorthosite composition (Neymark et al., 1993; Kempton et al., 1995; Vetrin, 1998), found in Devonian pipes (Kempton et al., 2001) and which are akin to the mafic granulites of the LGB, suggest that the granulite-facies metamorphism affected the rocks within a thick crustal section from c. 70 to c. 25 km depth (see Fig. 3.9-3). Tectonic decoupling of the continental crust occurred during the Palaeoproterozoic collision: a crustal slice c. 40 km thick was detached and overthrust, whereas the lowest crustal rocks were preserved in their original position (Mints et al., 2000a, submitted).

Geophysical data indicate that the continuation of the LGB beneath the platform cover of the East European Craton forms an extended arch-shaped system of belts approximately 2000 km long (Fig. 3.9-2). Near Moscow the thrust-nappe structure of these belts was recognised recently from reflection seismic profiling (Berzin and Mints, unpublished data). Petrological and geochronological studies of drill core samples yielded a 1.98 Ga magmatic age of the leuconorites and enderbites that were emplaced shortly before metamorphism. A preliminary estimate of peak metamorphic parameters of c. 1000°C and c. 10–12 kbar, as well as the age data (Bogdanova et al., 1999) match almost exactly the characteristics of the LGB assemblage in the Kola Peninsula.

The Taltson–Thelon orogenic belt of northern Canada, consisting of the Taltson Magmatic zone (TMZ) and Thelon orogen (TO), resembles the Lapland granulite belt in many aspects of age, composition and metamorphism. The TMZ is dominated by 1.99–1.96 Ga I-type and 1.95–1.93 Ga S-type granitoids, associated with granulite-facies metasedimentary rocks (6–8 kbar and > 900°C). It has been suggested that evolution of the TMZ involved subduction of oceanic crust beneath the Archaean Churchill Province at 1.99–1.96 Ga, followed by collision between this province and the 2.4–2.0 Ga Buffalo Head terrane (Hoffman, 1989c; McDonough et al., 1995; Ross and Eaton, 2002). However, there is new evidence that both I- and S-type granitoids had an exclusively intra-crustal origin (De et al., 2000). Together with the high-temperature metamorphism, this supports a within-plate (Chacko et al., 1994; Farquar et al., 1996) and possible plume-related origin for the TMZ. The TO was interpreted by Hoffman (1989c) as a product of dextral oblique collision between the Slave (foreland) and Rae (hinterland) Archaean provinces. Predominant granodiorites intruded into the high-grade country rocks (Hoffman, 1989c; Thompson, 1992), and the high-grade terranes of Boothia Peninsula–Somerset Island and Devon–Ellesmere Islands, which form the northern extension of the TO, are built of mainly Palaeoproterozoic ortho- and paragneisses (graphitic metasediments and marbles) with lenses of mafic and ultramafic granulites. The 1.9 Ga metamorphism at 740–900°C and 6–8 kbar was accompanied by syenitic magmatism and anatexis (Kitsul et al., 2000).

Similarly, a branching system of Palaeoproterozoic mobile belts in northeastern North America and southern Greenland, centred on the 1.87–1.85 Ga charnockitic Cumber-

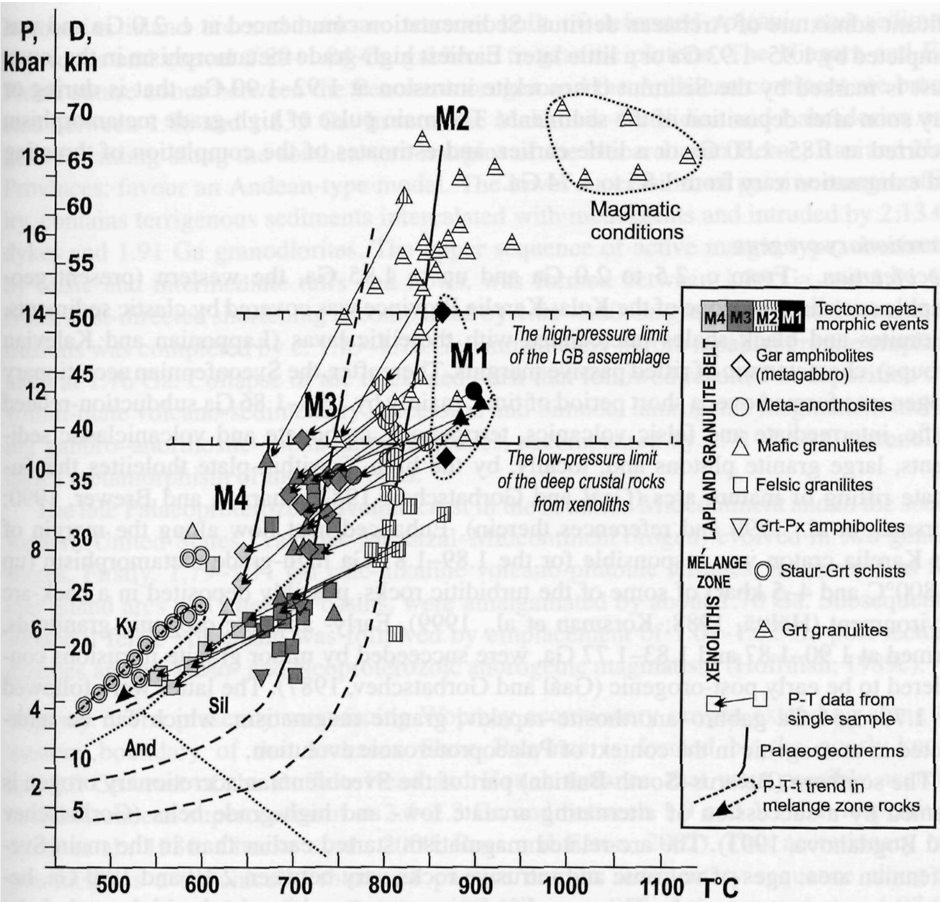


Fig. 3.9-3. *p-T* evolution of the Lapland granulite belt and corresponding deep crustal section (after Mints et al., submitted).

land batholith and comprising the Nagssugtoqidian, Rankain, Foxe and Torngat belts and some tectono-stratigraphic units (Lake Harbour Group, Narsajuaq arc, Ramsay River orthogneisses; St-Onge et al., 1999), can be reassessed in the light of recent geochronological and petrological studies (Taylor and Kalsbeek, 1990; Kalsbeek and Nutman, 1996; Van Kranendonk, 1996; Kalsbeek et al., 1998; Nutman et al., 1999; Scott, 1999; Jackson and Berman, 2000). These belts are formed mainly by granulite gneisses with inferred metamorphic temperatures having reached 950°C and pressures from c. 4 to c. 12 kbar. Protoliths of the lower parts of the metasedimentary sequences were predominantly platform- and rift-related rocks with subordinate evaporitic deposits, mafic and ultramafic volcanics and sills, and anorthositic bodies. The terrigenous metasediments were derived from 2.4-1.93 Ga juvenile Palaeoproterozoic precursors of unknown provenance with significant admixture of Archaean detritus. Sedimentation commenced at c. 2.0 Ga and was completed by 1.95–1.93 Ga or a little later. Earliest high-grade metamorphism in the lower crust is marked by the Sisimiut charnockite intrusion at 1.92–1.90 Ga, that is during or very soon after deposition of the sediments. The main pulse of high-grade metamorphism occurred at 1.85–1.80 Ga or a little earlier, and estimates of the completion of thrusting and exhumation vary from 1.85 to 1.74 Ga.

Accretionary orogens

Svecofennian. From c. 2.5 to 2.0 Ga and up to 1.95 Ga, the western (present geographic coordinates) slope of the Kola–Karelia Province was covered by clastic sediments, dolomites and black shales intercalated with tholeiitic lavas (Lapponian and Kalevian Groups), characteristic of rifted passive margins. Thereafter, the Svecofennian accretionary orogen was formed over a short period of time, mainly by 1.93–1.86 Ga subduction-related mafic, intermediate and felsic volcanics, terrigenous, carbonate and volcaniclastic sediments, large granite plutons and, locally, by the high-Ti within-plate tholeiites that indicate rifting of mature arcs (Gaál and Gorbatschev, 1987; Pharaoh and Brewer, 1990; Korsman et al., 1999 and references therein). Enhanced heat flow along the margin of the Karelia craton was responsible for the 1.89–1.81 Ga high-grade metamorphism (up to 800°C and 4–5 kbar) of some of the turbiditic rocks, possibly deposited in a back-arc environment (Hölttä, 1988; Korsman et al., 1999). Early- and late-orogenic granitoids, formed at 1.90–1.87 and 1.83–1.77 Ga, were succeeded by minor granite intrusions considered to be early post-orogenic (Gaál and Gorbatschev, 1987). The latter were followed by 1.70–1.54 Ga gabbro–anorthosite–rapakivi granite magmatism, which can be interpreted as anorogenic in the context of Palaeoproterozoic evolution.

The southern (Belarus–South-Baltian) part of the Svecofennian accretionary orogen is formed by a succession of alternating arcuate low- and high-grade belts (Gorbatschev and Bogdanova, 1993). The arc-related magmatism started earlier than in the main Svecofennian area; ages of volcanic and intrusive rocks vary between 2.10 and 1.80 Ga, becoming younger westwards. The granulite-facies metamorphism in the high-grade belts (up to 900°C and 8–10 kbar; Scridlaite and Motuza, 2001; Taran and Bogdanova, 2001b) occurred at 1.82–1.80 and 1.79–1.78 Ga and, like the arc-related magmatism, tends to become younger westwards. The 1.63–1.61 Ga metamorphic event is inferred to be linked to anorogenic magmatism. The mostly juvenile metaterrigenous component of the granulite–gneiss assemblages contains detrital zircons with ages between 2.45 and 1.98 Ga, and minor admixture of Archaean detritus (Claesson et al., 2001). Some of the detrital material was derived from a rather old Palaeoproterozoic source, which is unknown in this area. It is suggested that high-grade metamorphism was linked with back-arc and/or post-collisional extension of thickened crust. The thermal history includes a 1.55–1.45 Ga imprint related to anorogenic anorthosite–rapakivi granite magmatism (Bogdanova et al., 2001b). Incorporation of the Belarus–South-Baltian juvenile terrane into the main body of the Palaeoproterozoic crust started before 1.96 Ga (Claesson et al., 2001).

Penokean, Makkovik–Ketilidian, Labradorian and Yavapai–Mazatzal. The Penokean orogen is accreted to the southern edge of the Neoarchaean Superior craton (Fig. 3.9-2).

It contains island arc and back-arc terranes built of deformed volcanic and sedimentary rocks and coeval 1.89–1.84 Ga gabbroic to granitic plutons. The Niagara and Eau Pleine suture zones between the Penokean orogen and the Archaean continent are bracketed between 1.86 and 1.835 Ga. In turn, the Makkovik–Ketilidian and Labradorian orogens extending along the southeastern margins of the Neoarchaean Herne, Rae and Nain Provinces, favour an Andean-type model. The lower succession, of passive margin affinity, contains terrigenous sediments intercalated with metabasalts and intruded by 2.13 Ga dykes and 1.91 Ga granodiorites. The upper sequence of active margin type, dominated by felsic and intermediate tuffs and flows, was formed between 1.86 Ga and 1.81 Ga. Northwest-directed shortening accompanied by 1.81–1.80 Ga granodiorite and tonalite intrusions was completed by c. 1.79–1.76 Ga, immediately before rapakivi granite emplacement at 1.76 Ga. Collapse of the thickened crust that followed resulted in deposition of a within-plate volcano-sedimentary succession and bimodal anorogenic magmatism including gabbro–anorthosite–monzonite complexes, accompanied by 1.71–1.63 Ga granulite-facies metamorphism of the host rocks.

The late Palaeoproterozoic juvenile crust in the southern Midcontinent and in the southwestern United States (Yavapai–Mazatzal–Midcontinent orogen) evolved in two general stages. Firstly, 1.79–1.71 Ga calc-alkaline volcano-plutonic terranes, interpreted as former island arcs and inter-arc basins, were amalgamated by about 1.70 Ga. Subsequently, subaerial felsic volcanism was followed by emplacement of 1.64–1.62 Ga post-tectonic granites and by Meso- and Neoproterozoic anorogenic magmatism (Hoffman, 1989c).

Wopmay. The presently west-facing Wopmay accretionary orogen extending along the western boundary of the Archaean Slave Province is formed by the poorly-known 2.4–2.0 Ga crust of the Buffalo Head and Hottah terranes, that were possibly accreted to the Archaean supercontinent at 2.4–2.3 Ga and between 1.9 and 1.7 Ga, respectively (Goff et al., 1986; Chacko et al., 2000; Ross and Eaton, 2002). This accretionary orogen includes a tectonically shortened passive margin sedimentary sequence along the western boundary of the Slave Province. Shelf-type sedimentation started at c. 1.97 Ga, and at 1.90–1.88 Ga rift-related bimodal magmatism occurred mainly along the off-shelf boundary. Some magmatic arc terranes consisting of 1.95–1.91 and 1.88–1.86 Ga juvenile crust were accreted successively to the Archaean continent. Some of them are underlain by cryptic 2.4–2.0 Ga (possibly 2.3–2.1 Ga) crust. Accretion was followed by post-orogenic, 1.86–1.84 Ga syenogranites and 1.71 Ga anorogenic rift-related diorites, gabbros, and subordinate anorthositic and syenitic rocks in the Yukon at the northwestern edge of the Wopmay orogen (Thorkelson et al., 2001). The process as a whole was terminated by a final phase of deformation, from 1.84 to 1.66 Ga (Hoffman, 1989c; Ross et al., 1991).

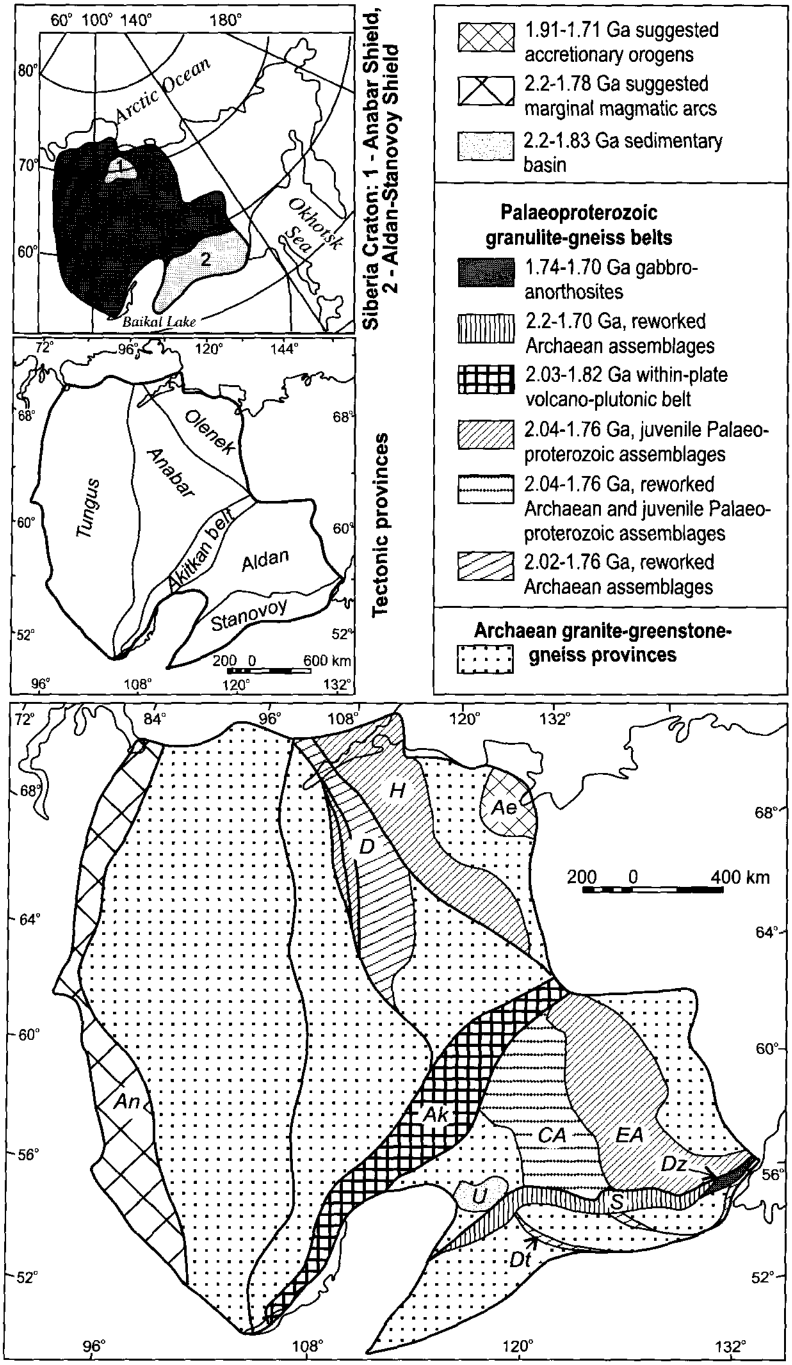
Geological Evolution of Siberia

The important difference between Siberia and Laurentia lies in the lack of Palaeoproterozoic low-grade volcano–sedimentary belts within the Siberian craton (Fig. 3.9-4), with the possible exception of the 2.03–1.82 Ga Akitkan belt (Gusev and Peskov, 1992; Rosen et al., 1994; Nozhkin, 1999). In contrast, Palaeoproterozoic granulite–gneiss belts are widely distributed within all tectonic provinces in Siberia. Rosen et al. (1994) argued that the origin of these belts was linked with the Palaeoproterozoic amalgamation of the Archaean microcontinents and collisional stacking of the crust. Frontal portions of these belts marked by blastomylonites, anorthosites and high-grade metamorphism have been directly interpreted as suture zones (Fig. 3.9-4). This concept coincides in the main with Hoffman's (1989c) model for the Palaeoproterozoic evolution of North America.

The Anabar and Olenek Provinces comprise Archaean and Palaeoproterozoic rocks, the latter of which form the Daldyn and Hapschan granulite–gneiss belts, characterised by magmatic ages of 2.55 Ga for gabbro–anorthosite and 2.42 Ga for granulite-facies mafic metavolcanics. Metagreywackes and metacarbonates of the Hapschan belt were deposited in a shelf-related setting of a passive continental margin between 2.44 and 2.08–1.97 Ga. Based on Sm–Nd data, the source areas were composed of mature Palaeoproterozoic continental crust, which is not observed at the present-day erosion level (Zlobin et al., 2002). The high-grade metamorphic overprints affecting both Archaean and juvenile Palaeoproterozoic assemblages occurred at 2.18, 1.97, 1.94–1.90 and 1.80–1.76 Ga. Predominantly subalkaline granitoids intruded at 1.84–1.80 Ga (Rosen et al., 1994, 2000).

Palaeoproterozic evolution in the Aldan Province began with emplacement of 2.49–2.40 Ga, within-plate, A-type granites (Rosen et al., 1994; Mints et al., 2000b and references therein). There are no precise data on other events during the early Palaeoproterozoic; however, recently acquired Nd model ages of Palaeoproterozoic metasedimentary granulites and igneous rocks range from 2.5 to 2.0 Ga (Kotov et al., 1995; Kovach et al., 1999; Rosen et al., 2000, 2002). The 9–12 km thick, 2.18–1.95 Ga Udokan Group, filling the intracontinental or passive margin sedimentary basin in the western part of the Aldan Province, consists predominantly of copper-bearing quartz arenites with intercalations of black shales, marine carbonates and molasse conglomerates. Emplacement of the Katugin alkaline granite at 2.01 Ga was related temporally and spatially to Udokan Group deposition. The calc-alkaline and subalkaline metavolcanic and metasedimentary granulites in the central and eastern parts of the Aldan Province were derived, based on Sm–Nd data, from both Archaean and Palaeoproterozoic sources. Soon after deposition, they underwent high-grade metamorphism (800–970°C at 7.0–10.7 kbar) together with the adjacent Archaean rocks. The 2.01–1.92 Ga granulite-facies metamorphism was broadly coeval with emplacement of a 2.04–2.01 Ga tonalite–trondhjemite complex and 1.99–1.90 Ga granitoids of various compositions. Later crust-forming episodes occurred after east–west compression as well as thrusting associated with intrusions of 1.92 Ga mafic–ultramafic dykes, 1.90–1.80 Ga A-granites and 1.87–1.77 Ga granulite-facies metamorphism. The final

Opposite: Fig. 3.9-4. The Palaeoproterozoic mobile belts in the Siberian craton (modified after Rosen et al., 1994). Palaeoproterozoic belts, tectonic zones and massifs: Ae, Aekit; An, Angara; CA, Central Aladanian; D, Daldyn; Dt, Dzheltulak; Dz, Dzugdzur gabbro-anorthosite; EA, East Aldanian (Uchur); H, Hapchan; S, Sutam; U, Udokan sedimentary basin.



Arctic Ocean
Okhotsk Sea
Baikal Lake
1
2
Siberia Craton: 1 - Anabar Shield, 2 - Aldan-Stanovoy Shield
Tungus
Anabar
Olenek
Akitkan belt
Aldan
Stanovoy
Tectonic provinces
200 0 600 km
1.91-1.71 Ga suggested accretionary orogens
2.2-1.78 Ga suggested marginal magmatic arcs
2.2-1.83 Ga sedimentary basin
Palaeoproterozoic granulite-gneiss belts
1.74-1.70 Ga gabbro-anorthosites
2.2-1.70 Ga, reworked Archaean assemblages
2.03-1.82 Ga within-plate volcano-plutonic belt
2.04-1.76 Ga, juvenile Palaeo-proterozoic assemblages
2.04-1.76 Ga, reworked Archaean and juvenile Palaeo-proterozoic assemblages
2.02-1.76 Ga, reworked Archaean assemblages
Archaean granite-greenstone-gneiss provinces
H
Ae
D
An
Ak
CA
EA
Dz
U
S
Dt
200 0 400 km

evolutionary stage saw 1.74–1.70 Ga within-plate magmatism (rapakivi-type batholiths, K-rich ash-flows, alkaline granites) on the eastern flank of the Aldan Province.

In the Stanovoy Province, directly south of the Aldan Province, juvenile Palaeoproterozoic volcano–plutonic rocks are poorly known. The 1.74–1.70 Ga Dzhugdzur gabbro-anorthosite massif is the only exception (Sukhanov and Zhuravlev, 1989; Neymark et al., 1992). Archaean granite–greenstone units contain fragmented linear belts and blocks of Archaean rocks, which bear the imprints of Palaeoproterozoic high-grade metamorphism at 2.2–2.0 Ga and 1.98–1.84 Ga (Mints, 2000b and references therein). The latter event preceded westwards-directed crustal shortening. The Palaeoproterozoic evolution terminated after 1.7 Ga with northwards overthrusting of the Stanovoy Province and its juxtaposition with the Aldan Province. The peak granulite-facies conditions (up to 1000–1100°C and 9.5–10.0 kbar; Karsakov, 1978), characteristic for the deepest crustal section (seen in outcrop within the Sutam unit, situated close to the boundary between the Aldan and Stanovoy Provinces), may have occurred at 1.98–1.84 Ga or later, synchronously with emplacement of the Dzhugdzhur gabbro-anorthosite massif and immediately before Aldan–Stanovoy juxtaposition.

Sm–Nd studies indicate that all or at least most of the Palaeoproterozoic juvenile magmatic rocks examined in the Siberian craton, including the gabbro–anorthosites, crystallised from melts strongly contaminated by Archaean crustal material (Kotov et al., 1995; Kovach et al., 1999; Rosen et al., 2000, 2002).

Geological Evolution of Other Continental Areas

The above discussion of the assemblages in the Laurentian and Siberian cratons illustrates the main lines of crustal evolution of, and most important events in, the Palaeoproterozoic. High-grade complexes formed in the early Palaeoproterozoic are also known from India and Antarctica (Raith et al., 1990, 1999; Harley, 1998; Asami et al., 2002). In contrast, both low-grade volcano-sedimentary and high-grade belts of 1.95–1.90 Ga and younger ages are distributed sporadically in various continental areas of Asia, Australia, Africa and South America (e.g., de Almeida et al., 2000; Martin et al., 2000; Teixeira et al., 2000; Ernst and Buchan, 2002a; Wei, 2002). The period from 2.44 to about 2.0 Ga was a quiescent one not only in Laurentia but also in African cratons (e.g., Eriksson et al., 1999) and possibly worldwide.

Granulite–Gneiss Belts: Geodynamic Interpretation

It is seen from the above review that the granulite–gneiss belts, besides their metamorphic grades, have a number of specific features, which are not shared by the low-grade volcano–sedimentary and volcano-plutonic belts. (1) The high-grade metamorphism is commonly predated or accompanied by crust-contaminated gabbro–anorthosite intrusions, high-temperature, "dry", within-plate granites and emplacement of enderbite–charnockite. (2) Lower parts of the high-grade sequences are usually formed by rift-related volcanics and meta-arenites derived from the Archaean basement. In contrast, the bulk of the upper metasediments ("khondalites" or "metagraywackes") was derived from juvenile Palaeoproterozoic rocks, whose provenance remains unknown or poorly constrained. The ages of detrital zircons and Sm–Nd data point to the existence of abundant juvenile, crust-contaminated felsic rocks of early to mid-Palaeoproterozoic ages, which are not only unknown in the present vicinity of the granulite–gneiss belts, but are rare in distant areas too. (3) The high-grade assemblages form thrust-nappe ensembles; inverted metamorphic zoning caused by heating from overthrusted hot tectonic slices can be observed in many para-autochthonous complexes. (4) The total thickness of the crustal sections that were affected by a single high-grade event may reach 50 kilometres. (5) Sediments in the uppermost part of the sequences were metamorphosed almost immediately after deposition and before transformation of the sedimentary basin into a thrust-nappe and fold belt. This suggests that intensive heat influx and related high-grade metamorphism affected the lower crust and then expanded into the volcano-sedimentary fill of the depositional basins. Final sedimentation, high-grade metamorphism and thrusting lasted some tens or at most, less than 50 My.

Ideas on the nature and tectonic and geodynamic significance of granulite–gneiss belts remain controversial. In many studies, granulite–gneiss belts have been interpreted as sutures or collisional orogens (e.g., section 3.8). The thermal modelling of metamorphic evolution within collisional orogens calls for significant heating of the thickened crust up to granulite-facies conditions (England and Thompson, 1986; Thompson and Ridley, 1987). On the other hand, the features of granulite–gneiss belts discussed here are in obvious contradiction with such a geodynamic interpretation. Harley (1992) argued that collision-related homogeneous lithospheric thickening can not explain the very high temperatures typical of most granulites for reasonable limitations of critical model parameters, such as thermal conductivity, crustal heat productivity and basal heat flux. Regional granulite metamorphism requires input of more heat than that available from a thickened crust-lithosphere system. The rather slow cooling deduced for many granulite areas requires that the heat source must be external to the crust undergoing metamorphism; in most regional granulite-facies assemblages the high-grade metamorphism and magmatism are probably both consequences of lithospheric-scale thermal processes (Harley, 1992). Harley's arguments are especially relevant in the light of the inferred enormous thickness of granulite-facies crustal sections.

Another key factor in the formation of granulite-facies terranes is low a H_2O conditions resulting in relatively anhydrous ("dry") mineral assemblages, not only in the metasedimentary granulites but in enderbite–charnockite magmas also (Touret and Hartel, 1990). An increased water activity, which is characteristic for supra-subduction settings, causes partial melting and temperature stabilisation at amphibolite-facies conditions. Regional desiccation of crustal rocks is unlikely in the supra-subduction environment, but may be linked to back-arc extension. Sandiford (1989b) and Harley (1989) considered the possibility of granulite-facies metamorphism during continental rifting and suggested that granulites are forming today beneath the North American Basin and Range province. Similarly, the Early Cretaceous mafic and intermediate granulites of the Fiordland terrane (New Zealand) associated with the mid-Palaeozoic Black Giant anorthosite pluton (Gibson and

Ireland, 1995, 1999) are localised in a back-arc extensional setting (Bradshaw, 1989). The observed abundance of rift-type assemblages in the lower portions of the Palaeoproterozoic granulitic sequences fits with the inferred rift- or back-arc-related origin of modern and young granulite assemblages. In summary, thus, the granulite-facies metamorphism was very likely caused by heat influx of mantle provenance and occurred before collision-related thrusting and thickening of the crust.

Recently published data on detrital zircon ages provide new insights into the origin of granulite successions. Two explanations for the common absence of the source rocks for the granulite-facies metasediments may be suggested. (1) These source rocks originated close to rapidly subsiding basins that were entirely eroded, following which the erosion products were transported to adjacent basins and, soon after deposition, underwent high-grade metamorphism. (2) The metasediments may, at least in part, have been ash-flow deposits, which filled extensive calderas and volcano-tectonic depressions. Although the first explanation seems appropriate for back-arc basins, it appears to be valid in some cases only, as most of the Palaeoproterozoic granulite belts are situated far from arc-type igneous assemblages. Moreover, in some areas with a number of extensive granulite–gneiss belts (e.g., in Siberia, India) the requisite volcano-sedimentary or volcano-plutonic belts are rare or absent. The second suggestion seems more attractive because many features of crustal state, geodynamic setting and lithology are equally characteristic for both granulite protoliths and ash-flow deposits. Most important among those features are: (1) close link with back-arc and rift-related settings (Yarmolyuk and Kovalenko, 1991); (2) great volume and high rate of eruption, caldera collapse and filling (Smith, 1979); (3) "dry" or water-undersaturated, high-temperature magmatic conditions (up to 940°C; e.g., Sutton et al., 2000); (4) correlation with granulite-facies metamorphism of mafic cumulates provided by deep crustal xenoliths (Smith et al., 1996); (5) common occurrence of ortho- and clinopyroxene, olivine and in some cases garnet crystals in the ash-flow assemblages and associated intrusives (e.g., Beddoe-Stephens and Mason, 1991; Sutton et al., 2000).

Thus, we consider that granulite–gneiss belts resulted from: plume-induced heating; magmatism; emergence of riftogenic basins and volcano-tectonic depressions, their filling with rift-type sediments and juvenile but strongly contaminated lavas and ash-flow deposits; high-grade recrystallisation of the lower- and mid-crustal assemblages including the intraplate and back-arc basin-fills; and final thrusting and exhumation of high-grade assemblages caused by collision-related tectonism. The granulite–gneiss assemblages form intraplate belts of regional extent and some local inclusions within both accretionary and collisional orogens.

Speculations on the Interaction of Palaeoproterozoic Plumes and Plate Tectonics

Reassessment of the nature of granulite-facies metamorphism leads to the recognition of the within-plate and plume-related origin of major granulite–gneiss belts. It seems probable that the start of plume-related evolution could be the same in both high- and low-grade mobile belts. Furthermore, commencement of Red Sea-type spreading resulted in rapid heat discharge via spreading ridges and cessation of the high-grade metamorphism in the adjacent continental crust. Although the real reasons for the different evolution of the belts remain unknown, it is clear that this difference played a fundamental role in crustal evolution. This new understanding will result directly in decreasing the number of orogenic belts that can be interpreted plausibly as sutures.

Considering the worldwide decrease of magmatic activity from c. 2.45 to 2.2 Ga, Condie (1994b) argued that deep-seated heat generation processes in the Earth could not have been significantly lower during that period. As the ocean ridges are the predominant conduits for heat loss, it is extremely unlikely that plate tectonics should have stopped and restarted again several times (see also section 3.6). In view of this, a feasible explanation for the inferred stability of the supercontinent during the early Palaeoproterozoic and for the scarcity of subduction events along its margins, may lie in the assumption of intensive recycling of juvenile oceanic and young arc-related crust within the oceanic hemisphere (cf. de Wit's intra-oceanic model; e.g., 1998).

Local post- and anorogenic events (bimodal magmatism including gabbro–anorthosite and rapakivi–granite intrusions, and high-grade metamorphism) in different places within a reborn Palaeoproterozoic supercontinent developed at various times after termination of the collisional and accretionary processes, generally after 1.7 Ga. In a certain sense, they can be interpreted as the beginning of a new Meso-Neoproterozoic evolutionary cycle.

Palaeoproterozoic tectonic and crust-forming processes were instigated mainly by 2.51–2.44 Ga and 2.0–1.95 Ga mantle plumes of global significance ("superplumes") (sections 3.2 and 3.3). The plume-related riftogenic and spreading processes within the continental areas can be attributed to "weak attempts" to disrupt the supercontinent (Mints, 1998). The fundamental change in the Earth's history at the Archaean–Palaeoproterozoic boundary was linked to the transition from Archaean "microplate tectonics" to Palaeoproterozoic "supercontinent tectonics" (or "micro-ocean tectonics" having in mind the limited size of the predominantly Red Sea-type oceans that originated within a partially disrupted supercontinent) (sections 3.4 and 3.6). The creation of the first supercontinent at the end of the Archaean (section 3.2), covering a significant part of the Earth's surface, must have played an essential role in the reorganisation of the convection cell system in the underlying mantle. From this perspective, the Palaeoproterozoic era witnessed only incipient breakup of the inferred supercontinent and revival of multi-cell convection in the mantle (see also section 3.6). On the other hand, the style of crustal evolution during the Palaeoproterozoic differs significantly not only from that in the Archaean, but also from that in the Phanerozoic.

We thus postulate that Palaeoproterozoic history can be subdivided into five periods: (1) 2.51–2.44 Ga superplume activity and displacement of Fennoscandia; (2) 2.44–2.0 (2.11) Ga quiescent within-plate development complicated by local plume- and plate tectonics-related processes (see also section 3.7); (3) a 2.0–1.95 Ga superplume event (see also sections 3.2 and 3.3); (4) 1.95–1.75 (1.71) Ga combined plume- and plate tectonics-related evolution, resulting in the partial disruption of the continental crust, and formation of accretionary orogens along some margins of the supercontinent, and rebirth of the supercontinent entity, and (5) $<$ 1.75 Ga post- and anorogenic magmatism and metamorphism.