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Thermal history of the Earth and its petrological expression

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ABSTRACT

Non-arc basalts of Archean and Proterozoic age have model primary magmas that exhibit mantle potential temperatures T_P that increase from 1350 °C at the present to a maximum of ~1500–1600 °C at 2.5–3.0 Ga. The overall trend of these temperatures converges smoothly to that of the present-day MORB source, supporting the interpretation that the non-arc basalts formed by the melting of hot ambient mantle, not mantle plumes, and that they can constrain the thermal history of the Earth. These petrological results are very similar to those predicted by thermal models characterized by a low Urey ratio and more sluggish mantle convection in the past. We infer that the mantle was warming in deep Archean–Hadean time because internal heating exceeded surface heat loss, and it has been cooling from 2.5 to 3.0 Ga to the present. Non-arc Precambrian basalts are likely to be similar to those that formed oceanic crust and erupted on continents. It is estimated that ~25–35 km of oceanic crust formed in the ancient Earth by about 30% melting of hot ambient mantle. In contrast, komatiite parental magmas reveal T_P that are higher than those of non-arc basalts, consistent with the hot plume model. However, the associated excess magmatism was minor and oceanic plateaus, if they existed, would have had subtle bathymetric variations, unlike those of Phanerozoic oceanic plateaus. Primary magmas of Precambrian ambient mantle had 18–24% MgO, and they left behind residues of harzburgite that are now found as xenoliths of cratonic mantle. We infer that primary basaltic partial melts having 10–13% MgO are a feature of Phanerozoic magmatism, not of the early Earth, which may be why modern-day analogs of oceanic crust have not been reported in Archean greenstone belts.

1. Introduction

Earth's present-day magmatism and tectonic activity are surface expressions of heat loss from mantle convection and heat gain from radioactive decay. This view has stimulated many attempts to understand how much hotter the early Earth was compared to the present time (Richter, 1985; Christensen, 1985; Korenaga, 2008a,b; Davies, 1980; 2009). While conclusions vary regarding the thermal state of the Earth billions of years ago, there is general agreement that melting of the mantle was more extensive, and this yielded thicker oceanic crust and hotter magmas (Sleep and Windley, 1982; Bickle, 1986; Abbott et al., 1994; Foley et al., 2003; Korenaga, 2006).

Archean komatiites have the highest MgO contents and eruption temperatures of any volcanic rock in the terrestrial record, evidence for a hotter early Earth (Bickle, 1986). But this interpretation is ambiguous because it has also been suggested that komatiites formed in hot mantle plumes (Jarvis and Campbell, 1983; Bickle, 1986; Campbell et al., 1989; Herzberg, 1992; Nisbet et al., 1993; Fan and Kerrich, 1997; Arndt et al., 1998; Herzberg et al., 2007; Arndt et al., 2008). Spatially associated non-arc basalts in Archean greenstone belts have trace element abundances that are strikingly similar to those of Phanerozoic oceanic plateau basalts (Puchtel et al., 1998; Kerrich et al., 1999; Hollings et al., 1999; Kerr et al., 2000; Polat and Kerrich, 2000; Arndt et al., 2001; Condie, 2003, 2005; Smithies et al., 2005; Kerrich and Polat, 2006), further evidence that has been used to support the mantle plume model. The problem is how can we distinguish lavas that formed from hot ambient mantle from those that formed in hot mantle plumes? This is an important problem to sort out if we wish to use the petrology of igneous rocks to constrain the thermal history of the Earth.

In this paper, we calculate the temperature and pressure conditions that were necessary to form non-arc basaltic rocks with well-defined ages in the Proterozoic and Archean. The temporal variation of these melting conditions is compared with secular cooling curves predicted by parameterized convection models (Korenaga, 2008a,b; Davies, 2009). Calculations are also performed on komatiites with Archean ages, and the results permit a fresh look at temperature differences between ambient mantle and mantle plumes in the early Earth. In addition to constraining the thermal history of the Earth, this information is necessary for shedding new light on old ideas concerning the composition and structure of the crust and lithosphere, and how they have changed with time.

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2. Methods

2.1. Geological samples

We have calculated mantle potential temperatures for mafic lava compositions with well-defined ages using a petrological method described below. We focus on lava samples that might have formed by melting ambient mantle, not anomalously hot or cold mantle. This condition excludes komatiites, although we work with them also as they are thought to be representative of ancient plumes. We exclude basalts that exhibit the geochemical characteristics of modern arcs, such as depletions in high field strength elements (Polat and Kerrich, 2000; Condie, 2005; Kerrich and Polat, 2006). That leaves us with the non-arc Precambrian basalts, those with MORB-like LREE depletions and a large group of basalts generally referred to as "oceanic plateau basalts", so called because of geochemical similarities with young oceanic plateau basalts (Kerrich et al., 1999; Polat and Kerrich, 2000; Condie, 2005; Kerrich and Polat, 2006). Our analysis uses an updated database of Condie (2003; 2005), lavas from the Cape Smith belt in Canada (Hynes and Francis, 1982; Picard, 1989), data used by Abbott et al., (1994; one sample of Phanerozoic age), and the University of Saskatchewan database of Kerrich and colleagues for basalts from the Archean Superior Province in Canada (Kerrich and Polat, 2006 and references therein). Table A1 contains original source references, original tectonic interpretations, and sample identification numbers so that our solutions can be replicated and compared with other petrological models. It provides primary magma solutions for lava compositions given by Green et al. (2000), Smithies et al. (2005), Kato and Nakamura (2003), Polat et al., (1999, 2008), Shchipansky et al. (2004), Naqv et al. (2006), Knoper and Condie (1988), Hanski et al. (2001), Volpe and Macdougall (1990), Wang et al. (2007), Hynes and Francis (1982), Picard (1989), Ahmad and Rajamani (1991), Bernard-Griffiths et al. (1986), Hollings and Kerrich (1999), Schulz (1982), Fan and Kerrich (1997).

We work with komatiites from the Alexo and Pyke Hill areas of the Abitibi greenstone belt (Arndt, 1986; Lahaye and Arndt, 1996; Puchtel et al., 2004; Fan and Kerrich, 1997; Sproule et al., 2002; Shore, 1996; Sobolev et al., 2007), and the Belingwe greenstone belt (Bickle et al., 1993; Puchtel et al., 2009; Sobolev et al., 2007). These occurrences were chosen because they are representative of komatiites with 2.7 Ga ages. The Alexo and Pyke Hill samples are similar to komatiites from the Belingwe greenstone belt (Puchtel et al. 2009) for which melt inclusion studies point to a hot and dry origin (Berry et al., 2008). We also work with 3.5 Ga komatiites from the Barberton greenstone belt (Smith and Erlank, 1982) which have been interpreted to have formed in mantle plumes (Herzberg, 1992; Arndt et al., 1998; Arndt et al., 2008) and subduction zones (Parman et al., 1997).

Archean komatiites are often associated with or interlayered with non-arc basalts (e.g., Arndt and Nesbitt, 1982; Redman and Keays 1985; Kerrich and Polat, 2006), but they occur as distinct lithostratigraphic units that do not grade into each other, and they are not related by fractional crystallization or melting (Arndt et al., 2008). Some thick basaltic komatiite flows contain gabbroic zones derived by fractional crystallization after ponding (e.g., Arndt, 1977), but most komatiites are dominated by olivine addition and subtraction, yielding whole rock compositions with 20–50% MgO. In contrast, most non-arc basalts are dominated by clinopyroxene fractionation (e.g., Lesher and Arndt, 1995) and, where aphyric, contain $<10\%$ MgO. In the Norseman-Wiluna greenstone belt in Western Australia the komatiites and associated non-arc basalts are both depleted in LREE (Redman & Keays, 1985; Lesher & Arndt, 1995), but are isotopically distinct (Lesher and Arndt, 1995). In the Abitibi greenstone belt, non-arc basalts exhibit flat rare earth element patterns similar to more modern oceanic plateau basalts (Kerrich et al., 1999), which contrasts with the komatiites which show strong depletions in the light rare earth elements (e.g., Lahaye and Arndt, 1996; Puchtel et al., 2009). We contribute to these distinctions by showing that the komatiites melted in a hotter mantle environment than the non-arc basalts.

2.2. Petrological method

The compositions of basaltic rocks of Phanerozoic age have been used to evaluate the MgO and FeO contents of their primary magmas, which are partial melt products of the mantle. Primary magma compositions are important to constrain because their MgO and FeO contents increase with mantle potential temperature T_P (Langmuir et al., 1992; Putirka, 2005; Herzberg et al., 2007; Herzberg and Asimow, 2008). They provide a petrological record of the thermal state of the mantle from which they formed (Herzberg et al., 2007). In this paper, we apply the same petrological protocols as used in previous studies with PRIMELT2 software (Herzberg and Asimow, 2008). The petrological theory and computational method have been described in detail elsewhere (Herzberg and O'Hara, 2002; Herzberg et al., 2007; Herzberg and Asimow, 2008).

PRIMELT2 was calibrated from both experiments and forward modeling on fertile peridotite KR-4003 (Walter, 1998; Herzberg and O'Hara, 2002; Herzberg and Asimow, 2008). All primary magma solutions are strictly valid for such a model source composition. It contains 8% FeO, the average for natural fertile and depleted peridotite occurrences (Herzberg, 1993; McDonough and Sun, 1995; Lyubetskaya and Korenaga, 2007). Uncertainties in the FeO content of peridotite can propagate to an uncertainty of ± 50–70 °C in T_P (Herzberg and Gazel, 2009). The inference that the Fe content of the Archean mantle was higher than it is today, drawn from the Fe content of picrites and komatiites (Francis et al., 1999), is not petrologically unique (Langmuir and Hanson, 1980; Herzberg and O'Hara, 2002; Herzberg, 2004a; Polat et al., 2006). Uncertainties in all other major elements for fertile peridotite do not propagate to significant variations in melt fraction and mantle potential temperature (Herzberg and O'Hara, 2002; Herzberg and Asimow, 2008). Melting of depleted peridotite propagates to calculated melt fractions that are too high, but with a negligible error in mantle potential temperature (Herzberg and O'Hara, 2002; Herzberg et al., 2007; Herzberg and Asimow, 2008). Fe_2O_3 has been calculated with $FeO/FeO_T = 0.9$. Melting is assumed to have been fractional, not batch, and all calculated primary magma compositions are aggregates of all melt increments that mix perfectly (i.e., accumulated fractional melting; Herzberg and O'Hara, 2002).

For each non-arc basalt primary magma solution, PRIMELT2 provides the olivine liquidus temperature T_{OL} at 1 atm and the mantle potential temperature T_P (Table A1, Appendix A). Both T_{OL} and T_P are dependent on the MgO content of the primary magma, and accuracy of T_{OL} is ± 31 °C at the 2σ level of confidence (Herzberg et al., 2007). Uncertainties in the partitioning of FeO and MgO between olivine and liquid propagate to an error of $\pm 2\%$ MgO in primary magma composition at the 2σ level of confidence (Herzberg and O'Hara, 2002). This translates to a nominal uncertainty in T_P of ± 44 °C (2σ). Uncertainties that arise by calculating FeO using $Fe_2O_3/TiO_2 = 0.5$ (Herzberg and Asimow, 2008) instead of $FeO/FeO_T = 0.9$ will increase the model MgO content by 2% MgO, and contribute to the overall uncertainty by $\pm 2\%$ MgO. The total uncertainty in MgO primary magma composition propagates to ± 60 °C (2σ) in T_P. PRIMELT2 uses solid state adiabatic gradients from Iwamori et al. (1995), which are similar to those of Stixrude and Lithgow–Bertelloni (2005). Adiabatic gradients used for melting are also similar to those of Iwamori et al. (1995) within the ± 60 °C uncertainty of calculated primary magma composition. Cerium contents of model primary magmas are less than 15 ppm, indicating less than 0.3% H_2O in the source, assuming $H_2O/Ce \sim 200$ (Dixon et al., 2002). This small amount of water is not sufficient to lower crystallization temperatures and compromise estimates of mantle potential temperature (Herzberg et al., 2007).

Different petrological models yield different parameterizations that link MgO and T_P, depending on assumptions concerning the thermodynamic properties of decompression melting (McKenzie, 1988; Iwamori et al., 1995; Langmuir et al., 1992). Agreement is excellent that primary magmas with 10–13% MgO required T_P in the 1300–1400 °C range. However, estimates of T_P required to produce a primary magma having >20% MgO can be discrepant by ~100 °C, as summarized in Fig. 5 of Herzberg et al. (2007). Clearly, more work needs to be done in order to obtain an absolute calibration of how T_P varies with MgO at extreme conditions of melting appropriate to both a hot early Earth and modern mantle plumes. However, it is important to note that most of the early T_P–MgO calibrations were parameterized from experimental data at low temperatures and pressures appropriate to low MgO MORB compositions. This is in contrast to the PRIMELT2 T_P–MgO calibration, which relies heavily on high T and P experimental data appropriate to the production of high MgO magmas considered in this work (Walter, 1998; Herzberg and O'Hara, 2002). Our T_P estimates reported here should be accurate to ±60 °C (2σ).

We have obtained 33 successful primary magma solutions for non-arc basalts, and the results are listed in Table A1 in Appendix A. A successful primary magma solution is internally consistent with experimental and parameterized primary magma compositions for fertile peridotite composition KR-4003 (Walter, 1998; Herzberg and O'Hara, 2002; Herzberg et al., 2007; Herzberg and Asimow, 2008). Melt fractions and pressures of melting inferred from a primary magma composition (Sections 3.3, 3.4) must display internal consistency in MgO–FeO–CaO–SiO_2 and projection space (Herzberg and Asimow, 2008). The 33 successful solutions is a small number compared with the ~1500 possible solutions from our global lava database. There are many important reasons why we have obtained so few successful solutions. First, the solution will fail if a non-arc basalt formed from an exotic source composition that differed substantially from KR-4003 used in the PRIMELT calibration. Second, most basaltic rocks do not preserve in their geochemistry the T–P conditions of melt formation in the mantle because they are highly differentiated, having typically crystallized olivine, clinopyroxene, and plagioclase. PRIMELT2 software is not useful in these cases because it only provides information about source temperature if the primary magma gained or lost only olivine before solidifying to a primitive rock. This condition is rarely achieved in nature, as first discussed by O'Hara (1968), and it is especially relevant to primary magmas that undergo assimilation and fractional crystallization (AFC) owing to interaction with the subcontinental lithosphere and continental crust. Third, many of the analyses are on lavas that had been metamorphosed, and this has likely compromised their compositions. Nevertheless, our primary magma solutions display good internal consistency of major elements using criteria discussed in Herzberg and Asimow (2008). Fourth, PRIMELT2 identifies magmas generated from pyroxenite and CO_2-rich peridotite sources, and excludes them. In summary, there are many petrological factors that can confound our ability to secure a meaningful primary magma composition, which is why we are left with so few successful solutions. It is fortunate, therefore, that these solutions span the Proterozoic and the late Archean with no major age gaps (Fig. 1).

For the komatiites, primary magma compositions could not be obtained with PRIMELT2 modeling because metamorphism has substantially modified CaO, a common problem (e.g., Herzberg, 1992; Lahaye and Arndt, 1996; Arndt et al., 2008). Therefore, an estimate was made of parental magma composition, which is the most magnesian erupted composition from which other temporally and spatially related lava compositions can be derived by fractional crystallization. A primary magma that exits the melting regime will be the same as a parental magma if it is not modified during transit from the melting regime to the surface. Should fractional crystallization of olivine occur in conduits or deeply buried sills, the MgO content of the parental

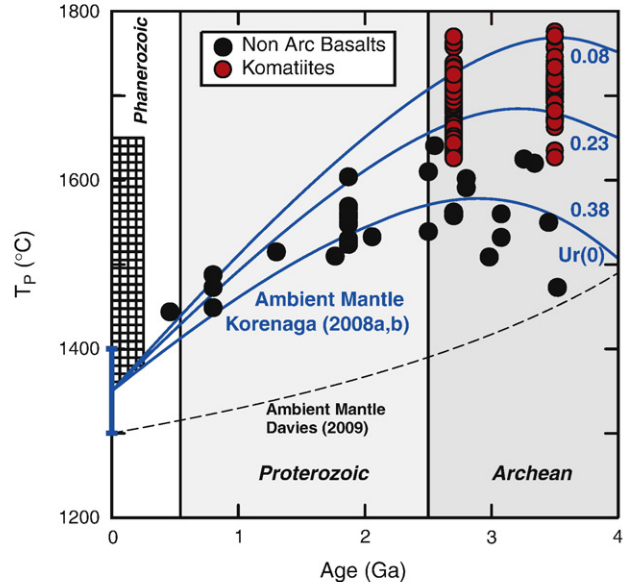

Fig. 1. Secular thermal Earth models for ambient mantle (blue curves; Korenaga, 2008a, b) compared with petrological estimates of mantle potential temperature (T_P) for non-arc lavas and komatiites with the ages indicated. Number next to each blue curve is the model present-day Urey ratio. We use the relation: T_P(°C) = 1463 + 12.74MgO − 2924/MgO (Herzberg et al., 2007; Herzberg and Asimow, 2008), where MgO is the primary magma MgO content given in Table A1. The range of T_P for plume-related magmas of Phanerozoic age is from Herzberg and Asimow (2008) and Herzberg and Gazel (2009).

magma will be less than that of the primary magma (see Herzberg et al., 2007 for discussion), yielding minimum bounds on inferred mantle potential temperature. In this work, we assume that the parental and primary magmas have the same compositions, and that they crystallized olivine having maximum observed Mg numbers; these are 93.0 for Belingwe (Sobolev et al., 2007), 94.5 for Alexo and Pyke Hill (Sobolev et al., 2007; Puchtel et al., 2004), and 94.0 for Barberton (Parman et al., 1997). Parental magma composition was determined for each lava by adding or subtracting olivine using 1 atm Fe–Mg partitioning from Toplis (2005) until the magma was in equilibrium with these olivine compositions. Berry et al. (2008) determined Fe^{2+}/Fe_T = 0.9 for Belingwe, and this value was assumed to hold for all komatiites.

2.3. Models of earth's thermal history

Thermal evolution models for Earth's mantle depend on the balance between heat production by radiogenic elements in the mantle and heat loss by mantle convection. Cosmochemical and geochemical constraints on the composition of Earth's primitive mantle, when combined with the composition model of continental crust, suggest that the radiogenic heat production in the mantle can account for only a small fraction of convective heat loss; the convective Urey ratio, which measures the relative contribution of mantle heat production to mantle heat flux, is estimated to be 0.23±0.15 at present (Korenaga, 2008b). This low Urey ratio is known to indicate an unrealistic thermal history called thermal catastrophe if a hotter mantle in the past is assumed to convect more rapidly (Christensen, 1985). Korenaga (2003) suggested that the partial dehydration melting of the convecting mantle beneath mid-ocean ridges must have created thicker dehydrated lithosphere, which may have slowed down plate tectonics, resulting in an inverse relationship between the temperature of the mantle and its convective vigor (i.e., more sluggish plate tectonics in the past), and thereby leading to more reasonable thermal evolution. Shown in Fig. 1 are three cooling curves spanning the plausible range of the present-day Urey ratio, based on the most recent version of such a model (Korenaga, 2008a) with the continental growth curve of Taylor and McLennan (1985). The calculations were performed by back-tracking the thermal history of the Earth from the present-day condition, which is defined by the potential temperature of ambient mantle that partially melts to produce oceanic crust along Earth's spreading ridges. At the present time, a mantle potential temperature (i.e., T_P) of 1350 ± 50 °C is required to produce primary basaltic magmas having 10–13% MgO (McKenzie and Bickle, 1988; Langmuir et al., 1992; Kinzler, 1997; Herzberg et al., 2007; Courtier et al., 2007; Lee et al., 2009; Gregg et al., 2009).

The low-Urey-ratio model with sluggish convection in the past predicts a thermal curve that is concave towards the time axis (Fig. 1), because in the past mantle heat flux was lower, internal heating was higher, and thus the Urey ratio was higher. In particular, the apex of the concave thermal curve corresponds to the time when the Urey ratio was unity. Prior to this time, internal heating exceeded surface heat loss, resulting in mantle warming. The concave form of the thermal evolution curve shown in Fig. 1 is a robust feature for low-Urey-ratio models, regardless of assumed continental growth curves. Different growth curves give virtually identical thermal histories up to ~3 Ga and start to diverge only in the early Archean (Korenaga, 2008a). Early continental growth in the Hadean (Armstrong, 1981; Harrison, 2009), for example, would have T_P that are ~100 °C higher at 4 Ga than the curves shown in Fig. 1 (Korenaga, 2008a), if the Hadean continental crust was as enriched in heat-producing elements as the modern analog and if the style of mantle convection in the Hadean was similar to contemporary plate tectonics.

Also shown in Fig. 1 is a secular cooling curve from Davies (2009), which is based on the conventional heat-flow scaling (i.e., more rapid plate tectonics in the past) and assumes a high present-day Urey ratio of 0.8. This is essentially the same as the traditional secular cooling model popularized in the early 1980s (e.g., Schubert et al., 1980; Davies, 1980). The high-Urey-ratio model with rapid convection in the past predicts a convex cooling curve (Fig. 1); mantle heat flux always exceeds internal heat production and the Urey ratio was lower in the past.

It is noted that low-Urey-ratio models with faster convection or high-Urey-ratio models with slower convection do not produce a sensible thermal history. For faster convection in the past to work, the present-day Urey ratio must be high (>0.7), which is in conflict with the geochemical budget of the Earth. More sluggish convection when the mantle was hotter may be counter-intuitive but appears to be consistent with what Precambrian geology indicates (Korenaga, 2006; Bradley, 2008).

3. Results

3.1. Secular heating and cooling

Fig. 1 shows that mantle potential temperatures inferred from petrological modeling of non-arc basalts of Archean and Proterozoic ages converge to the present-day ambient value, suggesting that these samples record the secular evolution of ambient mantle temperature. Compared to prior petrological constraints on mantle cooling (e.g., Abbott et al., 1994), the new estimates constrains more sharply the nature of cooling trend, which may be best characterized as more rapid cooling in the Proterozoic than in the Archean. This trend is consistent with the concave (towards the time axis) cooling curves predicted by the low-Urey-ratio model of Korenaga (2008a), suggesting that the estimated cooling trend is a physically realizable one for the ambient mantle. Both the petrological and parameterized convection models display a thermal maximum at 2.5–3.0 Ga. Petrological temperatures for Proterozoic basalts are well described by the present-day Urey ratios of 0.23 to 0.38, though results for Archean basalts are better characterized by a ±100 °C scatter in T_P around the curve for a Urey ratio of 0.38 (Fig. 1). This higher inferred Urey ratio and greater variability for the Archean lavas might be an artifact of local cooling owing to subduction/delamination, or by alteration during metamorphism. We also note that the calculated cooling curves assume the operation of plate tectonics similar to what we observed today throughout the last 4 Ga. The onset of plate tectonics in Earth history is a matter of great debate, and even if plate tectonics existed in the early Archean and the Hadean, its style may be different from modern plate tectonics. Different styles of mantle convection have different relations between mantle temperature and heat flow, so the reconstruction of thermal history prior to ~3 Ga must always be viewed with caution. For these reasons, we place an emphasis on the remarkable agreement between the petrological estimates and the theoretical prediction up to the late Archean.

Fig. 1 also shows less than 90 °C of ambient mantle cooling over the past 2.5 Ga in the secular cooling model of Davies (2009). Petrological T_P estimates of Archean and Proterozoic non-arc lavas are considerably higher, and would require a mantle plume model for their interpretation. However, we find this traditional high-Urey-ratio model problematic for the following reasons: 1) it is not obvious why T_P for late Proterozoic plume-interpreted lavas would be lower than many of those of Phanerozoic plumes, 2) it predicts the occurrence of MORB-like crust in the Proterozoic and early Archean, in contrast to observations (Bickle et al., 1994; Condie, 2005; see below), and 3) it is not obvious why plume temperatures should exhibit a thermal maximum at 2.5–3.0 Ga because both the mantle and the core cool down monotonically in the high-Urey-ratio model.

Archean komatiite parental magmas are generally higher in MgO than primary non-arc magmas (Fig. **2**), and these require hotter mantle to form (Fig. **1**). These results are consistent with the model that the komatiites formed in ancient mantle plumes (Jarvis and Campbell, 1983; Campbell et al., 1989; Herzberg, 1992; Nisbet et al., 1993; Fan and Kerrich, 1997; Arndt et al., 1998; Herzberg et al., 2007; Arndt et al., 2008). Our estimated parental magmas for the 2.7 Ga Alexo and Pyke Hill komatiites average 28% MgO and formed at $T_P = 1710$ °C, in excellent agreement with previous estimates (Puchtel et al., 2004; Arndt et al., 2008). Belingwe parental magmas contain on average 24% MgO, again in excellent agreement with previous

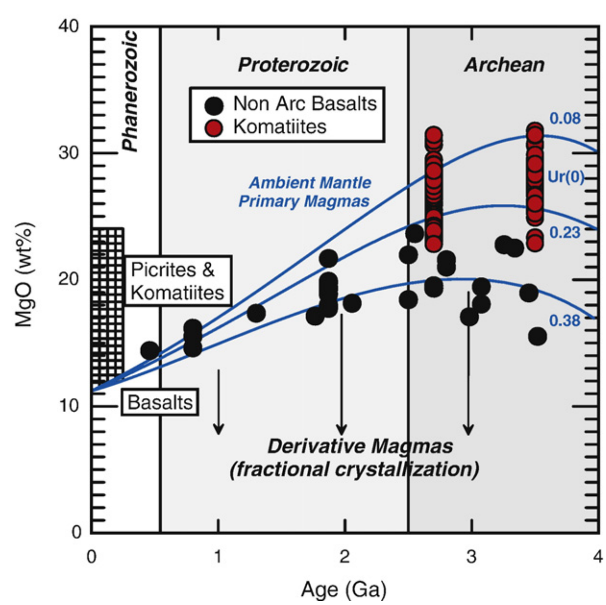

Fig. 2. Secular changes in the MgO contents of primary magmas of ambient mantle and parental magmas of Archean komatiites. The MgO contents can be recovered from mantle potential temperature using the relationship: $MgO = 58 - 0.0977T_P + 0.0000467T_P^2$ where T_P follows the secular cooling curves of Korenaga shown in Fig. 1. Number next to each blue curve is the model present-day Urey ratio. The range of MgO for plume-related magmas of Phanerozoic age is from Herzberg and Asimow (2008) and Herzberg and Gazel (2009).

estimates (Puchtel et al., 2009). However, the full range of FeO contents in the komatiites propagates to MgO that varies from 23 to 32% for komatiites with both 2.7 and 3.5 Ga ages (Fig. 2). This in turn propagates to inferred T_P that varies from ancient ambient mantle T_P to nearly 1800 °C (Fig. 1). Work on Phanerozoic plume magmas also reveals a range of MgO contents (Fig. 2), interpreted to reflect both plumes with variable T_P and the tapping of primary magmas from the hot axis and cooler periphery of a specific plume (Herzberg and Asimow, 2008; Herzberg and Gazel, 2009). However, we caution that the MgO range for Archean komatiites might not have the same petrological significance owing to possible shortcomings with the method of calculation as discussed above. In particular, MgO for parental magmas is the lower bound for MgO for primary magmas, as possibly reflected in the lower MgO for Belingwe komatiites. With these caveats in mind, we conclude that komatiites formed in Archean mantle plumes with T_P that was 100–200 °C higher than ambient mantle (Fig. **1**), a range that is possibly somewhat lower than those displayed by plumes in the Phanerozoic.

3.2. Composition of Archean–Proterozoic oceanic crust

Many of the Archean and Proterozoic samples we have modeled have been previously interpreted as plume-derived fragments of ancient oceanic crust, based on incompatible trace element ratios, evidence that shows striking similarities to those of Phanerozoic oceanic plateau basalts (Puchtel et al., 1998; Kerrich et al., 1999; Hollings et al., 1999; Kerr et al., 2000; Polat and Kerrich, 2000; Arndt et al., 2001; Condie, 2003, 2005; Smithies et al., 2005; Kerrich and Polat, 2006). Well-preserved pillow structures and/or turbidite sediments support a submarine environment for most greenstones (Polat, 2009). However, field evidence for some non-arc basalts suggests that they formed in autochthonous terranes (Bickle et al., 1994; Thurston, 2002; Ayer et al., 2002; Bleeker, 2002), possibly as continental rift or flood basalts (Bickle et al., 1994; Bleeker, 2002; Pearce, 2008). Some of the basalts for which we have successful solutions for mantle temperature were erupted on to continental crust (e. g., the Coonternah, Warrawoona, Sulfur Springs, Aravalli and Lapland sites, Table A1). Our work cannot contribute directly to this important debate. However, their convergence to present-day MORB T_P, together with the geochemically reasonable thermal history (Fig. 1), suggests that the non-arc basalts formed from hot ambient mantle, not hot mantle plumes. From this inference we conclude that they are representative of both magmatism in oceanic crust and continental basalts that erupted along rifts and adjacent areas during the Archean and Proterozoic.

The thermal model for ambient mantle permits a fresh look at the old problem of understanding the composition of ancient oceanic crust. Magmatism was associated with decompression melting of ambient mantle, producing primary magmas with 18–24% MgO near the Archean–Proterozoic boundary (Fig. 2), in agreement with Foley et al. (2003). Similar rocks are found in the Phanerozoic, but require for their formation the high temperatures that are characteristic of mantle plumes (Figs. 1 and 2). Although basaltic rocks having low MgO contents existed in the early Earth, these formed by fractional crystallization, not partial melting. Modern primary MORB melts contain 10–13% MgO and 6.5–8.0% FeO, and require T_P in the 1300–1400 °C range to form (McKenzie and Bickle, 1988; Langmuir et al., 1992; Kinzler, 1997; Herzberg et al., 2007). But our search for such primary magmas of oceanic crust in the Archean and the Proterozoic has been unsuccessful. However, we cannot eliminate the possibility that some may have been rejected by our calculation if their major elements had been remobilized. The lack of ancient primary MORB similar to modern primary MORB conflicts with the model of Davies (2009) shown in Fig. 1. We infer that primary basaltic melts with the MgO and FeO contents of modern primary MORB are a feature of Phanerozoic magmatism, not the early Earth (Fig. 2), which is why no

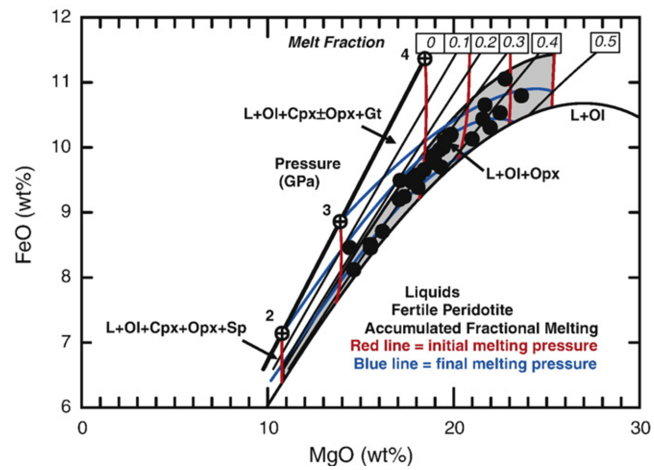

Fig. 3. MgO and FeO contents of model primary magmas for non-arc lavas of Archean and Proterozoic age compared with those of fertile peridotite. Black closed circles are primary magma solutions, from Table A1 in the Appendix. Black, blue and red lines are general solutions for melt fraction, pressure of final melting, and pressure of initial melting, respectively (Herzberg and O'Hara, 2002; Herzberg, 2004a; Herzberg and Asimow, 2008).

basaltic analogs of oceanic crust have been reported in Archean greenstone belts (Bickle et al., 1994; Condie, 2005).

3.3. Thickness of Archean–Proterozoic oceanic crust

Melt fractions for an assumed fertile peridotite source are provided by PRIMELT2 and given in Table A1 Appendix A for our primary magmas. These are also shown in Fig. 3, together with pressures of initial (P_I) and final (P_F) melting that were calculated by forward solutions to the equations for accumulated fractional melting (Herzberg and O'Hara, 2002; Herzberg, 2004a). Results show melt fractions that vary with the depth of the melting column (i.e., $\Delta P = P_I - P_F$) as shown in Fig. 4. The primary magmas of most non-arc lavas of Archean and Proterozoic ages formed by about 30% melting, and $\Delta P = 2$ to 3 GPa. Using a density of 2970 kg/m^3 (Korenaga, 2006), we can calculate crustal thickness of 25 to 35 km for the end-member

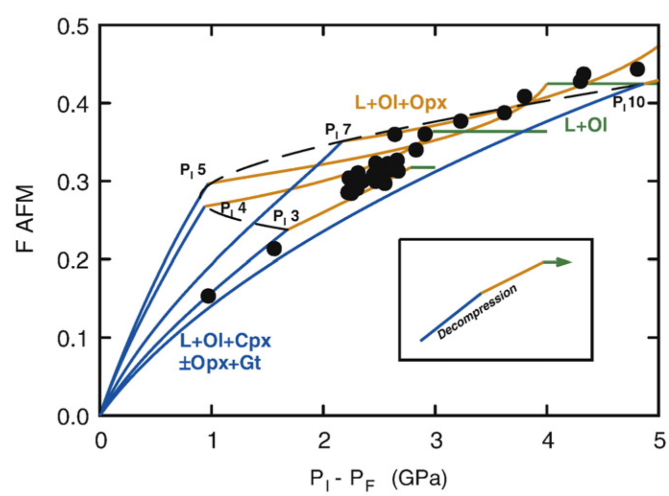

Fig. 4. Melt fraction for accumulated fractional melting as determined by the pressure range of melting. Black closed circles are primary magma solutions, from Table A1 in the Appendix. Blue, brown, and green lines are general solutions for decompression melting paths of lherzolite, harzburgite, and dunite phase assemblages, respectively, at the initial melting pressures indicated, from Fig. 3 and Herzberg and O'Hara (2002) and Herzberg (2004a). Inset is an example of changes in melt fraction with decompression from a specific initial melting pressure. Broken line tracks the transformation from lherzolite melting to harzburgite melting at 3 to 10 GPa. Instantaneous slopes depict melt productivity $\delta F/\delta P$.

case of a cylindrical melting column. This range is comparable to crustal thicknesses of the Ontong Java Plateau, which also formed by about 30% melting of peridotite (Herzberg, 2004b). It is inferred that 25–35 km of oceanic crust in the Archean–early Proterozoic is likely to have been normal, not anomalous, and uniformly thick (see also Hamilton, 1998).

3.4. Melt production in the Archean–Proterozoic and bathymetry

Most models of fractional melting assume that melting terminates at or near the lithosphere–asthenosphere boundary (Ellam, 1992; Langmuir et al., 1992; Humphreys and Niu, 2009), defining the final melting pressure. But in the early hot Earth, the pressure of final melting is effectively determined by a rapid reduction in melt productivity as controlled by the nature of the solid phases that contribute to melting (e.g., Asimow et al., 1997). This is illustrated in Fig. 5, a synthetic adiabatic melting path calculated from the MgO contents of the primary magmas and final melting pressures, as discussed in Fig. 5 and Herzberg and Gazel (2009). The temperature and pressure of final melting for any specific primary magma are approximately determined by the intersection of the adiabatic melting path and Opx stability (Fig. 5). The top of the melting column may indeed have been close to the surface, but melt productivity effectively shuts down when the only crystalline phase left in the residue is olivine (Fig. 4). This occurs because the adiabatic *T*–*P* melting path is approximately coincident with olivine liquidus temperature, and both solution and crystallization of olivine is not significant (Herzberg and O'Hara, 2002; Herzberg and Asimow, 2008).

For melt generation at very high T_P, the pressure of final melting is correspondingly high (Fig. 5). At pressures in excess of ~6 GPa, experimental data indicate a narrow temperature interval between the liquidus and solidus within which the stabilities of Opx, Cpx, and Gt are constrained (Herzberg and Zhang, 1996; Walter, 1998). This is a situation that is most relevant to the understanding of Archean komatiites. The MgO contents of our model komatiite parental magmas and inferred T_P are both higher than those of non-arc basalts (Figs. 1 and 2), consistent with the hot plume model. Additional support for a hot and dry plume model for komatiites was reported by Berry et al. (2008). Komatiites associated with non-arc basalts in Archean greenstone belts could have formed in melting columns over 200 km in depth, most of which would have been magmatically unproductive. And since many Archean komatiites were generated from depleted peridotite sources (Arndt, 1986; Bickle et al., 1993b; Chauvel et al., 1993; Arndt et al., 2008), magmatic productivity would have been even lower. Mass balance estimates indicate that they could have formed by about 50% melting of depleted peridotite (Herzberg, 2004a). This is certainly higher than 30–40% we get for ambient mantle (Fig. 4), and it would have added some local bathymetry to the Archean ocean floor. But this was not enough to make a Phanerozoic-style oceanic plateau, where about 8% melting of mantle peridotite is required to make 7 km of modern oceanic crust (Langmuir et al., 1992); in contrast, 30% melting is required to make 25–35 km of crust below the Ontong Java Plateau (Herzberg, 2004b; Richardson et al., 2000). Finally, komatiites are volumetrically minor, and constitute less than 5% of the volcanic rocks in most Archean greenstone belts (de Wit and Ashwal, 1997). The thermal evolution model with sluggish plate tectonics in the past predicts weaker plume activity in the past as well (Korenaga, 2008a). Therefore, komatiite magmatism was unlikely to have contributed significantly to the thickening of oceanic crust, evidence that does not support the existence of Phanerozoic oceanic plateau-like features in the early Earth.

3.5. Composition and origin of lithosphere in the Archean–Proterozoic

We can calculate the compositions of the complementary residues of primary magmas by simply rearranging the mass balance equation to $C_S=(C_o-FC_L)/(1-F)$ where C_o is the initial source composition, *F* the mass fraction of the aggregate melt (Table A1), C_L is the aggregate liquid composition (Table A1), and C_S is the composition of the residue (Herzberg, 2004b). We will assume that the initial source composition is similar to fertile peridotite KR-4003 (Walter, 1998; Herzberg, 2004a), for which melt fractions are obtained using PRIMELT2. Lithologically, the residues are mostly harzburgite, an expected consequence of ~30% melting, and they are very similar in

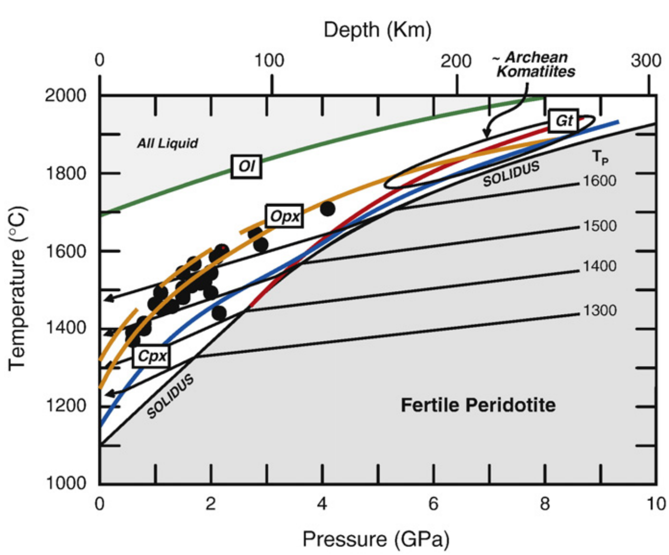

Fig. 5. Inferred temperatures and pressures at which fractional melting terminated for primary magmas of non-arc lavas with Proteozoic and Archean ages. For each primary magma the temperature and pressure is calculated using T_{OL} (°C) = $935+33\text{MgO}-0.37\text{MgO}^2+54P_F-2P_F^2$ where T_{OL} (°C) is the olivine liquidus temperature at the pressure of final melting P_F (Herzberg and Gazel, 2009). Colored curves labeled Ol, Opx, Cpx and Gt represent maximum temperatures where these phases are stable, for batch melting of fertile peridotite (from Fig. 13 of Herzberg and O'Hara, 2002). Broken line labeled Opx represents Opx stability for accumulated fractional melting. Solid state adiabatic gradients are from Iwamori et al. (1995), and are similar to those of Stixrude and Lithgow-Bertelloni (2005).

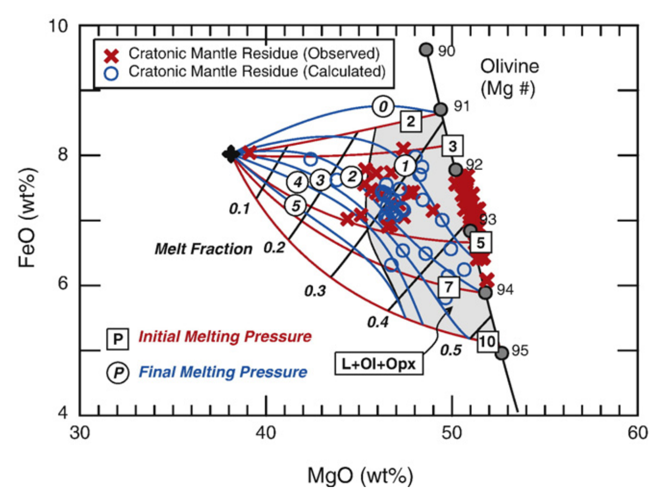

Fig. 6. Calculated residue compositions of primary magmas for non-arc lavas with Proteozoic and Archean ages. These are similar in composition to those observed for cratonic mantle (Boyd & Mertzman, 1987; Boyd et al., 1993; 1997; 1999; Lee & Rudnick, 1999; Bernstein et al., 2006). Initial and final melting pressures and melt fractions can be inferred from the red, blue, and black lines, respectively (Herzberg, 2004b), and are complimentary to those shown for primary magmas in Fig. 3. We do not consider xenoliths that are enriched in orthopyroxene by melt-rock reaction or other processes (Kelemen et al., 1998; Herzberg, 2004b; Canil and Lee, 2009) and xenoliths that have been enriched in garnet and pyroxene by metasomatic processes (Smith and Boyd, 1987; Griffin et al., 1989).

composition to xenoliths from the Kaapvaal, Siberian, Tanzanian, and West Greenland cratons (Fig. 6). These rocks have Archean–Proterozoic ages (Pearson, 1999; Carlson et al., 2005), and occupy a significant fraction of the lithospheric mantle below cratons to depths of around 250 km (Jordan, 1988; Boyd et al., 1997; Eaton et al., 2009).

Early interpretations that cratonic mantle formed as residues of hot plume magmatism (Herzberg, 1993, 1999) can now be discounted. Instead, the petrology is most consistent with a residual origin after extraction of hot magmas from ambient mantle below ancient ocean ridges and/or continents (Bickle, 1986; Herzberg, 2004a; Bernstein et al., 2006, 2007; this work). Mantle potential temperatures were typically 1500–1600 °C (Figs. 1 and 5). Melt fractions were ~0.3, but reached as high as 0.44 (Figs. 3 ,4 and 6). Initial melting pressures are inferred to have been mostly 3–5 GPa (Figs. 3 ,5, and 6). Final melting pressures were 1–2 GPa (Figs. 3 ,5, and 6), although dunite residues having Mg numbers of 92–93 likely formed by decompression to < 1 GPa (Fig. 6). Using these conditions, we estimate 85–135 km of lithospheric mantle residues were produced during the Archean and Proterozoic. This is in reasonable agreement with the estimates of Korenaga (2006), and considerably less than ~250 km below present-day Archean cratons. This thickness paradox can be resolved by tectonic thickening that accompanied underthrusting (Jordan, 1988; Herzberg, 1999; Canil, 2004), additional evidence that there existed some form of large scale lateral tectonic transport in the Archean (Korenaga, 2008a; Kerrich and Polat, 2006; Shirey et al., 2008). Thickening must have occurred early in order to explain the occurrence of diamonds with Archean and Proterozoic ages (Richardson et al., 1984; Shirey et al., 2002).

4. Discussion and conclusions

In the modern Earth a mantle potential temperature of 1350 ± 50 °C is required to yield ~7 km of MORB crust, having a bulk composition with 10–13% MgO and 6.5–8.0% FeO. Fragments of oceanic crust that formed throughout the Phanerozoic are preserved as ophiolites, and provide evidence that crust production was similar to today (e.g., Nicolas et al., 1994). But our search for primary magmas with low Phanerozoic MgO and FeO contents in rocks from the Archean and the Proterozoic has been unsuccessful. The ~1.9 Ga igneous rocks of the Cape Smith belt have been interpreted as oceanic crust formed along a rifted continental margin (Baragar, 1974; Hynes and Francis, 1982; Francis et al., 1983). However, our analysis reveals Cape Smith primary magmas with MgO contents and T_P that are both much higher than modern MORB (Figs. 1,2; Table A1). Similar results are obtained from the Winnepegosis komatiites of similar age located ~1500 km further to the southwest in the Trans-Hudson Orogen (M. Minifie, personal communication). We have not been able to obtain primary magma solutions for the 2 Ga Jormua ophiolite in Finland because of significant mobility of iron (see Peltonen et al. 1996). We infer that true basaltic partial melts having 10–13% MgO are a feature of Phanerozoic magmatism, not the early Earth, which is why modern-day analogs of oceanic crust have not been reported in Archean greenstone belts.

Petrological modeling of non-arc lavas with Archean and Proterozoic ages reveals secular changes in the conditions of melting in the mantle. Mantle potential temperature T_P increases from 1350 °C at the present time to a maximum of ~1500–1600 °C near the Archean–Proterozoic boundary (Fig. 1). The essential question surrounds the geodynamic setting of the mantle in which melting took place, whether it was hot ambient mantle or mantle plumes. Our model primary magma compositions are similar to those predicted by the thermal model of Korenaga (2008a,b), indicating that they formed from ambient mantle. An important implication is that these rocks are likely to be representative of those that formed in oceanic crust and erupted on continents. In contrast, the MgO contents of our model komatiite parental magmas and inferred T_P are both higher than those of non-arc basalts (Figs. 1 and 2), consistent with the mantle plume model. We interpret these results as ancient analogues of the modern Earth, with non-arc basalts and komatiites having formed from ambient mantle and mantle plumes, respectively.

The thermal model for ambient mantle permits a new look at the old problem of understanding the composition of ancient oceanic crust and lithosphere. We estimate that primary magmas of ambient mantle near the Archean–Proterozoic boundary contained ~18–24% MgO, formed by ~30% melting of peridotite at mantle potential temperature in the 1500–1600 °C range, and produced ~25–35 km of oceanic crust. These primary magmas left behind 85–135 km of complementary residues of harzburgite that are now found as xenoliths of cratonic mantle. Magma production in the early Earth was constrained by the mineralogy of the residues. Decompression melting of ambient mantle and plumes might have extended to near surface conditions, but melt productivity was negligible when the residuum mineralogy consisted of dunite. We infer that komatiitic magmatism contributed little to crustal thickness and that oceanic plateaus, if they existed, would have had subtle bathymetric variations, unlike those of Phanerozoic oceanic plateaus.

The consistency between the secular trend delineated by the model primary magmas of non-arc basalts and the thermal evolution model of Korenaga (2008a,b) provides support for a thermal budget of Earth characterized by a low Urey ratio and more sluggish mantle convection in the past. In this model, the form of the time–temperature curve is concaved towards the time axis (Fig. 1), reflecting early heating and later cooling. The Earth has been in a cooling mode throughout the Proterozoic and Phanerozoic. In contrast, we conclude that internal heating exceeded surface heat loss in deep Archean–Hadean time, resulting in mantle warming to T_P maxima at 2.5–3.0 Ga. Though our prediction for the Hadean is less secure as noted earlier, more sluggish convection thus lower heat flow in the Hadean may be consistent with the geotherm estimated from Hadean zircons (Harrison, 2009). And while we are equally insecure with the extrapolation of our temperatures derived from igneous rocks, the apparently improbable idea that thermal conditions of the modern Earth might also have existed in Hadean times (Harrison, 2009) is not unreasonable from a petrological point of view (Fig. 1).

In contrast, the traditional cooling model, revived by Davies (2009), has much cooler ambient mantle, and predicts the formation of modern MORB-like crust throughout the Proterozoic and Archean, which has so far not been observed. Davies (2009) criticized the model of Korenaga (2003) as being too sensitive to the adopted radius of curvature for plate bending, which is an important parameter for calculating heat-flow scaling. His argument, however, does not recognize that the heat-flow scaling is sensitive to the product of the radius of curvature and the effective lithospheric viscosity, both of which are not precisely known. The effective lithospheric viscosity has been estimated to be roughly in the range of 10^{22}–10^{24} Pa s (e.g., Gurnis et al., 2000). Even if the scaling is sensitive to the cube of the radius, therefore, the factor of 3 uncertainty in the radius can easily be compensated by two-orders-of-magnitude uncertainty in the lithospheric viscosity. The scaling of Korenaga (2003) thus does not require extreme conditions as Davies (2009) argued. A more serious problem is that the traditional high-Urey-ratio model is in conflict with the cosmochemical and geochemical constraints on the amount of radiogenic elements in Earth's mantle (McDonough and Sun, 1995; Lyubetskaya and Korenaga, 2007). It has also been shown that the existing models for Earth's primitive mantle composition are fairly robust even with the possibility of a hidden enriched geochemical reservoir (Korenaga, 2009); that is, the Urey ratio is still low even if the mantle is chemically layered. A possibility of a high Urey ratio due to lower time-averaged surface heat flux (instead of higher internal heat production) was also considered (e.g., Labrosse and Jaupart, 2007), but this notion has been discounted as being inconsistent with the geological record (Korenaga, 2007; Loyd et al., 2007).

The concave nature of secular cooling can only be achieved by having more sluggish plate tectonics in the past, which is required by the geochemical budget of the Earth and is also suggested by Precambrian geology. Our petrological inferences provide new support for such behavior of plate tectonics, contributing to an unusual confluence of different disciplines in the Earth sciences. In this study we assumed the heat-flow scaling of Korenaga (2003) which is based on mantle dehydration during mid-ocean ridge magmatism. However, there are other mechanisms that may lead to similar scaling (i.e., more sluggish plate tectonics when the mantle was hotter in the past), such as grain-growth kinetics (Solomatov, 2001) and gradual mantle hydration over Earth's history (Korenaga, 2008a). Though which mechanism would be most effective remains to be resolved, the new petrological constraints reported here suggest that the scaling of Korenaga (2003) captures reasonably well the ‘effective’ relation between mantle temperature and convective vigor. This agreement encourages a closer look at the dynamics of the early Earth using more varied physical models on an expanded database of igneous rocks.

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Appendix A. Supplementary data

Supplementary data associated with this article can be found, in the online version, at 10.1016/j.epsl.2010.01.022.

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