Geoscience Frontiers 7 (2016) 543–553

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Contents lists available at ScienceDirect

China University of Geosciences (Beijing)

Geoscience Frontiers

journal homepage: www.elsevier.com/locate/gsf

Focus paper

A great thermal divergence in the mantle beginning 2.5 Ga: Geochemical constraints from greenstone basalts and komatiites

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ARTICLE INFO

Article history:
Received 17 November 2015
Received in revised form
12 January 2016
Accepted 15 January 2016
Available online 2 February 2016

Keywords:
Archean greenstones
Mantle thermal evolution
Basalts
Mantle geochemistry
Komatiites

ABSTRACT

Greenstone basalts and komatiites provide a means to track both mantle composition and magma generation temperature with time. Four types of mantle are characterized from incompatible element distributions in basalts and komatiites: depleted, hydrated, enriched and mantle from which komatiites are derived. Our most important observation is the recognition for the first time of what we refer to as a *Great Thermal Divergence* within the mantle beginning near the end of the Archean, which we ascribe to thermal and convective evolution. Prior to 2.5 Ga, depleted and enriched mantle have indistinguishable thermal histories, whereas at 2.5–2.0 Ga a divergence in mantle magma generation temperature begins between these two types of mantle. Major and incompatible element distributions and calculated magma generation temperatures suggest that Archean enriched mantle did not come from mantle plumes, but was part of an undifferentiated or well-mixed mantle similar in composition to calculated primitive mantle. During this time, however, high-temperature mantle plumes from dominantly depleted sources gave rise to komatiites and associated basalts. Recycling of oceanic crust into the deep mantle after the Archean may have contributed to enrichment of Ti, Al, Ca and Na in basalts derived from enriched mantle sources. After 2.5 Ga, increases in $Mg^{\#}$ in basalts from depleted mantle and decreases in Fe and Mn reflect some combination of growing depletion and cooling of depleted mantle with time. A delay in cooling of depleted mantle until after the Archean probably reflects a combination of greater radiogenic heat sources in the Archean mantle and the propagation of plate tectonics after 3 Ga.

1. Introduction

To address the question of if and when Earth evolved from a stagnant lid to a plate tectonic regime, it is important to have an understanding of the chemical (Hofmann, 1988; Condie, 1994; Herzberg, 1995; Campbell, 2002) and thermal history of the planet (Davies, 2007; Labrosse and Jaupart, 2007; Nakagawa and Tackley, 2012; Van Hunen and Moyen, 2012; Hoink et al., 2013; Korenaga, 2013). A stagnant lid regime exists today on the Moon, Mars and probably on Venus, and is characterized by conductive and heat-pipe volcanic heat loss through a “one-plate” lithosphere. Although numerous papers have been published on this topic, we still have important outstanding questions. One issue not addressed is the compositional and thermal history of different types of mantle. Today we know that the mantle beneath ocean ridges is considerably cooler than the mantle source for oceanic island basalts such as Hawaii (Herzberg et al., 2007; Lee et al., 2009). Furthermore, geochemical and isotopic studies indicate the existence of several compositional reservoirs in the mantle (Hofmann, 1988). To better understand the thermal and tectonic history of the mantle, we must track these reservoirs through time.

One approach to this problem is to use basalts and komatiites, which are produced in the mantle and carry information on the thermal and compositional properties of their sources (Condie, 1994; Hofmann, 1997). Furthermore, these rocks occur in greenstones, which allow us to track these properties of the mantle to at least 3.8 Ga (Abbott et al., 1994; Herzberg et al., 2010). In this study, we make use of an extensive database of well-dated greenstone basalts and komatiites to track through time major element and mantle magma generation temperature (T_g) of oceanic mantle

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Peer-review under responsibility of China University of Geosciences (Beijing).

http://dx.doi.org/10.1016/j.gsf.2016.01.006
1674-9871/

domains. Incompatible trace element distributions are discussed in previous studies (Condie, 1994, 2003, 2015). Depleted mantle (DM) is sampled today by ocean ridge basalts and has an average mantle magma generation temperature (T_g) of 1350–1380 °C; enriched mantle (EM) is sampled at hot spots and has a T_g of 1450–1500 °C; komatiites are sampled in the Phanerozoic at only one location (Gorgona Island 90 Ma) and may come from plume tails with temperatures near 1600 °C (Campbell et al., 1989); and the fourth type of mantle, hydrated mantle (HM) characterizes convergent plate margins and has a present-day T_g similar to DM. The tectonic settings and mantle sources of modern oceanic basalts can be tracked with some degree of certainty to at least 2.2 Ga. The tight grouping of incompatible element ratios of non-hydrated mantle basalts $\geq$2.5 Ga suggests the mantle was well mixed by the late Archean (Condie, 2015).

The most exciting result of this study is that for the first time, we are able to track both thermal and compositional properties of depleted and enriched mantle through time and show that a great divergence in these properties occurred soon after the end of the Archean.

2. Methods

We limit our definition of basalt to samples with MgO of 7–17 wt.% and SiO_2 of 45–55 wt.% and komatiites are restricted to MgO of 17–35 wt.%, the upper limit imposed to eliminate rocks that may contain cumulus olivine. Using a smaller MgO range for komatiites (i.e., 20–30 wt.%) does not significantly change the median values upon which our interpretations are based. We group modern oceanic mantle into three categories (depleted (DM), enriched (EM), and hydrated (HM) mantle) based of a combination of geologic and incompatible element characteristics of greenstone basalts as summarized in Condie (2015, 1994) and to this we have added a fourth category, mantle sampled by komatiites (KM), which is rarely sampled after the end of the Archean (Arndt et al., 2008). These geochemical domains are hypothetical end members and as recorded by modern basalts and komatiites (Hart, 1988; Hofman, 1997; Stracke, 2012) and listed in the previous paragraph. The details of how each of these mantle domains is defined are given in Condie (2015) and are not repeated here. Also as discussed by Condie (2015), these mantle domains may exist in stagnant lid planets, and hence tracking them into the Archean and Hadean on Earth may not be equivalent to tracking plate tectonics on Earth into these early time periods. Below we discuss the major element characteristics of the basalts and komatiites through time and how they relate to mantle source compositions. Major element distributions are important in that they track (1) the degree of melting of mantle sources, and (2) the degree of depletion (with elements such as Ti) of the source with time. However, these changes may not track tectonic regimes prior to 2.5 Ga if Earth transitioned into a stagnant lid regime during this time (Van Hunen and Moyen, 2012; Condie, 2015).

Major elements are also used to calculate mantle magma generation temperatures (T_g) using the methods described in Lee et al. (2009). Because our approach in calculating primary magma compositions requires a reverse fractionation correction, samples were first filtered to include only those basalts with MgO of 7–17 wt.% in order to minimize the extent of fractional crystallization. We also eliminated samples that may contain cumulus minerals (chiefly olivine) as reported in the original publications. The primary magma composition is estimated by incrementally adding equilibrium olivine back into the magma, assuming Fe^{2+}/Mg exchange relationship as detailed in Herzberg and Asimow (2008), and Lee et al. (2009). This assumes that the magmas were saturated in olivine, not along a cotectic or with other phases. Selecting magmas with MgO of 7–17 wt.% minimizes these problems, but issues still remain for magmas undergoing pyroxene fractionation or derivation from pyroxenite sources. For the former, we use the filter based on Ca from Herzberg et al. (2007) and Herzberg and Asimow (2008), and for the latter, we select only samples with Fe/Mn ratios between 50 and 60. We terminate olivine addition when the olivine composition reaches a forsterite (Fo) content of 91. Magmas are assumed to be relatively un-oxidized, so an atomic Fe^{3+}/Fe^{T} of 0.1 is assumed. We recognize that there are different approaches to estimating primary magma composition. There is no doubt some uncertainty in assuming a fixed final forsterite content because this quantity varies with the extent of melting; our approach over-estimates temperatures if melting degrees are lower and under-estimates if melting degrees are higher than implied. Herzberg et al. (2007) simultaneously solved for temperature and melting degree in an attempt to reduce the arbitrariness of assuming a final forsterite content. Putirka (2005) used the same approach as ours, but chooses to terminate olivine addition at the highest forsterite content observed; this approach assumes that the magma is in equilibrium with the most depleted mantle residuum, but most magmas represent aggregate polybaric liquids so the average composition of the residues is more appropriate. There are thus inherent, but poorly constrained biases in each of these approaches and it is not clear whether any approach is superior. What we have done is to apply our approach consistently for all samples in order to evaluate whether any robust secular trends in temperature exist. The effects of variations in source composition, cotectic crystallization, magma mixing, and recharge (Lee and Bachmann, 2014) yield uncertainties in primary magma composition greater than our assumption of final forsterite content.

Temperatures of the primary magmas are estimated using MgO and SiO_2 thermobarometry following Lee et al. (2009). Considering all of the sources of error in the calculations, we consider the uncertainty range of our temperature calculations of ±50–100 °C. Because of difficulties in estimating equilibrium olivine composition and identifying samples with cumulus olivine, we do not calculate T_g for komatiites, but rather use published data of from Herzberg et al. (2007) and Herzberg and Asimow (2008). More detailed discussion of uncertainties of temperature calculations is given in Herzberg et al. (2007) and Lee et al. (2009).

Our calculated magma generation temperatures and pressures most likely represent an average temperature and pressure of pooled melts generated by decompression melting and therefore, reflect the average melting conditions in the mantle source. Strictly speaking, these temperatures do not correspond exactly to the mantle potential temperature (T_p) because latent heat absorption during adiabatic decompression melting causes a slight temperature decrease relative to the solid mantle adiabat. However, given the uncertainties in correcting for latent heat release, making this correction is not justified. For a low degree of melting (10–20%), the difference between melting temperatures and the solid mantle adiabat are small (<30 °C), and the differences between the melting temperatures and magma generation temperature are even smaller. For high degree melts (>30%), however, melting temperatures may under-estimate magma generation temperature by $\geq$100 °C.

3. Results

3.1. Major element distributions

Filtering our geochemical database for alteration and extensive fractional crystallization eliminated 40% of the basalts (from 5669 to 3414 samples) and 66% of the komatiites (from 3267 to 1118 samples). The filtered database with calculated mantle magma generation temperatures and depths of magma equilibration is given in Appendix 1.