Melt-rich lithosphere-asthenosphere boundary inferred from petit-spot volcanoes

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ABSTRACT
Young basaltic knolls have been discovered on the old oceanic lithosphere, namely petit-spot volcanoes. Based on their geochemical signatures, they have presumably originated from partial melts in the asthenosphere. However, there is no direct information on the depth provenance of petit-spot formation. Here we report new geothermobarometric data of rare mantle xenoliths discovered from petit-spot lavas exhibiting a geotherm much hotter than expected for the ca. 140 Ma seafloor on which petit-spots were formed. Such an anomalously hot geotherm indicates that melt porosity around the lithosphere-asthenosphere boundary (LAB) must be as high as a few percent. Such high melt porosity would be possible by continuous melt replenishment. Excess pressure induced by the outer-rise topography enables horizontal melt migration along the LAB and sustains a continuous melt supply to petit-spot magmatism. Given the general age-depth relationship of ocean basins, a melt-rich boundary region could also be a global feature.

INTRODUCTION
 Debates on the presence of partial melts in the asthenosphere have intensified recently. The occurrences of a seismic low-velocity zone and a high electric-conductive layer in upper mantle, for example, have provided supporting evidence for partial melting in the asthenosphere (e.g., Sifré et al., 2014), which is consistent with the experimentally determined peridotite solidus in the presence of H2O and CO2 (Wyllie, 2012). However, recent studies suggest that partial melts are not required to explain seismic observations (Faup and Jackson, 2005; Stixrude and Lithgow-Bertelloni, 2005; Priestley and McKenzie, 2006). In order to assess the presence of the melt, petit-spot volcanism deserves attention. Petit-spot volcanoes are tiny seamounts erupting off the fore-bulge of the downgoing oceanic plate beneath the sampling sites, but the geothermobarometric data of these mantle xenoliths imply a very hot geotherm corresponding to young (ca. 18 Ma) oceanic lithosphere (Fig. 1). Petit-spots are inevitably samples of asthenospheric partial melts beneath a mature oceanic lithosphere; a good understanding of how petit-spots actually form can provide entirely new insight into the physical state of the normal upper mantle.

So far, about a dozen mantle xenoliths have been recovered from petit-spot lavas (Yamamoto et al., 2009). The occurrence of the mantle xenoliths provides essential information on the host magma, such as the origin and ascent rate of the magma. In addition, these xenoliths can constrain how the oceanic lithosphere is affected by petit-spot formation. As long-standing volcanism at an old oceanic plate is expected to affect the geotherm, the thermal structure of the oceanic plate deduced from the mantle xenoliths could potentially test the occurrence of melt at the base of the lithosphere.

THERMAL STRUCTURE OF THE OCEANIC LITHOSPHERE
We used five xenoliths recovered from petit-spot basalts erupting on subducting Pacific Plate off the Japan Trench (Fig. DR1 in the GSA Data Repository1). To infer the thermal structure of the oceanic plate from which these xenoliths are supposed to have originated, we need geothermobarometric data of the mantle xenoliths. The temperature was estimated using the two-pyroxene thermometer proposed by Wells (1977). The range of temperature is 771–1129 °C (Tables DR1 and DR2 in the Data Repository). Regarding the pressure, we conducted a thorough investigation of the depth provenance of the xenoliths based on pressure information recorded in the mantle xenoliths such as thermodynamic geobarometry, mineralogy, and the residual pressure of fluid inclusions and their stability conditions. We summarize the essence of the investigation here. The xenoliths are spinel peridotites, which are stable in mantle at 1.0–2.2 GPa (O’Neill, 1981; Borghini et al., 2010). In addition, there are abundant CO2 inclusions in the xenoliths. Taking the stability conditions of CO2 fluid in olivine into consideration (Koziol and Newton, 1998), the present spinel peridotites should be derived from the pressure-temperature (P-T) conditions depicted by a shaded area in Figure 1.

There is geobarometry based on residual pressure of fluid inclusions, which can be applied to mantle xenoliths (e.g., Miller and Richter, 1982). The estimated pressure and temperature are shown in Figure 1 (see the Data Repository for data and analytical procedures). There is considerable validity to the pressure estimated by the fluid inclusion geobarometry because the P-T values are well supported by the stability conditions of both aluminum-bearing minerals and CO2 fluid in olivine. The estimated pressures have a positive correlation with the temperatures. The most plausible explanation for the correlation is that they reflect the geotherm beneath the sampling sites, but the geothermobarometric data of these mantle xenoliths imply a very hot geotherm corresponding to young (ca. 18 Ma) oceanic lithosphere (Fig. 1). Such a geotherm must be a localized feature, restricted to the vicinity of these petit-spots; if it represented a regional geotherm, an isostatic adjustment would have raised the seafloor by as much as ~2 km (Turcotte and Schubert, 2002). This view of a highly localized hot geotherm is also consistent with the spotty distribution of anomalously high heat-flow data reported from this region (Yamano et al., 2008).

GENERATION MECHANISM OF PETIT-SPOT MAGMA
The existence of localized thermal anomalies in the old oceanic lithosphere thus appears to be feasible, and as discussed below, two lines of reasoning suggest that such thermal anomalies may have profound implications for the origin of petit-spots as well as the nature of the lithosphere-asthenosphere boundary.

First, the amount of excess heat required to generate the thermal anomaly may be estimated as

\[ \Delta E = C_p \Delta T \rho_m HA, \]  

where

- \( C_p \) is the specific heat capacity of the upper mantle,
- \( \Delta T \) is the temperature anomaly,
- \( \rho_m \) is the density of the mantle,
- \( HA \) is the heat of melting.

This approach suggests that the thermal anomaly is sufficient to drive petit-spot magmatism. However, the exact mechanism for the generation of such anomalies is still under investigation. Further research is required to fully understand the physical and chemical processes that underlie the formation of petit-spot volcanoes and the evolution of the oceanic lithosphere.
Second, the formation of petit-spots requires melt migration through thick oceanic lithosphere, and a quantitative consideration suggests a rather specific circumstance, beyond a favorable stress state caused by nearby plate bending (Hirano et al., 2006). To avoid freezing within the lithosphere, melt has to travel sufficiently fast, or the Peclet number must be substantially greater than unity (Turcotte and Schubert, 2002):}

\[
P_{e} = \frac{v h}{k} \gg 1, \tag{2}
\]

where \( v \) is vertical melt velocity, \( h \) is a channel width, and \( k \) is thermal diffusivity \((10^{-6} \text{ m}^2 \text{ s}^{-1})\). A primary rate-limiting factor for melt ascent is the supply rate through the base of the lithosphere, which may be seen in the following mass conservation equation:

\[
v h - n \rho_{0} \delta_{m} \frac{k}{\mu} \frac{dp}{dx} \delta_{m}, \tag{3}
\]

where \( n \) is horizontal melt velocity, \( \rho_{0} \) is melt porosity in the asthenosphere, \( \delta_{m} \) is compaction length, \( k \) is permeability, \( \mu \) is melt viscosity \((10 \text{ Pa s})\), and \( dp/dx \) is the horizontal pressure gradient. The first term is vertical melt flux, and the second term is horizontal melt flux, which is expanded by Darcy’s law in the last expression. Partial melt in the asthenosphere would migrate upward by its own buoyancy and pool beneath the lithosphere. Though complete melt segregation is unlikely at low porosities, we can still expect a relatively melt-rich region due to compaction, the vertical scale of which can be estimated by the compaction length (McKenzie, 1984; Sparks and Parmentier, 1991). Horizontal melt migration is possible with a nonzero horizontal pressure gradient, and in the case of petit-spot magmatism, the outer-rise topography is the most important source for such excess pressure (Fig. 2), i.e.,

\[
\frac{dp}{dx} \left( \rho_{m} - \rho_{w} \right) g \Delta d L, \tag{4}
\]

where \( \rho_{w} \) is the density of sea water \((1000 \text{ kg m}^{-3})\), \( g \) is gravitational acceleration \((9.8 \text{ m s}^{-2})\), \( \Delta d \) is the excess topography of the outer rise \((-800 \text{ m})\),

\[
Y = \frac{\rho L^{3} g}{K}, \tag{5}
\]

where \( Y \) is vertical melt velocity, \( L \) is the length of a melt channel, \( K \) is the permeability of the lithosphere, and \( g \) is gravitational acceleration. The product of vertical melt velocity \( v \) and channel width \( b \) is constrained by this horizontal melt transport, though vertical melt velocity can increase by focusing melt transport so that petit-spot formation is spatially localized, at an interval of \( W \). The illustration is not drawn to scale; in particular, note that \( H \gg \delta_{0} > \Delta d \).
and $L$ is the horizontal spatial scale (~600 km). Matrix shear associated with the bending of oceanic lithosphere also results in a horizontal pressure gradient (McKenzie, 1984), which adds to the above pressure gradient. The pressure gradient due to bending, however, is on the order of $\eta V / L^2$, where $\eta$ is matrix viscosity and $V$ is plate velocity, being considerably smaller than the pressure gradient due to the outer-rise topography. At any rate, without continuous melt delivery from the asthenosphere, melt migration through the lithosphere would resemble the propagation of a buoyancy-driven melt-filled crack with constant volume, the thermal fate of which depends strongly on initial melt volume (Spence and Turcotte, 1990). Our formulation above may be regarded as a continuum approximation for discrete crack propagation. As already suggested (Hirano et al., 2006), we assume that vertical melt transport becomes prominent when the stress state at the base of lithosphere becomes extensional, i.e., far away from the outer rise.

Permeability is a function of porosity and grain size (Fig. 3A), and the compaction length is a function of permeability, melt viscosity, and matrix viscosity. Using values appropriate for asthenospheric mantle, the compaction length is found to be on the order of ~10 m to a few kilometers (Fig. 3B). By combining Equations 2–4, the Peclet number can be expressed as a function of porosity, and if melt migration is entirely two-dimensional (2-D; i.e., no variation along the horizontal axis perpendicular to plate motion), melt porosity has to be much higher than 10% to make the Peclet number greater than unity (see 2-D line in Fig. 3C). This estimate is insensitive to the channel width $h$, because the horizontal melt delivery rate constrains the total product of the channel width and vertical melt velocity, $vh$ (Equation 3). It is unlikely, however, that melt migration would remain 2-D, and any 3-D focusing can raise the Peclet number. If such 3-D effects eventually result in the average spacing, $W$, between petit-spots, the Peclet number would increase by a factor of $Wh / h$, for $h$ of 100 m and $W$ of 100 km, for example, surface eruption would become possible even with the porosity of ~3% (Fig. 3C). Moreover, from the very fact that mantle xenoliths (with a size of ~1 cm) were collected from petit-spots, vertical melt velocity must have become as high as ~1 cm s$^{-1}$ at some point in petit-spot evolution (Rutherford, 2008), which corresponds to the Peclet number of $10^4$–$10^6$ with a channel width of 1–100 m. For reasonable ranges of channel width (1–100 m) and petit-spot spacing (10–1000 km), the focusing factor $Wh / h$ can be as high as $10^6$, so it is possible to achieve such a high Peclet number if melt porosity is as high as a few percent (Fig. 3C); because of uncertainty in grain size and matrix viscosity, we cannot be more specific about required melt porosity, but it is difficult to achieve Peclet number of $10^4$ with melt porosity less than 1%. Also note that, for melts to flow seaward from the outer rise, the compaction length has to be greater than the outer rise excess topography (Fig. 2), which is possible only with high porosities (Fig. 3B); otherwise, overpressured melts would be entirely trapped within the flexed base of the lithosphere.

A likely scenario for petit-spot formation would therefore start with nearly complete failures of vertical melt migration by initially 2-D flow and gradually shift to more successful melt ascents by 3-D focusing. During this process, a column of lithosphere beneath a petit-spot would be heated up, resulting in an anomalous geotherm. Using Equation 3, we can estimate the volume of melt delivered per petit-spot by assuming the petit-spot spacing and the duration of melt delivery (Fig. 3D). For a plate velocity of 10 cm yr$^{-1}$ and one petit-spot per area of 100 x 100 km, the total melt volume delivered from the asthenosphere would be greater than that of the surface expression by 2–3 orders of magnitude for melt porosity of a few percent (Fig. 3D). The volcanic construct of a petit-spot thus represents only a small fraction of the total melt volume required to make surface eruption possible.

**DISCUSSION AND CONCLUSIONS**

A broad agreement between the two different estimates on the total melt volume involved, one based on excess heat anomaly and the other based on the dynamics of melt transport, is encouraging, but our preferred melt porosity (>1%) may be problematic; with volatile contents similar to the sources of mid-ocean-ridge basalt, only very small amounts of melt (<0.1%) are predicted to exist under asthenospheric conditions (Hirschmann, 2010). By vertical migration, however, melt can pond beneath the lithosphere, where melt porosity can be locally high. Indeed, this process is already assumed in our conceptual model by the use of compaction length (Fig. 2). The presence of the seismic G discontinuity, which is the stratum in mantle as detected by changes in seismic wave velocity (Kawakatsu et al., 2009; Revenaugh and Jordan, 1991; Gaherty et al., 1996), often associated with the lithosphere-asthenosphere boundary, requires a sharp and large velocity drop, which could be explained by such melt ponding. Without continuous replenishment, ponded melt would be frozen by a growing lithosphere (Hirschmann, 2010), but oceanic lithosphere does not simply grow, as inferred from the age-depth relationship of global seafloor. It has long been known that simple half-space cooling cannot explain the subsidence of the seafloor when the age of the seafloor exceeds ca. 70 Ma (Parsons and Sclater, 1977) (Fig. 1), clearly indicating that oceanic lithosphere is subject to excess heat supply from below. A recent statistical analysis of global age-depth data (Korenaga and Korenaga, 2008) shows that a majority of seafloor is affected by the formation of hotspots and oceanic plateaus, much broader areas than implied by the spatial extent of those volcanic features. The presence of petit-spots may reflect a regional feature specific to the northwestern Pacific (Obayashi et al., 2006), but the above considerations suggest that a thin, melt-rich layer could potentially exist at many other places beneath oceanic lithosphere;
recent geophysical studies (Schmerr, 2012; Naif et al., 2013) appear to be consistent with this view. Because of a limited number of so-far discovered petit-spots, it is difficult to judge the robustness of this inference, but the anomalous geotherm revealed by these rare mantle xenoliths provides a unique perspective to the ongoing debate on the nature of the lithosphere-asthenosphere boundary, through the fluid dynamics of petit-spot formation.

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