The geodynamics of mantle melting

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Basaltic volcanism is the result of partial melting of the Earth's mantle (Fig. 1A). Volcanic activity in arcs and along mid-ocean ridges and continental rifts is related to plate tectonics, through decompression melting of the shallow sublithospheric mantle in regions of lithospheric extension and thinning, and/or hydration of the mantle wedge above subducting slabs. In contrast, intraplate volcanism, responsible for thick continental flood basalts, has usually been linked to upwelling and decompression melting of large volumes of deep, hot mantle (e.g., mantle plume; Morgan, 1983; Richards et al., 1989). More recently, decompression melting in small-scale convection cells along thick lithospheric keels has been proposed to explain minor intraplate volcanism (Davies and Rawlinson, 2014).

One of the main differences between plate-tectonics–related melting of the sublithospheric mantle and partial melting in hot upwelling mantle material is the pressure (P) and temperature (T) at which the adiabat (i.e., geotherm in the convective mantle) intersects the mantle solidus (Fig. 1B). These cannot be obtained directly from the composition of basalts because they are affected by fractional crystallization and interaction with surrounding rocks, as well as the progressive evolution of the composition of the source. The geochemistry of the first-formed batch of melt as fertile mantle crosses its solidus (i.e., the primary melt) is a function the mantle potential temperature, T_P , as shown by forward thermodynamic modeling (Asimow, 1999; Asimow et al., 2001; Herzberg and O'Hara, 2002; Herzberg, 2004). T_P is the temperature measured at the earth surface from extrapolation of the mantle adiabat (McKenzie and Bickle, 1988), a convenient way to express the intersection temperature of adiabat and solidus. The composition of the elusive primary melt of a volcanic province derives from processing the most magnesian liquids (i.e., parental melts) inferred from a suite of basalts (Herzberg et al., 2007). Comparison of the calculated and modeled primary melt constrains the T_P from which the geodynamic setting of the partial melting can be deduced. By targeting primary melts, petrologists mitigate geochemical alteration of parental magmas through fractional crystallization and interaction with surrounding rocks during melt ascent (e.g., Herzberg et al., 2007). Keeping in mind the many pitfalls (Herzberg et al., 2007; Herzberg, 2011), data from present continental rifting and spreading ridges derive a T_P of less than 1400 °C

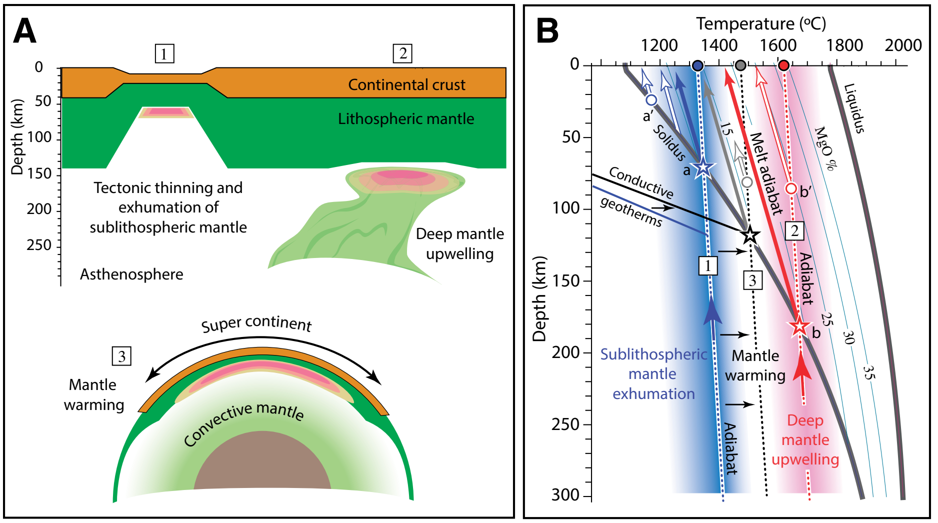

Figure 1. A: Geodynamic settings of mantle melting. B: Phase diagram, contoured for MgO, of mantle melting. 1—lithospheric thinning; 2—deep mantle upwelling; 3—mantle warming. During (1) and (2), the convective mantle is exhumed following an adiabat (blue and red dashed lines in B) intersecting the solidus at ~70 km and 180 km, respectively, with potential temperature (1330 °C and 1620 °C) shown by the blue and red dots at Earth's surface. Following mantle warming (3), the adiabat (black dashed line) intersects the solidus at ~120 km depth, with potential temperature (~1470 °C) shown by the gray dot. Primary melts are produced at solidus temperatures (blue, gray, and red stars). Should these melts be extracted as soon as they are produced, they would follow their respective melt adiabat (solid blue, gray, and red arrows). In the case of continental rifting leading to a mid-ocean ridge, melt is usually extracted at melt fraction of <1% (i.e., fractional melting). As latent heat escapes with the melt, the source rock cools and follows the solidus during its exhumation (path a-a′). As the source becomes progressively more residual, it produces parental melts with a range of composition (empty blue arrows). In mantle plumes, melt continuously reacts with the residue, with which it remains in equilibrium. As latent heat remains with the melt in the source, the upwelling mantle keeps following the adiabat (path b-b′). As the plume head slows down, melt segregates toward the top of the partially melted column, before escaping to the surface (i.e., batch or equilibrium melting) following the parental melt adiabat (empty red arrow). The pressure at which melt is extracted is the final melting pressure. Both fractional melting and equilibrium melting can occur in succession depending on strain rate, permeability, and the relative upwelling velocities of the residue and melt.

(Falloon et al., 2007; Herzberg et al., 2007). In this issue of *Geology*, Hole (2015, p. 311) shows that data from plume-related flood basalts point to a T_P of more than 1500 °C, in agreement with, for example, Thompson and Gibson (2000), Herzberg and O'Hara (2002), and Herzberg et al. (2007). The range of adiabats during tectonic extension and exhumation of the shallow mantle (blue region in Fig. 1B) and during deep mantle upwelling (red region in Fig. 1B), do not overlap. The concept of T_P can thus be used to constrain the geodynamic setting of mantle melting. In the second part of his paper, Hole focuses on the origin of the ca. 200 Ma Central Atlantic magmatic province (CAMP).

The CAMP extends over Africa, South America, Europe, and North America, covering ~10^7 km^2 (Marzoli et al., 1999). It postdates by 50 m.y. the final stages of Pangea's assembly. The peak igneous activity at 199–200 Ma pre-dates by ~10 m.y. the initial opening of the central Atlantic Ocean (Marzoli et al., 1999; Sahabi et al., 2004). The attribution of the CAMP to a mantle plume-head impinging the base of the lithosphere (Hill, 1991; Wilson, 1997; Leitch et al., 1998; Courtillot et al., 1999; Janney and Castillo, 2001; Ernst and Buchan, 2002) has been challenged on the basis that chemical and isotopic compositions of basalts point to shallow-mantle sources including lithospheric ones, enriched by earlier subduction (Bertrand et al., 1982; Bertrand, 1991; Puffer, 2001; Deckart et al., 2005; Verati et al., 2005). However, the origin of the CAMP

GEOLOGY, April 2015; v. 43; no. 4; p. 367–368 | doi:10.1130/focus042015.1

is still strongly debated because basaltic magmas change on their way to the surface.

Hole (2015) processes the composition of parental magmas of the CAMP to recover that of their primary melts, and concludes most CAMP primary magmas point to a T_P between 1400 °C (olivine normative basalts) and 1500 °C (quartz normative basalts). This temperature range is too hot for sublithospheric decompression melting related to lithospheric thinning, and too cold for a source in a hot upwelling mantle.

Continental lithosphere is thicker and relatively more stable than oceanic lithosphere, and thus impedes the removal of heat from the hotter convective mantle. As continents aggregate into supercontinents, they enlarge the wavelength of the convective pattern, further impeding heat loss, thus forcing large-scale warming of the sublithospheric mantle (Grigné et al., 2005). Modulation of mantle temperature below continents depends on the size, number, and distribution of continental plates (Coltice et al., 2007), and thus depends on plate tectonics: continents amalgamate to create supercontinents through the Wilson cycle, then break into smaller continental plates. Two- and three-dimensional numerical studies show that the T_P of the convective mantle can increase 10% to 15% (~130–200 °C) as plates aggregate into a supercontinent covering 20% to 35% of Earth's surface (Coltice et al., 2009), thus the T_P of partial melting following the formation of supercontinents is between that of plate tectonic processes (<1400 °C) and deep mantle upwelling (>1500 °C). The potential temperature determined by Hole from the CAMP's primary magmas using the most advanced model (PRIMELT2), combined with the observation that magmatism followed supercontinent aggregation, confirms that the CAMP was the result of large-scale mantle melting following the aggregation of the last supercontinent (Coltice et al., 2009).

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