Calculating melting temperatures and pressures of peridotite protoliths: Implications for the origin of cratonic mantle

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A B S T R A C T

The old, stable cores of continents – cratons – are underlain by thick and cold mantle keels, composed of melt-depleted and low density peridotite residues. The origins of these thick keels are debated. Were these thick keels formed in situ, by orogenic thickening, or by underplating of buoyant residual mantle? Key to this debate is determining the temperature and pressure at which the protoliths of cratonic peridotites melted (igneous protolith conditions) and comparing to their metamorphic (subsolidus) temperatures and pressures within the keel. This paper presents a method for explicit calculation of the temperatures and pressures at which the peridotite protoliths melted. The approach relies only on the bulk FeO and MgO of residual peridotites. A system of equations consisting of mass balance and new calibrations of Mg peridotite/melt partitioning and melt productivity is then solved simultaneously. The igneous protoliths of abyssal peridotites are found to have melted at effective pressures of 1–2 GPa and temperatures of 1300–1400 °C, within error of the magmatic temperatures and pressures of melt extraction inferred independently from the SiO₂ and MgO contents of mid-ocean ridge basalts. Archean cratonic peridotites, after filtering for the secondary effects of refertilization and orthopyroxene-metasomatism, give igneous protolith pressures and temperatures of 1–5 GPa (30–150 km) and 1400–1750 °C, similar to magmatic temperatures and pressures determined for Archean basalts thought to be representative of the thermal state of the Archean ambient mantle. Most importantly, cratonic peridotite protolith pressures and temperatures are shallower and hotter than their subsolidus equilibration pressures (3–7.5 GPa; 90–200 km) and temperatures (900–1300 °C), which reflects the recent thermal state of the cratonic lithosphere. Specifically, for individual samples with both melting and subsolidus thermobarometric constraints, we find that subsolidus pressures are 1–2 GPa (30–60 km) higher than their igneous protolith pressures although some of the deepest samples experienced minor increases in pressure. Collectively, these results support the suggestion that the building blocks of cratons were generated by hot shallow melting with a mantle potential temperature 200–300 °C warmer than the present. This shallowly generated mantle was subsequently thickened during orogenic episodes, culminating in the formation of a thick, stable craton. Whether such thickening has any modern analogs cannot be answered from this work alone.

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1. Introduction

The Archean cores of continents are underlain by unusually thick and cold mantle keels, which extend to depths of 150–250 km, far exceeding the thickness of oceanic lithospheres (Pollack and Chapman, 1977; Jordan, 1978; Boyd et al., 1985, 1997; Pollack, 1986; Boyd and Mertzman, 1987; Jordan, 1988; Rudnick et al., 1998; Griffin et al., 1999, 2003; Jaupart and Mareschal, 1999; O'Reilly et al., 2001; Gung et al., 2003; Lee, 2006;

Lee et al., 2011). Cold oceanic lithosphere subducts, so the thick cold thermal boundary layers underlying continents should be more convectively unstable. Yet radiogenic isotopic data seem to indicate that they have remained attached to the craton and isolated from the convecting mantle for billions of years (Walker et al., 1989; Carlson and Irving, 1994; Pearson et al., 1995a, 1995b; Carlson et al., 1999, 2005; Chesley et al., 1999; Pearson and Wittig, 2008). This has led to the suggestion that the negative thermal buoyancy of cratonic mantle is largely compensated by an intrinsic compositional buoyancy imparted by low density peridotites formed as the residues of high degrees of melt extraction (Boyd and McAllister, 1976; Jordan, 1978, 1988;

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Boyd, 1989). What is not known is under what conditions these peridotite building blocks melted and how they came to form thick lithospheric keels.

The peridotites that make up most of the lithospheric mantle beneath continents have likely gone through a long and tortuous history (Kelemen et al., 1992; Herzberg, 1993; Griffin et al., 1999, 2003; Xu, 2001; Pearson et al., 2003; Herzberg, 2004; Carlson et al., 2005; Ionov et al., 2005; Lee, 2006; Ionov and Hofmann, 2007; Simon et al., 2007; Pearson and Wittig, 2008; Wittig et al., 2008; Ionov et al., 2010; Lee et al., 2011; Aulbach, 2012; Doucet et al., 2012; Herzberg and Rudnick, 2012). Before being incorporated into the upper thermal boundary layer of the Earth, peridotites from the adiabatically convecting part of the mantle start off hot. Upon decompression, they may melt and become part of the upper thermal boundary layer, eventually cooling to subsolidus temperatures (Lee et al., 2005). Depending on the tectonic environment, these peridotites may also experience changes in pressure. For example, increases in P could occur during subsequent orogenic thickening whereas decreases in P would occur with subsequent surface erosion, lithospheric thinning, or gravitational spreading of the peridotite layer. Understanding the P–T history experienced by a peridotite rock will thus have important implications for understanding how continental lithosphere forms. Does continental lithosphere form by the thickening of oceanic lithospheres, thickening of arc lithospheres, *in situ* melting in a plume head, underplating of buoyant peridotite residues, or by some other process (Helmstaedt and Schulze, 1989; Pearson et al., 1995a; Kelemen et al., 1998; Herzberg, 1999, 2004; Shirey et al., 2002; Griffin et al., 2003; Carlson et al., 2005; Cooper et al., 2006; Lee, 2006; Griffin and O'Reilly, 2007; Pearson and Wittig, 2008; Lee et al., 2011; Aulbach, 2012; Herzberg and Rudnick, 2012)?

There are already many established thermobarometers for subsolidus conditions. These include calibrations of the two pyroxene solvus, Fe–Mg exchange between pyroxene and garnet and between olivine and garnet, Al concentration in orthopyroxene coexisting with garnet, and perhaps less robust, Ni partitioning between garnet and olivine, and Cr partitioning between spinel and olivine (Ellis and Green, 1979; Harley and Green, 1982; Griffin et al., 1989; Brey and Kohler, 1990; Krogh Ravna, 2000; Wan et al., 2008). These thermobarometers give constraints on the last recorded equilibration Ps and Ts of peridotite xenoliths just before they were entrained and brought to the surface via deeply derived volcanic eruptions (Rudnick et al., 1998; Lee et al., 2011). For these reasons, the contemporary (strictly speaking, the time at which the xenoliths were erupted) thermal state of continental lithosphere is generally well constrained for many parts of the continents. By contrast, constraints on the Ps and Ts at which the igneous protoliths of peridotites melted are less robust. The challenge in estimating protolith conditions is that one cannot use the thermobarometers described above because mineral assemblages and chemistries change as P and T change. Thus, it is nearly impossible to retain the signatures of the high Ts associated with melting in the mineral assemblage unless cooling occurs rapidly, which is generally not the case for peridotites entombed within the ancient cores of continents and cooling slowly over billions of years.

Protolith melting conditions must be based on the bulk rock composition and the assumptions that the peridotite's bulk composition reflects what remained after melt was extracted and that the residual peridotite has remained a closed system in terms of mass exchange. Based on this approach, protolith Ts may be reasonably constrained. Ts are estimated from the bulk Fe and Mg contents and the T-dependency of Fe and Mg partitioning between olivine and coexisting melt (Hanson and Langmuir, 1978; Herzberg, 1992, 1999; Pearson et al., 1995b). This may seem intractable given that all thermobarometers require the composition of at least two coexisting phases to be known, but in the case of peridotites, the

melt has long abandoned the peridotite residue so its composition can never be directly known. This problem was cleverly approached by taking advantage of the remarkable property that the partitioning of Fe and Mg between olivines, pyroxenes and melt is relatively insensitive to T (Hanson and Langmuir, 1978).

Current constraints on Ps of melting are more problematic. Several authors have attempted to use bulk rock Al or heavy rare earth element (HREEs) contents, reasoning that the presence of garnet during high P melting would retain Al and the HREEs more so than shallow melting in the absence of garnet (Walter, 1998, 1999, 2003; Canil, 2004; Lee, 2006; Pearson and Wittig, 2008; Lee et al., 2011). A potential complication in this approach is the growing view that many peridotites have experienced subsequent metasomatism (i.e., the assumption of closed system was violated), modifying their Al and HREE contents (as well as Fe and Mg) (Ionov et al., 2005; Simon et al., 2007; Aulbach, 2012). Only the most melt-depleted peridotites may thus be suitable for protolith thermobarometry if metasomatism is an issue. But if so, some of the most melt-depleted peridotites have melted to levels (30–50%) that would far exceed the point of garnet saturation (Bernstein et al., 2007), so low Al and HREE contents of highly melt-depleted peridotites may not be a sensitive indicator of P but rather an indicator of garnet exhaustion.

The alternative method for estimating the melting Ps and Ts of the igneous protoliths of cratonic peridotites is to examine the bulk FeO content of a peridotite (Herzberg, 1999, 2004; Ionov and Hofmann, 2007; Herzberg and Rudnick, 2012). At higher temperatures, Fe is more incompatible in the solid peridotite residuum, so bulk Fe content, as discussed above, can theoretically be used as a measure of T of melting (Pearson et al., 1995a). But given that solidus temperature increases with pressure, bulk Fe content, in conjunction with constraints from other elements, has also been used as a measure of P in several recent studies (Walter, 1998, 1999; Herzberg, 1999, 2004; Ionov and Hofmann, 2007; Herzberg and Rudnick, 2012). These studies, however, rely on graphical means of estimating P and T. Here, we build on the bulk FeO approach and provide a more quantitative means of extracting T and P.

2. Methodology

Estimating T and P requires at least two functions that relate T and P with a measurable compositional variable. One relationship is the elemental partitioning behavior between the peridotite solid residue and the melt, which depends on T and P. A second relationship involves the dependency of melt productivity on T and P, which, to first order, is controlled by the dependence of the solidus and liquidus Ts on P (Fig. 1A). That is, the amount of melt extracted from a peridotite depends on T and P. The intersection of these two equations in P–T space should, in theory, give a unique P–T of melting (Fig. 1B). However, this problem, as noted above, seems intractable because the melt that formed from this peridotite escaped long ago, so it is not possible to directly measure any partitioning relationships.

The above problem can be made tractable as follows. It is well-known that the MgO content of a melt, unlike most other elements, is buffered at a given T and P if the melt is in equilibrium with olivine (Albarède, 1992; Beattie, 1993; Herzberg and Zhang, 1996; Sugawara, 2000; Herzberg and O'Hara, 2002; Putirka, 2005, 2008; Herzberg et al., 2007; Herzberg and Asimow, 2008; Lee et al., 2009). Provided olivine is present, which is always the case during peridotite melting (at least in the uppermost mantle), the MgO content of the melt is buffered even if the composition of the bulk peridotite is not constant. To calibrate this relationship for peridotites, we used experimental melts coexisting with both olivine and orthopyroxene (valid for $P < 8$ GPa), the dominant