

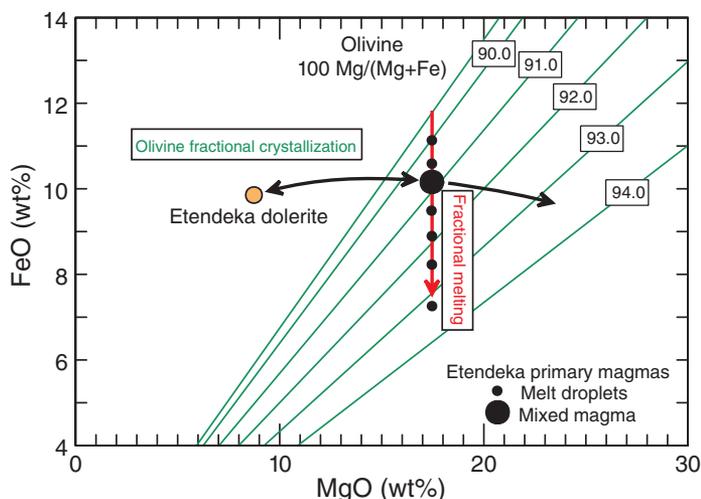
## Basalts as temperature probes of Earth's mantle

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The Earth is cooling down by convecting its heat from the interior to the surface, where it can be dissipated into space. Ambient mantle that rises to deliver its heat melts before it reaches the surface, and the basaltic magma forms new crust along oceanic ridges (Ahern and Turcotte, 1979; McKenzie and Bickle, 1988; Langmuir et al., 1992). Similarly, mantle plumes from the deep interior melt before reaching the surface, and the magma can form oceanic plateaus, islands, and seamounts, in addition to flood provinces on land (Coffin and Eldholm, 1994). The temperature that the solid adiabatically convecting mantle would be at the surface if it did not melt is called the potential temperature ( $T_p$ ; McKenzie and Bickle, 1988).  $T_p$  is a convenient measure of the temperature of the mantle and its variations. The question is, how do we measure the temperature of the mantle?

Basalts can be used as temperature probes of the mantle because their compositions change with variations in  $T_p$  (Langmuir and Hanson, 1980; Langmuir et al., 1992; Putirka, 2005; Herzberg et al., 2007; Herzberg and Asimow, 2008). A petrological method for estimating  $T_p$  is illustrated in Figure 1, using an olivine dolerite rock from the Paraná-Etendeka large igneous province (LIP) in Namibia as an example. The method is based on the understanding that higher mantle temperatures cause mantle peridotite to partially melt to form primary magmas with higher MgO contents. As reviewed by White (2010), there exist many petrological models, and they



**Figure 1.** MgO and FeO contents of magmas and the compositions of olivines that they would crystallize. Primary magma compositions are those expected for peridotite melting. In all cases,  $\text{FeO} = 0.9\text{FeOT}$ , and olivine compositions are calculated with  $K_D (\text{Fe/Mg})^{\text{ol}}/(\text{Fe/Mg})^{\text{m}} = 0.31$ . The Etendeka dolerite is sample JVT-09-21 reported by Keiding et al. (2011), and the mixed primary magma is computed from PRIMELT2 (Herzberg and Asimow, 2008). Small melt droplets are mostly illustrative of fractional melting, but those with the lowest FeO contents are from Keiding et al. (2011). Olivine forsterite contents ( $F_o$ ) used in this article are interchanged with Mg number,  $100\text{Mg}/(\text{Mg} + \text{Fe})$  in cation%.

all attempt to calibrate the MgO content of a primary magma with either mantle  $T_p$  or the temperature at which the magma starts to crystallize olivine ( $T_{\text{OL}}$ ).  $T_p$  and  $T_{\text{OL}}$  are positively correlated, and both can be used as thermometers of the mantle. But Keiding et al. (2011, p. 1095 in this issue of *Geology*) contribute significantly to basalt thermometry by demonstrating that petrological models that ignore the physics of primary magma formation are particularly susceptible to large uncertainties.

A significant problem is that primary magmas never erupt to the surface because they are modified by fractional crystallization (O'Hara, 1968). If olivine ( $\text{Mg,Fe}$ ) $_2\text{SiO}_4$  was the only fractionating mineral, then in principle the primary magma could be restored by simply adding olivine back into the rock. This rock must be a solidified melt, and it can have a variable olivine phenocryst content. The restoration procedure is called an inverse model because it attempts to invert the process of fractional crystallization, and it is the restored primary magma that provides temperature information. The problem is that one needs to know how much olivine to add. In the example shown in Figure 1, olivine is added to an Etendeka dolerite dike composition provided by Keiding et al., and the backtracking calculation is stopped when the magma composition is in equilibrium with observed olivine phenocrysts having maximum forsterite contents, this being  $F_{o_{93.3}}$ . This backtracking is a popular petrological method, and the Etendeka result is a primary magma having ~24% MgO. Such a primary magma would have required a  $T_p$  of ~1650 °C; ~300 °C higher than ambient mid-oceanic ridge basalt (MORB)-producing mantle, and similar to estimates originally reported by Thompson and Gibson (2000) for the Etendeka dolerite dikes. The problem with this method of calculation is that it ignores the compromising effects of the partial melting process.

Mantle peridotite partially melts at low melt fractions, and the small melt droplets are efficiently removed by buoyancy-driven draining from the residue (Ahern and Turcotte, 1979; McKenzie, 1984), a process called fractional melting. The residue is continuously melted in small amounts during decompression, and these melt droplets can develop very low FeO contents with MgO that can remain mostly unchanged (Fig. 1). In the Etendeka case, all melt droplets can have ~17.5% MgO and be in equilibrium with olivine having  $F_{o_{89.5-93.3}}$ . At some stage, these melt droplets mostly mixed to produce an "aggregate" or "mixed" primary magma composition which would begin to crystallize olivine with ~ $F_{o_{91}}$  and form the Etendeka dolerite sample shown in Figure 1. However, mixing of these melt droplets was not complete, and Keiding et al. have shown that some were trapped and preserved as inclusions in olivine with  $F_{o_{93.3}}$  (Fig. 1). Melt droplets with low FeO and 17.5% MgO and in equilibrium with  $F_{o_{93.3}}$  olivine indicate a mantle  $T_p$  of 1520 °C. This is substantially lower than the 1650 °C given above for a primary magma having 24% MgO, also in equilibrium with  $F_{o_{93.3}}$  olivine. Both estimates are substantially higher than the 1350 °C for ambient MORB-producing mantle, and both support a mantle plume thermal anomaly for the Paraná-Etendeka LIP (Thompson and Gibson, 2000). However, in other cases, an uncertainty of 130 °C is more serious because it can be too high to distinguish whether a basalt formed from ambient mantle or anomalous mantle, casting doubt on any geodynamic interpretation. This type of inverse model is fundamentally flawed because a wide range of primary magma compositions can be in equilibrium with  $F_{o_{93.3}}$  olivine, and  $T_p$  is poorly constrained.

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An important outcome of the study by Keiding et al. is that inferences about mantle  $T_p$  can have large errors using an incorrect petrological procedure. The method in question is the reconstruction of a primary magma composition to a targeted olivine composition, in this case  $Fo_{93.3}$  (Fig. 1). Keiding et al. show that this commonly used petrological method can result in a serious overestimation of  $T_p$  if the primary magma composition is not well mixed. Future studies should explore and develop other petrological methods for computing  $T_p$  and integrate the results with melt inclusion studies. For example, PRIMELT2 software (Herzberg and Asimow, 2008) is one such option because this method does not use olivine phenocryst composition for computing primary magma MgO content and  $T_p$ . It is well beyond the scope of this Research Focus article to describe the working of PRIMELT2; however, the application of this method to Etendeka whole-rock sample JVT-09–21 reported by Keiding et al. yields a primary magma with 17.5% MgO and  $T_p = 1520$  °C, in excellent agreement with those inferred from melt inclusions in equilibrium with  $Fo_{93.3}$  olivine (Fig. 1).

There are many reasons why it is important to know the temperature of Earth's mantle. On a planetary scale, the thermal state of the Earth's mantle, together with its chemistry, lithology, and physical structure, are attributes that define the solid Earth as a whole. At a more magnified view, most petrological models agree that  $T_p$  for LIPs and ocean islands are typically higher than those for oceanic ridges; evidence that supports the mantle plume model (Putirka, 2005; Putirka et al., 2007; Herzberg et al., 2007; Herzberg and Asimow, 2008; Herzberg and Gazel, 2009; Lee et al., 2009; White, 2010). However,  $T_p$  is also highly variable for these intraplate occurrences (Herzberg and Asimow, 2008; Herzberg and Gazel, 2009), possibly owing to variable amounts of recycled crust in mantle plumes. Plumes that carry more compositionally dense eclogite/pyroxenite will upwell at a lower vertical velocity, they diffuse more heat because the travel time is longer, and they cool down.  $T_p$  is even variable for individual intraplate occurrences, consistent with magmatic tapping of both the hot core and cooler peripheries of a mantle plumes.

While there are many examples of natural variations in mantle temperature, some variations in  $T_p$  that have been reported in the literature are also artifacts of the petrological method (White, 2010). Different petrological models yield absolute temperatures that can differ by more than 100 °C in some cases, and the mixing problem discussed by Keiding et al. is just one of several factors that contribute to discrepant results. Other common problems include the way primary magmas are calibrated to provide temperature information, and difficulties encountered in inverting the fractional crystallization process of a basalt to obtain its primary magma composition. Future research should seek to resolve these disparate petrological temperature estimates, and to provide a well-calibrated basaltic temperature probe of Earth's mantle. Progress in this petrological direction can be integrated with geochemical studies to sharpen our ability to distinguish the magmatic products of mantle plumes from those of lithosphere flexure and fracture (Hofmann and Hart, 2007; Hirano and Koppers, 2007).

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