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Key Points:

- Slab-derived melt and depleted peridotite can interact to form phlogopite $\pm$ amphibole-bearing peridotite $\pm$ magnesite at cratonic mantle
- Slab-derived melt (3–10 wt.%) and mantle interaction may causally be linked to Mid-Lithospheric Discontinuity (MLD) with V_s reductions $\leq \sim 6\%$
- MLD at greater depths (>100 km) is more likely to be explained by presence of phlogopite ($\pm$ carbonate minerals).

Supporting Information:

- Supporting Information S1

Correspondence to:

S. Saha,
Sriparna.Saha@rice.edu

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High Pressure Phase Relations of a Depleted Peridotite Fluxed by CO_2 - H_2O -Bearing Siliceous Melts and the Origin of Mid-Lithospheric Discontinuity

Sriparna Saha¹ , Rajdeep Dasgupta¹ , and Kyusei Tsuno¹
¹Department of Earth, Environmental and Planetary Sciences, Rice University, Houston, TX, USA

Abstract We present phase equilibria experiments on a depleted peridotite (Mg# 92) fluxed with variable proportions of a slab-derived rhyolitic melt (with 9.4 wt.% H_2O , 5 wt.% CO_2), envisaging an interaction that could occur during formation of continents by imbrication of slabs/accretion of subarc mantles. Experiments were performed with 5 wt.% (*Bulk 2*) and 10 wt.% (*Bulk 1*) melt at 950–1175°C and 2–4 GPa using a piston-cylinder and a multi-anvil apparatus, to test the hypothesis that volatile-bearing mineral-phases produced during craton formation can cause reduction in aggregate shear-wave velocities (V_s) at mid-lithospheric depths beneath continents. In addition to the presence of olivine, orthopyroxene, clinopyroxene, and garnet/spinel, phlogopite (*Bulk 1*: 3–7.6 wt.%; *Bulk 2*: 2.6–5 wt.%) at 2–4 GPa, and amphibole (*Bulk 1*: 3–9 wt.%; *Bulk 2*: 2–6 wt.%) at 2–3 GPa ($\leq 1050^\circ\text{C}$) are also present. Magnesite (*Bulk 1*: ~ 1 wt.% and *Bulk 2*: ~ 0.6 wt.%) is present at 2–4 GPa ($< 1000^\circ\text{C}$ at 3 and $< 1050^\circ\text{C}$ at 4 GPa) and its thermal breakdown coincides with the visual appearance of trace-melt. However, an extremely small fraction of melt is inferred at all experiments based on the knowledge of fluid-saturated peridotite solidus and the difference between bulk H_2O and total H_2O stored in the hydrous phases. Calculated mineral end-member volume-proportions were used to calculate V_s of the resulting assemblage at experimental conditions and along representative continental geotherms (surface heat flow of 40–50 mWm^{-2}). We note that reactive crystallization of phlogopite $\pm$ amphibole by infiltration of 3–10% slab-derived hydrous-silicic melt can cause up to 6% reduction in V_s and that the estimated reduction in V_s increases with increasing melt:rock ratio. The presence of phlogopite limits amphibole-stability, making phlogopite a more likely candidate for MLDs at >100 km depth.

1. Introduction

The long-term stability of continental lithospheric mantle (CLM) in general and cratons in particular, for at least as much as 2–3 billion years, has puzzled scientists for long. Overall, continental mantle is characterized by low heat flow and high seismic velocities compared to oceanic mantle. However, a peculiar feature of cratons that has recently been identified by various high-resolution seismic studies is a substantial reduction in seismic shear wave velocity (V_s) mostly by 2–10% and up to 24% globally at depths of 60–150 km (Abt et al., 2010; Fischer et al., 2010; Sodoudi et al., 2013; Wolbern et al., 2012). Given that thickness of the high velocity mantle keel beneath cratons typically exceeds ~ 200 km (as determined by xenolith thermobarometry, surface heat flux measurements, and surface wave tomography), it is unlikely that the observed velocity reductions can be attributed to the lithosphere-asthenosphere boundary (LAB), which in tectonically active areas occurs at depths ~ 100 km (Fischer et al., 2010; Lee et al., 2011; Romanowicz, 2009). Instead this velocity perturbation (in cratons) is observed at mid lithospheric depths and has been termed as the Mid Lithospheric Discontinuity (MLD) (Abt et al., 2010; Fischer et al., 2010; Rader et al., 2015; Selway et al., 2015; Sodoudi, 2013; Wolbern et al., 2012). MLD occurrences have been documented globally and a range of seismological observations provide evidence for this layer of reduced seismic velocity at shallow depths of CLM. Evidence for its occurrence has been found under different cratons, including: the Canadian Shield (Miller & Eaton, 2010); Australia (Ford et al., 2010); Kalahari craton (Savage & Silver, 2008; Sodoudi, 2013); and the North American craton (Abt et al., 2010; Foster et al., 2014; Hopper et al., 2014). Some studies have even reported existence of multiple MLDs, where the deeper MLD tends to dip beneath the shallower MLD, particularly in zones where Proterozoic terranes have collided with Archean CLM (Hopper et al., 2014; Hopper & Fischer, 2015; Lekik & Fischer, 2014; Sodoudi, 2013).

The origin of MLD is highly debated and different mechanisms have been proposed to explain it, including: (1) the presence of partial melt (Kumar et al., 2012; Thybo & Perchuc, 1997); (2) elastically accommodated grain boundary sliding (EAGBS) (Karato, 2012; Karato et al., 2015); (3) compositional layering i.e., change in Mg# with depth (Yuan & Romanowicz, 2010); (4) change in the seismic anisotropy of minerals (Rychert & Shearer, 2009); and (5) presence of hydrous mineral phases produced through metasomatism (Hacker et al., 2003a, 2003b; Rader et al., 2015; Savage & Silver, 2008; Selway et al., 2015). Although each of the above proposed mechanisms is attractive, a universally applicable explanation for the origin of MLD is lacking at present. For example, velocity reductions associated with compositional layering (depletion) are too low to account for the MLD (Lee, 2003; Selway et al., 2015). Similarly, though anisotropic layering can explain the existence of MLD in certain regions, in general such layering cannot explain velocity reductions associated with MLD globally (Ford et al., 2010; Hopper et al., 2014; Sodoudi et al., 2013; Wirth & Long, 2014). Moreover, though EAGBS has also been proposed as one of the mechanisms that can lower V_s , not only the magnitude of velocity drop associated with it is unclear but also the effect of grain size on velocity reduction remains unresolved (Hopper et al., 2014; Karato et al., 2015; Selway et al., 2015).

Models in which MLDs have been argued to represent the top of a solidified/frozen-in melt layer are well known and have become popular with the recognition that the isobaric nature of the carbonated peridotite solidus over a wide range of temperature can explain the relatively limited depth range in which most MLD observations have been reported (Abt et al., 2010; Dasgupta, 2018; Dasgupta et al., 2013; Hopper & Fischer, 2015; Rader et al., 2015). Subsequently, if this carbonate-rich melt contains dissolved water, it can also interact with the surrounding depleted mantle leading to formation of hydrous accessory phases such as phlogopite or amphibole and can cause reductions in seismic velocities (Hacker et al., 2003a, 2003b; Rader et al., 2015; Selway et al., 2015).

Although the explanation of MLD due to the presence of volatile-bearing partial melts and/or volatile-bearing mineral phases is appealing, this model needs to be evaluated in light of the compositions and modal abundances of the volatile bearing phases that are required to explain the velocity reductions associated with MLD. In particular, if the presence of partial-melt or volatile-bearing (accessory) phases is the cause of MLD (Rader et al., 2015; Selway et al., 2015), it is not clear what process of volatile-addition can lead to their stability and whether or not the multitude of tectonic processes and melt/volatile-addition events can be linked to specific features of MLDs. It remains unclear what mode of volatile-addition leads to the stability of hydrous and/or carbonate minerals as opposed to stability of partial melts.

Though metasomatism as a globally applicable explanation for MLD occurrence may be questionable (Karato et al., 2015), it is important to evaluate the possibility of such volatile metasomatism/infiltration events during cratonization. While such volatile rich layers might have formed in the Archean cratonic lithospheric mantle (or asthenosphere), the possibility of low degree partial melts from the asthenosphere infiltrating into the cratonic lithosphere, adding to the volatile rich layers over time cannot be annulled (Hansen et al., 2015; Hopper & Fischer, 2015; Rader et al., 2015). This might lead one to question, if a causal link exists between MLDs and the origin of cratons. Compositional similarity between continental subarc mantle and cratonic xenoliths favor accretion and orogenic thickening of arcs to be one of the popular models for formation of CLM (Kay et al., 2005; Lee, 2006; Lee et al., 2011). Similar arguments regarding the origin of the CLM have been made by seismic studies as well (Bostock, 1997, 1998; Miller & Eaton, 2010). If this indeed is the dominant mode of origin of CLM, fluids/hydrous melts released from the subducting slab can interact with the depleted mantle peridotitic rocks of the mantle wedge during cratonization. Such metasomatic interactions can refertilize the mantle, as evidenced in some mantle xenoliths that show orthopyroxene enrichment and low Mg/Si ratios (Boyd, 1989; Kelemen et al., 1992, 1998; Pearson & Wittig, 2013). Not only has the occurrence of such volatile-bearing phases (amphibole, phlogopite, and carbonate minerals) been reported in xenoliths derived from the cratonic lithosphere (Lee et al., 2000; Yaxley et al., 1998) but several experimental studies show that presence of H_2O and CO_2 in the metasomatizing fluid/melt phase in subduction environment can stabilize hydrous/carbonate minerals in the overlying mantle wedge (Mallik et al., 2015, 2016; Pirard & Hermann, 2015a, 2015b; Tumietti et al., 2013).

To evaluate the exact depth range over which $CO_2 + H_2O$ -bearing partial melts and/or hydrous or carbonate minerals might be stabilized in a peridotitic environment associated with craton formation, it is important to understand the near solidus phase relations of a depleted peridotite + $CO_2 + H_2O$ at P - T conditions that are relevant for the reported MLD occurrences. Phase relations of depleted mantle peridotite, i.e., high

Mg# ($>\sim 90.5$), low CaO ($<\sim 2$ wt.%), and low alumina ($<\sim 2.5$ wt.%) fluxed by CO_2 - H_2O are however lacking at present. Previous experimental studies on mantle metasomatism or flux melting of peridotites involving volatiles have either considered the effect of one volatile such as CO_2 (Brey et al., 2007; Dasgupta et al., 2007; Dasgupta & Hirschmann, 2006, 2007; Falloon & Green, 1989; Ghosh et al., 2014; Hirose, 1997; 2013; Mallik & Dasgupta, 2013, 2014; Mallik et al., 2015) or H_2O (Balta et al., 2011; Conceicao & Green, 2004; Falloon & Danyushevsky, 2000; Fumagalli et al., 2009; Green, 1973; Grove et al., 2006; Kawamoto & Holloway, 1997; Konzett & Ulmer, 1999; Niida & Green, 1999; Pirard & Hermann, 2015a; Tenner et al., 2012; Till et al., 2012) and the studies involving both volatiles have been carried out on fertile peridotite compositions (Dvir & Kessel, 2017; Olafsson & Eggler, 1983; Thibault et al., 1992; Tumati et al., 2013; Wallace & Green, 1988, 1991) (Figure 1 and supporting information Table S1). The recent models of peridotite melting in the presence of $\text{CO}_2 + \text{H}_2\text{O}$ are also based on fertile peridotite composition (Dasgupta, 2018; Dasgupta et al., 2013). Few studies that have explored the metasomatism of peridotite by slab-derived silicate melts have been done on H_2O -bearing systems, mostly using fertile peridotite compositions, and in the context of mantle wedge melting and subduction zone magmatism (Mallik et al., 2015, 2016; Pirard & Hermann, 2015a, 2015b). A comprehensive study evaluating the influence of mixed CO_2 - H_2O volatiles on a depleted mantle composition is lacking at present.

This study aims to constrain the phase relations of a cratonic peridotite composition (low CaO, high Mg#) fluxed by a volatile-bearing ($\text{CO}_2 + \text{H}_2\text{O}$) silicic melt at P - T conditions relevant to mid depths of the CLM (~ 60 – 150 km) in order to explore the potential link between observed seismic velocity perturbations and mantle mineralogy and potential presence of partial melts, with particular emphasis on the stability and composition of different volatile-bearing phases. The motivation behind exploring the phase relations of depleted peridotite and CO_2 - H_2O -bearing silicic melt mixture is to assess how mantle properties will be affected for craton formation via amalgamation of subduction modified wedge mantles or by slab imbrication. Among various volatile species carriers, hydrous silicic melt is particularly relevant for intermediate to warm subduction zones of ancient past (Dasgupta et al., 2013; Duncan & Dasgupta, 2017; Tsuno & Dasgupta, 2011). The ultimate goal of this study is to constrain the phase compositions and modal proportions of different volatile-bearing phases and apply the results to geophysical and petrologic properties of CLM.

2. Methods

2.1. Starting Material

The depleted peridotite composition (DP) used in this study is the average composition of the global cratonic xenoliths with Mg# 92 (Lee et al., 2011; Luet et al., 2015; Maier et al., 2012; Pearson & Wittig, 2013). The volatile-bearing silicic melt (SM) that has been used as the metasomatic agent is similar to a partial melt composition derived from the average composition of global subducting sediments (GLOSS) obtained at 3.5 GPa and 900°C in a study by Hermann and Spandler (2008). Based on the amount of carbon that can be dissolved in a model slab-derived rhyolitic melt at subarc depths (Duncan & Dasgupta, 2014, 2015; Ghiorso & Gualda, 2015), 5 wt.% CO_2 was added to the partial melt composition along with 9.4 wt.% H_2O to understand the effect of mixed volatiles on the mineral phase stability as a consequence of subduction influence during craton formation. To understand how the proportion of the infiltrating silicic slab-derived melt would affect the modes or composition of the volatile-bearing phases, two different bulk compositions (*Bulk 1*: 90 wt.% DP + 10 wt.% SM and *Bulk 2*: 95 wt.% DP + 5 wt.% SM) were investigated. We chose our bulk compositions such that bulk $\text{H}_2\text{O} < \sim 1$ wt.% so that the experiments are fluid undersaturated or produce very small fraction of excess fluid/hydrous melt near the solidus (Mandler & Grove, 2016; Tsuno & Dasgupta, 2012). Bulk starting compositions, *Bulk 1* and *Bulk 2* along with the composition of the base peridotite (DP) and the volatile-bearing silicic melt (SM) are reported in Table 1. The experimental bulk compositions (*Bulk 1* and *2*) are also compared with natural peridotite xenoliths and previous experimental compositions in Figure 1. It is evident from Figure 1 that bulk compositions in our study plot toward the high Mg# and low CaO, low Al_2O_3 end of the composition space. *Bulk 1* and *Bulk 2* have different K_2O contents, which might affect the stability or abundance of the potassic hydrous phases of our interest.

To prepare the starting mixes, reagent grade oxides and carbonates were first dried by firing SiO_2 , TiO_2 , Al_2O_3 overnight at 1000°C , Fe_2O_3 at 800°C , MnO at 400°C , CaCO_3 at 200°C , and K_2CO_3 and Na_2CO_3 at 110°C . P_2O_5 was kept in a desiccator. For synthesis of the depleted peridotite (DP) composition, various oxides and

Figure 1. (a) CaO, (b) Al_2O_3 , and (c) K_2O contents (in wt %) of bulk starting compositions used in this study (solid yellow squares) plotted as a function of Mg# along with peridotite compositions used in previous experimental studies in volatile bearing systems. Also plotted for reference are the cratonic peridotite xenolith compositions (grey circles) (Lee et al., 2011; Luguét et al., 2015; Maier et al., 2012; Pearson & Wittig, 2013). CO_2 -bearing, nominally dry experiments are represented by solid blue circles (Brey et al., 2007; Dasgupta & Hirschmann, 2006; Falloon & Green, 1989; Ghosh et al., 2014; Hirose, 1997); H_2O -bearing studies are represented in red (Balta et al., 2011; Conceicao & Green, 2004; Falloon & Danyushevsky, 2000; Fumagalli et al., 2009; Fumagalli & Poli, 2005; Green, 1973; Grove et al., 2006; Kawamoto & Holloway, 1997; Konzett & Ulmer, 1999; Mandler & Grove, 2016; Niida & Green, 1999; Parman & Grove, 2004; Pirard & Hermann, 2015a; Tenner et al., 2012; Till et al., 2012) and mixed H_2O and CO_2 -bearing studies are represented in yellow (Dvir & Kessel, 2017; Olafsson & Eggler, 1983; Thibault et al., 1992; Tumiati et al., 2013; Wallace & Green, 1988, 1991).

carbonates were added in predetermined proportion in an agate mortar and ground under ethanol for an hour to homogenize the mix. The mixture was then left overnight and for ethanol to evaporate. The powder was then collected in a clean Au-lined alumina crucible and the mix was decarbonated and reduced at 1000°C in a CO-CO₂ gas mixing furnace for 24 h at logfO₂~ FMQ-2. The rhyolitic sediment melt mix (SM) was prepared in two steps. The first step was similar to the preparation of the DP mix, where the oxides SiO₂, TiO₂, Fe₂O₃, MnO, MgO, P₂O₅ were added in the predetermined proportion, and ground under ethanol for 45 min to homogenize it. After the ethanol evaporated, the mixture was fired at 1000°C in a CO-CO₂ gas mixing furnace for 24 h (at logfO₂~ FMQ-2) in order to reduce Fe³⁺ to Fe²⁺ followed by addition of Na₂CO₃, CaCO₃, K₂CO₃, and Al(OH)₃ so that the desired proportion of these oxides are added to the mix along with 5 wt.% CO₂ and 9.4 wt.% H₂O. The mix was ground again in ethanol for an hour to homogenize it and then left overnight to dry. It was later collected and stored in a glass vial in a 110°C drying oven.

The starting mixes (*Bulk 1* and *Bulk 2*) were then prepared by weighing out predetermined proportions of the depleted peridotite and the rhyolitic sediment melt mix, homogenizing the mixes in an agate mortar under ethanol, and finally drying the mixes by storing in an oven. Homogeneous mixture of siliceous melt and depleted peridotite was prepared to simulate grain scale, porous infiltration of melt in a peridotite matrix similar to previous studies (Mallik & Dasgupta, 2014; Mallik et al., 2016).

2.2. Experimental Procedure

Experiments were performed using a piston cylinder (PC) and an 1100 ton Walker-style multi-anvil (MA) apparatus in the Experimental Petrology Laboratory at Rice University, Houston. The PC experiments were done using a half-inch BaCO₃/MgO pressure media and graphite furnace assembly following the calibration and procedure described in previous studies (Duncan & Dasgupta, 2015; Eguchi & Dasgupta, 2017; Tsuno & Dasgupta, 2011). The MA experiments were conducted using a 18 mm MgO-Al₂O₃-SiO₂ Walker-style castable assembly, following the calibration reported in Ding et al. (2014) and adopted in subsequent studies (Li et al., 2016; Tsuno & Dasgupta, 2015). A Type-C thermocouple oriented axially with respect to the heater and located next to the capsule was used to monitor and control the temperature for both PC and MA experiments. Pressure and temperature uncertainties are estimated to be ± 0.1 GPa, 12°C for PC and ± 0.3 GPa, 10°C for the MA experiments.

Gold capsules (2 mm outer diameter) were used to contain the homogeneous starting mixes in both PC and MA experiments. The packed capsules were welded shut using a graphite arc welder or PUK welding machine. Weight loss due to welding measured for each capsule with *Bulk 1* or *Bulk 2* was ≤ 0.6% and 0.04% relative to the unwelded capsule for the graphite arc welder and the PUK welder, respectively. Experiments using *Bulk 1* were performed at 950°C at 2 GPa, 950–1,175°C at 3 GPa, and 950–1150°C at 4 GPa. As evident from Figure 2, these *P-T* conditions are relevant for the CLM (Boyd et al., 1993, 1997; Carswell et al., 1979; Ehrenberg, 1982; Kopylova et al., 1999; Lee & Rudnick, 1999; Nixon et al., 1981; ; Pearson et al., 1995; Rudnick et al., 1994; Winterburn et al., 1990). In order to study the change in modes and compositions of the volatile-bearing phases as a function of melt:rock ratio, three additional experiments were performed at 950°C at 2 GPa, 1050°C at 3 GPa, and 1100°C at 4 GPa with *Bulk 2*. To constrain the effects of volatile-bearing siliceous melt metasomatism on depleted peridotite, a melt-free experiment with DP was also performed at 3 GPa and 950°C (Table 2). In all experiments, the desired pressure was attained first and then the samples were heated to the temperature of interest at a rate of 100°C/min. Run durations for the experiments varied from 3 to 7 days depending on the temperatures of interest, with lower temperature experiments typically run longer.

The experiments were terminated by cutting off power to the heater. After slow decompression of the assemblies, the locations of the capsules with respect to the thermocouples were verified for every experiment. The recovered capsules were mounted in epoxy, then exposed by grinding using 240–1200 grit silicon carbide paper, and polished on a dry nylon/velvet microcloth using 1–3 μm diamond powder. Repeated vacuum impregnation with low viscosity epoxy was performed to minimize sample loss while polishing. Any liquid lubricant such as water or organic solvents was avoided during polishing to minimize loss of fragile carbonate and hydrous mineral and/or melt phases.

2.3. Analysis of the Run Products

The polished experimental samples were imaged and analyzed using JEOL JXA-8530F HyperProbe Electron Probe Microanalyzers at NASA Johnson Space Center (JSC) and Rice University, Houston, TX. The 3 GPa

Table 1
Starting Compositions (in wt %) Used in This Study

Starting composition	SiO ₂	TiO ₂	Al ₂ O ₃	Cr ₂ O ₃	FeO*	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	CO ₂	H ₂ O	Total	Mg#
^a DP	45.04	0.09	1.17	0.4	6.99	0.12	45.02	0.92	0.07	0.17	0.02	0	0	100	91.98
^b SM	62.93	0.3	12.34	0	0.75	0.03	0.34	1.02	3.22	4.44	0.21	5	9.41	100	44.68
^c Bulk 1	46.83	0.11	2.29	0.36	6.37	0.11	40.55	0.93	0.38	0.6	0.04	0.5	0.94	100	91.9
^d Bulk 2	45.93	0.1	1.73	0.38	6.68	0.12	42.78	0.92	0.23	0.39	0.03	0.25	0.47	100	91.94

Note. Major element compositions of starting mixes used in this study. FeO* indicates all Fe reported as FeO. ^aBase-depleted peridotite (DP). ^bRhyolitic sediment melt mix (SM) is the average composition of partial melt from global subducting sediments derived at 3.5 GPa and 900°C from the experimental study of Hermann and Spandler (2008) with 5 wt % CO₂ added to take into account the amount of dissolved CO₂ slab-derived sediment may contain based on CO₂ solubility experiments of Duncan and Dasgupta (2014, 2015). ^cBulk 1 mixture consists of 90 wt % of DP and 10 wt % of SM. ^dBulk 2 consisting of 95 wt % DP and 5 wt % SM.

experiments for *Bulk 1* were analyzed at JSC, while the 2 and 4 GPa experiments have been analyzed at Rice University, along with all the experiments for *Bulk 2*. Mineral phases were identified using energy dispersive spectroscopy (EDS) and concentrations of major and minor elements were determined using WDS spectroscopy. Melt was identified texturally by its occurrence along mineral grain boundaries and triple junctions. Accelerating voltage of 15 kV and ZAF correction was used for the WDS analyses. Silicate phases such as olivine, orthopyroxene, clinopyroxene, and garnet were analyzed using a 1 μ m diameter beam and 20 nA beam current, whereas phlogopite was analyzed using a 5 μ m wide beam and a beam current of 5 nA. Amphibole was analyzed using a 2–3 μ m wide beam depending on the size of the grain and a beam current of 20 nA. For magnesite, a defocused beam of 2–5 μ m was used. Almandine (Al), biotite (K), chromite (Cr),

Figure 2. The pressure-temperature conditions of our experiments. The solid square symbols represent the experiments with starting composition *Bulk 1*. The solid squares with a * beneath them represent experimental *P-T* conditions with starting composition *Bulk 2*. The dotted grey lines represent conductive geotherms that correspond to different surface heat flow. The green-shaded region marks the *P-T* conditions beneath continents based on xenolith thermobarometry data from Lee et al. (2011). The grey band indicates the depth range where MLD occurrences are reported.

Table 2

Summary of Experiments With Phase Assemblages and Available Phase Proportions as Obtained by Mass Balance of Oxides

Run #	P (GPa)	T (°C)	Run duration (days)	ol	opx	cpx	gt	spinel	amph	Phl	mgs	melt	$\sum r^2$
<i>DP</i>													
B338	3	950	6	48.5(6)	41.5(9)	0.9(7)	9.1(4)	–	–	–	+	–	0.9(1)
<i>Bulk 1</i>													
B405	2	950	6	49.3(5)	35.5(3)	2.4(6)	–	+	6.7(4)	5.1(2)	0.98(4)	–	0.07(1)
B342	3	950	6	50.5(8)	33.1(9)	0.05(2)	+	–	9.4(1)	6.0(3)	1.0(0)	–	0.4(3)
B373	3	975	7	49.6(4)	36.3(7)	0.1(8)	+	–	7.5(1)	5.6(2)	1.0(0)	–	0.14(6)
B352	3	1000	5	45.6(1)	44.7(7)	0.01(4)	+	–	3.6(6)	5.8(2)	–	+	0.05(3)
B353	3	1025	6	49.5(5)	39.4(3)	0.1(9)	+	–	5.3(8)	5.7(8)	–	+	0.2(2)
B363	3	1050	5	50.6(8)	36.4(9)	1.9(8)	+	–	5.9(6)	5.2(9)	–	+	0.04(2)
B402	3	1100	5	47.0(8)	40.1(7)	3.2(6)	2.1(6)	–	–	7.6(3)	–	+	0.07(1)
B379	3	1175	6	41.4(8)	44.1(3)	3.3(6)	0.3(8)	–	–	3.3(7)	–	+	0.23(8)
MA131	4	950	3	44.5(5)	38.9(7)	3.9(4)	5.8(8)	–	–	5.9(1)	1.00(0)	–	0.7(5)
MA136	4	1000	4	48.3(8)	36.8(7)	2.7(4)	5.7(4)	–	–	5.4(4)	1.00(0)	+	0.20(5)
MA124	4	1050	5	47.5(3)	40.0(3)	2.7(1)	4.7(2)	–	–	5.1(1)	0.00(0)	+	0.11(3)
MA119	4	1100	3	51.1(8)	35.3(3)	2.6(3)	5.8(4)	–	–	5.2(4)	–	+	0.5(1)
MA125	4	1150	3	47.2(5)	39.8(7)	3.9(4)	3.7(4)	–	–	5.4(3)	–	+	0.47(6)
<i>Bulk 2</i>													
B413	2	950	6	60.8(6)	27.2(6)	2.89(5)	–	+	6.5(7)	2.6(3)	0.60(5)	–	0.28(7)
B410	3	1050	5	61.9(4)	29.3(4)	4.9(3)	+	–	2.2(5)	3.5(2)	–	+	0.16(4)
MA171	4	1100	5	62.2(8)	26.6(4)	3.7(3)	2.4(3)	–	–	5.0(4)	–	+	0.12(2)

Note. Numbers in the parentheses represent the $\pm 1\sigma$ error determined by propagating the errors in each oxide by Monte Carlo simulations ($n = 20$). For example, 59.8(4) is 59.8 ± 0.4 wt. %; “+” sign indicates the presence of a phase, but due to small size the phase could not be analyzed and therefore not included in the mass balance calculations. Absence of a phase is denoted by “–”. opx: orthopyroxene; cpx: clinopyroxene; phl: phlogopite; ol: olivine; amph: amphibole; mgs: magnesite; gt: garnet; melt: visible presence of partial melt. $\sum r^2$ —sum of residual squares obtained from mass balance of oxides. In order to check the robustness of our mass balance calculations, SEM images (for run # MA 131) were processed using the software ImageJ to estimate phase abundances. The modal abundance for each phase based on textural analysis lies within one standard deviation of values obtained by mass balance of oxides.

diopside (Ca), jadeite (Na), olivine (Si, Fe Mg), rhodonite (Mn), and rutile (Ti) were used as analytical standards. Olivine, diopside, orthopyroxene, almandine, biotite, and dolomite of known compositions were analyzed in each probe session as secondary standards to ensure accuracy of the analytical data.

3. Results

Phase assemblage obtained at various P - T conditions are depicted in Figure 3 and documented in Table 2 along with the experimental conditions and phase proportions (estimated by mass balance on a volatile and melt free basis). Textures of typical experiments are presented in Figure 4. The available major element oxide compositions of the experimental phases are reported in Tables (3–5) and supporting information Tables S2–S5.

3.1. Approach to Equilibrium

Reversal experiments have not been performed in this study but several lines of reasoning can be used to assess maintenance of closed system and approach toward thermodynamic equilibrium in our experiments. (1) Given that trace melt was observed only in a subset of our experiments, all experiments were mass balanced on a melt-free basis. Sum of squared residuals for all oxides using the average mineral compositions and the bulk compositions vary between 0.04 and 0.7, which suggests maintenance of closed system during the experiments. (2) The duration of our experiments ranges from 3 to 7 days, which is also comparable to the duration of previous experiments performed at similar temperature range and with similarly volatile-rich peridotite bulk compositions, where approach to equilibrium was demonstrated (Mandler & Grove, 2016). (3) Minerals produced in our experiments are compositionally uniform from core to rim and no evidence of chemical zoning is found. This approach to establish chemical equilibrium was adopted in previous studies on melting of volatile-bearing peridotite systems (Condamine et al., 2016; Condamine &

Médard, 2014; Mallik et al., 2015; Parman & Grove, 2004). (4) The difference between the nominal temperatures at which the experiments were performed and the corresponding temperatures calculated using the two-pyroxene thermometer of Brey and Kohler (1990) for most of the experiments in this study vary from 13 to 119°C, indicating approach to equilibrium as demonstrated in a study by Mallik et al. (2015). The experiments that show higher deviation from the nominally measured temperature (between 140 and 205°C), when mass balanced, show low sum of squared residuals ($\Sigma r^2 = 0.07\text{--}0.9$) (Table 2). (5) Finally, the partition coefficients of K between amphibole and phlogopite in our study range between 0.14 and 0.27. These values are similar to those obtained in previous experimental studies (0.08–1.23) (Condamine & Médard, 2014; Fumagalli et al., 2009; Konzett & Ulmer, 1999; Mallik et al., 2015; Mengel & Green, 1989; Pirard & Hermann, 2015a).

3.2. Phase Assemblage and Texture

Mineral proportions calculated by mass balance of oxides and are reported in Table 2. Comparison with modal abundance obtained from SEM image analysis (for Run# MA 131) shows that both methods yield identical results within error. This ensures the robustness of the mass balance calculations. The experiment performed with DP at 3 GPa and 950°C resulted in an assemblage of olivine (~49 wt.%), orthopyroxene (~41 wt.%), clinopyroxene (~1 wt.%), and garnet (~9 wt.%). Volatile-bearing melt added experiments, on the other hand, resulted in an assemblage of olivine + orthopyroxene + clinopyroxene + spinel (at 2 GPa)/ garnet (at 3 GPa) + phlogopite + amphibole (2 and 3 GPa and $\leq 1050^\circ\text{C}$) $\pm$ magnesite $\pm$ melt (Table 2). As evident from Figure 3, the magnesite-out boundary is between 975 and 1000°C at 3 GPa and between 1050 and 1100°C at 4 GPa for *Bulk 1*. The visibly clear appearance of melt coincided with the magnesite-out

Figure 3. Experimental phase assemblage observed in this study for starting compositions *Bulk 1* (polygons) and *Bulk 2* (polygons marked by * beneath them) in pressure-temperature space. Each sector of the polygon represents a phase. The absence of a phase is marked by white sector, while the presence of a phase is marked by grey sector. Orthopyroxene, clinopyroxene, olivine, and phlogopite are present in all experiments. Amphibole is absent in the 4 GPa runs and above 1,050°C at 3 GPa. The magnesite-out boundary coincides with the first visual appearance of partial melts. The observed phase assemblages for *Bulk 1* and *Bulk 2* are identical.

Figure 4. Back scattered electron images of experiments performed in this study with the 90:10 (*Bulk 1*) and 95:5 (*Bulk 2*) mixture of depleted peridotite and hydrous rhyolitic partial melt at 2, 3 and 4 GPa. (a) B352 (3 GPa, 1,000°C): Grains of amphibole surrounded by a matrix of olivine grains; (b) MA136 (4 GPa, 1,000°C): close association of garnet with olivine and accicular orthopyroxene grains; (c) B413 (2 GPa, 950°C): Grains of olivine, orthopyroxene, and clinopyroxene distributed homogenously throughout the capsule; (d) B405 (2 GPa, 950°C): Magnesite along with phlogopite and orthopyroxene in experiment with no obvious presence of partial melt; (e) B410 (3 GPa, 1,050°C): Phlogopite and small grains of garnet surrounded by olivine and interstitial orthopyroxene, with traces of melt along the grain boundaries; (f) MA 171 (4 GPa, 1,050°C): traces of melt along grain boundaries between olivine grains.

boundary. Although textural detection of melt was only possible at magnesite-absent conditions, extremely small fraction of melt is inferred at all experimental conditions based on the expected location of the fluid-saturated peridotite solidus and deficit of H₂O between the bulk compositions and those stored in hydrous phases. For the identical *P-T* condition at which DP was run, *Bulk 1* resulted in ~50.5 wt.% olivine, ~33 wt.% orthopyroxene, trace quantities of cpx and garnet, ~6 wt.% phlogopite, ~9.5 wt.% amphibole, and ~1 wt.% magnesite (Table 2). At 2 GPa and 950°C, *Bulk 1* yielded an assemblage of olivine (~49.3 wt.%), orthopyroxene (~35.3 wt.%), clinopyroxene (~2.4 wt.%), amphibole (~6.6 wt.%), and phlogopite (~5.1 wt.%) with trace quantities of magnesite (~0.98 wt.%) and spinel, whereas at the same condition, *Bulk 2* resulted in an

Table 3
Composition of Amphibole as Analyzed Using EPMA

Run# <i>n</i>	B405 11	B342 7	B373 7	B352 5	B353 4	B363 8	B413 5	B410 9
SiO ₂	48.1(6)	47.7(6)	47.8(5)	47.7(5)	47.5(6)	47(1)	47.5(9)	48.8(5)
TiO ₂	0.61(5)	0.54(5)	0.54(7)	0.53(5)	1.5(1)	1.15(5)	0.9(4)	0.70(1)
Al ₂ O ₃	10.3(2)	11.3(2)	11.3(2)	11.6(3)	11.9(4)	12.9(4)	11.4(9)	10.0(6)
Cr ₂ O ₃	1.1(5)	1.1(5)	1.1(2)	1.4(1)	1.7(1)	1.4(2)	1.5(4)	0.6(3)
FeO	3.5(2)	3.1(2)	3.1(1)	3.0(1)	3.3(1)	3.2(2)	3.2(4)	3.1(3)
MnO	0.06(1)	0.06(1)	0.06(2)	0.08(2)	0.08(2)	0.06(2)	0.06(2)	0.04(2)
MgO	20.5(3)	19.2(3)	19.2(3)	19.3(4)	18.1(7)	18.8(5)	19.4(6)	21.5(8)
CaO	8.5(3)	9.11(3)	9.2(4)	8.2(4)	7.6(4)	7.2(1)	8.3(8)	9.0(4)
Na ₂ O	3.9(3)	3.96(3)	4.0(2)	4.44(5)	4.4(3)	4.4(7)	4.1(4)	3.4(1)
K ₂ O	0.71(9)	1.00(9)	1.0(1)	1.13(1)	1.20(8)	1.20(1)	1.0(3)	0.64(5)
P ₂ O ₅								
H ₂ O	2.78(1)	2.78(1)	2.75(1)	2.68(1)	2.73(3)	2.7(4)	2.74(2)	2.75(1)
Sum	100.0(1)	100.0(2)	100.0(4)	100(2)	100(1)	100.0(1)	100.0(9)	100.0(4)
Mg#	91.3(7)	91.3(7)	91.7(3)	92.1(2)	90.8(4)	91.3(4)	91.57(1)	92.5(5)

Note. *n* represents the number of points considered from EPMA analyses to calculate the mean and standard deviation. Mg# = molar MgO/(MgO + FeO*) × 100 and FeO* implies that all iron is assumed to be present as FeO; numbers in the parentheses represent the ±1σ error are based on replicate microprobe analyses and are reported in the last digit cited; for example, 48.1(6) should be read as 48.1 ± 0.6 wt %. H₂O content is calculated by stoichiometry.

assemblage of olivine (~60.8 wt.%), orthopyroxene (~27.2 wt.%), clinopyroxene (~2.89 wt.%), amphibole (~6.5 wt.%), phlogopite (~2.6 wt.%), magnesite (~0.6 wt.%), and trace amount of spinel. No major variation in the modal proportions in the metasomatized peridotite is observed with temperature at either 3 or 4 GPa for each bulk composition. However, the abundance of mineral phases show systematic change as a function of melt:peridotite ratio, with greater abundance of olivine and lesser proportion of orthopyroxene, phlogopite, amphibole, and phlogopite in the experiments with lower melt:peridotite ratio (Table 2).

Olivine exists as euhedral to subhedral grains and is closely associated with amphibole, where it is present (Figure 4a). At 4 GPa, grains of garnet are euhedral/subhedral and are associated with olivine and orthopyroxene (Figure 4b), while in some of the 3 GPa experiments garnets are <1 micron in diameter. Nominally melt-free experiments, across all investigated pressures, are those that contain magnesite (950°C at 2 GPa, ≤1000°C at 3 GPa, and <1050°C at 4GPa), and show lack of mineral gradation along the length of the capsule (Figures 4c and 4d). The experiments at 1050–1175°C at 3 GPa and 1050–1150°C at 4 GPa exhibit gradation in mineralogy across the length of the capsule. In these experiments, obvious presence of trace melt is observed along grain boundaries and as tiny pockets in triple junctions of mineral grains (Figures 4e and 4f). Intergranular residual orthopyroxene is present between larger orthopyroxene grains toward the top of the capsule (hot end) with the residual orthopyroxene more intimately associated with phlogopite and clinopyroxene towards the bottom of the capsule, which is known to be the cold end in our assembly (Tsun & Dasgupta, 2011). The presence of trace amount of magnesite is observed in the base depleted peridotite composition experiment (Run# B338), which can be attributed to carbon contamination due to use of the graphite arc welder during capsule preparation for this experiment (Mallik et al., 2015).

3.3. Major Element Compositions of Phases

This section describes the compositional estimates of different phases observed in our experiments—olivine, orthopyroxene, clinopyroxene, garnet, amphibole, phlogopite, magnesite, and melt.

3.3.1. Olivine

Olivine is forsteritic (with 7–8% fayalite component; supporting information Table S2). Forsterite (Fo) content of olivine in the depleted base peridotite (Run #B338) is 92.8. Addition of melt slightly lowers the Fo content of olivine in *Bulk 1* to 92.5 at 2 GPa, while at 3 GPa, it ranges from 91.6 to 92.0, and from 92.1 to 92.6 at 4 GPa. No systematic variation in FeO* (where all Fe is expressed as FeO) and MgO content of olivine with temperature or pressure is observed. In *Bulk 2*, (with 5 wt.% melt added), olivine Fo ranges from 91.6 to 91.8 and FeO* contents are higher (8.0–8.2 wt.%) compared to the depleted peridotite.

Table 4
Composition of Phlogopite as Analyzed Using EPMA

Run# <i>n</i>	B405 8	B342 11	B373 6	B352 16	B353 14	B363 3	B402 5	B379 7	MA131 17	MA136 5	MA124 8	MA119 12	MA125 10	B413 7	B410 21	MA171 10
SiO ₂	44.1(8)	41.4(9)	40.5(4)	41.0(9)	40.3(9)	40.3(8)	39.3(3)	40.1(3)	41.2(9)	42.2(5)	40.6(5)	42.2(4)	42.3(6)	43.1(7)	40.6(7)	42.9(4)
TiO ₂	1.3(1)	0.6(1)	0.74(4)	0.7(2)	1.4(3)	1.27(6)	0.7(1)	0.88(5)	0.47(7)	0.66(6)	0.74(5)	0.56(5)	0.82(6)	0.95(4)	1.1(2)	0.8(3)
Al ₂ O ₃	14.7(2)	15.6(8)	16.1(4)	15.5(1)	15.7(6)	15.4(1)	14.1(6)	15.6(6)	14.1(9)	12.9(6)	13.6(3)	13.6(4)	13.9(6)	16.8(1)	17.2(9)	14.2(6)
Cr ₂ O ₃	0.51(4)	0.6(3)	0.8(2)	1.0(2)	0.9(4)	1.4(7)	1.1(3)	1.8(1)	0.7(2)	0.56(1)	0.8(3)	0.94(5)	0.6(2)	0.34(8)	0.6(2)	0.7(3)
FeO	3.4(3)	3.3(1)	3.04(5)	3.3(2)	3.6(3)	3.5(2)	4.0(1)	3.1(1)	3.0(4)	2.9(1)	2.9(1)	2.4(1)	2.95(9)	3.41(8)	3.8(2)	2.8(5)
MnO	0.01(2)	0.02(2)	0.02(2)	0.02(2)	0.02(2)	0.02(3)	0.05(4)	0.02(2)	0.03(3)	0.03(2)	0.03(1)	0.03(3)	0.02(2)	0.01(1)	0.02(2)	0.04(3)
MgO	21.3(1)	24.7(6)	24.9(4)	24.5(9)	23.9(5)	23.9(2)	26.6(5)	23.8(5)	26.0(7)	24.8(4)	26.6(6)	25.5(4)	25.0(3)	20.0(5)	22.5(5)	24.4(4)
CaO	0.02(2)	0.03(2)	0.02(2)	0.05(7)	0.02(1)	0.02(1)	0.2(3)	0.01(1)	0.1(3)	0.03(2)	0.01(1)	0.02(3)	0.05(6)	0.03(3)	0.03(3)	0.01(1)
Na ₂ O	0.21(3)	1.0(2)	1.1(1)	0.9(1)	1.0(2)	0.7(2)	0.7(3)	0.58(5)	0.7(2)	1.4(4)	0.7(3)	0.6(1)	0.5(1)	1.1(1)	0.7(3)	0.50(9)
K ₂ O	10.3(5)	8.3(3)	8.34(9)	8.5(6)	8.9(4)	9.1(1)	9.03(9)	9.6(1)	9.3(5)	10.2(3)	9.6(2)	9.6(2)	9.5(3)	9.9(4)	9.1(8)	9.4(4)
P ₂ O ₅		0.01(1)	0.05(3)	0.02(3)	0.02(3)						0.02(3)	0.01(2)	0.02(4)			
H ₂ O	4.16(2)	4.38(1)	4.42(1)	4.47(2)	4.34(1)	4.37(1)	4.26(2)	4.39(2)	4.44(1)	4.36(1)	4.38(1)	4.40(1)	4.34(1)	4.35(1)	4.45(1)	4.21(1)
Sum	100.0(1)	100.0(2)	100.0(4)	100(2)	100(1)	100.0(1)	100.0(9)	100.0(4)	100.0(1)	100(1)	100.0(5)	100.0(7)	100.0(7)	100.0(5)	100.0(7)	100(1)
Mg#	91.82(1)	92.9(2)	93.50(3)	92.8(4)	92.2(5)	92.4(4)	92.37(1)	93.3(2)	93.8(9)	93.9(3)	94.1(3)	94.5(2)	93.7(2)	91.29(1)	91.30(6)	93.92(1)

Note. *n* represents the number of points considered from EPMA analyses to calculate the mean and standard deviation. Mg# = molar MgO/(MgO + FeO*) × 100 and FeO* implies that all iron is assumed to be present as FeO; numbers in the parentheses represent the $\pm 1\sigma$ error are based on replicate microprobe analyses and are reported in the last digit cited; for example, 44.1(8) should be read as 44.1 $\pm$ 0.8 wt.%. H₂O content is calculated by stoichiometry.

3.3.2. Orthopyroxene

Orthopyroxene compositions in our study are enstatitic with Mg# $\geq$ 90 (supporting information Table S3). In the depleted base peridotite, Mg# of orthopyroxene is 90.9, which is low compared to Mg# of orthopyroxene in *Bulk 1* and 2. Mg# of orthopyroxene in *Bulk 1*, increases with pressure (92.3 at 2 GPa, varies from 91.9 to 92.9 at 3 GPa and 92.9 to 93.5 at 4 GPa). At a given *P-T* condition, the proportion of enstatite in *Bulk 1* is higher (92.1–92.4 wt.% at 3 GPa) when compared to the depleted peridotite (~90.9 wt.%). No systematic variation in enstatite content of orthopyroxene with temperature or pressure is observed. For *Bulk 1*, Al₂O₃ content in orthopyroxene is ~2.2 wt.% at 2 GPa, varies from 0.93 to 3.8 wt.% at 3 GPa, and 0.7 to 1.8 wt.% at 4 GPa where garnet mode increases. Mg# of orthopyroxene in *Bulk 2* varies from 91.3 to 92.3, with Al₂O₃ content ~1.7–2.1 wt.%.

Table 5
Composition of Magnesite as Analyzed Using EPMA

Magnesite <i>n</i>	B405 5	B342 10	B373 8	MA131 11	MA136 8	B413 6
SiO ₂	0.06(3)	0.2(2)	0.4(6)	0.3(3)	0.2(2)	0(0)
TiO ₂	0.03(6)	0.02(4)	0.02(4)	0.01(2)	0.004(6)	0.02(3)
Al ₂ O ₃	0(0)	0.09(8)	0(0)	0.2(2)	0.1(2)	0.5(9)
Cr ₂ O ₃	0.01(8)	0.01(3)	0.06(8)		0.04(5)	0.03(2)
FeO*	2.6(1)	4.3(7)	2.9(8)	4.4(8)	3.0(9)	3.5(9)
MnO	0.05(5)	0.16(6)	0.10(3)	0.16(6)	0.13(5)	0.12(5)
MgO	44.4(3)	43.5(9)	40(0)	40(0)	40(0)	40(0)
CaO	0.7(1)	1.5(5)	1.0(6)	1.5(3)	1.2(4)	1.2(3)
Na ₂ O	0.01(2)	0.04(3)	0.04(3)	0.03(3)	0.01(2)	0.03(4)
K ₂ O	0.01(1)	0.03(2)	0.03(2)	0.03(2)	0.02(3)	0.1(1)
P ₂ O ₅	0.1(1)	0.06(8)	0.06(8)	0.08(7)	0.03(5)	0.07(8)
CO ₂	51.3(9)	50.1(7)	50.0(0)	50.0(3)	51.1(6)	49.9(7)
sum	100.0(3)	100.0(6)	100.0(9)	100.0(1)	100.0(7)	100.0(8)
Mg#	96.8(1)	94.7(8)	96.4(7)	94.6(2)	96.38	95.3(2)

Note. *n* represents the number of points considered from EPMA analyses to calculate the mean and standard deviation. Mg# = molar MgO/(MgO + FeO*) × 100 and FeO* implies that all iron is assumed to be present as FeO; numbers in the parentheses represent the $\pm 1\sigma$ error are based on replicate microprobe analyses and are reported in the last digit cited; for example, 41.8(6) should be read as 41.8 $\pm$ 0.6 wt.%. CO₂ content is calculated by stoichiometry.

3.3.3. Clinopyroxene

Clinopyroxene is magnesian and diopside-rich in all experiments (supporting information Table S4). Mg# of clinopyroxene in the depleted base peridotite is 89.7 with 1.5 wt.% Al_2O_3 and 21 wt.% CaO. In *Bulk 1*, Mg# is ~ 90.9 at 2 GPa, varies from ~ 89 to 91 at 3 GPa, and ~ 91 to 92 at 4 GPa. Al_2O_3 contents of clinopyroxenes are as high as ~ 13 wt.% (at 1000°C), which drop to ~ 4 wt.% (1025°C); At 4 GPa, in presence of abundant garnet Al_2O_3 contents are lower, ranging between 2.9 and 3.1 wt. %. In *Bulk 2*, Mg# of clinopyroxene varies between 91 and 92. At a given *P-T* condition, Al_2O_3 contents are higher in *Bulk 2* (when compared to *Bulk 1* for 2–3 GPa) while at 4 GPa, it is comparable in both *Bulk 1* and 2 (~ 2.8 wt.%).

3.3.4. Garnet

Garnet is pyrope-rich (~ 71 to 72%; supporting information Table S5), with variable proportion of almandine (~ 12 –13%) and grossular (~ 14 –15%). Mg# of garnets in the base depleted peridotite is low (~ 80.5) compared to the melt added experiments; in *Bulk 1*, Mg# ~ 82.5 at 3 GPa (garnet in several 3 GPa runs could not be analyzed, due to their small grain size), and varies from ~ 84 to 85 at 4 GPa. CaO content varies from 4.8 to 5.9 wt.% and is lower compared to the depleted base peridotite (~ 9.1 wt.%). The 4 GPa experiment with *Bulk 2* shows similar garnet composition as with *Bulk 1* at same pressure.

3.3.5. Amphibole

Following the classification scheme of Leake et al. (1997, 2003) for calcic amphiboles ($\text{Ca}_B \geq 1.50$; $(\text{Na} + \text{K})_A \geq 0.50$, where A and B are cation sites), amphibole compositions in our experiments are pargasitic (Table 3), i. e., Si ~ 6.0 to 6.5 and $[\text{Mg}/(\text{Mg} + \text{Fe})] \sim 0.89$ –0.92 (where Si, Mg, and Fe are expressed in terms of atoms per formula unit based on 23 oxygens). Mg# of amphibole varies from 91.3 to 92.1 in *Bulk 1* and 91.5 to 92.5 in *Bulk 2* (Table 3) and do not show much compositional variability.

3.3.6. Phlogopite

Phlogopite has variable proportion of annite end-member (Table 4). In *Bulk 1*, Mg# varies between 92.9 and 93.5 at 3 GPa (with 6.7 to 7.6 wt.% of annite). At 4 GPa, Mg# is higher, and ranges from 93.8 to 94.5 (with 5.5 to 6.3 wt.% annite). Al_2O_3 content decreases with increasing pressure (from 15.6–16.1 wt.% at 3 GPa to 12.9–14 wt.% at 4 GPa). At a given *P-T* condition, Mg# of phlogopite crystallized in *Bulk 2*, is higher compared to *Bulk 1* and has lower Al_2O_3 contents.

3.3.7. Magnesite

Mg# of magnesite for *Bulk 1* is 96.8 at 2 GPa and varies from 94 to 96 at both 3 and 4 GPa, with Mg# increasing with increasing temperature at each pressure. At the same *P-T* condition (2 GPa, 950°C), Mg# of magnesite_{ss} produced in *Bulk 2*, is lower compared to that produced from *Bulk 1* (Table 5).

3.3.8. Melt

Because of the presence along grain boundaries and in very small pools at triple junctions of the grains ($< 1 \mu\text{m}$), the melt could not be analyzed using WDS. However, based on our textural observation that there is breakdown of magnesite after melt visually appears, we can infer that the melt could be hydrous-carbonatitic or at least likely has high dissolved CO_2 content (discussed further in section 4.4).

4. Discussion

4.1. Melt-Rock Reaction and Reactive Recrystallization

Though mass balance calculations can provide an insight on how mineral abundances vary as a function of change in *P-T* conditions for a given bulk composition, to constrain the effect of metasomatic interaction between the depleted peridotite and the silicic melt, phase proportions for systems with and without melt added must be known. The following reaction captures how mineralogy of a depleted peridotite changes owing to infiltration by 10 wt.% of the slab-derived hydrous and CO_2 -bearing silicic melt at the 3 GPa and 950°C, for a unit gram of melt introduction

where mineral abbreviations are the same as in Table 2.

Equation (1) quantifies the amount of volatile-bearing phases that would be produced as a function of unit mass of melt consumption. This is vital in reconciling modal abundance of these hydrous phases in light of permissible chemistry of the cratonic mantle. Moreover, the mechanism of volatile addition could also influence mineral stability, as is evidenced by variable stability of amphibole and phlogopite in ultramafic lithologies of various fertility and alkali/ H_2O ratios (Mallik et al., 2016; Mandler & Grove, 2016). In the following

section, we thus compare the stability of hydrous/carbonate phases from this study with previous experimental studies to understand how phase stability can be linked to the observed geophysical characteristics of cratons.

4.2. Stability of Hydrous Silicate and Carbonate Mineral Phases: A Comparison With the Previous Studies

4.2.1. Stability of Carbonates

Magnesite is the stable carbonate phase in the bulk compositions investigated in this study at all pressures and its breakdown coincides with the textural appearance of trace partial melt. The chief difference between this study and previous experiments on peridotite + $\text{CO}_2 \pm \text{H}_2\text{O}$ systems is that for all previous experiments on fertile ($\text{Mg\#} < 90.5$) systems, magnesite solid solution either appears at pressures ≥ 2.7 GPa with dolomite being the stable carbonate mineral phase at lower pressures at the solidus (Dasgupta & Hirschmann, 2007; Falloon & Green, 1989; Wallace & Green, 1988) or dolomite solid solution coexisting with magnesite to 2.5 GPa (Tumiati et al., 2013). In contrast, for higher $\text{Mg\#}$ (~ 92) and low-calcic depleted peridotites investigated here, magnesite is the only stable carbonate mineral phase near the solidus of peridotite at pressure at least as low as 2 GPa (Figure 5). Stabilization of magnesite to lower pressures (~ 2 GPa) in low-Ca, high $\text{Mg\#}$ systems investigated in this study suggests that it is the more relevant carbonate end-member in the cratonic mantle. Future studies will need to investigate whether the well-known carbonated peridotite solidus "ledge" (Falloon & Green, 1989) shifts in location for low-calcic, depleted peridotite compositions relevant for CLM because the first carbonate phase to appear may be magnesite rather than dolomite as observed in previous experiments with relatively higher CaO and lower $\text{Mg\#}$.

The magnesite-out boundary in our study is encountered between 975 and 1000°C at 3 GPa and between 1000 and 1050°C at 4 GPa. This is higher than the magnesite-out boundary of $\sim 930^\circ\text{C}$ at 3 GPa observed in Wallace and Green (1988) (Figure 5) that employed a peridotite + $\text{CO}_2 + \text{H}_2\text{O}$ composition with $\text{Mg\#}$ of 87.6, despite the fact that the bulk H_2O content of our experiment are higher than ~ 0.3 wt.% bulk H_2O of Wallace and Green (1988). This highlights the potential importance of depleted peridotite composition in expanding the stability field of magnesite. Olafsson and Eggler (1983), however, bracketed the carbonate-out boundary of their peridotite + CO_2 (~ 0.7 wt.%) + H_2O (~ 0.3 wt.%) composition at $\sim 1100^\circ\text{C}$ at 3 GPa, which is higher than that observed in this study. The reason for the higher carbonate-out boundary in the Olafsson and Eggler (1983) study, however, cannot be assessed given that study provided no information about the type of carbonate being present in their experiments. Comparison with magnesite-out boundary from Dasgupta and Hirschmann (2006) and Falloon and Green (1989) reveals that magnesite out boundary in nominally water-free peridotite + CO_2 systems are higher by as much as $\sim 100^\circ\text{C}$ relative to that observed in peridotite + $\text{CO}_2 + \text{H}_2\text{O}$ -bearing systems of the present study (Figure 5). Because cratonic mantle metasomatism by a CO_2 - H_2O -bearing rhyolitic/kimberlitic/carbonatitic melt would result in coexistence of carbonates and hydrous phases, the presence of magnesite would be restricted to shallower depths compared to nominally dry systems. On the other hand, carbonated partial melts would become stable at shallower depths (lower temperatures) in hydrous systems compared to that in nominally dry systems and therefore a carbonated melt rather than a mineral carbonate may become more relevant for MLD occurrences at shallower (< 100 km) depths.

4.2.2. Stability of Amphibole

The upper stability limit of amphibole in ultramafic lithologies of variable fertility has been widely investigated and is shown in Figure 6 (Fumagalli et al., 2009; Green, 1973; Grove et al., 2006; Mandler & Grove, 2016; Mengel & Green, 1989; Millhollen et al., 1974; Mysen & Boettcher, 1975; Niida & Green, 1999; Olafsson & Eggler, 1983; Tumiati et al., 2013; Wallace & Green, 1991). The upper stability limit of amphibole is controlled by compositional factors such as bulk alkali and H_2O contents and can make the pressure stability field of amphibole vary by as much as ~ 2 GPa (Dasgupta, 2018; Fumagalli et al., 2009; Mandler & Grove, 2016; Niida & Green, 1999).

However, amphibole stability cannot be constrained completely by these compositional variables and if the flux of Na and H_2O are coupled with that of K, amphibole can coexist with phlogopite, similar to what is observed in our experiments. This would mean that in potassic systems, not all the alkali would get incorporated in amphibole and therefore the correlation of alkali content in amphibole and its pressure stability may no longer hold. Indeed, the empirical barometer proposed by Mandler and Grove (2016) based on the alkali contents of amphibole does not reproduce the experimental pressures of our experiments, within the

Figure 5. *P-T* space showing dolomite-magnesite transition or magnesite-in boundary (at higher-*P*) from different studies on peridotite + CO₂ ± H₂O systems (dotted lines) compared to the stability of magnesite observed in our study (blue stars) (B83: Brey et al. (1983); WG83: Wallace and Green (1988); FG89: Falloon and Green (1989); DH06: Dasgupta and Hirschmann (2006); Tm13: Tumiati et al. (2013)). Brey et al. (1983) delineate the reaction enstatite + dolomite (low-*P*) = diopside + magnesite (high-*P*). Carbonated peridotite solidi from previous experimental studies (grey solid lines) are also plotted (WG88: Wallace and Green (1988); FG89: Falloon and Green (1990); DH06: Dasgupta and Hirschmann (2006); DH13: Dasgupta and Hirschmann (2013); Tm13: Tumiati et al. (2013) compared with the magnesite-out boundary from our study. Magnesite is present in our study at pressures as low as 2 GPa and dolomite is absent. This is in contrast with previous studies in more calcic, fertile peridotite systems, where dolomite solid solution continues to be the stable carbonate phase at pressure as high as 3 GPa at the solidus. The green field in the background represents the field of cratonic peridotite xenoliths from Figure 2. The histogram on the right side shows the frequency distribution of MLD occurrences in cratons as a function of depth (Rader et al., 2015).

error of the calibration. This is because of the presence of phlogopite, which is expected to affect the correlation between pressure and amphibole alkali content, as already noted by Mandler and Grove (2016). Otherwise, the amphibole-out boundary bracketed in our study (Figure 6) is broadly consistent with the previous studies that have bracketed amphibole stability of peridotite systems, with the boundary being close to those expected for bulk compositions that are either H₂O-fluid undersaturated or with very little excess fluid (bulk H₂O < 1 wt.%).

4.2.3. Stability of Phlogopite

Phlogopite stability in broadly peridotitic bulk compositions has been investigated in numerous previous experiments, that show that it is stable up-to depths as high as 150–200 km and temperature as high as ~1400°C at 2–4 GPa (Allen et al., 1972; Barton & Hamilton, 1979; Conceicao & Green, 2004; Condamine & Médard, 2014; Elkins-Tanton & Grove, 2003; Enggist et al., 2012; Enggist & Luth, 2010; Esperança & Holloway, 1987; Konzett & Ulmer, 1999; Kushiro, 1969; Kushiro et al., 1967; Mallik et al., 2015; Mitchell, 1995; Modreski & Boettcher, 1972; Sato et al., 1997; Yoder & Kushiro, 1969) (Figure 6). The stability of phlogopite in our experiments at all *P-T* conditions is consistent with previous experimental studies and can potentially be more relevant for MLD occurrences that have been reported at >100 km depths (Ford et al., 2010; Foster et al., 2014; Savage & Silver, 2008; Sodoudi, 2013; Wittlinger & Farra, 2007).

4.3. Comparison of Volatile-Bearing Mineral Compositions With Previous Peridotite Experiments and Natural Peridotite Xenoliths

In order to reliably relate the petrologic constraints obtained from our experiments to seismological observations reported at MLD depths, it is necessary to evaluate how compositions of the volatile-bearing phases

Figure 6. Upper stability limits of amphibole (Fumagalli et al., 2009; Green, 1973; Grove et al., 2006; Mandler & Grove, 2016; Mengel & Green, 1989; Millhollen et al., 1974; Mysen & Boettcher, 1975; Niida & Green, 1999; Olafsson & Eggler, 1983; Tumati et al., 2013; Wallace & Green, 1991) and phlogopite (Allen et al., 1972; Barton & Hamilton, 1979; Conceicao & Green, 2004; Condamine & Médard, 2014; Elkins-Tanton & Grove, 2003; Enggist et al., 2012; Enggist & Luth, 2010; Esperança & Holloway, 1987; Konzett & Ulmer, 1999; Kushiro, 1969; Kushiro et al., 1967; Mallik et al., 2015; Mitchell, 1995; Modreski & Boettcher, 1972; Sato et al., 1997; Yoder & Kushiro, 1969) from previous studies depicted as the red and purple stippled band, respectively. The width of the stability limit bands results from stability limit variation of the hydrous phases as a function of bulk composition. For amphibole, the higher P - T end of the band is mostly for peridotite bulk compositions low in H_2O (<1 wt %) whereas the lower P - T end of the band derives from studies with high bulk H_2O , i.e., with a significant amount of excess fluid. The purple stars with arrows represent the lower bound of phlogopite stability limit from our study whereas the orange thick line represents the amphibole stability limit from our study. The green field in the background represents the field of cratonic peridotite xenoliths from Figure 2. The histogram on the right side shows the frequency distribution of MLD occurrences in cratons as a function of depth (Rader et al., 2015).

obtained in our experiments compare with natural continental mantle samples and previous experimental studies. This is important considering the fact that diverse chemistry of these mineral phases can affect their physical properties such as bulk modulus and shear modulus that are critical parameters in determination of seismic velocities.

4.3.1 Amphibole

Amphiboles crystallized in peridotitic systems are mostly calcic ($Ca_B \geq 1.50$; $(Na + K)_A \geq 0.50$) or soldic-calcic ($((Ca + Na)_B \geq 1.34$; $0.67 \leq Na_B < 1.34$) in composition and exhibit compositional variability ranging from paragonitic to edenitic and in some cases approach the K-richterite end-member (Konzett & Ulmer, 1999) for ultrapotassic bulk compositions. In Figure 7, variation of Ca with total alkali content in amphibole is represented as a function of bulk alkali and bulk CaO contents for experimental amphibole compositions from this study and previous experimental studies (Condamine & Médard, 2014; Fumagalli et al., 2009; Fumagalli & Poli, 2005; Mallik et al., 2015, 2016; Mandler & Grove, 2016; Mengel & Green, 1989; Niida & Green, 1999; Tumati et al., 2013; Wallace & Green, 1991). Amphibole compositions found in cratonic peridotite xenoliths are also plotted for reference. We note that amphibole compositions in this study lie toward alkali-rich and less calcic end of the spectrum of amphiboles found in previous experimental studies and cratonic peridotite xenoliths. Deviations from this trend are observed in amphiboles from the experimental studies of Niida and Green (1999), Fumagalli et al. (2009), and Pirard and Hermann (2015a, 2015b), respectively, primarily due to extremely calcium poor and/or strongly alkali-rich nature of the bulk compositions in these studies (Figure 7). The spread in the Ca-alkali parameter space observed for the amphibole compositions in this study can be

Figure 7. Variation of Ca content in amphibole as a function of total alkali content in amphiboles from this study (squares) and previous experimental studies (circles) plotted as a function of (a) bulk CaO and (b) bulk alkali content of volatile-bearing peridotite (Condamine & Médard, 2014; Fumagalli et al., 2009; Fumagalli & Poli, 2005; Green et al., 2014; Mallik et al., 2015; Mengel & Green, 1989; Tumiati et al., 2013; Wallace & Green, 1991). The light yellow-orange field in the background represents amphibole compositions from cratonic peridotite xenoliths (Beard et al., 2007; Frey & Green, 1974; Gamble & Kyle, 1987; Griffin, 1973; Griffin et al., 1988; Glaser et al., 1999; Ionov, 2007; Lucassen et al., 2005; Messbahi et al., 2015; Scott et al., 2014; Witt & Seck, 1987; Xu et al., 1993, 2013).

explained by the variability in the bulk alkali contents (~ 0.98 wt.% in *Bulk 1* and ~ 0.62 wt.% in *Bulk 2*). We also note that the field of cratonic peridotites xenoliths exhibit Ca contents that are higher than experimental amphibole compositions and this could be attributed to refertilization of the cratonic peridotites by a high-Ca-bearing carbonatitic or kimberlitic fluid. Variation in bulk CaO and alkali contents can thus lead to variation in amphibole compositions. By comparing the bulk composition-dependent change of amphibole composition in our and previous experimental studies with amphibole from continental mantle xenoliths, we show that silicic melt metasomatism would likely generate amphiboles that are alkali-rich and relatively Ca-poor.

4.3.2. Phlogopite

In Figure 8, we show that phlogopite compositions from this study do not vary much when plotted in a K-Al composition space, and are similar to compositions reported in previous experimental studies in peridotite + $\text{H}_2\text{O} \pm \text{CO}_2$ -bearing systems (Conceicao & Green, 2004; Condamine et al., 2016; Condamine & Médard, 2014; Falloon & Green, 1989; Fumagalli et al., 2009; Mallik et al., 2015, 2016; Mengel & Green, 1989; Thibault et al., 1992; Tumiati et al., 2013). For phlogopite compositions found in this study, K content varies between 0.7 and 0.85 while Al varies from 1.12 to 1.3. The correlation between the Al and K contents of phlogopite observed toward the low K and low Al end of the composition space has been reported previously by Fumagalli et al. (2009) who attributed it to substitution-exchange reactions as a consequence of changing P - T conditions. We also note that experimental phlogopite compositions fall within field of phlogopite compositions reported in cratonic peridotite xenoliths samples (Figure 8). K contents of ~ 0.9 a.p.f.u in phlogopite can be attributed to bulk $\text{K}_2\text{O} > 0.5$ and low Al_2O_3 contents in phlogopite is caused by low Al_2O_3 contents in the bulk. While bulk K_2O and Al_2O_3 can produce variability in phlogopite chemistry, phlogopite compositions in experimental studies and cratonic peridotite xenoliths match and therefore seismic shear wave velocity would mostly be affected by elastic parameters for phlogopite end-member.

4.3.3. Magnesite Solid Solution

Rader et al. (2015) showed that in addition to the presence of hydrous phases, a high modal proportion of carbonate phases (10–25%) could also account for the velocity reductions observed at MLD. Though proportions of magnesite in our experiments is only ≤ 1 wt.%, it could still contribute toward reduction in V_s in addition to the effects of phlogopite and amphibole. Therefore it is relevant to evaluate how the composition of magnesite solid solution from our study compares with those found in previous experimental studies. Compared to previous experimental studies on peridotite + $\text{CO}_2 \pm \text{H}_2\text{O}$ systems with lower Mg# (Mg# of ~ 87 –90;

Figure 8. Variation of K cations with Al cations in phlogopite is represented as a function (a) bulk K_2O and (b) bulk Al_2O_3 for compositions found in this study (solid squares) and are compared to phlogopite compositions found in previous experimental studies (solid circles) (Conceicao & Green, 2004; Condamine & Médard, 2014; Esperança & Holloway, 1987; Fumagalli et al., 2009; Mallik et al., 2015, 2016; Mengel & Green, 1989; Thibault et al., 1992; Tumiaty et al., 2013; Wendlandt & Eggler, 1980). Phlogopite compositions found in cratonic peridotite xenoliths are plotted as a light yellow-orange field in the background (Bachinski & Sltvtpsonl, 1984; Baptiste et al., 2012; Beard et al., 2007; Beccaluva et al., 2004; Conticelli & Peccerillo, 1989; Dawson & Smith, 1988; Downes et al., 2004; Edgar et al., 1989; Ehrenberg, 1982; Grégoire et al., 2002; Giuliani et al., 2016, 2014; Griffin et al., 1999; Hopp et al., 2008; Konzett et al., 2013; Lee et al., 2000; Lee & Rudnick, 1999; Litasov et al., 2000; Mukhopadhyay & Manton, 1994; Pike et al., 1980; Rudnick et al., 1994; Wartho and Kelley, 2003; Zhang et al., 2000).

Dasgupta & Hirschmann, 2006, 2007; Falloon & Green, 1989; Wallace & Green, 1988), magnesite solid solutions in this study are more magnesian ($Mg\# \sim 94\text{--}96$) primarily because of higher $Mg\#$ of the bulk composition (Dasgupta & Hirschmann, 2007; Wallace & Green, 1988). We, however note that when magnesite coexists with dolomite, in relatively fertile compositions, such as in Tumiaty et al. (2013), $Mg\#$ of magnesite-solid solution is higher (~ 98) than that found in our study. Therefore, the aggregate seismic velocity, if affected by the presence of a carbonate mineral phase, will be most affected by the elastic constants of the magnesite end-member rather than that of calcite, dolomite, ankerite, or siderite.

4.4. The near-Solidus Phase Relations of Peridotite + H_2O + CO_2 : Comparison With Previous Studies

The presence of trace melt along grain boundaries prevented the quantification of the melt phase composition but predictions about the nature of the near solidus melt can be made based on H_2O mass balance, assuming maintenance of close system during our experiments. The bulk H_2O concentration calculated from modal proportion of phases, at different experimental P - T conditions indicate a deficit of ~ 0.5 wt.% H_2O in all the experiments. This implies that H_2O present in phlogopite and amphibole cannot account for all the H_2O present in the system, and it is likely that some amount of this excess H_2O would be present in a separate phase. Therefore, the true solidus of our composition must be fluid-saturated (Dasgupta, 2018). Given the location of H_2O -rich fluid-present solidus of peridotite (Dasgupta, 2018; Grove et al., 2006; Kawamoto & Holloway, 1997; Till et al., 2012), the missing H_2O at our experimental conditions is likely to be present as a trace hydrous silicate melt. The “apparent” solidus obtained in this study coincides with the magnesite-out boundary, however. Considering that there is no other carbon-bearing phase observed, it is likely that after the breakdown of magnesite all the CO_2 in the system is incorporated in small volume melt present. Hence we infer that the trace melt at the nominally melt-free experimental conditions was a hydrous silicate melt, which evolves to a visually observable carbonated hydrous silicate melt after the magnesite-out boundary.

Solidus locations in complex volatile-bearing systems are controlled by different factors such as bulk composition and the volatile content. For similar amount of volatiles present in the system, bulk compositions with higher $Mg\#$ tend to have a higher solidus temperature as compared to low $Mg\#$ compositions as evidenced from comparison with previous studies (Olafsson & Eggler, 1983; Wallace & Green, 1988). The higher

Mg# of our bulk composition can explain the higher temperature at which the apparent solidus is observed compared to the study of Wallace and Green (1988) (supporting information Table S1). In contrast, the apparent solidus obtained in this study is lower compared to the C-O-H-bearing peridotite systems examined by Olafsson and Eggler (1983) and Thibault et al. (1992) possibly owing to higher concentration of bulk H₂O in this study, which diminishes the temperature of onset of main melting.

4.5. Aggregate Seismic Velocity of the Solid Peridotite Assemblage Modified by Reaction With Volatile-Bearing Slab Melts

Given the dynamic stability of volatile-rich partial melts is questionable, in the following section, we have calculated the aggregate shear wave velocity of the mineral assemblage produced in our experiments in order to test whether these velocities can explain velocities recorded at MLD depths in cratons.

Chemistry and proportions (wt.%) of the phases obtained by mass balance calculations were used to obtain modal abundance (vol.%) of the different constituent mineral end-members. To convert mass fractions of various end-members for a given experimental assemblage to volume fractions of end-members, density estimates of respective end-members at the *P-T* of interest were used based on Abers and Hacker (2016). The MS Excel macro of Abers and Hacker (2016) was also used to calculate aggregate shear wave velocity at the *P-T* conditions of the experiments (Table 6). Although we used the elastic constants for most end-members from the Abers and Hacker (2016)'s compiled database, we chose more recent estimates for key (hydrous) phases like amphibole and phlogopite. Bulk modulus, shear modulus, and their *P-T* derivatives for phlogopite were taken from Chheda et al. (2014). Bulk and shear modulus for pargasite were taken from Brown and Abramson (2016) while their *P-T* derivatives are from the database of Holland and Powell (1998). We note that using updated elastic constants for phlogopite and amphibole, yielded higher aggregate

Table 6
Summary of the Seismic Velocity Calculations for the Different Experimental Runs in This Study

Run#	B405	B342	B373	B352	B353	B363	B402	B373	MA131	MA136	MA124	MA119	MA125	B413	B410	MA171
<i>P</i> (GPa)	2.0	3.0	3.0	3.0	3.0	3.0	3.0	3.0	4.0	4.0	4.0	4.0	4.0	2.0	3.0	4.0
<i>T</i> (°C)	950	950	975	1,000	1,025	1,050	1,100	1,175	950	1,000	1,050	1,100	1,150	950	1,000	1,050
^a End-members used (vol %)																
Pyrope	0.00	0.00	0.00	0.00	0.00	0.00	1.41	3.62	3.88	3.79	2.75	3.95	2.57	0.00	0.00	1.97
Forsterite	44.62	46.63	45.98	42.22	45.88	47.17	44.35	47.18	42.64	46.42	45.50	49.16	45.28	56.25	56.43	59.44
Fayalite	2.65	3.03	3.00	2.59	3.05	3.14	2.97	3.00	2.65	2.61	2.87	2.93	2.65	3.66	3.72	4.10
Enstatite	35.33	31.17	34.19	41.09	36.12	34.02	38.42	37.52	37.73	35.50	38.48	34.03	38.27	25.23	28.02	24.89
Ferrosilite	2.37	1.57	1.93	2.90	2.89	2.36	2.57	2.44	2.06	2.16	2.34	2.02	2.32	1.96	2.00	1.51
Diopside	2.18	0.05	0.09	0.01	0.09	1.62	3.01	0.09	3.63	2.54	2.55	2.44	3.61	2.67	4.52	3.43
Pargasite	7.00	9.79	7.86	3.75	5.55	6.23	0.00	0.00	0.00	0.00	0.00	0.00	0.00	6.92	2.36	0.00
Phlogopite	4.80	6.69	5.87	6.37	6.41	5.45	7.27	6.15	6.32	5.87	5.51	5.47	5.30	2.67	2.95	4.67
Magnesite	1.06	1.07	1.08	1.07	0.00	0.00	0.00	0.00	1.10	1.10	0.00	0.00	0.00	0.64	0.00	0.00
Sum	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
^b Physical properties calculated with Hashin-Shtrikman average																
ρ (g/cm ³)	3.19	3.21	3.22	3.22	3.22	3.22	3.22	3.22	3.27	3.26	3.26	3.26	3.25	3.21	3.24	3.27
V_p (km/s)	7.71	7.76	7.77	7.73	7.71	7.74	7.69	7.70	7.95	7.94	7.90	7.90	7.84	7.82	7.91	7.97
V_s (km/s)	4.42	4.40	4.41	4.39	4.38	4.39	4.36	4.37	4.50	4.49	4.47	4.46	4.43	4.48	4.50	4.50
V_p/V_s	1.75	1.76	1.76	1.76	1.76	1.76	1.76	1.76	1.77	1.77	1.77	1.77	1.77	1.75	1.76	1.77
Poissons	0.26	0.26	0.26	0.26	0.26	0.26	0.26	0.26	0.26	0.27	0.26	0.27	0.27	0.26	0.26	0.27
<i>K</i> (GPa)	106.64	110.42	110.59	109.44	109.34	109.89	108.61	108.98	118.41	118.06	116.73	116.80	114.78	110.64	115.06	119.33
<i>G</i> (GPa)	62.26	62.17	62.58	62.18	61.72	62.13	61.31	61.45	66.10	65.79	65.15	64.88	63.85	64.30	65.78	66.26
^c V_s _Reference	4.67	4.69	4.70	4.69	4.68	4.67	4.65	4.62	4.75	4.72	4.70	4.68	4.66	4.67	4.69	4.70
^d % drop	5.35	6.18	6.17	6.40	6.41	6.00	6.24	5.41	5.26	4.87	4.89	4.70	4.94	4.07	4.05	4.26

^avol.% of mineral end-members are calculated using mineral phase proportions (in wt.%) obtained by mass balance, end-member calculations based on mineral chemistry, and density of various mineral end-members calculated at the *P-T* conditions of experiments (see text for details). The presence of melt has been ignored while calculating aggregate shear wave velocities with the notion that the entire melt fraction may be dynamically unstable and the residual mineral assemblage would be responsible for affecting aggregate shear wave velocity of the assemblage. ^bDensity, V_p , V_s , shear modulus, bulk modulus are calculated using the Hashin-Shtrikman averaging scheme following Abers and Hacker (2016) and Hacker et al. (2003a) (details in text). ^c V_s _Reference refers to V_s , for a depleted peridotitic composition calculated at the experimental *P-T* conditions. Mineral modes used for calculating reference V_s , at 2 GPa are from Rudnick et al. (2004); at 3 GPa and 4 GPa mineral modes from the depleted peridotite run is used (run # B338). ^d% drop in V_s is calculated as $[(V_{s_ref} - V_{s(P,T)})/V_{s_ref}] \times 100$ at each *P-T* condition.

shear wave velocities as compared to values obtained using elastic constants in the compiled database of Abers and Hacker (2016).

Table 6 shows that the estimated velocity reductions vary between 4.7 and 6.4% for *Bulk 1* and up to ~4% *Bulk 2* and are similar to the mid to lower end of the globally observed range of V_s reductions at MLDs.

Seismic velocities were also calculated along a representative continental geotherm with an equivalent surface heat flux of 44 mWm^{-2} . Velocity profiles, based on these calculations along the continental geotherm, are depicted in Figure 9b and the P - T conditions of these calculations along the 44 mWm^{-2} geotherm are marked in Figure 9a. Mineral modes from experimental conditions closest to the P - T conditions along the geotherm were used for the calculations. We note that the estimated aggregate shear wave velocity (V_s) shows a systematic variation with the degree of melt infiltration and decreases as the proportion of infiltrating melt increases. The calculated aggregate shear wave velocities at the same P - T conditions along the geotherm were also compared with the aggregate velocities for unmetasomatized peridotites and reductions in V_s (expressed in percent) are plotted as a function of the degree of melt infiltration in Figure 10. Mineral modes used for calculating reference V_s , at 2 GPa are from Rudnick et al. (2004) (44.7 wt.% olivine; 37.4 wt.% orthopyroxene; 16.1 wt.% clinopyroxene, and 1.8 wt.% spinel); at 3 and 4 GPa are calculated using mineral modes from the depleted peridotite run# B338 from this study (48.5 wt.% olivine; 41.5wt.% orthopyroxene; 9.1wt.% garnet; 0.9 wt.% clinopyroxene). The reference V_s at 2 GPa was also calculated excluding the high mode of clinopyroxene (~16 wt.%) and normalizing the total to 100 to consider the low mode of clinopyroxene (~0.9 wt.%) found in our experiment at 3 GPa (B338). We find that the aggregate velocity is same as when 16 wt.% clinopyroxene is included in the calculation. From Figure 10, based on the trend of estimated velocity reduction (owing to presence of hydrous and carbonate phases) as a function of melt:peridotite ratio, we suggest that introduction of 3–10% slab derived H_2O - CO_2 -bearing silicic melt into a depleted peridotite composition can produce up to 6% reduction in the aggregate seismic shear wave velocity. This range of velocity reduction matches the lower to middle end of the range of velocity reductions observed at MLD. To match up the higher end of the observed velocity reductions (~10%), additional mechanisms might be required. For example, in our seismic velocity calculations, we have not considered the effect of presence of melt on V_s , but it can also contribute toward a decrease in the seismic shear wave velocity, either by the direct presence along grain-edge tubules or via effects on grain boundary sliding (Karato et al., 2015; Kumar et al., 2012; Thybo & Perchuc, 1997).

Figure 9. Aggregate shear wave velocity (V_s) calculated for volatile-free depleted peridotite and the solid assemblages from two H_2O - CO_2 -bearing compositions studied here along a 44 mWm^{-2} conductive geotherm at 2, 3, and 4 GPa at P - T conditions marked by red stars in (a) is represented in (b). V_s for *Bulk 1* and *Bulk 2* are represented by red and green squares, respectively, while V_s for depleted peridotite is represented by black squares. It can be observed that the aggregate shear wave velocity decreases with increase in the degree of melt infiltration owing to greater abundance of hydrous phases.

Figure 10. Percentage reduction in aggregate shear wave velocity (V_s) at P - T conditions along a 44 mW^{-2} geotherm, plotted as a function of the mass fraction of the slab-derived hydrous ($+\text{CO}_2$ -bearing) melt interacting with a depleted peridotite at 2, 3, and 4 GPa. The most observed range of V_s reductions at MLD is represented by the grey-shaded band. The plot shows that amphibole and phlogopite formation owing to 5–10 wt % siliceous hydrous slab melt and peridotite interactions can result in reduction in seismic shear wave speeds by ~ 3 –6%. We note, the computed reduction in aggregate shear-wave velocity is entirely due to the stable mineral phases and does not take into account the additional effects of trace melts.

5. Concluding Remarks

We show that interaction of a slab-derived H_2O - CO_2 -bearing silicic melt with the depleted peridotite can produce seismically slow mineral phases such as phlogopite, amphibole, and magnesite that can cause reductions in aggregate seismic shear wave velocity, V_s . Such silicic-slab-derived melts could have been generated in ancient subduction zones at temperatures between 700 and 900°C and pressures of up to 5 GPa and interacted with the overlying wedge peridotite that is depleted from prior melt-extraction event (Syracuse et al., 2010; Van Keken et al., 2002), leading to reactive crystallization of such volatile-bearing phases (Figure 11). Coexistence of amphibole and phlogopite likely lowers the stability of amphibole,

Figure 11. Schematic showing reduction in aggregate shear wave velocities at mid lithospheric depths in cratons as (a) as a function of depth (b) presence of amphibole $\pm$ phlogopite at depths ≤ 100 km and phlogopite up to ~ 150 km and (c) infiltration of slab-derived silicic melt in ancient subduction zones results in reactive crystallization of phlogopite at depths greater than 100 km and both amphibole and phlogopite at shallower depths. Our study suggests that accretion and thickening of subarc mantle that is modified by slab-derived melt can lead to the formation of amphibole and phlogopite at 70–150 km depths and as a consequence give rise to a reduction in shear wave speeds.

causally linking MLDs at greater depths (> 100 km) to the presence of phlogopite $\pm$ carbonates only. In addition, though slab-derived melt and mantle interaction may causally be linked to lowering of seismic velocities introduction of a modest amount of melt does not result in formation of hydrous minerals of sufficient abundance to explain the entire range of the velocity reductions reported at MLDs and the role of additional mechanisms in contributing to decrease in V_S may need to be considered. However, a key suggestion arising from our work is that craton formation via accretion and thickening of slab-modified, depleted, subarc wedge peridotite may inevitably lead to stabilization of a hydrous mineral phase-bearing horizon at midlithospheric depths. Given the thermal structure of downgoing slabs, the release of hydrous silicic melt to the overlying mantle wedge is likely to happen at depths similar to those of MLDs. Thus, our study suggests that formation of amphibole or phlogopite in cratonic mantle could happen early, during craton formation rather than via more recent metasomatic alterations. This suggestion, if correct, can reconcile why MLDs are pervasive feature in cratons.

Infiltration by slab-derived silicic melt is however one of the several distinct scenarios of cratonic mantle metasomatism. How other modes of volatile introduction affect the relative stability and compositions of the volatile-bearing phases and their relative contribution toward the velocity reduction need to be constrained in future studies. We also recognize that the observed velocity reduction at midlithospheric depths in cratons might be the cumulative effect of different mechanisms acting in tandem. For example, though we have performed the seismic velocity calculations for melt-free assemblages, presence of even a very small amount (<1 wt.%) of melt along grain boundaries can also cause significant velocity reduction (Faul et al., 2004). Future studies should also therefore incorporate the effects of both volatile-bearing mineral phases as well as partial melts on the aggregate seismic velocity. What the present study highlights, however, is that particular attention to major element bulk compositions of peridotite needs to be given to constrain both the most appropriate composition and proportions of mineral phases as well as the exact domain of partial melt stability versus volatile-bearing mineral phases.

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References

- Abers, G., & Hacker, B. R. (2016). A MATLAB toolbox and Excel workbook for calculating the densities, seismic wave speeds, and major element composition of minerals and rocks at pressure and temperature. *Geochemistry, Geophysics, Geosystems*, 17, 616–624. <https://doi.org/10.1002/2015GC006171>
- Abt, D. L., Fischer, K. M., French, S. W., Ford, H. A., Yuan, H., & Romanowicz, B. (2010). North American lithospheric discontinuity structure imaged by Ps and Sp receiver functions. *Journal of Geophysical Research*, 115, 1–24. <https://doi.org/10.1029/2009JB006914>
- Allen, J. C., Modreski, P. J., Haygood, C., & Boettcher, A. L. (1972). The role of water in the mantle of the Earth. 24th International Geological Congress, Montreal, Section 2, 231–240.
- Bachinski, S. W., & Simpson, E. L. (1984). Ti-phlogopite of the Shaw's Cove minette: A comparison with micas of other lamprophyres, potassic rocks, kimberlites and mantle xenoliths. *American Mineralogist*, 69, 41–56.
- Balta, J. B., Asimow, P. D., & Mosenfelder, J. L. (2011). Hydrous, low-carbon melting of garnet peridotite. *Journal of Petrology*, 52(11), 2079–2105. <https://doi.org/10.1093/petrology/egr040>
- Baptiste, V., Tommasi, A., & Demouchy, S. (2012). Deformation and hydration of the lithospheric mantle beneath the Kaapvaal craton, South Africa. *Lithos*, 149, 31–50. <https://doi.org/10.1016/j.lithos.2012.05.001>
- Barton, M., & Hamilton, D. L. (1979). The melting relationships of a madupite from the Leucite Hills, Wyoming, to 30 Kb. *Contributions to Mineralogy and Petrology*, 69(2), 133–142. <https://doi.org/10.1007/BF00371856>
- Beard, A. D., Downes, H., Mason, P. R. D., & Vetrin, V. R. (2007). Depletion and enrichment processes in the lithospheric mantle beneath the Kola Peninsula (Russia): Evidence from spinel lherzolite and wehrlite xenoliths. *Lithos*, 94(1–4), 1–24. <https://doi.org/10.1016/j.lithos.2006.02.002>
- Beccaluva, L., Bianchini, G., Bonadiman, C., Siena, F., & Vaccaro, C. (2004). Coexisting anorogenic and subduction-related metasomatism in mantle xenoliths from the Betic Cordillera (southern Spain). *Lithos*, 75(1–2), 67–87. <https://doi.org/10.1016/j.lithos.2003.12.015>
- Bianchini, G., Beccaluva, L., Bonadiman, C., Nowell, G., Pearson, G., Siena, F., et al. (2007). Evidence of diverse depletion and metasomatic events in harzburgite-lherzolite mantle xenoliths from the Iberian plate (Olot, NE Spain): Implications for lithosphere accretionary processes. *Lithos*, 94(1–4), 25–45. <https://doi.org/10.1016/j.lithos.2006.06.008>
- Bostock, M. G. (1997). Anisotropic upper-mantle stratigraphy and architecture of the Slave craton. *Nature*, 390(6658), 392–395.
- Bostock, M. G. (1998). Mantle stratigraphy and evolution of the Slave province. *Journal of Geophysical Research*, 103(B9), 21183–21200. <https://doi.org/10.1029/98JB01069>
- Boyd, F. R. (1989). Compositional distinction between oceanic and cratonic lithosphere. *Earth and Planetary Science Letters*, 96(1–2), 15–26. [https://doi.org/10.1016/0012-821X\(89\)90120-9](https://doi.org/10.1016/0012-821X(89)90120-9)
- Boyd, F. R., Pearson, D. G., Nixon, P. H., & Mertzman, S. A. (1993). Low-calcium garnet harzburgites from southern Africa: their relations to craton structure and diamond crystallization. *Contributions to Mineralogy and Petrology*, 113(3), 352–366. <https://doi.org/10.1007/BF00286927>
- Boyd, F. R., Pokhilenko, N. P., Pearson, D. G., Mertzman, S. A., Sobolev, N. V., & Finger, L. W. (1997). Composition of the Siberian cratonic mantle: Evidence from Udachnaya peridotite xenoliths. *Contributions to Mineralogy and Petrology*, 128(2–3), 228–246. <https://doi.org/10.1007/s004100050305>
- Brey, G., Brice, W. R., Ellis, D. J., Green, D. H., Harris, K. L., & Ryabchikov, I. D. (1983). Pyroxene-carbonate reactions in the upper mantle. *Earth and Planetary Science Letters*, 62(1), 63–74. [https://doi.org/10.1016/0012-821X\(83\)90071-7](https://doi.org/10.1016/0012-821X(83)90071-7)

- Brey, G. P., Bulatov, V. K., Giris, A. V., & Lahaye, Y. (2007). Experimental melting of carbonated peridotite at 6–10 GPa. *Journal of Petrology*, 49(4), 797–821. <https://doi.org/10.1093/petrology/egn002>
- Brey, G. P., & K Hler, T. (1990). Geothermobarometry in four-phase lherzolites II. New thermobarometers, and practical assessment of existing thermobarometers. *Journal of Petrology*, 31(6), 1353–1378. <https://doi.org/10.1093/petrology/31.6.1353>
- Brown, J. M., & Abramson, E. H. (2016). Elasticity of calcium and calcium-sodium amphiboles. *Physics of the Earth and Planetary Interiors*, 261, 161–171. <https://doi.org/10.1016/j.pepi.2016.10.010>
- Carswell, D. A., Clarke, D. B., & Mitchell, R. H. (1979). The petrology and geochemistry of ultramafic nodules from pipe 200, Northern Lesotho. In F. R. Boyd and H. O. A. Meyer (Eds.), *The mantle samples: Inclusion in Kimberlites and Other Volcanics* (pp. 145–155). Washington, DC: American Geophysical Union. <https://doi.org/10.1029/SP016p0127>
- Chheda, T. D., Mookherjee, M., Mainprice, D., Dos Santos, A. M., Molaison, J. J., Chantel, J., et al. (2014). Structure and elasticity of phlogopite under compression: Geophysical implications. *Physics of the Earth and Planetary Interiors*, 233, 1–12. <https://doi.org/10.1016/j.pepi.2014.05.004>
- Conceicao, R. V., & Green, D. H. (2004). Derivation of potassic (shoshonitic) magmas by decompression melting of phlogopite plus pargasite lherzolite. *Lithos*, 72(3–4), 209–229. <https://doi.org/10.1016/J.Lithos.2003.09.003>
- Condamine, P., & Médard, E. (2014). Experimental melting of phlogopite-bearing mantle at 1 GPa: Implications for potassic magmatism. *Earth and Planetary Science Letters*, 397, 80–92. <https://doi.org/10.1016/j.epsl.2014.04.027>
- Condamine, P., Medard, E., & Devidal, J. L. (2016). Experimental melting of phlogopite-peridotite in the garnet stability field. *Contributions to Mineralogy and Petrology*, 171(11), 1–26. <https://doi.org/10.1007/s00410-016-1306-0>
- Coticelli, S., & Peccerillo, A. (1990). Petrological significance of high-pressure ultramafic xenoliths from ultrapotassic rocks of Central Italy. *Lithos*, 24(4), 305–322. [https://doi.org/10.1016/0024-4937\(89\)90050-9](https://doi.org/10.1016/0024-4937(89)90050-9)
- Dasgupta, R. (2018). Volatile bearing partial melts beneath oceans and continents: Where, how much, and of what compositions? *American Journal of Science*, 318(1), 147–165. <https://doi.org/10.2475/01.2018.06>
- Dasgupta, R., & Hirschmann, M. M. (2006). Melting in the Earth's deep upper mantle caused by carbon dioxide. *Nature*, 440(7084), 659–662. <https://doi.org/10.1038/nature04612>
- Dasgupta, R., & Hirschmann, M. M. (2007). Effect of variable carbonate concentration on the solidus of mantle peridotite. *American Mineralogist*, 92(2–3), 370–379. <https://doi.org/10.2138/am.2007.2201>
- Dasgupta, R., Hirschmann, M. M., & Smith, N. D. (2007). Partial melting experiments of peridotite + CO₂ at 3 GPa and genesis of alkaline ocean island basalts. *Journal of Petrology*, 48(11), 2093–2124. <https://doi.org/10.1093/petrology/egm053>
- Dasgupta, R., Mallik, A., Tsuno, K., Withers, A. C., Hirth, G., & Hirschmann, M. M. (2013). Carbon-dioxide-rich silicate melt in the Earth's upper mantle. *Nature*, 493(7431), 211–215. <https://doi.org/10.1038/nature11731>
- Dawson, J. B., & Smith, J. V. (1988). Metasomatized and veined upper-mantle xenoliths from Pello Hill, Tanzania: Evidence for anomalously-light mantle beneath the Tanzanian sector of the East African Rift Valley. *Contributions to Mineralogy and Petrology*, 100(4), 510–527. <https://doi.org/10.1007/BF00371380>
- Ding, S., Dasgupta, R., & Tsuno, K. (2014). Sulfur concentration of Martian basalts at sulfide saturation at high pressures and temperatures - Implications for deep sulfur cycle on Mars. *Geochimica et Cosmochimica Acta*, 131, 227–246. <https://doi.org/10.1016/j.gca.2014.02.003>
- Downes, H., Macdonald, R., Upton, B. G. J., Cox, K. G., Bodinier, J. L., Mason, P. R. D., et al. (2004). Ultramafic xenoliths from the Bearpaw Mountains, Montana, USA: Evidence for multiple metasomatic events in the lithospheric mantle beneath the Wyoming craton. *Journal of Petrology*, 45(8), 1631–1662. <https://doi.org/10.1093/petrology/egh027>
- Duncan, M. S., & Dasgupta, R. (2014). CO₂ solubility and speciation in rhyolitic sediment partial melts at 1.5–3.0 GPa - Implications for carbon flux in subduction zones. *Geochimica et Cosmochimica Acta*, 124, 328–347. <https://doi.org/10.1016/j.gca.2013.09.026>
- Duncan, M. S., & Dasgupta, R. (2015). Pressure and temperature dependence of CO₂ solubility in hydrous rhyolitic melt - Implications for carbon transfer to mantle source of volcanic arcs via partial melt of subducting crustal lithologies. *Contributions to Mineralogy and Petrology*, 169(6), 1–19. <https://doi.org/10.1007/s00410-015-1144-5>
- Duncan, M. S., & Dasgupta, R. (2017). Rise of Earth's atmospheric oxygen controlled by efficient subduction of organic carbon, *Nature Geoscience*, 10, 387–392. <https://doi.org/10.1038/NGEO2939>
- Dvir, O., & Kessel, R. (2017). The effect of CO₂ on the water-saturated solidus of K-poor peridotite between 4 and 6 GPa. *Geochimica et Cosmochimica Acta*, 206, 184–200. <https://doi.org/10.1016/j.gca.2017.02.028>
- Edgar, A. D., Lloyd, F. E., Forsyth, D. M., & Barnett, R. L. (1989). Origin of glass in upper mantle xenoliths from the quaternary volcanics of Gees, West Eifel, Germany. *Contributions to Mineralogy and Petrology*, 103(3), 277–286. <https://doi.org/10.1007/BF00402915>
- Eguchi, J., & Dasgupta, R. (2017). CO₂ content of andesitic melts at graphite-saturated upper mantle conditions with implications for redox state of oceanic basalt source regions and remobilization of reduced carbon from subducted eclogite. *Contributions to Mineralogy and Petrology*, 172(2–3), 1–16. <https://doi.org/10.1007/s00410-017-1330-8>
- Ehrenberg, S. N. (1982). Petrogenesis of garnet lherzolite and megacrystalline nodules from the thumb, navajo volcanic field. *Journal of Petrology*, 23(4), 507–547. <https://doi.org/10.1093/petrology/23.4.507>
- Elkins-Tanton, L. T., & Grove, T. L. (2003). Evidence for deep melting of hydrous metasomatized mantle: Pliocene high-potassium magmas from the Sierra Nevada. *Journal of Geophysical Research*, 108(B7), 2350. <https://doi.org/10.1029/2002JB002168>
- Enggist, A., Chu, L., & Luth, R. W. (2012). Phase relations of phlogopite with magnesite from 4 to 8 GPa. *Contributions to Mineralogy and Petrology*, 163(3), 467–481. <https://doi.org/10.1007/s00410-011-0681-9>
- Enggist, A., & Luth, R. W. (2010). Breakdown of phlogopite in presence of carbonate: Metasomatic agents in the sub-continental lithospheric mantle. In *Geo convention* (pp. 2–5). Calgary, Alberta, Canada.
- Esperança, S., & Holloway, J. R. (1987). On the origin of some mica-lamprophyres: experimental evidence from a mafic minette. *Contributions to Mineralogy and Petrology*, 95(2), 207–216. <https://doi.org/10.1007/BF00381270>
- Falloon, T. J., & Danyushevsky, L. V. (2000). Melting of refractory mantle at 1.5, 2 and 2.5 GPa under anhydrous and H₂O-undersaturated conditions: Implications for the petrogenesis of high-Ca boninites and the influence of subduction components on mantle melting. *Journal of Petrology*, 41(2), 257–283. <https://doi.org/doi.org/10.1093/petrology/41.2.257>
- Falloon, T. J., & Green, D. H. (1989). The solidus of carbonated, fertile peridotite. *Earth and Planetary Science Letters*, 94(3–4), 364–370. [https://doi.org/10.1016/0012-821X\(89\)90153-2](https://doi.org/10.1016/0012-821X(89)90153-2)
- Falloon, T. J., & Green, D. H. (1990). Solidus of carbonated fertile peridotite under fluid-saturated conditions. *Geology*, 18(3), 195–199. [https://doi.org/10.1130/0091-7613\(1990\)018<0195](https://doi.org/10.1130/0091-7613(1990)018<0195)
- Faul, U. H., Fitz Gerald, J. D., & Jackson, I. (2004). Shear wave attenuation and dispersion in melt-bearing olivine polycrystals: 2. Microstructural interpretation and seismological implications. *Journal of Geophysical Research*, 109, B06201. <https://doi.org/10.1029/2003JB002407>
- Fischer, K. M., Ford, H. A., Abt, D. L., & Rychert, C. A. (2010). The lithosphere-asthenosphere boundary. *Annual Review of Earth and Planetary Sciences*, 38(1), 551–575. <https://doi.org/10.1146/annurev-earth-040809-152438>

- Ford, H. A., Fischer, K. M., Abt, D. L., Rychert, C. A., & Elkins-Tanton, L. T. (2010). The lithosphere-asthenosphere boundary and cratonic lithospheric layering beneath Australia from Sp wave imaging. *Earth and Planetary Science Letters*, 300(3–4), 299–310. <https://doi.org/10.1016/j.epsl.2010.10.007>
- Foster, K., Dueker, K., Schmandt, B., & Yuan, H. (2014). A sharp cratonic lithosphere-asthenosphere boundary beneath the American Midwest and its relation to mantle flow. *Earth and Planetary Science Letters*, 402(C), 82–89. <https://doi.org/10.1016/j.epsl.2013.11.018>
- Frey, F. A., & Green, D. H. (1974). The mineralogy, geochemistry and origin of lherzolite inclusions in Victorian basanites. *Geochimica et Cosmochimica Acta*, 38(7), 1023–1059. [https://doi.org/10.1016/0016-7037\(74\)90003-9](https://doi.org/10.1016/0016-7037(74)90003-9)
- Fumagalli, P., & Poli, S. (2005). Experimentally determined phase relations in hydrous peridotites to 6.5 GPa and their consequences on the dynamics of subduction zones. *Journal of Petrology*, 46(3), 555–578. <https://doi.org/10.1093/petrology/egh088>
- Fumagalli, P., Zanchetta, S., & Poli, S. (2009). Alkali in phlogopite and amphibole and their effects on phase relations in metasomatized peridotites: A high-pressure study. *Contributions to Mineralogy and Petrology*, 158(6), 723–737. <https://doi.org/10.1007/s00410-009-0407-4>
- Gamble, J. A., & Kyle, P. R. (1987). The origins of glass and amphibole in spinel—wehrlite xenoliths from Foster Crater, McMurdo Volcanic Group, Antarctica. *Journal of Petrology*, 28(5), 755–779. <https://doi.org/10.1093/petrology/28.5.755>
- Ghiorsso, M. S., & Gualda, G. A. R. (2015). An H₂O–CO₂ mixed fluid saturation model compatible with rhyolite-MELTS. *Contributions to Mineralogy and Petrology*, 169(6), 1–30. <https://doi.org/10.1007/s00410-015-1141-8>
- Ghosh, S., Litasov, K., & Ohtani, E. (2014). Phase relations and melting of carbonated peridotite between 10 and 20 GPa: A proxy for alkali- and CO₂-rich silicate melts in the deep mantle. *Contributions to Mineralogy and Petrology*, 167(2), 964. <https://doi.org/10.1007/s00410-014-0964-z>
- Giuliani, A., Phillips, D., Kamenetsky, V. S., & Goemann, K. (2016). Constraints on kimberlite ascent mechanisms revealed by phlogopite compositions in kimberlites and mantle xenoliths. *Lithos*, 240–243, 189–201. <https://doi.org/10.1016/j.lithos.2015.11.013>
- Giuliani, A., Phillips, D., Maas, R., Woodhead, J. D., Kendrick, M. A., Greig, A., et al. (2014). LIMA U–Pb ages link lithospheric mantle metasomatism to Karoo magmatism beneath the Kimberley region, South Africa. *Earth and Planetary Science Letters*, 401, 132–147. <https://doi.org/10.1016/j.epsl.2014.05.044>
- Glaser, S. M., Foley, S. F., & Günther, D. (1999). Trace element compositions of minerals in garnet and spinel peridotite xenoliths from the Vitim volcanic field, Transbaikalia, Eastern Siberia. *Lithos*, 48(1–4), 263–285. [https://doi.org/10.1016/S0016-7037\(99\)80015-9](https://doi.org/10.1016/S0016-7037(99)80015-9)
- Green, D. H. (1973). Experimental melting studies on a model upper mantle composition at high pressure under water-saturated and water-undersaturated conditions. *Earth and Planetary Science Letters*, 19(1), 37–53. [https://doi.org/10.1016/0012-821X\(73\)90176-3](https://doi.org/10.1016/0012-821X(73)90176-3)
- Green, D. H., Hibberson, W. O., Rosenthal, A., Kovacs, I., Yaxley, G. M., Falloon, T. J., et al. (2014). Experimental study of the influence of water on melting and phase assemblages in the upper mantle. *Journal of Petrology*, 55(10), 2067–2096. <https://doi.org/10.1093/petrology/egu050>
- Grégoire, M., Bell, D., & Le Roex, A. (2002). Trace element geochemistry of phlogopite-rich mafic mantle xenoliths: their classification and their relationship to phlogopite-bearing peridotites and kimberlites revisited. *Contributions to Mineralogy and Petrology*, 142(5), 603–625. <https://doi.org/10.1007/s00410-001-0315-8>
- Griffin, W. L. (1973). Lherzolite nodules from the Fen alkaline complex, Norway. *Contributions to Mineralogy and Petrology*, 38(2), 135–146. <https://doi.org/10.1007/BF00373877>
- Griffin, W. L., O'Reilly, S. Y., & Stabel, A. (1988). Mantle metasomatism beneath western Victoria, Australia: II. Isotopic geochemistry of Cr-diopside lherzolites and Al-augite pyroxenites. *Geochimica et Cosmochimica Acta*, 52(2), 449–459. [https://doi.org/10.1016/0016-7037\(88\)90100-7](https://doi.org/10.1016/0016-7037(88)90100-7)
- Griffin, W. L., Shee, S. R., Ryan, C. G., Win, T. T., & Wyatt, B. A. (1999). Harzburgite to lherzolite and back again: Metasomatic processes in ultramafic xenoliths from the Wesselton kimberlite, Kimberley, South Africa. *Contributions to Mineralogy and Petrology*, 134(2–3), 232–250. <https://doi.org/10.1007/s004100050481>
- Grove, T., Chatterjee, N., Parman, S., & Medard, E. (2006). The influence of H₂O on mantle wedge melting. *Earth and Planetary Science Letters*, 249(1–2), 74–89. <https://doi.org/10.1016/j.epsl.2006.06.043>
- Hacker, B. R., Abers, G. A., & Peacock, S. M. (2003a). Subduction factory 1. Theoretical mineralogy, densities, seismic wave speeds, and H₂O contents. *Journal of Geophysical Research*, 108(B1), 2029. <https://doi.org/10.1029/2001JB001127>
- Hacker, B. R., Peacock, S. M., Abers, G. A., & Holloway, S. D. (2003b). Subduction factory 2. Are intermediate-depth earthquakes in subducting slabs linked to metamorphic dehydration reactions? *Journal of Geophysical Research*, 108(B1), 2030. <https://doi.org/10.1029/2001JB001129>
- Hansen, S. M., Dueker, K., & Schmandt, B. (2015). Thermal classification of lithospheric discontinuities beneath USArray. *Earth and Planetary Science Letters*, 431, 36–47. <https://doi.org/10.1016/j.epsl.2015.09.009>
- Hermann, J., & Spandler, C. J. (2008). Sediment melts at sub-arc depths: An experimental study. *Journal of Petrology*, 49(4), 717–740. <https://doi.org/10.1093/petrology/egm073>
- Hirose, K. (1997). Partial melt compositions of carbonated peridotite at 3 GPa and role of CO₂ in alkali-basalt magma generation. *Geophysical Research Letters*, 24(22), 2837–2840. <https://doi.org/10.1029/97GL02956>
- Holland, T. J. B., & Powell, R. (2004). An internally consistent thermodynamic data set for phases of petrological interest. *Journal of Metamorphic Geology*, 16(3), 309–343. <https://doi.org/10.1111/j.1525-1314.1998.00140.x>
- Hopp, J., Trierloff, M., Brey, G. P., Woodland, A. B., Simon, N. S. C., Wijbrans, J. R., et al. (2008). 40Ar/39Ar-ages of phlogopite in mantle xenoliths from South African kimberlites: Evidence for metasomatic mantle impregnation during the Kibaran orogenic cycle. *Lithos*, 106(3–4), 351–364. <https://doi.org/10.1016/j.lithos.2008.09.001>
- Hopper, E., & Fischer, K. M. (2015). The meaning of midlithospheric discontinuities: A case study in the northern U.S. craton. *Geochemistry Geophysics, Geosystems*, 18, 1541–1576. <https://doi.org/10.1002/2014GC005684>
- Hopper, E., Ford, H. A., Fischer, K. M., Lekic, V., & Fouch, M. J. (2014). The lithosphere-asthenosphere boundary and the tectonic and magmatic history of the northwestern United States. *Earth and Planetary Science Letters*, 402, 69–81. <https://doi.org/10.1016/j.epsl.2013.12.016>
- Ionov, D. A. (2007). Compositional variations and heterogeneity in fertile lithospheric mantle: Peridotite xenoliths in basalts from Tariat, Mongolia. *Contributions to Mineralogy and Petrology*, 154(4), 455–477. <https://doi.org/10.1007/s00410-007-0203-y>
- Karato, S. I. (2012). On the origin of the asthenosphere. *Earth and Planetary Science Letters*, 321–322, 95–103. <https://doi.org/10.1016/j.epsl.2012.01.001>
- Karato, S. I., Ogluogji, T., & Park, J. (2015). Mechanisms and geologic significance of the mid-lithosphere discontinuity in the continents. *Nature Geoscience*, 8, 509–514. <https://doi.org/10.1038/ngeo2462>
- Kawamoto, T., & Holloway, J. R. (1997). Melting temperature and partial melt chemistry of H₂O-saturated mantle peridotite to 11 gigapascals. *Science*, 276(5310), 240–243. <https://doi.org/10.1126/science.276.5310.240>
- Kay, S. M., Godoy, E., & Kurtz, A. (2005). Episodic Arc Migration, crustal thickening, subduction erosion and magmatism in the south Central-Andes. *Geological Society of America Bulletin*, 117(1), 67–88.

- Kelemen, P. B., Dick, H. J. B., & Quick, J. E. (1992). Formation of harzburgite by pervasive melt/rock reaction in the upper mantle. *Nature*, 358(6388), 635–641. <https://doi.org/10.1038/358635a0>
- Kelemen, P. B., Hart, S. R., & Bernstein, S. (1998). Silica enrichment in the continental upper mantle via melt/rock reaction. *Earth and Planetary Science Letters*, 164(1–2), 387–406. [https://doi.org/10.1016/S0012-821X\(98\)00233-7](https://doi.org/10.1016/S0012-821X(98)00233-7)
- Konzett, J., & Ulmer, P. (1999). The stability of hydrous potassic phases in Iherzolitic mantle—an experimental study to 9.5 GPa in simplified and natural bulk compositions. *Journal of Petrology*, 40(4), 629–652. <https://doi.org/10.1093/petroj/40.4.629>
- Konzett, J., Wirth, R., Hauzenberger, C., & Whitehouse, M. (2013). Two episodes of fluid migration in the Kaapvaal Craton lithospheric mantle associated with Cretaceous kimberlite activity: Evidence from a harzburgite containing a unique assemblage of metasomatic zirconium-phases. *Lithos*, 182, 183–184. <https://doi.org/10.1016/j.lithos.2013.10.005>
- Kopylova, M. G., Russell, J. K., & Cookenboo, H. (1999). Petrology of peridotite and pyroxenite xenoliths from the Jericho kimberlite: implications for the thermal state of the mantle beneath the Slave craton, northern Canada. *Journal of Petrology*, 40(1), 79–104. <https://doi.org/10.1093/petroj/40.1.79>
- Kumar, P., Yuan, X., Kind, R., & Mechie, J. (2012). The lithosphere-asthenosphere boundary observed with USArray receiver functions. *Solid Earth*, 3(1), 149–159. <https://doi.org/10.5194/se-3-149-2012>
- Kushiro, I. (1969). Stability of amphibole and phlogopite in the upper mantle. *Carnegie Institution of Washington, Yearbook*, 68, 245–247.
- Kushiro, I., Syono, Y., & Akimoto, S. (1967). Stability of phlogopite at high pressures and possible presence of phlogopite in earth's upper mantle. *Earth and Planetary Science Letters*, 3, 197–203. [https://doi.org/10.1016/0012-821X\(67\)90036-2](https://doi.org/10.1016/0012-821X(67)90036-2)
- Leake, B. E. (1997). Nomenclature of amphiboles: Report on the subcommittee on amphiboles of the International Mineralogical Association, Commission on New Minerals and Mineral Names. *American Mineralogist* 82, 1019–1037. <https://doi.org/10.1180/minmag.1997.061.405.13>
- Leake, B. E. (2003). Nomenclature of amphiboles: Additions and revisions to the International Mineralogical Association's amphibole nomenclature. *Canadian Mineralogist*, 41, 1355–1370.
- Lee, C.-T., Rudnick, R. L., McDonough, W. F., & Horn, I. (2000). Petrologic and geochemical investigation of carbonates in peridotite xenoliths from northeastern Tanzania. *Contributions to Mineralogy and Petrology*, 139(4), 470–484. <https://doi.org/10.1007/s004100000144>
- Lee, C.-T. A. (2003). Compositional variation of density and seismic velocities in natural peridotites at STP conditions: Implications for seismic imaging of compositional heterogeneities in the upper mantle. *Journal of Geophysical Research*, 108(B9), 2441. <https://doi.org/10.1029/2003JB002413>
- Lee, C.-T. A. (2006). Geochemical/petrologic constraints of the origin of cratonic mantle. In K. Benn, J. Mareschal, and K. Condie (Eds.), *Archaeo geodynamics and environment* (pp. 89–114). Washington, DC: American Geophysical Union.
- Lee, C.-T. A., Luffi, P., & Chin, E. J. (2011). Building and destroying continental mantle. *Annual Review of Earth and Planetary Sciences*, 39(1), 59–90. <https://doi.org/10.1146/annurev-earth-040610-133505>
- Lee, C.-T. A., & Rudnick, R. L. (1999). Compositionally stratified cratonic lithosphere: Petrology and geochemistry of peridotite xenoliths from Labait tuff cone, Tanzania. In J. J. Gurney et al. (Eds.), *Proceedings of 7th International Kimberlite Conference*, Cape Town, South Africa: Red Roof Design 7, 503–521.
- Lekić, V., & Fischer, K. M. (2014). Contrasting lithospheric signatures across the western United States revealed by Sp receiver functions. *Earth and Planetary Science Letters*, 402, 90–98. <https://doi.org/10.1016/j.epsl.2013.11.026>
- Li, Y., Dasgupta, R., Tsuno, K., Monteleone, B., & Shimizu, N. (2016). Carbon and sulfur budget of the silicate Earth explained by accretion of differentiated planetary embryos. *Nature Geoscience*, 9, 781–785. <https://doi.org/10.1038/ngeo2801>
- Litasov, K. D., Foley, S. F., & Litasov, Y. D. (2000). Magmatic modification and metasomatism of the subcontinental mantle beneath the Vitim volcanic field (East Siberia): Evidence from trace element data on pyroxenite and peridotite xenoliths from Miocene microbasalt. *Lithos*, 54(1–2), 83–114. [https://doi.org/10.1016/S0024-4937\(00\)00016-5](https://doi.org/10.1016/S0024-4937(00)00016-5)
- Lucassen, F., Franz, G., Viramonte, J., Romer, R. L., Dulski, P., & Lang, A. (2005). The late cretaceous lithospheric mantle beneath the Central Andes: Evidence from phase equilibria and composition of mantle xenoliths. *Lithos*, 82(3–4), 379–406. <https://doi.org/10.1016/j.lithos.2004.08.002>
- Luguet, A., Behrens, M., Pearson, D. G., König, S., & Herwartz, D. (2015). Significance of the whole rock Re–Os ages in cryptically and modally metasomatised cratonic peridotites: Constraints from HSE–Se–Te systematics. *Geochimica et Cosmochimica Acta*, 164, 441–463. <https://doi.org/10.1016/j.gca.2015.06.016>
- Maier, W. D., Peltonen, P., McDonald, I., Barnes, S. J., Barnes, S. J., Hatton, C., et al. (2012). The concentration of platinum-group elements and gold in southern African and Karelian kimberlite-hosted mantle xenoliths: Implications for the noble metal content of the Earth's mantle. *Chemical Geology*, 302–303, 119–135. <https://doi.org/10.1016/j.chemgeo.2011.06.014>
- Mallik, A., & Dasgupta, R. (2013). Reactive infiltration of MORB-eclogite-derived carbonated silicate melt into fertile peridotite at 3GPa and genesis of alkalic magmas. *Journal of Petrology*, 54(11), 2267–2300. <https://doi.org/10.1093/petrology/egt047>
- Mallik, A., & Dasgupta, R. (2014). Effect of variable CO₂ on eclogite-derived andesite and Iherzolite reaction at 3GPa—Implications for mantle source characteristics of alkalic ocean island basalts. *Geochemistry, Geophysics, Geosystems*, 15, 1533–1557. <https://doi.org/10.1002/2013GC005162>
- Mallik, A., Dasgupta, R., Tsuno, K., & Nelson, J. (2016). Effects of water, depth and temperature on partial melting of mantle-wedge fluxed by hydrous sediment-melt in subduction zones. *Geochimica et Cosmochimica Acta*, 195, 226–243. <https://doi.org/10.1016/j.gca.2016.08.018>
- Mallik, A., Nelson, J., & Dasgupta, R. (2015). Partial melting of fertile peridotite fluxed by hydrous rhyolitic melt at 2–3 GPa: Implications for mantle wedge hybridization by sediment melt and generation of ultrapotassic magmas in convergent margins. *Contributions to Mineralogy and Petrology*, 169, 48. <https://doi.org/10.1007/s00410-015-1139-2>
- Mandler, B. E., & Grove, T. L. (2016). Controls on the stability and composition of amphibole in the Earth's mantle. *Contributions to Mineralogy and Petrology*, 171(8–9), 68. <https://doi.org/10.1007/s00410-016-1281-5>
- Mengel, K., & Green, D. H. (1989). Stability of amphibole and phlogopite in metasomatized peridotite under water-saturated and water-undersaturated conditions. In J. Ross et al. (Eds.), *Kimberlites related rocks: Their composition, occurrence, origin, and emplacement*, Geological Society of Australia Special Publication (Vol. 1, pp. 571–581). Victoria, BC: Blackwell.
- Messbahi, H., El, J. L., Bodinier, A., Vauchez, J. M., Dautria, H., Ouali, C. J., et al. (2015). Short wavelength lateral variability of lithospheric mantle beneath the Middle Atlas (Morocco) as recorded by mantle xenoliths. *Tectonophysics*, 650, 34–52. <https://doi.org/10.1016/j.tecto.2014.11.020>
- Miller, M. S., & Eaton, D. W. (2010). Formation of cratonic mantle keels by arc accretion: Evidence from S receiver functions. *Geophysical Research Letters*, 37, L18305. <https://doi.org/10.1029/2010GL044366>
- Millhollen, G., Irving, A., & Wyllie, P. (Eds.) (1974). Melting interval of peridotite with 5.7 per cent water to 30 kilobars. *Journal of Geology*, 82(5), 575–587.

- Mitchell, R. H. (1995). Melting experiments on a sanidine phlogopite lamproite at 4–7 GPa and their bearing on the sources of lamproitic magmas. *Journal of Petrology*, 36(5), 1455–1474. <https://doi.org/10.1093/petrology/36.5.1455>
- Modreski, P., & Boettcher, A. (1972). The stability of phlogopite + enstatite at high pressures: A model for micas in the interior of the earth. *American Journal of Science*, 272(9), 852–869. <https://doi.org/10.2475/ajs.272.9.852>
- Mukhopadhyay, B., & Manton, W. I. (1994). Upper-mantle fragments from beneath the Sierra Nevada Batholith: Partial fusion, fractional crystallization, and metasomatism in a subduction related ancient lithosphere. *Journal of Petrology*, 35(5), 1417–1450. <https://doi.org/10.1093/petrology/35.5.1417>
- Mysen, B. O., & Boettcher, A. L. (1975). Melting of a hydrous mantle: I. Phase relations of natural peridotite at high pressures and temperatures with controlled activities of water, carbon dioxide, and hydrogen. *Journal of Petrology*, 16(3), 520–548. <https://doi.org/10.1093/petrology/16.1.520>
- Niida, K., & Green, H. (1999). Stability and chemical composition of pargasitic amphibole in MORB pyrolite under upper mantle conditions. *Contributions to Mineralogy and Petrology*, 135(1), 18–40. <https://doi.org/10.1007/s004100050495>
- Nixon, P. H., Rogers, N. W., Gibson, I. L., & Grey, A. (1981). Depleted and fertile mantle xenoliths from southern African kimberlites. *Annual Review of Earth and Planetary Sciences*, 9(1), 285. <https://doi.org/10.1146/annurev.ea.09.050181.001441>
- Olafsson, M., & Eggler, D. H. (1983). Phase relations of amphibole, amphibole-carbonate, and phlogopite-carbonate peridotite: petrologic constraints on the asthenosphere. *Earth and Planetary Science Letters*, 64(2), 305–315.
- Parman, S. W., & Grove, T. L. (2004). Harzburgite melting with and without H₂O: Experimental data and predictive modeling. *Journal of Geophysical Research*, 109, B02201. <https://doi.org/10.1029/2003JB002566>
- Pearson, D. G., & Wittig, N. (2013). The formation and evolution of cratonic mantle lithosphere: Evidence from mantle xenoliths. In R. W. Carlson and H. D. Holland (Eds.), *The mantle and core*, Treatise on Geochemistry (Vol. 3, pp. 225–292). Oxford, UK: Elsevier.
- Pearson, D. G., Shirey, S. B., Carlson, R. W., Boyd, F. R., Pokhilenko, N. P., & Shimizu, N. (1995). Re–Os, Sm–Nd, and Rb–Sr isotope evidence for thick Archaean lithospheric mantle beneath the Siberian craton modified by multistage metasomatism. *Geochimica et Cosmochimica Acta*, 59(5), 959–977. [https://doi.org/10.1016/0016-7037\(95\)00014-3](https://doi.org/10.1016/0016-7037(95)00014-3)
- Pike, J. E. N., Meyer, C. E., & Wilshire, H. G. (1980). Petrography and chemical composition of a suite of Ultramafic xenoliths from Lashaine, Tanzania. *Geological Notes*, 88(3), 343–352. <https://doi.org/10.1086/628512>
- Pirard, C., & Hermann, J. (2015a). Experimentally determined stability of alkali amphibole in metasomatised dunite at sub-arc pressures. *Contributions to Mineralogy and Petrology*, 169(1), <https://doi.org/10.1007/s00410-014-1095-2>
- Pirard, C., & Hermann, J. (2015b). Focused fluid transfer through the mantle above subduction zones. *Geology*, 43(10), 915–918. <https://doi.org/10.1130/G37026.1>
- Rader, E., Emry, E., Schmerr, N., Frost, D., Cheng, C., Menard, J., et al. (2015). Characterization and petrological constraints of the mid-lithospheric discontinuity. *Geochemistry, Geophysics, Geosystems*, 16, 3484–3504. <https://doi.org/10.1002/2015GC005943>
- Romanowicz, B. (2009). Thickness of the tectonic plates. *Science*, 324(5926), 474–476. <https://doi.org/10.1126/science.1172879>
- Rudnick, R. L., Gao, S., Ling, W. L., Liu, Y. S., & McDonough, W. F. (2004). Petrology and geochemistry of spinel peridotite xenoliths from Hanguoba and Qixia, North China craton. *Lithos*, 77(1–4), 609–637. <https://doi.org/10.1016/j.lithos.2004.03.033>
- Rudnick, R. L., McDonough, W. F., & Orpin, A. (1994). *Northern Tanzanian peridotite xenoliths: A comparison with Kapvaal peridotites and inferences of metasomatic reactions*. Paper presented at Proceedings of the 5th International Kimberlite Conference Araxa, Braz, pp. 336–353.
- Rychert, C. A., & Shearer, P. M. (2009). A global View of the Lithosphere-Asthenosphere Boundary. *Science*, 324(5926), 495–448. <https://doi.org/10.1126/science.1169754>
- Sato, K., Katsura, T., & Ito, E. (1997). Phase relations of natural phlogopite with and without enstatite up to 8 GPa: implication for mantle metasomatism. *Earth and Planetary Science Letters*, 146(3–4), 511–526. [https://doi.org/10.1016/S0012-821X\(96\)00246-4](https://doi.org/10.1016/S0012-821X(96)00246-4)
- Savage, B., & Silver, P. G. (2008). Evidence for a compositional boundary within the lithospheric mantle beneath the Kalahari craton from S receiver functions. *Earth and Planetary Science Letters*, 272(3–4), 600–609. <https://doi.org/10.1016/j.epsl.2008.05.026>
- Scott, J. M., Waight, T. E., van der Meer, Q. H. A., Palin, J. M., Cooper, A. F., & Munker, C. (2014). Metasomatized ancient lithospheric mantle beneath the young Zealandia microcontinent and its role in HIMU-like intraplate magmatism. *Geochemistry, Geophysics, Geosystems*, 15, 3477–3501. <https://doi.org/10.1002/2014GC005300>
- Selway, K., Ford, H., & Kelemen, P. (2015). The seismic mid-lithosphere discontinuity. *Earth and Planetary Science Letters*, 414, 45–57. <https://doi.org/10.1016/j.epsl.2014.12.029>
- Sodoudi, F. (2013). LAB and other lithospheric discontinuities below Cratons. In *EGU General Assembly* (EGU2013–4378, Vol. 15, p. 4378). Vienna, Austria: EGU General Assembly.
- Sodoudi, F., Yuan, X., Kind, R., Lebedev, S., Adam, J. M. C., Kästle, E., et al. (2013). Seismic evidence for stratification in composition and anisotropic fabric within the thick lithosphere of Kalahari Craton. *Geochemistry, Geophysics, Geosystems*, 14, 5393–5412. <https://doi.org/10.1002/2013GC004955>
- Syracuse, E. M., van Keken, P. E., Abers, G. A., Suetsugu, D., Bina, C., Inoue, T., et al. (2010). The global range of subduction zone thermal models. *Physics of the Earth and Planetary Interiors*, 183(1–2), 73–90. <https://doi.org/10.1016/j.pepi.2010.02.004>
- Tenner, T. J., Hirschmann, M. M., & Humayun, M. (2012). The effect of H₂O on partial melting of garnet peridotite at 3.5 GPa. *Geochemistry, Geophysics, Geosystems*, 13, Q03016. <https://doi.org/10.1029/2011GC003942>
- Thibault, Y., Edgar, A. D., & Lloyd, F. E. (1992). Experimental investigation of melts from a carbonated phlogopite ilmenite: Implications for metasomatism in the continental lithospheric mantle. *American Mineralogist*, 77, 784–794.
- Thybo, H., & Perchuc, E. (1997). The seismic 8° discontinuity and partial melting in continental mantle. *Science*, 275(5306), 1626–1629. <https://doi.org/10.1126/science.275.5306.1626>
- Till, C. B., Grove, T. L., & Withers, A. C. (2012). The beginnings of hydrous mantle wedge melting. *Contributions to Mineralogy and Petrology*, 163(4), 669–688. <https://doi.org/10.1007/s00410-011-0692-6>
- Tsuno, K., & Dasgupta, R. (2011). Melting phase relation of nominally anhydrous, carbonated pelitic-eclogite at 2.5–3.0 GPa and deep cycling of sedimentary carbon. *Contributions to Mineralogy and Petrology*, 161, 743–763. <https://doi.org/10.1007/s00410-010-0560-9>
- Tsuno, K., & Dasgupta, R. (2012). The effect of carbonates on near-solidus melting of pelite at 3GPa: Relative efficiency of H₂O and CO₂ subduction. *Earth and Planetary Science Letters*, 319–320, 185–196. <https://doi.org/10.1016/j.epsl.2011.12.007>
- Tsuno, K., & Dasgupta, R. (2015). Fe–Ni–Cu–C–S phase relations at high pressures and temperatures - The role of sulfur in carbon storage and diamond stability at mid- to deep-upper mantle. *Earth and Planetary Science Letters*, 412, 132–142. <https://doi.org/10.1016/j.epsl.2014.12.018>
- Tumati, S., Fumagalli, P., Tiraboschi, C., & Poli, S. (2013). An experimental study on COH-bearing peridotite up to 3.2 GPa and implications for crust-mantle recycling. *Journal of Petrology*, 54(3), 453–479. <https://doi.org/10.1093/petrology/egs074>

- Van Keken, P. E., Kiefer, B., & Peacock, S. M. (2002). High-resolution models of subduction zones: Implications for mineral dehydration reactions and the transport of water into the deep mantle. *Geochemistry, Geophysics, Geosystems*, 3(10), 1056. <https://doi.org/10.1029/2001GC000256>
- Wallace, M. E., & Green, D. H. (1991). The effect of bulk rock composition on the stability of amphibole in the upper mantle: Implications for solidus positions and mantle metasomatism. *Mineralogy and Petrology*, 44(1–2), 1–19. <https://doi.org/10.1007/BF01167097>
- Wallace, M. E., & Green, D. H. (1988). Experimental determination of primary carbonatite magma composition. *Nature*, 335(6188), 343–346. <https://doi.org/10.1038/335343a0>
- Wartho, J., & Kelley, S. P. (2003). 40Ar/39Ar ages in mantle xenolith phlogopites: Determining the ages of multiple lithospheric mantle events and diatreme ascent rates in southern Africa and Malaita, Solomon Islands. *Geological Society Special Publications* 220(1), 231–248. <https://doi.org/10.1144/GSL.SP.2003.220.01.14>
- Wendlandt, R. F., & Eggler, D. H. (1980). The origins of potassic magmas: 2. Stability of Phlogopite in natural spinel lherzolite and in the system KAlSiO₄-MgO-SiO₂-CO₂ at high pressures and high temperatures. *American Journal of Science*, 280(5), 421–458. <https://doi.org/10.2475/ajs.280.5.421>
- Winterburn, P. A., Harte, B., & Gurney, J. J. (1990). Peridotite xenoliths from the Jagersfontein kimberlite pipe: I. Primary and primary-metasomatic mineralogy. *Geochimica et Cosmochimica Acta*, 54(2), 329–341. [https://doi.org/10.1016/0016-7037\(90\)90322-C](https://doi.org/10.1016/0016-7037(90)90322-C)
- Wirth, E. A., & Long, M. D. (2014). A contrast in anisotropy across mid-lithospheric discontinuities beneath the central United States-A relic of craton formation. *Geology*, 42(10), 851–854. <https://doi.org/10.1130/G35804.1>
- Witt, G., & Seck, H. A. (1987). Temperature history of sheared mantle xenoliths from the West Eifel, West Germany: Evidence for mantle diapirism beneath the rhenish massif. *Journal of Petrology*, 28(3), 475–493. <https://doi.org/10.1093/petrology/28.3.475>
- Wittlinger, G., & Farra, V. (2007). Converted waves reveal a thick and layered tectosphere beneath the Kalahari super-craton. *Earth and Planetary Science Letters*, 254(3–4), 404–415. <https://doi.org/10.1016/j.epsl.2006.11.048>
- Wolbern, I., Rumpker, G., Link, K., & Sodoudi, F. (2012). Melt infiltration of the lower lithosphere beneath the Tanzania craton and the Albertine rift inferred from S receiver functions. *Geochemistry, Geophysics, Geosystems*, 13, Q0AK08. <https://doi.org/10.1029/2012GC004167>
- Xu, R., Liu, Y., Tong, X., Hu, Z., Zong, K., & Gao, S. (2013). In-situ trace elements and Li and Sr isotopes in peridotite xenoliths from Kuandian, North China Craton: Insights into Pacific slab subduction-related mantle modification. *Chemical Geology*, 354, 107–123. <https://doi.org/10.1016/j.chemgeo.2013.06.022>
- Xu, Y.-G., Ross, J. V., & Mercier, J.-C. (1993). The upper mantle beneath the continental rift of Tanlu, eastern China: evidence for the intra-lithospheric shear zones. *Tectonophysics*, 225(4), 337–360. [https://doi.org/10.1016/0040-1951\(93\)90304-3](https://doi.org/10.1016/0040-1951(93)90304-3)
- Yaxley, G. M., Green, D. H., & Kamenetsky, V. (1998). Carbonatite metasomatism in the southeastern Australian lithosphere. *Journal of Petrology*, 39(11–12), 1917–1930. <https://doi.org/10.1093/ptro/39.11-12.1917>
- Yoder, H. S., & Kushiro, I. (1969). Melting of a hydrous phase: Phlogopite. *American Journal of Science*, 267–A, 558–582.
- Yuan, H., & Romanowicz, B. (2010). Lithospheric layering in the North American craton. *Nature*, 466(7310), 1063–1068. <https://doi.org/10.1038/nature09332>
- Zhang, M., Suddaby, P., Reilly, S. Y. O., Norman, M., & Qiu, J. (2000). Nature of the lithospheric mantle beneath the eastern part of the Central Asian fold belt: mantle xenolith evidence. *Tectonophysics*, 328(1–2), 131–156. [https://doi.org/10.1016/S0040-1951\(00\)00181-5](https://doi.org/10.1016/S0040-1951(00)00181-5)