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Formation and evolution of Precambrian granulite terranes: A gravitational redistribution model

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ABSTRACT

This paper proposes a revision of the gravitational redistribution model suggested by Leonid Perchuk for the formation, evolution, and exhumation of Precambrian high-grade terranes (HGTs) located between granite-greenstone cratons. Such HGTs are separated from greenstone belts by crustal-scale shear zones up to 10 km wide and several hundred kilometers long. Pelite samples far (>~50 km) from the bounding shear zones show coronitic and symplectitic textures that reflect a decompression-cooling (DC) pressure-temperature (P-T) path. On the other hand, samples from within ~50 km of the bounding shear zones are characterized by textures that reflect an isobaric or near-isobaric cooling (IC) path. Local mineral equilibria in the schists from the shear zones record hairpin-shaped clockwise P-T loops. The results of a numerical test of the gravitational redistribution model show the following plausible scenario: The diapiric rise of low-density, hot granulite upward in the crust causes the relatively high-density, predominantly mafic upper crust in the adjacent greenstone belt, consisting of metabasalt and komatiite, to move downward (subducted), cooling the base of the granulites along the intervening syntectonic shear zone. This causes (1) the formation of local convection cells that control the movement of some of the ascending granulite blocks near the contact with the cratonic rocks, and (2) near-isobaric cooling (IC) of the granulite blocks in the vicinity of the boundary with the colder wall rocks. Cooling of granulite blocks farther away from the contact is not arrested, and they ascend to the Earth's surface, recording DC P-T paths. In general, the results of numerical modeling provide support for the buoyant exhumation mechanism of granulites owing to gravitational redistribution within the metastable relatively hot and soft early Precambrian crust that was subjected to high-temperature (HT) and ultrahigh-temperature (UHT) metamorphism.

*posthumous

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INTRODUCTION

Many Precambrian HGTs (high grade terranes) comprise rocks that formed in the lower crust and were subsequently displaced to the Earth's surface; hence they might be expected to have recorded evidence of the prograde, peak (separate pressure and temperature peaks may exist along a P-T path), and retrograde conditions in their mineral assemblages. However, a record of the prograde and peak stages is rarely preserved, whereas the retrograde stage related to the exhumation of the HGT is commonly well recorded by both the geological structures and mineral assemblages. Apart from rare exceptions (e.g., Perchuk et al., 1985; Zeh et al., 2004), the absence of petrological evidence for the prograde stage of high-grade metamorphism is systematic and is considered to be due to the increase in temperature and therefore in entropy, providing the energy for the structural and textural homogenization at peak temperature conditions (e.g., Perchuk, 1986, 1989; Vernon et al., 2008). In contrast, the declining temperature characteristic of the retrograde stage of metamorphism provides a more suitable thermodynamic environment for the preservation of structural and textural records and mineral compositions. Correct interpretation of such retrograde records is the only way to establish and quantitatively evaluate a geodynamic model of the exhumation history of an HGT (Perchuk and van Reenen, 2008; Perchuk, this volume; Mahan et al., this volume, and references therein).

In the last decade the issue of ultrahigh-temperature (UHT) metamorphism has occupied the pages of some geological journals (e.g., Harley, 1998; Santosh et al., 2004; Sajeev and Santosh, 2006; Kelsey, 2008; Brown and White, 2008; Santosh and Omori, 2008a, 2008b, and references) as a result of the discovery of very high temperature Neoarchean metamorphic rocks (>1000 °C) that were found in Canada (e.g., Arima and Barnett, 1984), Antarctica (e.g., Harley, 1985; Harley and Motoyoshi, 2000), South Africa (e.g., Tsunogae et al., 2004), India (Sajeev et al., 2004), and Russia (Fonarev et al., 2006). Three major models have been proposed to explain the UHT metamorphism in the Neoarchaean: (1) episodic assembly and disruption of supercontinents, (2) plume activity, and (3) UHT conditions in the Neoarchean backarc zones in the course of subduction (Brown, 2006, 2008).

Among the diverse hypotheses for the formation, evolution, and exhumation of Precambrian HGTs (see a review in Thompson, 1990) the most popular are the collision-subduction models (e.g., Ellis, 1980, 1987; England and Thompson, 1984; Perchuk et al., 1985; Sandiford and Powell, 1986; Bohlen, 1987; De Wit et al., 1992; Treloar et al., 1992; Samsonov et al., 2005; Brown, 2006; Lopez et al., 2006). With few exceptions (e.g., Perchuk, 1989, 1991; van Reenen et al., 1987, 1990; Perchuk et al., 1996, 2000b; Dirks, 1995; Percival et al., 1997; Bennett et al., 2005) these models were discussed without detailed consideration of the evolution of adjacent greenstone belts against which the HGTs were frequently juxtaposed. For the particularly well-studied Limpopo HGT, Roering et al. (1992a) proposed the so-called "pop-up" model, in which the HGT exhumation was inferred to be the deep crustal response to continental collision. Although this model was able to explain some features of the Limpopo Complex, aspects of the relationship between the HGT and the cratonic wall rocks remained open.

More generally, plate tectonic models are applied to greenstone belts on the basis of petrochemical and/or geochemical data. However, occurrences of adakite or sanukitoid bodies in greenstone belts do not provide an unequivocal basis for the conclusion of a typical subduction origin (Condie, 2005; Condie and Pease, 2008, and references therein) because the magmatic rocks may have been derived from lower crust composed of garnet-bearing metagabbroids (Hamilton, 1998). Moreover, there has so far been only one report of *true Neoarchean eclogite,* which is from the Belomorian mobile belt, along the northern portion of the Karelian Craton in Russia (Volodichev et al., 2004). In this case, the Neoarchean age was proven for coexisting zircon and garnet, based on the U-Pb method and rare earth element (REE) partitioning (Rubatto, 2002; Rubatto and Hermann, 2007).

Indeed, there are several important observations that are not in accord with collisional models and that concern a number of HGTs (including UHT examples) worldwide. (The first three examples below constitute negative evidence and point toward the absence of first-order features, which should otherwise be broadly present.)

Sediments resulting from erosion of orogenic belts produced by putative Precambrian collisions have not been documented around the HGTs. For example, no sediments from the ca. 1.9 Ga Lapland HGT (Fig. 1A) have been found in either the Karelian or the Inari craton of Kola-Fennoscandia (e.g., Kozlov et al., 1990; Perchuk et al., 2000a, and references therein). Similarly, no sedimentary material from the ca. 2.6 Ga Limpopo HGT has been found within the Precambrian successions of the adjacent Kaapvaal and Zimbabwe Cratons of southern Africa (Fig. 1B), although a large amount of sediments, ~5×10^6 km^3, should have accumulated within the restricted period of ca. 2.69–2.65 Ga (van Reenen and Smit, 1996; Kröner et al., 1999; Kreissig et al., 2000, 2001; Dorland et al., 2004).

In addition, granulite clasts derived from the Lapland HGT have not been observed in rare early Precambrian conglomerates of the Kola Peninsula (e.g., Glebovitskii et al., 1996). Moreover, no Archean zircons have been found so far within the Mesoproterozoic Pretoria Sedimentary Formation, whereas the ca. 2 Ga detrital zircon is very common in these rocks (Dorland et al., 2004). This suggests that, in contrast to the Paleoproterozoic continental crust, no significant mountains existed or eroded above Archean HGTs.

Published stratigraphies of Archean cratons (De Wit and Ashwal, 1997, and references therein; Perchuk, 1989, 1991; Hart et al., 1990; Gerya et al., 1997) show that they are mainly composed of orthogneisses ("gray gneisses") and greenstone belt materials. The latter are mainly granite-gneiss with subordinate banded iron formation (BIF), metapelite, metabasite, metakomatiite, and marble. In contrast, xenoliths from younger kimberlite pipes that intruded granulite-facies rocks are mainly composed of garnet-bearing metabasites and peridotites (e.g., Griffin and O'Reilly, 1987a, 1987b). This may indicate that the lower crust beneath HGTs is dominated by (ultra)mafic rocks (Specius, 1998). For example, the early Paleozoic (534 Ma) (Allsopp et al., 1995) Venetia diamondiferous kimberlites that intrude the Central Zone of the Limpopo granulite complex contain ~50% mafic and ultramafic lower crustal xenoliths (Pretorius and Barton, 1997; Barton and Gerya, 2003). Similar compositional characteristics of the lower crust are typical for xenoliths from the Paleozoic Elovy diatreme (e.g., Kempton et al., 1995) that intruded the boundary between the Belomorian Complex and the Tanaelv Belt (Fig. 1A). In addition, Perchuk et al. (1999) described inclusions of relict epidote and amphibole preserved in clinopyroxene of high-grade metabasite in the giant Tanaelv shear zone (see Fig. 1A) that are similar to the widespread assemblages of low-grade metamorphic rocks from the Karelian greenstone belt. The garnet-rich metabasites from the Tanaelv Shear Zone record a P-T of ~12 kbar, 750–800 °C (Fonarev et al., 1994). Thus, the mafic greenstone material was metamorphosed at depths of ~35 km and subsequently was exhumed from the lower crust.

The granulites are typically thrusted onto the cratonic rocks, displaying intrusive-like or harpolith geometry in cross section (a *harpolith* is a large sickle-shaped igneous intrusion injected into previously deformed rocks; Tomkeieff, 1983) (Fig. 1A). This geometry has been constrained from geophysical data, including seismic reflection studies across Limpopo and Lapland HGTs (De Beer and Stettler, 1992; Durrheim et al., 1992; Pozhilenko et al., 1997). This also suggests that marginal parts of granulite-facies terranes may have been significantly transported laterally during exhumation so that they no longer overlie the lower-middle crust from which they were presumably derived.

Crustal-scale shear zones a few kilometers wide and hundreds of kilometers long separate the HGTs from the adjacent cratonic rocks; metamorphic temperature zonation in the cratonic rocks is well documented across these shear zones (e.g., van Reenen and Smit, 1996; Perchuk and Krotov, 1998; Perchuk et al., 1996, 2000b).

The Kaapvaal and Zimbabwe Cratons, between which the Limpopo HGT is located, both exhibit similar lithologies and rock chemistries to those of the granulites, suggesting that the latter are high-grade equivalents of the adjacent greenstone belt

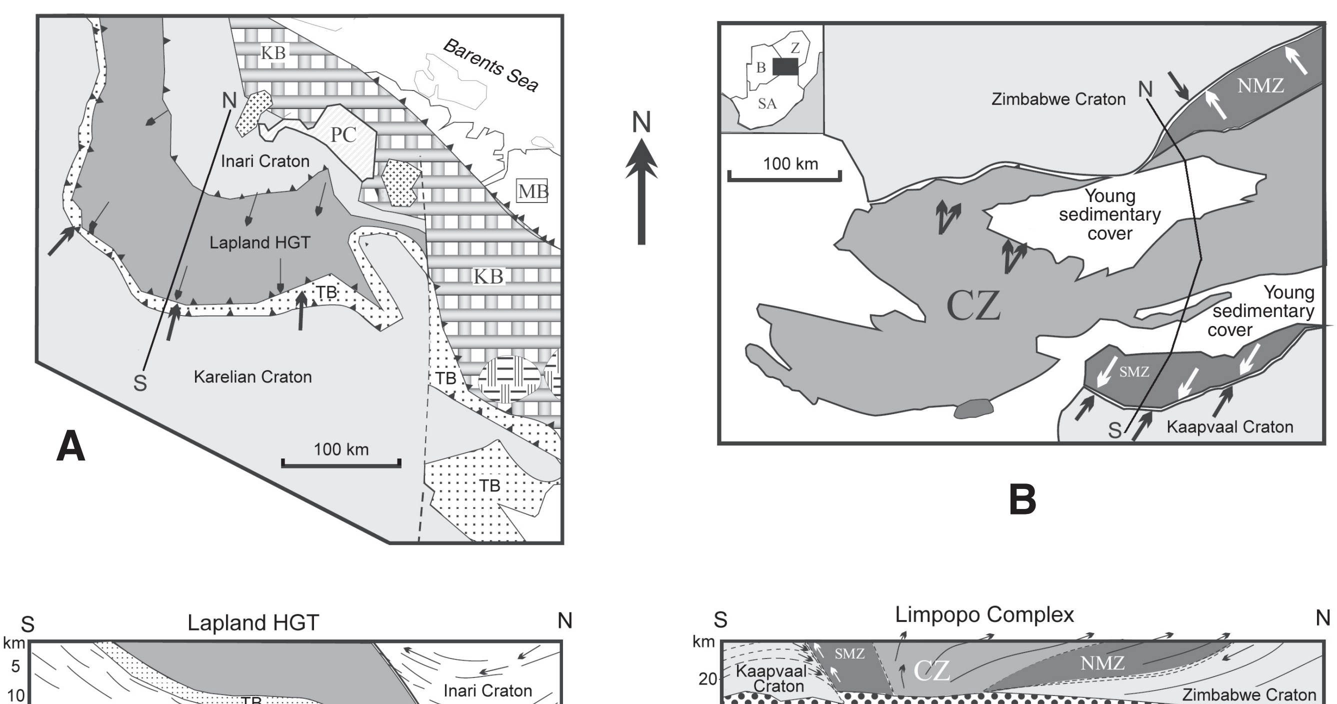

Figure 1. Schematic geological maps (A, B) and cross sections (C, D), based on geophysical data (de Beer and Stettler, 1992; Durrheim et al., 1992; Nguuri et al., 2001; Pozhilenko et al., 1997) for the Lapland and the Limpopo HGTs situated between cratons (taken from Perchuk et al., 2000a). See discussion in text. (A) Lapland HGT: PC—Pechenga Green Stone Block; KB—Kola granulite Block (ca. 2.7 Ga); TB—Tanaelv Shear Zone (1.9 Ga). (B) Limpopo HGT: NMZ—Northern Marginal Zone; CZ—Central Zone; SMZ—Southern Marginal Zone. Inset map: SA—South Africa; B—Botswana; Z—Zimbabwe; black rectangle—Limpopo HGT. Note that the crust beneath the Limpopo HGT is thinner than beneath adjacent cratons, suggesting elevation of the mantle (de Beer and Stettler, 1992). Results of seismic tomography, however, allow an interpretation as garnet-bearing metabasites and metakomatiites (Nguuri et al., 2001; Barton and Gerya, 2003).

rocks (e.g., Petrova and Levitskii, 1986; Petrova et al., 2001; Perchuk, 1989, 1991; van Reenen et al., 1990; Kreissig et al., 2000, 2001). Similar relationships were described for the Lapland HGT, situated between the Karelian and Inari Cratons (e.g., Perchuk et al., 1999, 2000a), and the Sharyzhalgay HGT at the SW shore of Lake Baikal (Petrova and Levitskii, 1986; Perchuk, 1989).

Metamorphic ages derived for the granulites are systematically younger than the greenstone belts of the adjacent cratons. Perchuk (1989) was the first to suggest this relationship as a geochronological-geodynamic rule. Subsequently Kozhevnikov and Svetov (2001) compared metamorphic ages for >50 HGTs and adjacent greenstone belts and did not find any exception to this rule.

These observations provided the basis for an alternative, gravitational redistribution model for Archean HGTs (Perchuk, 1989, 1991; Perchuk et al., 2001). In this model, the Earth's crust is considered as a multilayered system composed of layers of different density that is metastable in the gravity field under relatively low-temperature conditions because of its overall high effective viscosity (e.g., Ramberg, 1981). This viscosity, however, is exponentially lowered with increasing temperature, so that under high-grade conditions gravitational redistribution is triggered in the inherently unstable crustal configuration (e.g., Gerya et al., 2004). The driving forces of the proposed model are thus based on the principle of gravitational instability of density- and viscosity-layered materials within the Earth's crust. Gravitational redistribution is presumably triggered by an enhanced mantle-derived, fluid-heat flow; in contrast to the common term *heat flow* as a conductive heat transfer, the term *fluid-heat flow* defines both heat conduction and heat advection assisted by mantle-derived fluids and melts (Perchuk, 1976, 1989, 1991), which possibly can be related to mantle plume activity.

The gravitational redistribution model has, in particular, been applied to the origin of the Lake Baikal HGT (Perchuk, 1989); the Limpopo, Lapland, and Yenisey HGTs; and the adjacent greenstone belts (Bibikova et al., 1993; Perchuk et al., 1996, 1999, 2000a, 2000b, 2001; Perchuk and Krotov, 1998; Smit et al., 2001; Kreissig et al., 2000, 2001; Gerya and Maresch, 2004), and presumably it also applies to the Aldan HGT (Perchuk et al., 1985; Smelov, 1995) in eastern Siberia, and the Ouzzal HGT in the western Hoggar, Algeria (Ouzegane et al., 2003). The present chapter summarizes a large volume of natural data and compares them systematically to test the *gravitational redistribution model,* focusing on relationships between and within rocks from HGTs and adjacent cratons with particular focus in the context of the seven crucial points listed above. The most detailed testing was done using the abundant natural data from several well-studied granulite complexes between cratons, such as the Limpopo HGT complex of South Africa (Fig. 1B), the Yenisey Range HGT (Perchuk et al., 1989; Smit et al., 2000; Gerya and Maresch, 2004), the Lapland granulite complex of the Kola Peninsula and Fennoscandia (Fig. 1A), and the Aldan HGT (Perchuk et al., 1985; Smelov, 1995) in eastern Siberia. Apart from employing traditional geological, petrological, geochemical, and geophysical evidence, we have also used results of two-dimensional numerical modeling to more rigorously determine the critical geodynamic constraints (Perchuk et al., 1992; Gerya et al., 2000, 2001, 2002, 2004).

It should also be pointed out that crustal thickening and gravitational redistribution processes are not mutually exclusive and can complement each other (e.g., Gerya et al., 2004). Therefore, the main focus of this chapter is not to disprove collisional models proposed for the origin of granulites but rather to bring to attention gravitational mechanisms of high-grade rock movement and exhumation that were possibly dominant in the hot Precambrian crust (Perchuk, 1989, 1991).

MODEL TESTING BY NATURAL OBSERVATIONS

Structural and Lithological Tests

The critical question to be addressed here is: Do structural and seismic data reflect the tectonic history of an HGT as the result of gravitational redistribution? If yes, two major aspects should be addressed for testing geological aspects of the model: (A) the style of deformation within granulite complexes, and (B) the relationships of the complexes to adjacent granite-greenstone belts. These are discussed separately below.

MacGregor (1951) was the first to suggest gravitational redistribution as a major driving force in the formation of granite-greenstone belts. His proposal was based on structural studies of relationships between tonalite intrusions and their host metavolcanic sequences of felsic to ultramafic composition. He suggested that gravity-driven movement of this sequence was triggered by heating and partial melting of the lower crust, resulting in the formation of granite domes among greenstone rocks. On geological maps these domes exhibit oval to circular shapes (Figs. 2A, 2B) that were interpreted to reflect upward, almost vertical movement of granitic material into the overlying carapace of amphibolite- and greenschist-facies rocks that are inferred to have moved simultaneously downward.

Similar structural features are also typical for the HGTs that occur among greenstone belts. If the exhumation of an HGT resulted from gravitational redistribution of material, the sheath folds should be common regional structures that resulted from oppositely directed dome-, diapir-, and plume-like material movements driven by gravity. The sheath folds should not necessarily be vertically oriented, since horizontal movements are dominant at culminate stages in large-scale, laterally spreading, buoyancy-driven flows (e.g., Ramberg, 1981). Complementary (i.e., oppositely directed) structures are represented by narrow, intensely deformed gneissic rims that commonly surround the sheath folds (Figs. 2C, 2D). The Limpopo HGT is perhaps the best example, providing a continuous cross section from the Kaapvaal Craton in the south to the Zimbabwe Craton in the north. This cross section allows direct observation of the change in deformational style from that typical of greenstone terranes to a macro-mélange of highly attenuated granulite-facies rocks that compose the Limpopo HGT. In general, the tectonic style in granite-greenstone belts is characterized by granite domes surrounded by strongly deformed amphibolite- to greenschist-facies rocks. In contrast, the deformational style of the HGT is characterized by complex fold patterns that include mega–sheath folds formed at high grade, the shapes of which are similar to structural domes (Figs. 2C, 2D). For example, the high-grade Central Zone of the Limpopo Belt, despite its polymetamorphic history (Perchuk et al., 2006), clearly preserves the oval to cylindrical style of folding of heterogeneous multilayered crust at high-temperature granulite-facies conditions. Increasing temperature is inferred to have led to a decrease in viscosities of the hot rocks of different densities and, as the result, to the dominance of buoyancy forces. Therefore we should expect to observe this type of structure in many Archean HGTs. According to this interpretation, the cylindrical type of the sheath folds must predominate in the structural pattern. This was recently shown to be the case for the Central Zone of the Limpopo HGT (Perchuk et al., 2008). Another discussion of the structural aspects related to the gravitational redistribution model is given by Smit et al. (this volume).

In many areas around the world, the Precambrian granulite complexes have tectonic contacts with lower grade wall rocks (Kozhevnikov and Svetov, 2001). Detailed geological mapping in the contact zones between HGTs and greenstone belts (including orientations of the principal structural elements such as linear and planar fabrics, as well as asymmetric structures that permitted kinematic analysis) of contact zones between HGT and GSB (granite-greenstone belts), resulted in concluding that the zones commonly have a typical thrust origin (Kozlov et al., 1990; Roering et al., 1992b; Marker, 1991; Pozhilenko et al., 1997; van Reenen and Smit, 1996; Perchuk and Krotov, 1998; Perchuk et al., 1996, 1999, 2000b; Kreissig et al., 2000, 2001; Smit et al., 2000). Systematic structural analyses of these boundaries have resulted in the conclusion that the granulite complexes overrode the granite-greenstone regions along the steep to moderately dipping crustal-scale shear zones that separated the greenstone belt from the HGT. The maps and sections in Figure 1 illustrate the Tanaelv Shear Zone separating the Lapland HGT from the

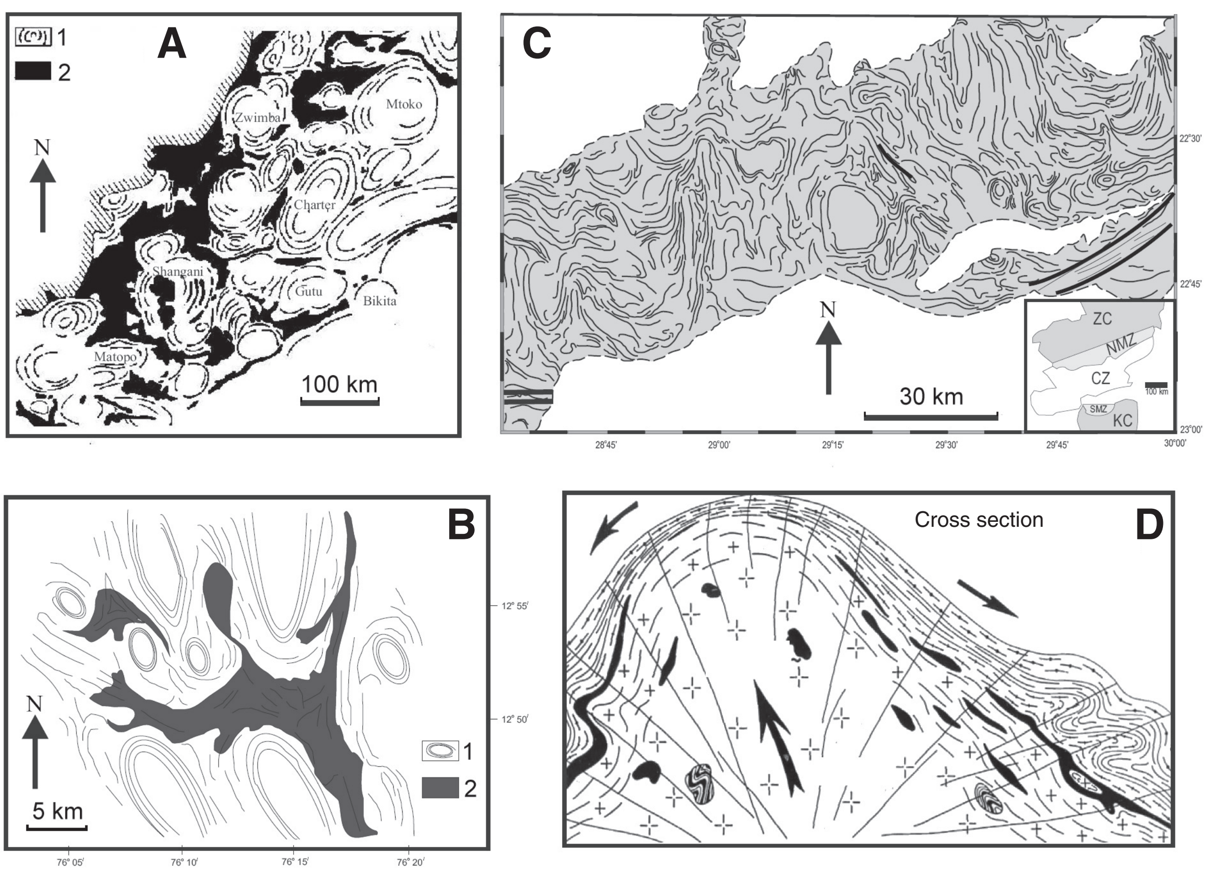

Figure 2. Morphologies of dome structures in Archean granite-greenstone belts (A, B) and high-grade terranes (C, D). (A, B) Granite domes (1—white) within greenstone belts (2—black) of the Zimbabwe Craton, southern Africa (A), and the Western Dharwar Craton, India (B) (MacGregor, 1951). (C) Sheath folds in the Central Zone of the Limpopo HGT, southern Africa (van Reenen et al., 2004). (D) Vertical cross section of a small granite cupola within the Sharyzhalgay granulite complex, SW Lake Baikal (Perchuk et al., 1992). First-order similarities between structural patterns of greenstone belts (A, B) and high-grade terranes (C, D) suggest that gravity-driven tectonics originally proposed for greenstone belts (A, B; e.g., MacGregor, 1951; Ramberg, 1981) may also be relevant for granulite-facies terranes (C, D; Perchuk, 1989, 1991).

Karelian greenstone belt (Fig. 1A), and the Hout River Shear Zone separating the Limpopo HGT from the Kaapvaal granite-greenstone terrane (Fig. 1B). In all cases, the contacting rocks on either side of the shear zone underwent synchronous but oppositely directed movement. Moreover the displacement was synchronous with the formation of sheath folds in gneisses and migmatites in the mid-crust.

Thus, we conclude that structural data from several HGTs and adjacent greenstone belts strongly support synchronous, but oppositely directed, movement of granulites (upward) on the one hand, and host greenstones (downward) on the other.

Geochronological Test

For years, high-grade rocks were considered as the basement of the continental crust and to be older and to have been metamorphosed before the overlying greenstone belts (e.g., De Wit et al., 1992). The gravitational redistribution model, however, suggests the opposite relationships—i.e., that the low-grade metamorphism in the greenstone belt should be older than the high-grade metamorphism in the HGT. The critical question is, therefore: Do the geochronological data document a coherent geochronological history of major geological events that reflect gravitational emplacement of the HGT? A large volume of geochronological data exists for many HGTs and hosting GSBs. Kozhevnikov and Svetov (2001) summarized metamorphic age data for 59 HGT + GSB pairs and clearly showed that in *all* of the pairs the granulites are younger than the contacting greenstones. Table 1 exemplifies this rule with data on the Limpopo and Lapland HGTs. In contrast, the rocks from shear zones separating the GSB from the HGT have similar or identical metamorphic ages with the HGT. For example, the Lapland HGT (1.9 Ga) is much younger than both the Karelian and the Inari adjacent GSBs (more than 2.7 Ga) and shows the same age as mica schists from the Tanaelv Shear Zone (Table 1). The Limpopo HGT (2.69 Ga) is younger than the GSB from both the adjacent Zimbabwe and Kaapvaal Cratons (more than 2.75 Ga), but little difference in age exists for the formation of the Hout River Shear Zone (2.67 Ga), which separates the South Marginal Zone (SMZ) of the Limpopo HGT from the Kaapvaal Craton (Table 1). Similar isotopic age data show rocks from the Northern Marginal Zone (NMZ) (2.62–2.71 Ga) and the adjacent Zimbabwe Craton (more than 2.7 Ga) (Table 1). Thus, geochronological data support an idea that HGTs have been emplaced within GSBs along shear zones much later than the formation of the cratons.

Textural Test for Exhumation

Hundreds of retrograde reaction textures and corresponding P-T paths have been published for the Precambrian HGTs, but there are only a few examples of prograde textures documented in granulites. These include data from the Aldan Shield (Perchuk et al., 1985), Palni Hill Ranges of southern India (Raith et al., 1997), and the Central Zone of the Limpopo Complex (Zeh et al., 2004; Perchuk et al., 2008). Because the exponential increase in diffusion rates above 700–750 °C leads to mineralogical homogenization and microstructural recrystallization, the majority of reaction textures preserved in granulites formed after the peak metamorphic T (e.g., Perchuk et al., 1985; Perchuk, 1989; Harley, 1989).

Metapelites in many HGTs commonly show evidence (Fig. 3) for two mineral reactions (e.g., Harley, 1989):

$$\text{Grt} + \text{Qtz} \rightarrow \text{Opx} + \text{Crd}, \tag{1}$$

i.e., $2(Mg, Fe)_3Al_2Si_3O_{12} + 3SiO_2 = 2(Mg, Fe)_2Si_2O_6 + (Mg, Fe)_2Al_4Si_5O_{12}$,

and

$$\text{Grt} + \text{Qtz} + \text{Sil} \rightarrow \text{Crd}, \tag{2}$$

i.e., $2(Mg, Fe)_3Al_2Si_3O_{12} + 5SiO_2 + 4Al_2SiO_5 = 3(Mg, Fe)_2Al_4Si_5O_{12}$.

Depending on change of P-T parameters, these reactions can be displaced to either side, producing well-developed and distinctive textures (e.g., Harley, 1989; Perchuk et al., 1985,

TABLE 1. U-Pb METAMORPHIC AGES OF ROCKS FROM SHEAR ZONES SEPARATING HIGH-GRADE TERRANES FROM GREENSTONE BELTS

Rock	Zone/Group	Age (Ga)	References
Lapland area (Russia, Finland)			
Ky-bearing mica schist after GSB	Shear zone, western part of the Tanaelv Belt (Finland)	1.9	Bernard-Griffiths et al. (1984)
Ky-bearing mica schist after GSB	Korva Tundra Group (shear zone), the Tanaelv Belt (Russia)	1.91	Volodichev (1990)
Grt amphibolite (sample Lap34)	Kandalaksha Group (shear zone), the Tanaelv Belt (Russia)	1.911	Perchuk et al. (2006)
Opx-Bt plagiogneiss (Lap-9)	Pados, southern part of the Lapland HGT (Russia)	1.91	
Granulitic gneiss	Tupaya Guba, southern part of the Lapland HGT	1.916	Bibikova et al. (1993)
Gneisses-tonalites	Karelian Craton (Belomorides)	2.9–2.7	Kozlov et al. (1990)
Limpopo area (southern Africa)			
Ky-bearing mica schist after KVC	Hout River Shear Zone, at the contact with granulite body	2.689	Kreissig et al. (2000, 2001)
Granulitic metapelite	Southern Marginal Zone (SMZ) of the Limpopo granulite belt	2.691	
Charno-enderbite	Matok Pluton (SMZ)	2.671	Barton et al. (1992)
Metavolcanics	Kaapvaal Craton	2.75–3.54	de Wit et al. (1992)

Note: HGT—high-grade terrane; GSB—greenstone belt; KVC—Kaapvaal Craton; Ky—kyanite; Grt—garnet; Opx—orthopyroxene; Bt—biotite.

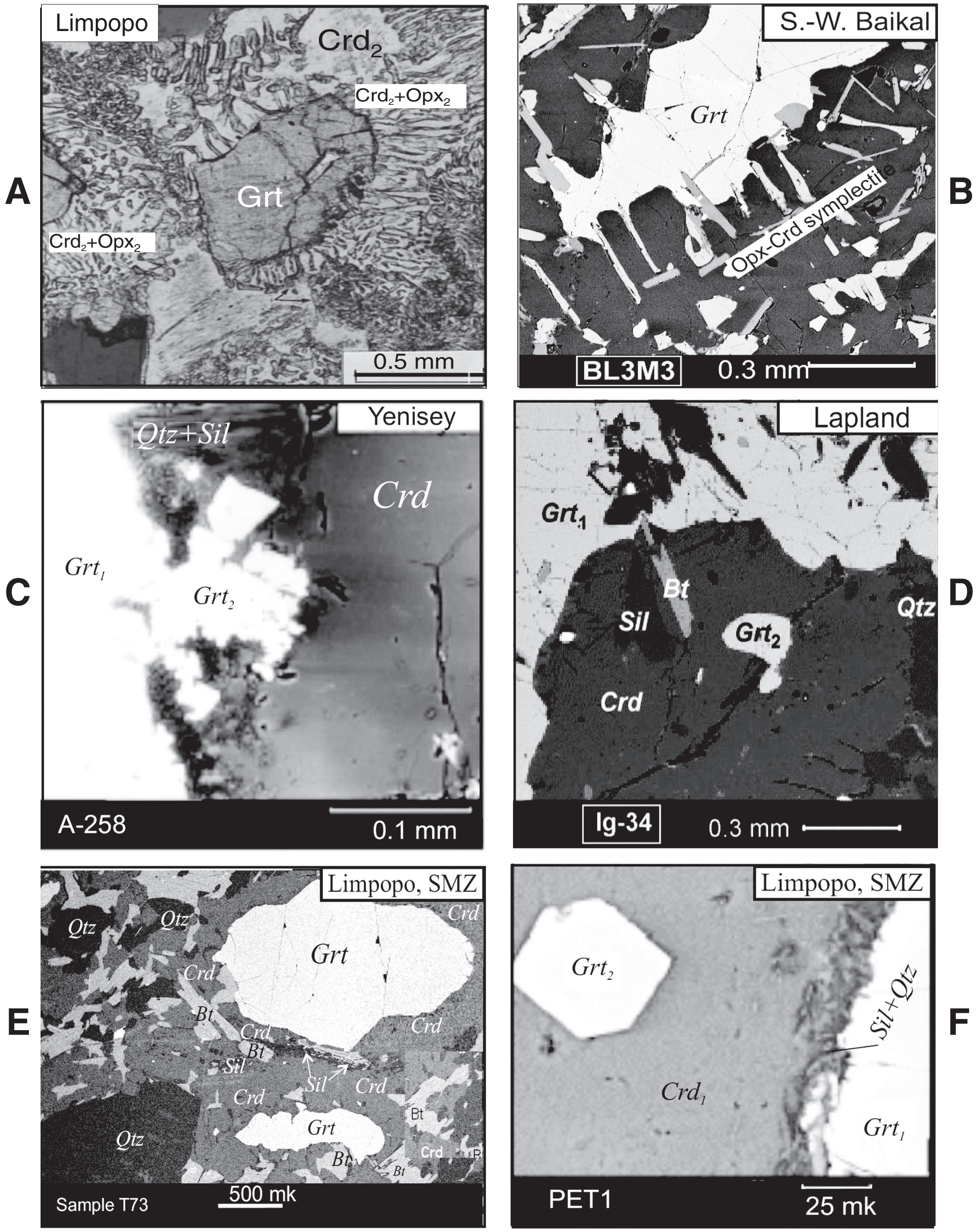

Figure 3. Typical reaction textures *Grt* + *Qtz* → *Opx* + *Crd* (reaction 1), *Grt* + *Sil* + *Qtz* → *Crd* (reaction 2), and *Crd* → *Grt* + *Sil* + *Qtz* (reaction 3) occurring in high-grade metapelites are recorded in Precambrian HGTs worldwide. Grt_1 (*Grt*) and Crd_1 (*Crd*) are stable at the metamorphic peak, and Grt_2 formed from the breakdown of Crd_1 during isobaric cooling (IC) and compression cooling (CC). Crd_2 recrystallized from Crd_1 during the growth of Grt_2. (A) Texture resulting from reaction 1, composed of three zones: (i) *Grt* core → (ii) *Crd-Opx* symplectitic zone → (iii) *Opx* rim, CZ, Limpopo HGT, southern Africa (compiled from Perchuk et al., 2006). (B) The same as (A): Sharyzhalgay HGT, SW Lake Baikal, eastern Siberia (from Perchuk, 1989). (C) Reaction texture 3 in metapelite from the Atamanovskaia Group, the Kanskiy HGT, Yenisey River, eastern Siberia. (D) Reaction texture 3 in metapelite from northern portion of the Lapland HGT, Kola Peninsula, Russia (from Perchuk et al., 1999). (E) Texture resulting from reaction 2; i.e., rim of *Crd* around *Grt* containing *Sil* + *Qtz* in sample from the 2 Ga Baklykraal regional shear structure; CZ, Limpopo HGT, southern Africa (from van Reenen et al., 2004). (F) Texture resulting from reaction 3 in metapelite from the Petronela Shear Zone, SMZ, Limpopo HGT (Smit et al., 2001); new euhedral Grt_2 and *Sil* + *Qtz* intergrowth appeared within *Crd* matrix at the contact with Grt_1: $X_{Mg}^{Grt1} < X_{Mg}^{Grt2}$.

1989, 1996; Raith et al., 1997). Decompression cooling (DC) of an HGT appears the most common style of exhumation type in Precambrian granulite-facies terranes. Metapelites from the SMZ of the Limpopo HGT show textures that reflect *both* reactions (1) and (2). Orthopyroxene coronas and *Opx* + *Crd* symplectites are well developed between garnet and quartz (reaction 1, Figs. 3A, 3B). At a relatively high Ca content in garnet, a plagioclase zone can be formed between the *Opx* corona and the *Crd-Opx* symplectite zone. The commonly observed growth of cordierite at the contact of garnet with sillimanite *(Sil)* and quartz *(Qtz)* is also documented in the metapelites (reaction 2).

On the other hand, the reaction:

$$\text{Crd} \rightarrow \text{Grt} + \text{Qtz} + \text{Sil} \qquad (3)$$

produces textures that are composed of a very fine grained (<100 μm) skeletal intergrowth of garnet, quartz, and sillimanite on the rims of cordierite (Fig. 3C). This texture was first observed in metapelites of the Yenisey Range in eastern Siberia (Fig. 3C; Perchuk et al., 1989) and then in similar rocks of the northern part of the Lapland HGT (Fig. 3D; Perchuk et al., 1999, 2000a) and in the SMZ of the Limpopo Complex (Figs. 3E, 3F; Perchuk et al., 1996). An important observation is that samples with the reaction texture (3) were only collected close (<~50 km) to the boundaries of the HGTs with greenstone-belt wall rocks, whereas samples with reaction textures (1) and (2) commonly occur farther away (>~50 km) from the boundaries.

P-T History

P-T parameters were uniformly calculated for both the high-grade rocks and the rocks from shear zones separating HGTs from the cratons using methods and approaches discussed in detail elsewhere in this volume (Perchuk, this volume).

Granulite Complexes

In the rocks from the HGTs studied we found two groups of P-T paths: decompression cooling (DC) and near-isobaric cooling (IC) paths that are strongly correlated with the two groups of reaction textures noted above, and which are typical for many *different* granulite facies complexes (Harley, 1989). However, both kinds of P-T path occur in a single HGT depending upon their location within the terrane (e.g., Perchuk et al., 1996, 1999, 2000a). For example, in the Limpopo HGT the samples preserving a record of the DC PT path (Fig. 4A) occur far (>~50 km) from the boundary with the craton, whereas samples located close to the boundary (<~50 km) also exhibit IC paths (Fig. 4B). This can be explained if some crustal blocks were exhumed sufficiently slowly so they could equilibrate with the adjacent crust during their emplacement (DC path), whereas the emplacement of others was arrested at crustal levels of ~13–15 km (IC path). We infer that the large temperature gradient between the hot granulite and the cooler wall rocks promoted heat flow from the Limpopo HGT toward the greenschist-facies footwall of the Kaapvaal Craton. This resulted in isobaric cooling of the granulite (IC portion of the P-T path). Hydration reactions (involving the formation of cummingtonite, anthophyllite, gedrite, secondary biotite, water-bearing cordierite, etc.) that characterize many samples from the southern part of the SMZ (van Reenen, 1986) reflect the involvement of water-rich fluids in the exhumation process (Perchuk et al., 1996). This suggestion was later supported by numerical modeling (Gerya et al., 2000) and additional petrological observations in other HGTs (e.g., Perchuk et al., 1999, 2000a; Smit et al., 2000; Gerya and Maresch, 2004).

Rarely, granulites near contacts with cool wall rock may preserve evidence for a compression-cooling (CC) segment of the P-T path in place of an IC segment (Fig. 4B). Figures 4E and 4F, determined for mineral assemblages from the Kanskiy Complex, Yenisey Range, eastern Siberia (Gerya and Maresch, 2004), are inferred to reflect differential vertical movements within parts of the granulite body during its emplacement in the mid-crust. Both the IC and CC portions of P-T paths therefore characterize complications in the flow-cooling patterns of marginal zones of the HGT caused by their thermomechanical interactions with colder and stronger cratonic rocks.

Shear-Zone Wall Rocks

The P-T loop for the footwall rocks in the Tanaelv Shear Zone underlying the Lapland HGT (Fig. 1A) was first derived by Perchuk and Krotov (1998) on the basis of the local equilibrium *Bt* + *Ky* + *Qtz* = *Ms* + *Grt* within the studied samples. Figure 5 exemplifies all stages of the evolution of these garnet-bearing schists from the chlorite-staurolite and kyanite-staurolite zones in the shear zone. Figure 5 documents both the prograde and retrograde stages of the evolution of the shear zone rocks in the footwall in terms of *Grt* morphology, chemical zoning, and P-T path. The rotated garnets with quartz-rich inclusion trails (Fig. 5A) are associated with the prograde stage of metamorphism, whereas snowball (Fig. 5B) and inclusion-free (Fig. 5C) garnet reflects peak and subsequent retrograde metamorphism. Similar textural relationships occur in the Hout River Shear Zone, separating the Limpopo HGT from the adjacent Kaapvaal Craton Perchuk et al., 2000b). Figures 5D–5F demonstrates movement of the shearing rocks downward to the mid-crust and subsequent return toward the surface along with the simultaneous exhumation of granulite (see Table 1 for isotopic data indicating the contemporaneity of decompressional metamorphism in the HGT and prograde metamorphism in the subjacent footwall shear zone).

P-T paths in Figures 5D–5F were calculated using the *Bt-Ky-Qtz-Grt-Ms* geothermobarometer (Perchuk, 1973, 1977), which was recalibrated using new thermodynamic data (Perchuk and Krotov, 1998; Perchuk et al., 2000b). These P-T paths define very tight, hairpin-shaped loops, indicating that the cool subducting plate, composed of greenstone material, and the hot ascending granulite diapir moved along virtually the same P-T gradient. This result is, in a way, similar to the constraint imposed upon a subducting oceanic plate in which both the prograde (downward)

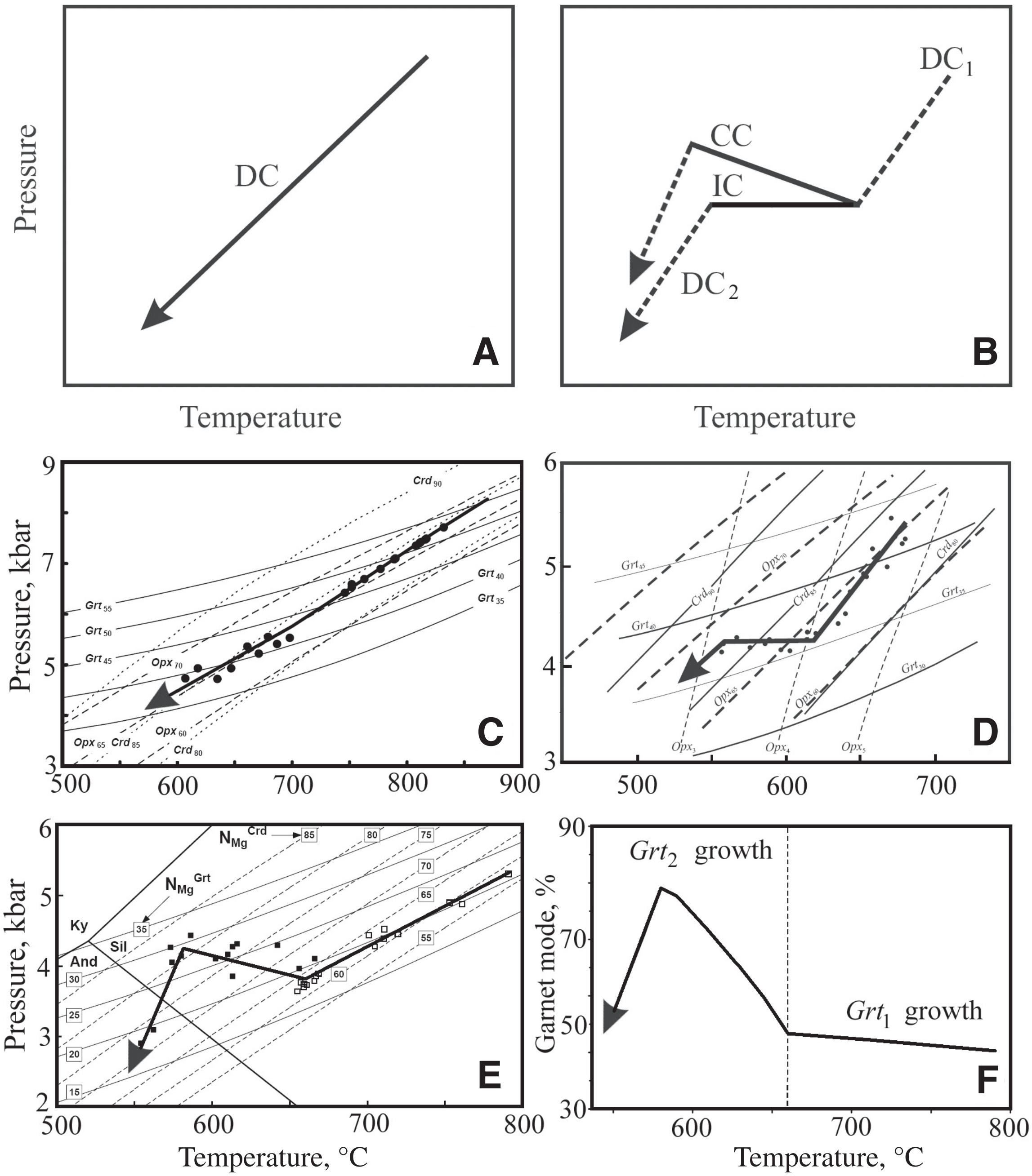

Figure 4. Typical styles of P-T paths for monometamorphic HGTs. (A) Typical decompression-cooling (DC) path. (B) isobaric (IC) and compression-cooling (CC) portions of the DC P-T paths (see Perchuk, this volume). (C) DC P-T path for metapelite DR45 from the SMZ of the Limpopo HGT. (D) DC_1-IC-DC_2 segments of P-T path for metapelite LW7 collected from the most marginal part of the SMZ of the Limpopo Complex. Both C and D after Perchuk et al. (1996). (E) P-T path derived for granulite A-275 is from the Kanskiy HGT (Yenisey Range, eastern Siberia); open rectangles indicate P-T data for the early generation of garnet (*Grt*$_1$), and solid rectangles indicate those for the later generation (*Grt*$_2$). (F) Calculated equilibrium garnet modes [atomic ratio 100(Mg + Fe)$_{\text{in garnet}}$/(Mg + Fe)$_{\text{in rock}}$] along the P-T path in E. Diagrams E and F after Gerya and Maresch (2004). Relative uncertainties of P-T paths (±0.5 kbar and ±25 °C) correspond to the scattering of individual P-T points in diagrams C, D, and E. See text for discussion.

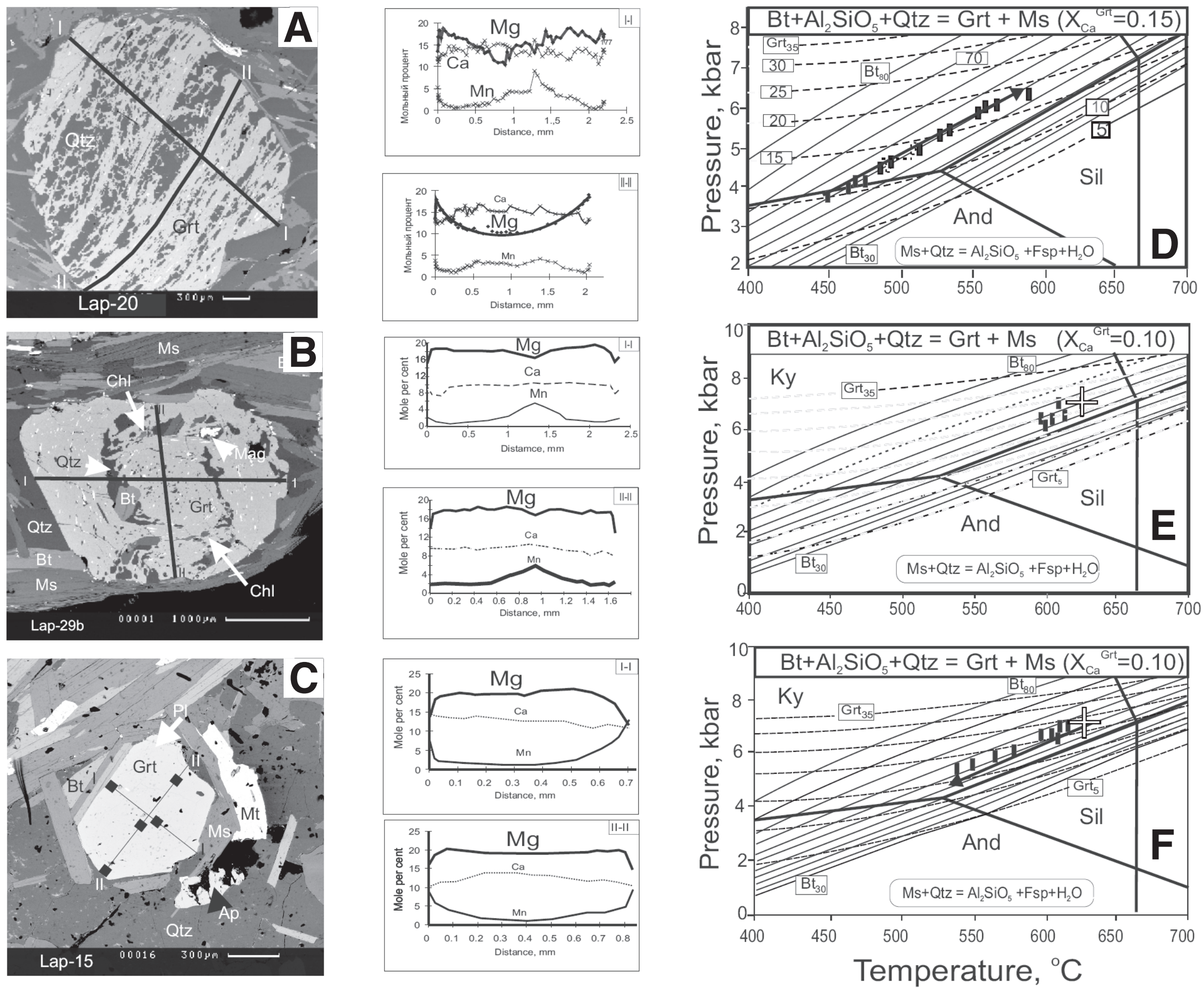

Figure 5. Textural, chemical, and P-T evolution of the mineral assemblage *Bt + Ky + Qtz + Grt + Ms* in mica schists from the Tanaelv Shear Zone, separating the Lapland HGT from the Karelian Greenstone Belt (modified from Perchuk and Krotov, 1998). (A–C) Backscattered electron images of garnet from mica schist, illustrating morphology of the *Grt* porphyroblasts and their textural characteristics in the prograde (A), peak (B), and retrograde (C) stages of metamorphic evolution. *Rotated* garnet porphyroblasts from the chlorite-staurolite zone characterize the prograde stage. *Isometric* inclusion-rich snowboll *Grt* porphyroblasts occur in the biotite-kyanite zone and indicate both partly prograde and peak stages of metamorphism. *Euhedral* inclusions-free Grt porphyroblasts from the chlorite-staurolite zone indicate peak to retrograde stages of metamorphism. Lines indicated as I–I, II–II, etc., correspond to the profiles along which detailed microprobe analyses were done (see second column in the figure). Each path in the (D–F) diagrams relates to (A–C) textural, chemical, and morphological evolution of the *Grt* porphyroblasts shown in (A–C), from prograde (A), through peak (B), to retrograde (C) P-T conditions. Crosses on the P-T diagrams indicate precision of P-T estimates (±0.5 kbar and ±25 °C). See text for discussion.

and the retrograde (upward) histories are similar but reversed (e.g., Cloos, 1982).

P-T diagrams in Figure 6 demonstrate that the peak metamorphic conditions for the footwall rocks coincide with the P-T minimum recorded by the granulite assemblages. Considering the similar ages of metamorphism in both cases (Table 1), we infer that this convergence of P-T conditions implies that the P-T loops for the footwall rocks resulted from the emplacement of the granulite body into the upper crust.

Geophysical Test

The gravitational redistribution (or diapiric) model assumes that the gravitational redistribution of material occurred within the normal-thickness continental crust, produced intrusion-shaped geometries caused by the reduced viscosity of the hot granulite bodies, and was triggered by the activity of large mantle plumes (Perchuk, 1989; Perchuk et al., 2001). This suggestion is exemplified by seismic data for both the Limpopo (e.g., De Beer and Stettler, 1992; Durrheim et al., 1992; Nguuri et al., 2001) and Lapland (Pozhilenko et al., 1997) granulite complexes (Fig. 1). The interpretation of the seismic profiles, coupled with structural and geochronological data, suggests that the Limpopo HGT resembles the shape of a laccolith or harpolith (Tomkeieff, 1983) emplaced between the Kaapvaal and Zimbabwe Cratons (Fig. 1B). The geometry of the Lapland granulite complex, between the Karelian and Inari greenstone belts, is a mirror image of the Limpopo HGT. On the southern and western margins the Tanaelv listric shear zone dips northward under the Lapland HGT at an angle ranging from 60° to 12°. However, the northern boundary of the Lapland HGT with the Inari greenstone belt is still an unresolved problem: According to the structural and seismic data from Finland (Marker, 1991; Mints et al., 1996; Pozhilenko et al., 1997) the northern boundary of the Lapland HGT dips steeply NE, whereas the Russian seismic data suggest that the contact is vertical to steeply S-dipping. A cross section through the Lapland HGT also suggests a harpolith shape (Fig. 1C). Thus, both the Limpopo and Lapland HGTs form bodies which resemble crustal-scale harpoliths that plausibly reflect intrusive-like mechanisms of the emplacement of the intervening HGTs.

In reality, the mantle-derived fluid-heat flow that triggers gravitational redistribution in our model may also associate with either horizontal or vertical movement of crustal material by other processes such as thrust thickening, extensional thinning, erosion, magmatic underplating, or crustal delamination. For example, if the formation of the Limpopo HGT resulted from the collision (horizontal movement) of the Kaapvaal and Zimbabwe Cratons (Treloar et al., 1992), the locations of the plate boundaries are expected to be imaged by deep-seated seismic profiles across the “craton-orogen-craton” system. However, the detailed seismic tomographies of this portion of the African continent show the mantle lithosphere to be a single unit, and there is no seismic discontinuity at the Limpopo HGT (James et al., 2001). Consequently, if no subsequent overprinting affected this lithosphere, it was acting as a single tectonic unit during the exhumation of the Limpopo HGT.

MODEL TESTING BY NUMERICAL EXPERIMENTS

Gravitational mechanism of granulite exhumation was extensively studied numerically (Perchuk et al., 1992, 1999; Gerya et al., 2000, 2004), which is reviewed below. The first numerical testing of a gravitational redistribution model for granulites was performed by Perchuk et al. (1992). Using a mechanical numerical model and a gravitationally metastable model set-up with a rhythmic multilayered crust (e.g., Perchuk 1989) comprising rocks of different density and viscosity, and assuming that the gravitational redistribution process was triggered by a mantle-derived fluid-heat flow (e.g., England and Thompson, 1984; Perchuk, 1976, 1991; Hoernes et al., 1995; Pili et al., 1997), Perchuk et al. demonstrated that the growth dynamics of multiwavelength gravity structures (polydiapirs) (Weinberg and Shmelling, 1992) is accelerated by a chain-reaction mechanism (Perchuk, 1991). The modeling showed that the formation of some granulite complexes as the result of the continuous evolution of an initial metastable multilayered craton structure triggered by fluid-heat flow is indeed feasible. Similar results can also be obtained with more realistic temperature- and stress-dependent rheologies (Fig. 7) with a gravitational redistribution process being triggered in the lower crust where effective viscosity for all rock types is the lowest because of the high temperature. The effective viscosity of the upper crust (cf. Figs. 7A, 7B) mainly regulates the potential for the penetration of large, strongly internally deformed granulite bodies into the

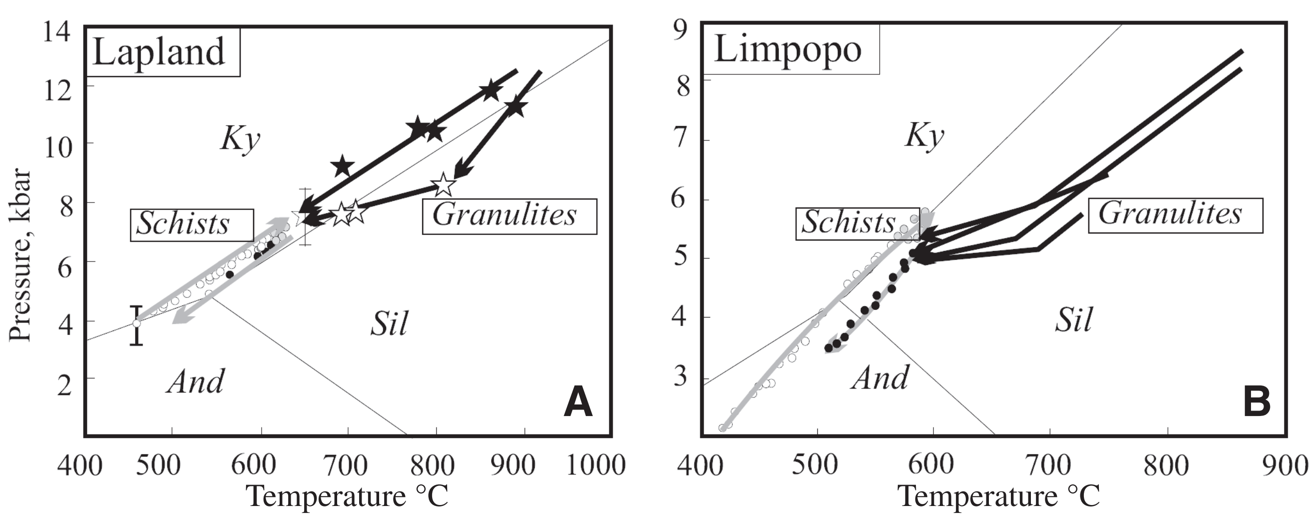

Figure 6. Integrated P-T paths, reflecting simultaneous tectonic histories of the high-grade terranes and schists from the footwall shear zones separating granulite bodies from cratonic wall rocks. (A) Lapland; (B) Limpopo; modified from Perchuk et al., 2000b. Note that the burial and exhumation branches of paths for mica schists are nearly identical. Also note that P-T scales in A and B are different.

shallower levels of the crust. Detailed petrological data for granulite complexes and adjacent greenstone belts (Perchuk et al., 1996, 1999, 2000b; Perchuk and Krotov, 1998, and summaries in this chapter) created a basis for thermomechanical numerical modeling (including simulation of metamorphic P-T paths) for the generation, exhumation, and emplacement of granulite complexes within greenstone belts (Gerya et al., 2000, 2004).

Based on the typical stratigraphy for most well-known greenstone belts (e.g., De Wit et al., 1992; De Wit and Ashwall, 1997; Perchuk et al., 2000a), a rhythmic layered succession was used (Gerya et al., 2000) in the initial design (Fig. 8A, for 0 m.y.). This figure shows a general scenario for the gravitational redistribution of rocks in the Earth's crust (see Gerya et al., 2000, 2004, for the choice of numerical values and sensitivity analysis). Both the

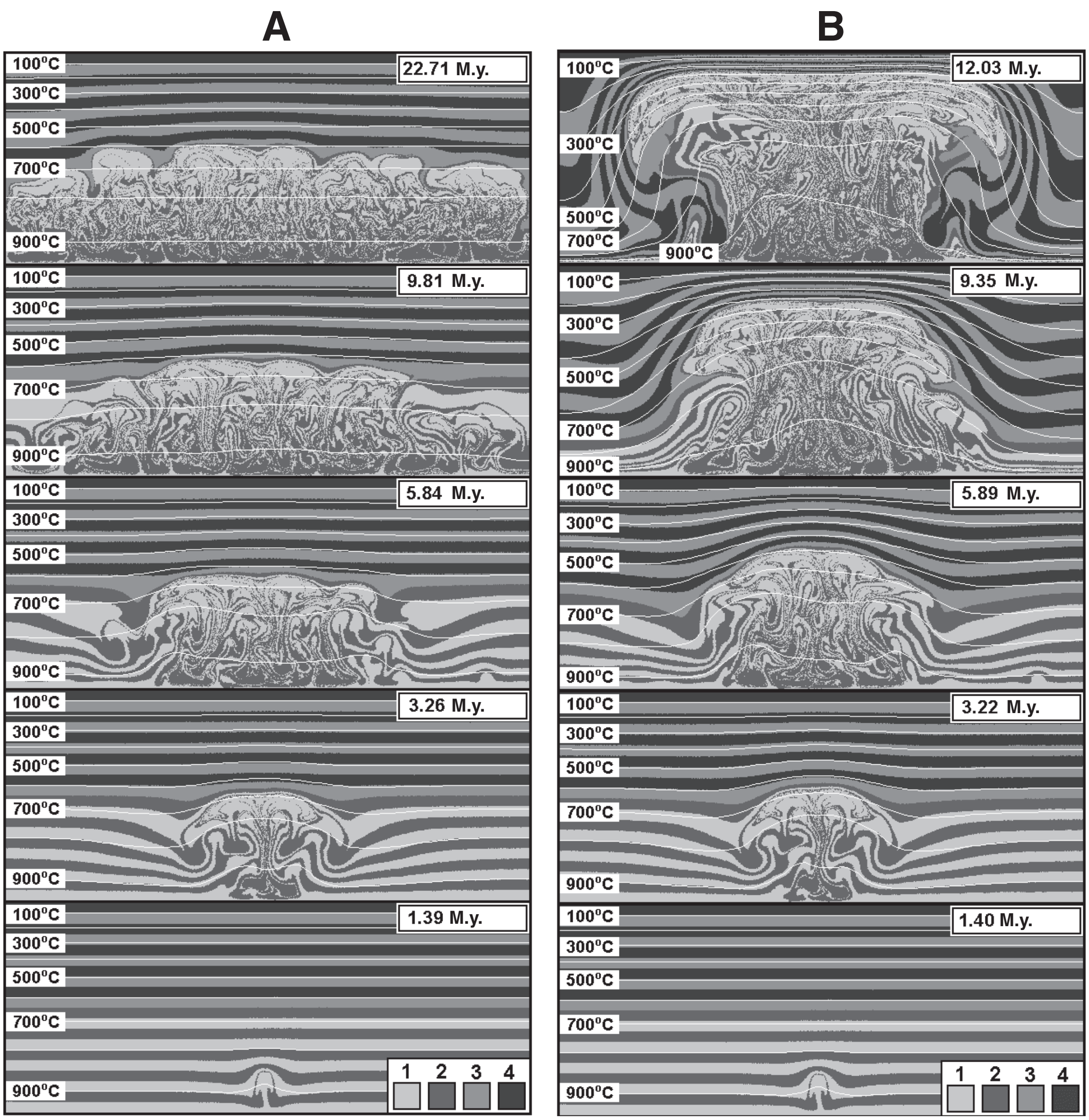

Figure 7. Results of numerical modeling for the gravitational redistribution process in multilayered crust, forming either a highly deformed granulite layer in the lower-middle crust (A) or a crustal-scale granulite diapir penetrating into the upper crust (B). Model size is 100 × 40 km. Color code corresponds to different lithologies of upper (1, 2) and lower (3, 4) crust with the following properties (see Gerya et al., 2000, 2004, for the choice of numerical values and sensitivity analysis): 1 and 3, density = 2820 kg/m^3, thermal conductivity = 2 W/m/K, radiogenic heat production = 9×10^{-10} W/kg, heat capacity = 1100 J/kg, flow law = wet granite (Ranalli, 1995); 2 and 4, density = 2870 kg/m^3, thermal conductivity = 2 W/m/K, radiogenic heat production = 3×10^{-10} W/kg, heat capacity = 1100 J/kg, flow law = diorite (Ranalli, 1995). Maximum model viscosity contrast between the upper and lower crust is 10^4 for A and 10^3 for B. Instability is initiated by small thermal perturbations (placed at the bottom of the model). No lateral forcing is applied in the models.

presence of three major initial layer sequences (rhythms) in the modeled stratigraphic succession and the difference in densities between layers of each rhythm provide acceleration of the process of gravitational redistribution (Perchuk et al., 1992). The interaction between the rhythms and individual layers allows a rapid large-scale flow over the entire sequence (Fig. 8A, 1.7–8.9 m.y.; see also Fig. 7B). The model development shows thrusting of the hot, flattened granulitic diapir onto the colder, upper crustal cratonic rocks. Figure 8B demonstrates oppositely directed vertical movement of mafic and felsic material from different crustal levels: granulites moving up to the surface, and cold metabasites and metakomatiites of the upper part of the craton moving down under hot granulites along the shear zone separating them.

The numerical modeling of gravitational redistribution allows the calculation of P-T-time paths for both the granulites and the cratonic rocks (Figs. 9, 10), which then can be compared with natural data. Numerical experiments (Gerya et al., 2000) showed that the geometry of the modeled P-T paths (Fig. 9) compares closely with petrological P-T paths (Figs. 4, 6). Figure 8B shows oppositely directed movement of mafic and felsic material from different crustal levels: While granulites move up toward the surface, metabasaltic and metakomatiitic cratonic rocks move down and cool the granulites along the shear zone separating the two terranes. This causes a local convective cell that changes the movement of some uprising granulite fragments: The square marker in Figure 8B (1.7–2.5 m.y.) moves from 700 °C to 600 °C toward the downward moving, cooler greenstone plate. In the case of the Limpopo HGT the upward movement of the two Marginal Zones along the contact with the adjacent cratons was accompanied by a narrow zone of amphibolite-facies retrogression of the uprising granulites (see Figs. 4, 6; van Reenen, 1986; van Reenen et al., this volume). This retrogression was caused

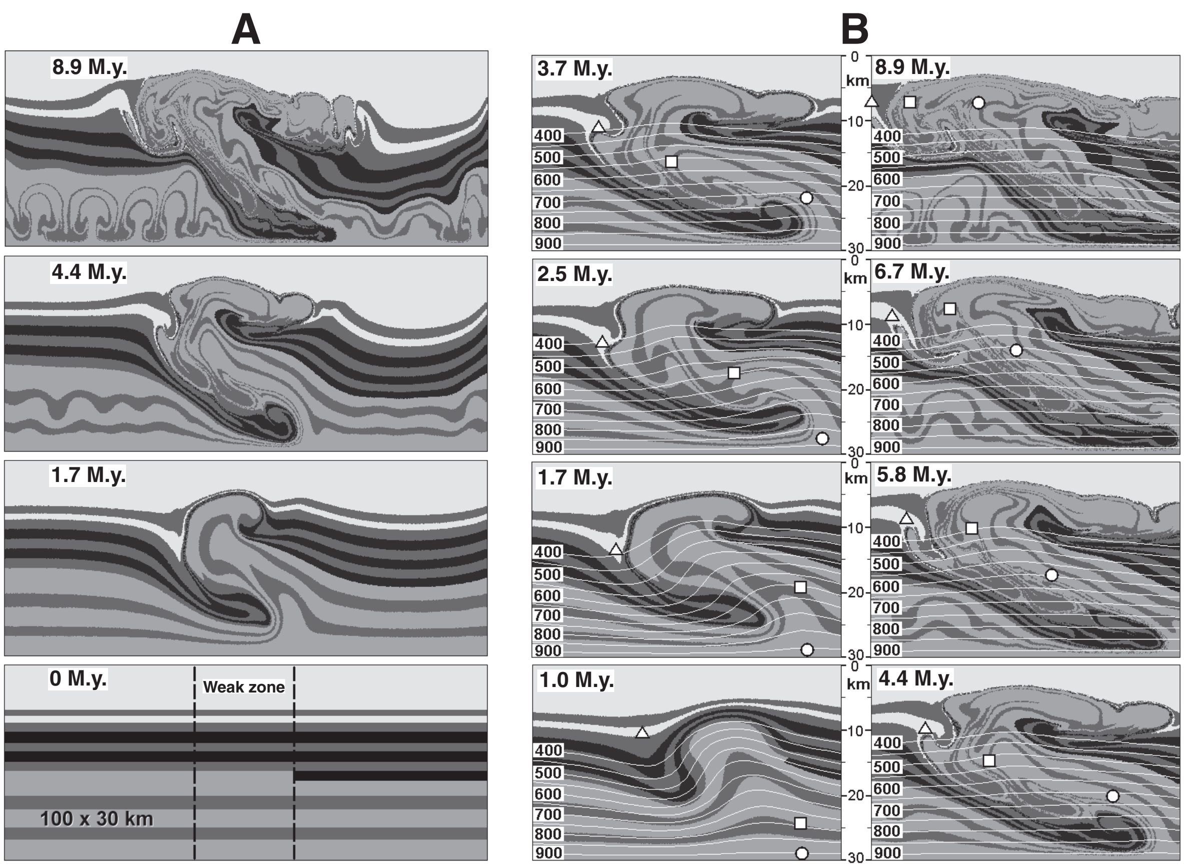

Figure 8. Results of numerical modeling (Gerya et al., 2000) of the buoyant exhumation of a granulite complex: (A) Overall model development; (B) enlargement of the granulite diapir. Model design in A: Size, 100 × 30 km; grid resolution, 100 × 30 nodes, 500 × 150 markers; weak zone mimics preexisting tectonic boundary between two cratons; instability is initiated owing to different numbers of ultramafic layers in the cratons; no lateral forcing is applied in the model. Rock types and properties (see Gerya et al., 2000, for the choice of numerical values and sensitivity analysis): sediments (white, ρ = 2700 kg/m^3, $\eta = 10^{19}$ Pa s, where ρ is density and η is viscosity), felsic granulites (light gray, ρ = 2800 kg/m^3, $\eta = 10^{19}$ Pa s), metabasites (dark gray, ρ = 3000 kg/m^3, $\eta = 10^{19}$ Pa s at T >600 °C in granulite sequence, $\eta = 10^{21}$ Pa s at T <600 °C in greenstone sequence), metakomatiites (black, ρ = 3300 kg/m^3, $\eta = 10^{21}$ Pa s), weak tectonic zone (dashed, $\eta = 10^{19}$ Pa s for all rock types). Heat conductivity of rocks, 4 W/m/K; isobaric heat capacity of rocks, 1100 J/kg/K. Symbols (triangle, circle, and square) in B show movement of representative rock units with P-T-time paths displayed in Figure 9.

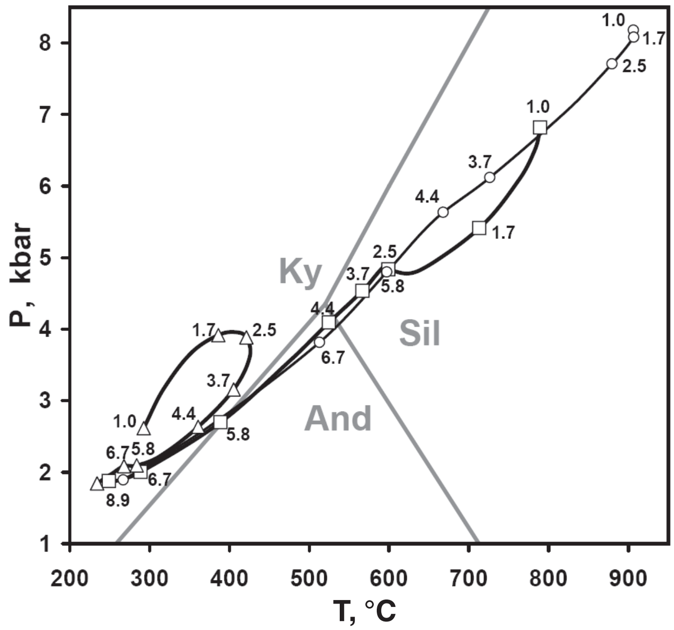

Figure 9. Results of numerical modeling of P-T-time paths for representative cratonic rock units (triangle) and two markers in the granulite fragments (circle and square) shown in Figure 8 (compare with Figs. 4–6). See Gerya et al. (2000) and Gerya and Maresch (2004) for comparison of natural and numerical P-T paths.

by H_2O-rich fluids that resulted from dehydration reactions in the downward movement of the greenstones (see movement of the triangle marker in Fig. 8B) in the greenstone lithologies plate. These movements result in a near-isobaric portion of the P-T path for the upper granulite fragment (see steplike lines in Fig. 4B and kinked P-T path with squares in Fig. 9). On the other hand, a high-temperature, deep-seated fragment moves from the 900 °C isotherm directly to a near-surface level (see circle marker in Fig. 8B), producing a decompression-cooling P-T path (see straight line in Fig. 4A and linear P-T path with circles in Fig. 9).

An important detail is illustrated in Figure 8A (8.9 m.y.): The low-grade greenstone block of the craton on the right side of the model is bent upward by the flow of exhuming granulite, and its leftmost part is exhumed close to the middle of the flattened granulite complex. If this feature is realistic, one should be able to observe comparable features among granulites (i.e., higher grade greenstone rocks that were metamorphosed under amphibolite to epidote-amphibolite facies). A possible example is greenstone rocks that occur in the Central Zone of the Limpopo HGT, part of the so-called Venetia Klippen Complex, for which a well-constrained P-T loop with maximum P-T estimates of 6 kbar and 650 °C was inferred (Zeh et al., 2004).

According to the P-T paths in Figures 5 and 6, the model in Figure 8 must reflect the circulation of some of the cratonic wall rocks within shear zones separating cratons from granulite complexes. Indeed, triangle markers in Figure 8B trace the movement of the rocks, repeating a clockwise P-T loop for the cratonic rocks shown in Figures 5 and 6. Overall the variability of P-T paths in the numerical model (Fig. 10) is quite large and coincides well with variations in P-T trajectories discovered within and outside of each studied granulite complex and associated greenstone belt (e.g., Perchuk et al., 1989, 1996; Perchuk and Krotov, 1998; Gerya et al., 2004; see also Perchuk, this volume).

Seismic profiling, electric studies, and gravity data show that the continental crust beneath some granulite terranes is thinner than that under the adjacent cratons (e.g., Fig. 11A). The correct position of the Moho boundary beneath such crustal-scale complexes, however, is unknown (e.g., Griffin and O'Reilly, 1987a, 1987b) presumably because their present lower parts are composed of garnet-bearing mafic and ultramafic rocks (Pretorius and Barton, 1995; Specius, 1998) whose rheologies and densities are similar to those of the upper mantle. Figure 11A illustrates a N-S cross section through the Limpopo HGT compared with the results of numerical modeling of gravitational redistribution of the Archean greenstone belts. Figure 11B shows a NE-SW cross section through the Lapland HGT compared with the results of numerical modeling (Perchuk et al., 1999) of gravitational redistribution of rocks within the Archean Karelian-Inari Cratons. We contend that the morphologies of these two examples, for which good-quality seismic data are available (De Beer and Stettler, 1992; Pozhilenko et al., 1997; James et al., 2001; Nguuri et al., 2001), are in accord with the gravitational redistribution mechanism of HGT formation.

Thus the results of numerical modeling suggest that granulite-facies terranes can be formed (and then exhumed) within essentially normal-thickness (35–40 km), hot continental crust affected by the activity of mantle-derived fluid-heat flow. Such a process results in the formation of granulitic bodies with intrusion-like shapes (Figs. 8, 11) through a mechanism of gravitational redistribution of material in the crust during a time period of ~10 m.y. The results of our modeling are also in accordance with the conclusion of England and Thompson (1984) and Henry et al. (1997) that *thick* (60–70 km) continental crust in many cases is not necessarily the most appropriate setting for the formation of granulite-facies terranes. Thickening of such crust initially results in a relatively cold geotherm, which is more suitable for the origin of eclogite and HP granulite along a prograde PT path, followed by crustal thinning (e.g., owing to strong mantle plume activity) that in turn leads to the formation of garnet amphibolite (Perchuk, 1977, 1989). In some cases (e.g., Tibet) such eclogites can subsequently undergo granulite-facies conditions (e.g., Henry et al., 1997).

DISCUSSION

Granulite-facies terranes occur in a variety of geological settings of different age. However, the majority of them were formed in the Precambrian, when the Earth's crust was hotter than today. Thompson (1990) reviewed several tectonic models (collision followed by erosional or extensional crustal thinning, multiple episodes of crustal thickening, magmatic underplating, and crustal delamination) and concluded that there are limited possibilities, if they exist at all, to form and exhume lower crustal granulites into the upper Earth's crust during a single tectonic cycle, and most granulite-facies terranes were exposed tectonically. Harley (1989), on the basis of a study of 90 P-T paths from different HGTs, concluded that no single universal tectonic model could be proposed for the genesis of granulites. Nevertheless, despite the fact that eclogites as markers of subduction-collision mechanisms occur extremely rarely in Precambrian continental crust in general, and for the Archean, in particular, many petrologists still believe the plate tectonic models. This may be the result of systematic comparison of the Precambrian granulites with significantly younger granulites from Phanerozoic fold belts, but the rheology of these rocks was very different.

Recently Sizova et al. (2010), based on a numerical modeling study, demonstrated that there should be major transitions in geodynamic styles back in Earth history related to the cooling of sub-lithospheric mantle (see also, e.g., Hamilton, 1998; Brown, 2006). These authors identified a first-order transition from a "no-subduction" tectonic regime through a "pre-subduction" tectonic regime to the modern style of subduction. The first transition is gradual, and occurs at upper mantle temperatures between 250 and 200 K above present-day values, whereas the second transition is more abrupt and occurs at 175–160 K above present-day values. The link between geological observations and model

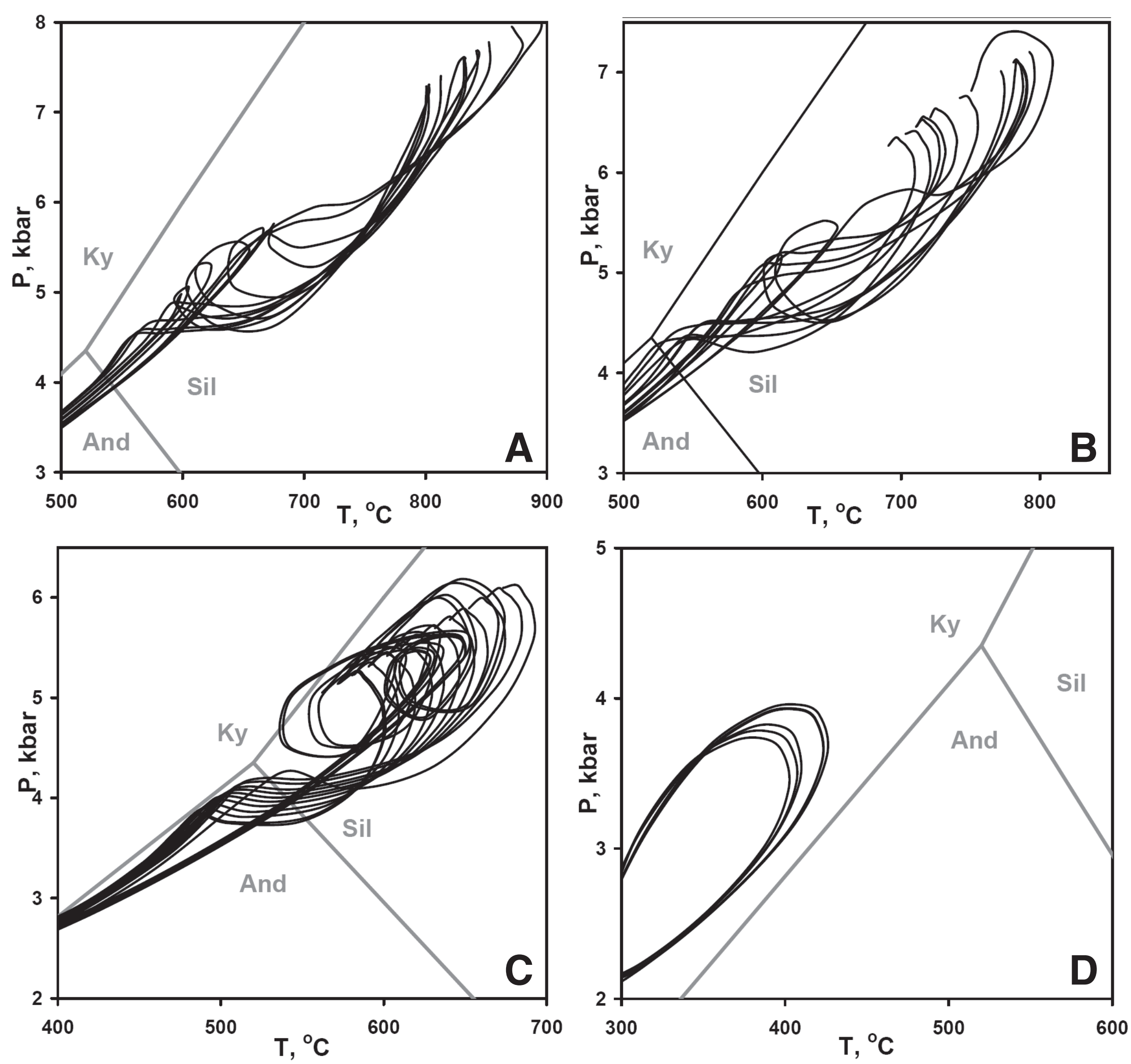

Figure 10. Characteristic shapes of P-T paths for (A) high- and (B) medium-temperature granulites in the hanging wall of the shear zone for the model shown in Figure 8. Characteristic P-T paths for (C) mid-crustal and (D) upper crustal cratonic footwall rocks for the model shown in Figure 8 (compare with Figs. 4–6). See Gerya et al. (2000) and Gerya and Maresch (2004) for comparison of natural and numerical P-T paths.

results suggests that the transition to the modern plate tectonics regime might have occurred during Mesoarchean–Neoarchean time (ca. 3.2–2.5 Ga). For the "pre-subduction" tectonic regime (upper-mantle temperature 175–250 K above present) the plates are weakened by intense percolation of melts derived from the underlying hot melt–bearing sub-lithospheric mantle. In such cases, convergence does not produce self-sustaining, one-sided subduction, but rather results in shallow underthrusting of the oceanic plate under the continental plate. Further increase in the upper-mantle temperature (>250 K above present) causes a transition to a "no subduction" regime in which horizontal movements of small deformable plate fragments are accommodated by internal strain, and even shallow underthrusts do not form under convergence. Taking this into account, one can expect that the geodynamic origin of many Precambrian granulites may have differed markedly from their modern analogues.

Inherent gravitational instability of hot continental crust was confirmed on the basis of modeling of in situ rock properties using a Gibbs free-energy minimization approach (Gerya et al., 2001, 2002, 2004). This modeling showed that regional metamorphism of granulite-facies crust may critically enhance the decrease of crustal density with depth. This leads to a gravitational instability of hot continental crust, resulting in regional doming and diapirism. Two types of crustal models were studied: (1) lithologically homogeneous crust, and (2) lithologically heterogeneous, multilayered crust. Gravitational instability of relatively homogeneous continental crust is related to a vertical density contrast developed during prograde changes in mineral

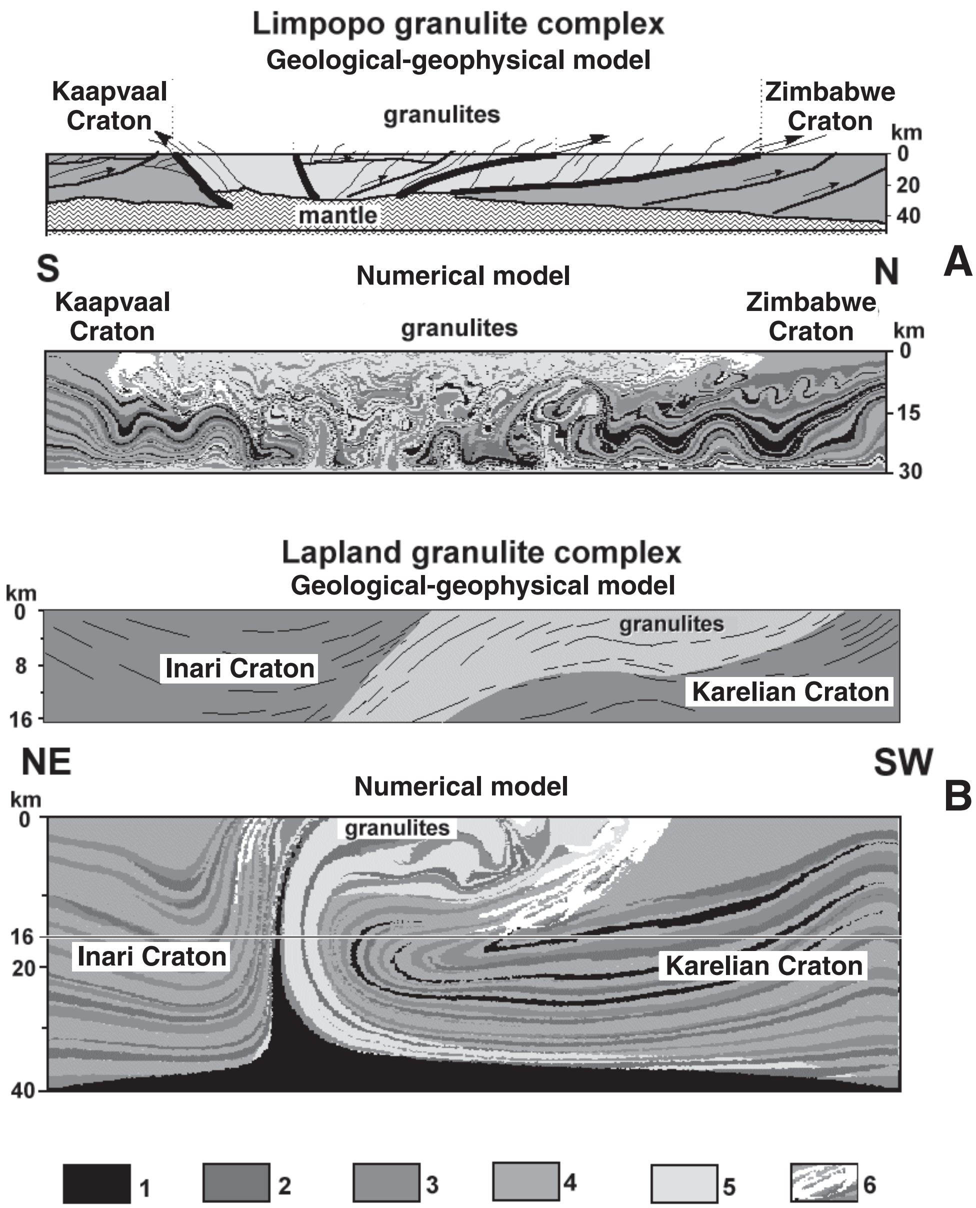

Figure 11. Geological-geophysical cross sections of (A) the Limpopo HGT (Roering et al., 1992a), and (B) the Lapland granulite complex (Pozhilenko et al., 1997) compared with results of numerical experiments (Perchuk et al., 1999) conducted with the gravitational redistribution models. In the presented models, granulite-facies rocks were originally in the lower portion of a metastable layered 30–40-km-thick crust. No lateral forcing is applied in the models. Color code for numerical models and rock properties (see Perchuk et al., 1999; Gerya et al., 2000, for the choice of numerical values and sensitivity analysis): 1 = ultramafic rocks ($\rho = 3300$ kg/m^3, $\eta = 10^{21}$ Pa s); 2 = metabasic rocks ($\rho = 3000$ kg/m^3, $\eta = 10^{21}$ Pa s); 3 = andesitic rocks ($\rho = 2800$ kg/m^3, $\eta = 10^{21}$ Pa s); 4 = felsic and sedimentary rocks ($\rho = 2700$ kg/m^3, $\eta = 10^{19}$ Pa s); 5 = felsic, partially molten granulites ($\rho = 2650$ kg/m^3, $\eta = 10^{19}$ Pa s); 6 = weak tectonic zones in rocks of various lithologies ($\eta = 10^{19}$ Pa s).

assemblages and the thermal expansion of minerals with increasing temperature. Gravitational instability of lithologically heterogeneous multilayered crust is related to an initial density contrast of dissimilar intercalated layers enhanced by high-temperature phase transformations. In addition, the thermal regime of heterogeneous crust strongly depends on the pattern of vertical interlayering: A strong positive correlation between temperature and the estimated degree of lithological gravitational instability is indicated. It has also been shown (Gerya et al., 2004) that exponential lowering of viscosity with increasing temperature, in conjunction with prograde changes in metamorphic mineral assemblages during thermal relaxation after collisional thickening of the crust, provides a positive feedback mechanism leading to regional doming and diapirism that contribute to the exhumation of high-grade metamorphic rocks.

It is important to mention that collisional (i.e., driven by external forces) and gravitational (i.e., driven by internal forces) mechanisms of rock deformation are not mutually exclusive. Gravitationally unstable crust is expected to result, above all, from collisional events involving initially stable sections of continental crust where regional thrusting, multiple stacking, and regional folding occur (e.g., the double-stacked crust of England and Thompson, 1984; Le Pichon et al., 1997). This suggests a strong causal and temporal link between external collisional and internal gravitational mechanisms of rock deformation in high-grade metamorphic regions. Collisional mechanisms should operate during the early prograde stages of a tectono-metamorphic cycle, causing thickening of the crust and a corresponding increase in radiogenic thermal supply, whereas gravitational mechanisms should dominate during the later thermal peak and retrograde stages, providing an important factor for regional doming and diapirism that contributes to the exhumation of high-grade rocks (Figs. 7, 8). This also extends validity of gravitational redistribution models for relatively hot young orogens where gravitational tectonics is activated (e.g., Ramberg, 1981) owing to thickening and heating of initially cold continental crust (Burg and Gerya, 2005; Gerya et al., 2004, 2008; Faccenda et al., 2008).

For this discussion it should be recalled that Ramberg (1981) extended the diapiric model of MacGregor (1951) originally proposed for granite-greenstone belts to explain the structural pattern of many orogenic belts. More recently this model has been invoked to explain the formation of granite domes within greenstone belts (Hunter and Stowe, 1997). In essence, an increase in temperature leads to a decrease in both density and viscosity of the rocks in the lower-middle crust, thus triggering their movement within the gravity field. This softening of crustal rocks may generate their upward movement if the density and viscosity contrasts are appropriate (e.g., Fig. 7). This simple idea is also the background of the gravitational redistribution model suggested for granulites (Perchuk, 1989, 1991; Perchuk et al., 1992). Crustal (radioactive decay) and mantle heat flow (England and Thompson, 1984; Sandiford and Powell, 1986; Harley, 1989; Thompson, 1990; Pili et al., 1997; Gerya et al., 2002, 2004) are both considered to be the driving forces for the crustal heating process. The gravitational redistribution model suggests a strong mantle-derived fluid-heat flow as the trigger of high-grade metamorphism and both horizontal and vertical movements of rocks within convecting continental crust. We can easily determine the high temperature of the metamorphic process from mineral assemblages, but the contribution of the fluid component of the flow can only be proven on the basis of petrological and isotopic data. For example, Pili et al. (1997) demonstrated the existence of CO_2 flow by stable isotopic data obtained for the rocks from regional-scale shear zones in the Madagascar granulites of similar age.

Because most of the Precambrian HGTs lie between granite-greenstone belts, we have to consider their genetic connection in terms of evolution of the upper mantle as the source of heat and fluids. It is clear that the heat cannot be introduced in the giant metamorphic system by conductivity only. Therefore, crustal radiogenic decay (e.g., Gerya et al., 2001, 2002; Sizova et al., 2010) and mantle-derived fluid flows are necessary physical mechanisms to supply a deep-seated metamorphic system with heat and volatiles, thereby supporting high-temperature conditions. In relation to this, partially molten sub-lithospheric mantle plumes are expected to have played crucial roles in producing magmatism and metamorphism in the Paleoarchean, and even Hadean, crust (3.8–4.4 Ga), which follows from discovery of the 4.4 Ga detrital zircon derived from granite in the Yilgarn Craton of Western Australia (Wilde et al., 2001). These plumes produced large amounts of mafic and ultramafic magmas that could have initiated growth of the continental crust on the one hand (e.g., Bogatikov and Sharkov, 2008), and that partially melted the crust, producing tonalite-trondhjemite-granodiorite cupolas on the other (MacGregor, 1951; Ramberg, 1981; Hunter and Stowe, 1997). Large amounts of sub-lithospheric melts produced by these plumes are expected also to have notably weakened the Precambrian lithosphere, causing distinct tectonic styles (such as pre-subduction and no-subduction regimes) with pronounced horizontal movements (Sizova et al., 2010). Plume tectonics should induce horizontal flows in the rheologically weakened overlying lithosphere and create areas of assembled, deforming cratonic crust between mantle upwellings (Fig. 12). Such crustal assembly areas (where HGTs were subsequently generated) are precursors of modern collision zones, but in contrast to them they are not characterized by greatly thickened crust and high topography, and they are associated with gentle two-sided mantle downwellings rather than with asymmetric subduction and collision of rigid lithospheric plates (e.g., Sizova et al., 2010). Figure 12 illustrates this idea, showing both processes beneath two cratons and the HGT located between them. Ascent of the large plumes is accompanied by their partial and then total melting, which provides growth of the crust by mafic and ultramafic magma additions. At earlier stages the plumes provide enough heat for melting of ancient crust, resulting in felsic magma production. All these provide a magmatic basis for the formation of granite-greenstone terranes. At the later stages these terranes could be assembled in inter-plume areas

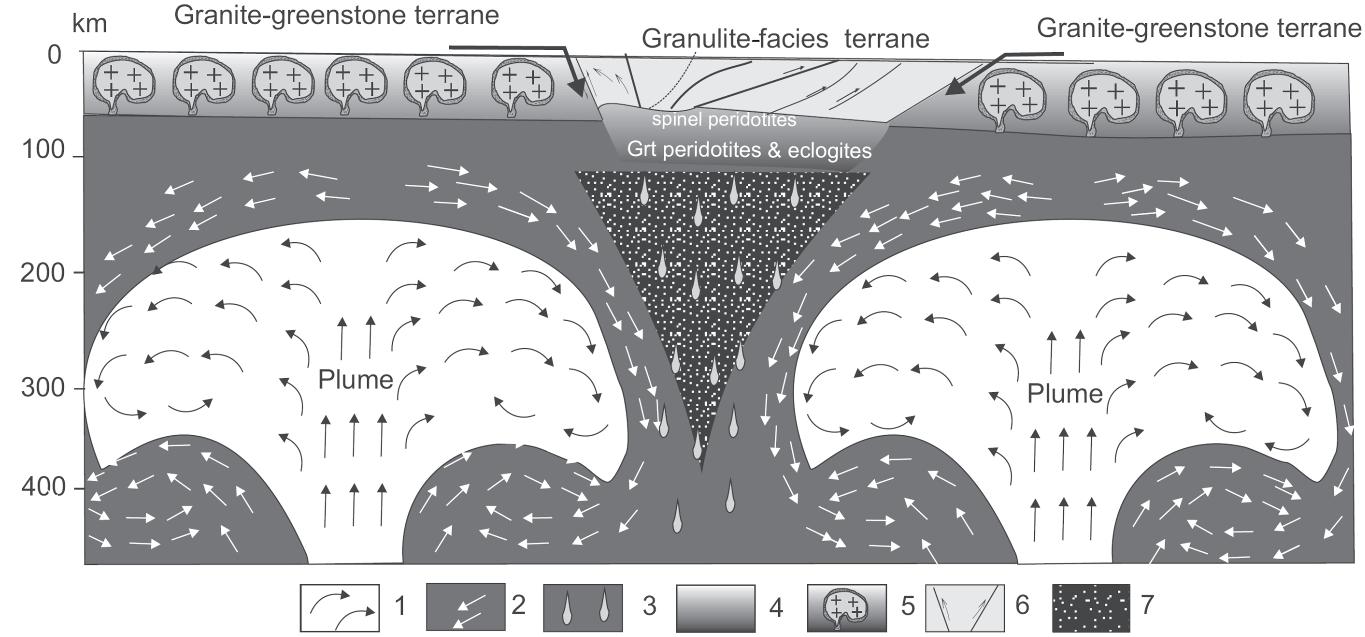

Figure 12. Conceptual scheme for the formation of the Precambrian continental crust resulting from plume tectonics: (1) arrows reflecting movement directions of plume material; (2) upper mantle rocks and direction of their movements in response to the plume activity; (3) crustal eclogites descending into the deep mantle; (4) continental crust composed mainly of volcanic rocks in layered successions, as discussed in the text (see also Perchuk, 1989); (5) ascending TTG (tonalite-trondhjemite-granodiorite) cupola within the continental crust, resulting from the hot plume activities; (6) HGT with shear zones (straight gneisses, Smit and van Reenen, 1997) recording its exhumation; (7) upward-directed heat-fluid flow that triggers HG metamorphism and exhumation of granulites within the continental crust from mechanisms of gravitational redistribution (fluids are presumably released owing to eclogitization reactions and crystallization of residual mantle melts in the downwelling zone). Dimension and shape of the granulite complex approximately correspond to those of the Limpopo HGT (Fig. 11A).

and involved in the gravitational redistribution (Fig. 12) triggered by remnant fluid-heat flows separated from descending and crystallizing mantle plume material (Fig. 12). The conceptual scenario of plume tectonics delineated in Figure 12 obviously needs further assessment and testing on the basis of both field observation and numerical modeling, including detailed studies of mutual structural and metamorphic aspects of granulite terranes and adjacent greenstone belts (e.g., Roering et al., 1992b; Smit et al., 2000; Bennett et al., 2005; van Reenen et al., this volume).

CONCLUSIONS

Precambrian HGTs can be formed and exhumed during a single cycle within a hot continental crust that can be relatively thin (35–40 km thick). The driving forces of the process are (1) a potential gravitational instability of an initial metastable stratigraphic/tectonic succession, and (2) gravitational redistribution of the crustal rocks owing to a mantle-derived fluid-heat flow (a mantle plume). This conclusion is based on the following points:

1. Many HGTs have a laccolith or harpolith shape in vertical cross section. Hence, such complexes could represent large diapirs intruded into the upper crust.
2. The HGTs exhibit DC and DC-IC retrograde paths compatible with a single episode exhumation. DC paths are >~50 km from the footwall of the HGT, whereas IC-DC paths are <~50 km from the footwall.
3. The greenstone belts in the footwall adjacent to the granulite complexes show *strongly non-isobaric* metamorphic zoning that is time- and mechanism-related to the emplacement of the granulite diapir. The peak metamorphic conditions of the prograde path for the footwall cratonic rocks coincide with the P-T minimum recorded by the granulite assemblages. This convergence of P-T conditions implies that the P-T loops for the footwall rocks resulted from the emplacement of the granulite diapir into the upper crust.
4. Collisional and gravitational mechanisms of rock deformation are not mutually exclusive: Collisional mechanisms may operate during the early prograde stages of a tectono-metamorphic cycle, whereas gravitational mechanisms should dominate during the later thermal peak and retrograde stages.
5. The gravitational redistribution model is well supported by multidisciplinary tests in many HGTs, including those in the Limpopo and Lapland HGTs.

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