The Significance of Seismic Wavespeed Minima and Thermal Maxima in the Mantle and the Role of Dynamic Melting

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Abstract

It is widely assumed that the boundary layer above the core is the source of intraplate volcanoes such as Hawaii, Samoa and Yellowstone and that the sub-plate boundary layer at the top of the mantle is thin and entirely subsolidus. In fact, this layer is thicker and has higher expansivity, buoyancy and insulating power than the lower one, and may have higher potential temperatures. The observed seismic structure of the low-velocity zone (LVZ) including attenuation, anisotropy, sharp boundaries and a reduction of both compressional and shear moduli can be taken as strong evidence for the ubiquitous presence of melt in the upper mantle. If the LVZ contains as little as 1-2% melt then it is the most plausible and accessible source for mid-plate magmas; deeply rooted active upwellings are unnecessary. The upper boundary layer is also the most plausible source of ancient isotopic signatures of these magmas and their inclusions.

Introduction

Seismological arguments for a melt component in the upper mantle extend back nearly 50 years (1-3). It was realized that, while normal P-T gradients in the mantle cause a minimum of seismic wavespeeds at depths between about 100 and 200 km, the actual properties of this region require something more. By 1970 geophysicists had shown that the missing ingredient is probably athermal, such as stress- and volatile-enhanced melting (1-6). Low velocity zones in the mantle are now routinely attributed to dehydration melting although CO_2 and dynamic melting may play important roles. The theory for the interaction of elastic waves with partially molten rock was well developed prior to 1970 (1, 5, 7), and provides a self-consistent explanation of reduced seismic velocities, increased anisotropy, and high attenuation of seismic waves in the low velocity zone. An elastic wave can be viewed as a time dependent stress and temperature perturbation to the thermodynamic state of the medium (1, 7, 8). In regions near the melting point it can promote a small amount of crystallization or melting. The associated volume change introduces a strain in the medium, which lowers the ratio of stress to strain resulting in a reduced seismic velocity. This magnifies the effect of a small amount of melt on wavespeeds beyond that given by simple static volume averaging (1, 5, 9).

The precipitous change in elastic velocities at the onset of partial melting explains the sharp variable-depth boundaries of the LVZ, the G and the L discontinuities. The global G discontinuity at the top of the LVZ is an abrupt 7-8% reduction in short-period shear-wave speeds, which has been attributed to sub-horizontal melt-rich lamellae (10). Comparable drops are inferred for compressional waves. The existence of converted phases from the lid-LVZ interface provides evidence that the interface is sharp, not diffuse as is expected at the lithosphere-asthenosphere boundary, which represents the transition from long-term strength to weakness on geological time scales. Waveform modeling indicates seismic velocity drops of up to 10–20% (11). Beneath the central Pacific, shear wave reverberations imply abrupt 5-14 % velocity drops at depths that vary from 66 to 80 km (12), and a further 8–9% gradual decrease from the top of the LVZ to about 160 km (13,14). Below continental shields, the upper boundary of the LVZ is depressed to 150 km.

If G and L represent boundaries of the partial melt zone (9, 12-14), the implication is that volatiles have drained upwards and the melt content decreases with depth below the axis of the LVZ (Fig. 1). Melts can be shear-driven and compaction-driven as well as buoyancy-driven. In other words, volcanoes can be, and probably are, the result of stresses and fractures associated with plate tectonics rather than localized 'hotspots'. This tectonic paradigm has been challenged recently by arguments that melts do not occur in the LVZ, that they drain out quickly, or that the effect is too small to explain the seismic observations (15-17). These arguments ignore the importance of interactions of stress waves with small melt fractions (5, 18) in the transmitting media, interactions of migrating melts with the matrix and the role of impermeable interfaces.

The competing paradigm is that only mantle plumes are hot enough to produce melt and to have sufficiently deep roots to supply magmas with ancient-enriched and "undegassed" mantle isotope signatures. These geochemical inferences are based on implausible assumptions about ambient mantle, the geotherm, the mode of mantle convection, and the nature and physical state of the surface boundary layer (19), which is usually ignored.

Analog experiments

The response of a partially molten solid to the transmission of elastic waves is different from that of a fluid-saturated but unreactive porous solid (1, 8). Several studies have determined the variation of elastic wave velocities across the melting point (1, 17, 18-23) (Fig. 2). Note the abrupt drops in wavespeeds at the eutectic temperature. Similar effects may explain the G - or Gutenberg discontinuity, and other LVZs associated with phase changes in the deeper mantle, but a solid-melt interface is not required) (7, 8). Additional details on relevant analog experiments and a discussion of deep LVZs are in the Appendix.

Geochemistry

The upper boundary layer of the mantle is composed primarily (>97%) of refractory high-melting lithologies, e.g. dunite or harzburgite, that collect at the top of the mantle because they are buoyant and which may trap fragments of ancient mantle, including high $^3He/^4He$ gases or low-U,Th inclusions (9, 24, 25). The interleaved low rigidity layers probably account for 1-2 % of the volume, and may be pyroxenite, magma-mush lenses, sills, metasomatic lamellae or fine-grained shear zones, which are large-scale versions of the veins in metasomatized 'lithosphere'. Trace "exotic" components, such as sulfides, oxides, carbonatites and fluid inclusions collect in the surface layers because of their solubility, density, mobility and volatility and appear to resolve the various U-Pb-He-heatflow, ocean island basalt and delayed core-formation paradoxes (19, 24, 26-28). Although the highest potential temperatures in the mantle are likely to coincide with the LVZ (24, 29), low-melting fertile lithologies such as dense eclogites and pyroxenites collect at the base of the transition zone. The potential temperature of any volume in the mantle is the temperature that that volume would have if it were compressed or expanded to some reference pressure. It is often erroneously assumed in the petrological literature that this is the extension of an actual sub-plate adiabatic geotherm.

The canonical models of mantle geochemistry and petrology assume that, except under ridges (melting due to decompression melting), trenches (melting due to water induced melting) and hotspots (melting due to anomalous temperature), the upper mantle is entirely subsolidus (5, 30, 31) and that the boundary layer above the core is the most plausible source of intraplate volcanoes such as Hawaii, Samoa and Yellowstone (30, 32-34). The so-called basal mélange is assumed to be the largest, hottest, most accessible, least degassed and most "primordial" (defined as having $^3He/^4He$ ratios higher than an average mid-ocean-range basalt MORB) part of the mantle. From classical physics and logical points of view (19) these assumptions are based on a number of questionable premises: 1) upper mantle geotherms cannot cross the solidi of mantle minerals, and cannot exceed ~1400 ^oC or temperatures assigned to sub-ridge mantle (31, 35-38), 2) high $^3He/^4He$ ratios in mid-plate magmas relative to those in average MORB are due to excess ^3He (not low ^4He), and 3) most of the mantle supports an adiabatic gradient and is not cooling with time. However, geotherms depend on thermal history and on the distribution of radioactive elements. Internal heating, and by inference, secular cooling modulated by internal heating, lead to a temperature maximum in the shallow mantle, and a subadiabatic gradient throughout most of the mantle (29, 39-41). This thermal max results in an inverted geotherm, or overshoot, which probably coincides with the LVZ (Fig. 1). The base of the surface boundary layer and the base of the mantle cannot be assumed to be non-cooling isotherms (31, 41).

High $^3He/^4He$ ratios in mid-plate magmas relative to those in average MORB cannot be taken as evidence that they have a deep origin (9, 24, 26). Basalt isotope chemistry is controlled by mixing of components (26), but both mixing relations and isotope evolution trajectories are indifferent as to the location, depth, size and absolute compositions of the components. This means that the mixture of a high $^3He/^4He$-low ^3He component, long resident in the shallow mantle, and a high ^3He-low $^3He/^4He$ component such as MORB is indistinguishable from the signatures attributed to deep undegassed or primordial mantle sources.

The Geotherm

Curve fitting assumptions, adjustments of the data, anelastic corrections, experimental errors and theoretical oversights have led to the view that ambient midplate upper mantle is colder and less variable than a straightforward analysis of bathymetric and seismic data would indicate (18-21). Cambridge geotherms, for example, are required to converge at depth, and to be asymptotic to the MORB adiabat (31, 35). In other words, temperatures are forced to bend as they approach 1400 °C or an a priori adiabat. Such constrained temperatures are more than 300 °C lower than would be inferred from the same data without this enforced bend (36, 37). This has led to the view that temperatures that are higher than the constrained adiabats are anomalous and require deep mantle sources.

Both vertical and lateral temperature gradients are high in conduction boundary layers. Basalts derived from 150 km depth in a mature boundary layer will, in general, be hotter than those derived from 60 km depth but may involve lower extents of melting (42). A given isotherm deepens with time. There are several lines of evidence that sub-boundary layer temperatures, however, increase away from ridges toward plate interiors. The central Pacific has much lower upper mantle wavespeeds and higher inferred temperatures than other oceans and than global reference models (43). Near-ridge mantle below 200 km depth has higher shear velocities, on average, than midplate upper mantle (24, 43, 44). Bathymetry data yields mantle temperatures for ‘normal’ regions of the north Pacific that are ~200°C higher than the constrained temperatures (16, 31, 38, 45). Subsidence rates and residual bathymetry of the younger portions of plates imply that sub-ridge mantle is denser, on average, and in the southern oceans, than mantle under older plates. Geophysical data are consistent with petrological inferences that near-ridge mantle and midplate mantle temperatures differ, on average, by >100°C but also show that the higher temperatures are widespread and reflect ambient mantle under plates while lower temperatures appear to preferentially occur under young oceanic plates or to be sampled by midplate volcanoes that extract magma from the upper parts of the boundary layer. Hawaii, therefore, is not a localized thermal anomaly associated with a plume; it is a small-scale sample of a deep portion of ambient midplate boundary layer mantle (24, 46, 47). Its location and magma output are controlled by a step in the thickness of the plate at the wide Molokai fracture zone (48), and by stresses in the plate, not by conditions at the core-mantle boundary.

The maximum depth of melting, and inferred source temperature, of oceanic magmas increases, and the fraction of melt decreases, with plate age (42, 46). The P-T-depth-age trajectories calculated for midplate magmas track the 1500-1600°C cooling half-space isotherms, rather than the predicted 1300°C horizontal isotherm (24, 45). Petrological and seismic data are consistent with midplate magmas being extracted from within and near the base of the boundary layer, which has maximum temperatures some ~200°C higher than generally assumed for ambient mantle. On the other hand the low temperatures, low diffusivity, and high strength and buoyancy of the outer part of the shell mean the upper boundary layer is also the most plausible place to preserve ancient enriched, isotopic signatures. These considerations turn canonical geochemical models on their heads (30, 32-34, 49) and remove what have been considered as geochemical paradoxes and inconsistencies between geochemical and geophysical models.

Discussion

It is significant that the magnitude of the anisotropy in the LVZ is about the same as the drop in wavespeed at the lid-LVZ boundary (the G-discontinuity). This supports the sheared multiphase aggregate model with low-rigidity oriented lamellae (10, 24, 50). In such a model, melt segregates into sub-horizontal layers that lubricate plate motion. This interpretation is further supported by observations of strong heterogeneity above 200 km depth and the numerous reflections, both S- wave and P-wave between 100 and 200 km depth. The number of reflections in this depth range exceeds the number observed near 400 km depth (51). Variations in V_P of 7.8%, in Vs of 14.5% and in Vp/Vs of 7.2% correspond to very high temperature variations, 850 to 960 K, or small amounts of melt.

Anisotropy and strong reflections at the boundaries of the LVZ are best explained by large lenses of aligned/segregated melt accumulations [LLAMA] (9, 10, 24, 50). The interface (the G-discontinuity) and the L-discontinuity correspond to solidi, solidification or zone-refined fronts or volatile-no-volatile boundaries. Others have equated the much thinner mechanical boundary layer and the lithosphere to the lid (16) and sought to explain rheological transitions in terms of water content, grain size or thermal/compositional effects (36, 52, 53). The sharp boundaries, anisotropy and bulk modulus of the LVZ cannot be explained with these effects or with simple static volume averaging schemes.

Melt-rich layers in a sheared mélange are tilted relative to the shear plane (54) This explains shear-wave splitting and the apparent tilts of low wavespeed features in the mantle. It has been argued that that gravity will effectively drain out all the melt along these tilted interfaces and that gravity–driven fluid dynamic instabilities will destroy any such layers that survive (15). However, the melt channels are probably not continuous, they react with the matrix and they are constantly reforming in the shear field.

Volume changing solid-solid and liquid-solid phase transformations can be effective in lowering elastic moduli (7, 55). Even if such volume changes are small, their effect on the stress-strain relation can be large. The main question is, can a high frequency seismic wave change the thermodynamic state fast enough so that volume changes occur in the medium? Since the effect is measured at laboratory frequencies (1, 21), the answer is probably yes. The fact that the absorption band overlaps the seismic band (9) supports this conclusion.

If melt is concentrated in thin lamellae, only about 1-2 vol. % is required to account for the amplitudes of the observed reflections and anisotropy (9, 10, 24). This is greater than the static equilibrium melt fraction at the mantle temperatures and compositions that are assumed in petrological models (16). However, the BL may be hotter or of different composition than in the experiments (6, 19, 46). The melt fraction in LLAMA may also be enhanced by accumulation of melts that have migrated or been sheared in from greater depths or zone-refined in from above.

High-resolution seismic imaging (10, 56-59) and rock physics experiments (50, 54), give strong support to the type of model discussed here. A heterogeneous sheared boundary layer source model not only explains a variety of geophysical data but has provided evidence that midplate volcanoes are often underlain by higher wavespeeds at shallow depths than occur under ridges and some volcano-free areas (43, 44, 57), suggesting that melt extraction may have created high wavespeed areas. Large offsets between deep low wavespeed mantle features and volcanoes are commonly observed suggesting that they may be unrelated.

Summary

The causes of low velocities, high attenuation and anisotropy in the Earth's upper mantle have recently become controversial. The plate tectonic paradigm for intraplate volcanism, which involves stress release of indigenous melts in the LVZ, has been challenged with arguments that assert that melts do not occur in the LVZ or that if they do they drain out quickly or that even if they don't drain out the effect is too small to explain the seismic observations (15-17). The theory used, however, does not explain the laboratory data on partially molten materials that motivated the idea in the first place (1, 21). This apparent paradox is due to the neglect of the pertinent physics. Models that claim to rule out upper mantle melts assume that a melt phase is an unreactive component that can be accounted for by volume averaging and conclude that small amounts of melt do not have a large effect on seismic velocities (15, 16). These studies ignore experiments and theory that show the ability of small amounts of melt to dramatically modify physical properties, including permeability, through chemical effects (7, 8). The interaction between seismic waves and melt, and of migrating melts with the matrix, are significant. The effects of fluids are also not entirely microscopic; the presence of large lamellae and sills are consistent with observed seismic anisotropy in the LVZ (10, 24).

Taken as a whole, the seismic structure of the LVZ, including attenuation, sharp boundaries, and anisotropy are best explained using a dynamic melting model. The implication is that the LVZ contains melt and that it is not only a plausible source, but it is the most likely source for mid-plate and large igneous province volcanism (4, 60, 61). Sources deeper than ~300 km may be too cold (24, 29, 39), to explain the hottest Hawaiian basalts (46).

LVZs that occur sporadically near 400 and 700 km depth (63, 64) may have similar explanations (2, 23) and do not require recycling of water and crustal components to the base of the mantle and back through a water filter [see S12]. Many geochemical and geodynamic models assume whole mantle convection at the outset, and ignore the effects of dynamic melting and CO_2 when interpreting LVAs (62, 64).

Appendix 1 - Dynamic melting

It is now well established that seismic waves can interact with phase changes and that this lowers seismic wavespeeds (1, 7, 8). Several experimental studies of elastic properties during the melting process (1, 18, 22, 23) were used in earlier conclusions about the role of melts in the LVZ. One investigation (1) used a salt-water mixture that produced a binary melting relation (Fig. 2). Two compositions were studied, 1% and 2% NaCl solutions, that yielded a factor of two difference in the amounts of melt (F) at a given temperature. It was found that at the eutectic temperature of the 1% [2%] system, the formation of 3% [6%] liquid, was accompanied by a shear modulus drop of about 25% [60%] and a bulk modulus drop of 8% [30%]. Furthermore, for the same melt content, the bulk modulus decrease in the low-salt system, relative to the unmelted solid, was about 1/3 and the shear modulus decrease was 1/2 of that for the high-salt system. In other words, wavespeeds are not simply related to melt content.

Laboratory (1, 7, 8, 18, 21-23) and seismic results (10-12, 14) challenge the standard static two-phase aggregate models (15) for partial melting in three respects: 1) The shear, longitudinal and bulk moduli all abruptly decrease at the onset of melting and by more than that due to the geometric effects of such small melt volumes, 2) The moduli drops are not determined only by the fraction of melt present and the melt geometry. The dynamic melting model (21) is compatible with these observations since it is the pressure derivative, dF/dP, not F, that dictates the amount of decrease, and 3) The observed velocity changes for P waves and S waves are comparable, both at the top and the base of the LVZ and the inferred changes in bulk modulus, are significant (8, 21-23). The onset of melting of peridotite (21) is associated with a relative drop in P velocity that is equal to the relative drop in shear velocity. A seismic wave interacting with a part of the mantle at its solidus is slowed more than a wave interacting with an equal amount of melt at higher temperatures (1, 7, 8, 18, 21-23), consistent with dynamic melting, which depends on *dF/dP*. This interaction also explains the large effect on bulk modulus and the anelasticity of the LVZ. Hence, laboratory studies support the dynamic melting model but are incompatible with inert two-phase aggregate models (15).

While there are a variety of mechanisms that can explain decreasing seismic wavespeeds with depth (9, 24, 35, 36, 65), the logically dubious argument is sometimes made that since such mechanisms exist, the partial melt explanation can be ruled out (15). Many of these mechanisms, however, are concerned only with relaxation of the shear modulus and do not explain the seismic discontinuities at the boundaries of the LVZ, the strong seismic anisotropy in the LVZ and the laboratory experiments on dynamic melting that led to the suggestion in the first place (1, 4). At the other extreme, some studies assume that any LVZ is due to dehydration melting and confirms the presence of large amounts of water at depth in the mantle.

Some arguments against partial melting refer to laboratory studies that claim to observe large reductions in wavespeeds at subsolidus conditions. However, these studies actually involved melting at grain boundaries or unacounted for losses to the apparatus (18, 65). Minimum LVZ wavespeeds have been overestimated and large subsolidus temperature derivatives for Vs have been used in the arguments against the need for partial melting in the LVZ (15, 35, 36); low-resolution tomographic models do not recover the lowest wavespeeds in the boundary layer (13, 56). Tomography, by its nature, averages over large volumes. Melting in the LVZ of the upper boundary layer appears to be unavoidable, even for the low ambient temperatures and relatively refractory compositions that have been adopted in laboratory experiments (16, 21, 46).

Two of the most recent arguments that have been used against the presence of melt in the shallow mantle assume an unsheared homogeneous matrix with no permeability barriers, and extraordinarily efficient melt extraction mechanisms (15) ; 1) melts in the mantle do not wet grain boundaries and hence the ability of partial melt to influence physical properties is limited, and 2) it difficult to retain melt in a gravity field; gravity will effectively drain out all the melt and gravity–driven fluid dynamic instabilities will destroy melt-rich layers. The same arguments, if valid, could be used against the water filter model (62), that assumes that melts accumulate at 410 and 700 km depths.

Arguments to the effect that partial melt models for the low velocity zone can be ruled out based on the wettability of grain boundaries, the aspect ratio of thin film melts and dihedral angles are viewing the phenomenon at the wrong scale. In a polyphase mélange the melt pockets are sheared into discontinuous lamellae that have the same effects on long wavelength seismic waves as grain boundary films but which are independent of wetting angles and surface tension (10, 24).

Appendix 2 - The Transition Zone water filter model and dehydration melting

In addition to the low-velocity zone at the top of the mantle, others have been detected at depths of roughly 350-370, 400-410 and 600-700 km (46, 62-66). These have been explained by CO_2 and the accumulation of eclogite, or other crustal components. In the water filter model (62) it is assumed that broad upwelling currents dehydrate and melt as they pass through the 410-km discontinuity, leaving water, melt and imputitities behind. Lower mantle LVZs are attributed to diffuse downwellings that dehydrate as they sink below 650 km (64), whole mantle convection being assumed. Transition zone properties, however, are consistent with cold mantle accumulating above, depressing the 650-km discontinuity, displacing older warmer mantle upwards, elevating the 410; they are not consistent with whole mantle convection with throughgoing slabs and hot plumes (19). Alternative explanations of deep LVAs are CO_2, segregated basalt, metastability, underplating and interaction of the seismic waves with phase changes. None of these require whole mantle convection, deep slab penetration or transport of water into the lower mantle and then back again to 410 km (9, 24).

Low-velocity anomalies (LVAs) are often simply attributed to excess temperature or water content, small grain size or decompression and dehydration melting, but the actual situation is much more complex and requires mechanisms for causing these phenomena. A horizontal LVZ can be due to the effects of CO_2, ponding under a permeability barrier or a negative Clapyron slope boundary, shearing, metastable phases or the dynamic effects discussed in this paper. Deep LVAs can form at solid-solid phase boundaries and do not require the presence of either water or of melt (7, 8).

The transition-zone water filter and dehydration melting models (62, 64) assume that water is the main 'impurity' that lowers melting points and that the transition zone is the major water reservoir in the mantle. Melts and impurities accumulate above and below the transition zone but not in the shallow LVZ (15). In these models, the global mantle flow pattern is dominated by slab-related localized downwelling currents and diffuse upwelling flow. This is the precise opposite of the mantle plume model, which assumes narrow focused upwellings (plumes) and diffuse downflow. Alternatively, ancient ambient depleted mantle at the base of the TZ is forced up by the downward flux of subducting slabs and becomes the passive depleted upwellings that fuel midocean ridges and near-ridge hotspots. Those that rise midplate interact with the surface boundary layer and pick up the impurities and chemical components that define midplate basalts (24). Volatiles and impurities are sheared into the surface boundary layer and this, not the regions above the TZ or the core, is the source of Hawaii, Samoa, Yellowstone and other intraplate volcanoes.

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Figure Captions

Figure 1. Seismic velocities in the LVZ (panel A; modified from refs. 14 and 43) are a function of T and the number-density of melt-rich lamellae. The velocities V_{SV} of SV waves are mainly controlled by the low-velocity melt-rich or low-rigidity lamellae illustrated schematically in panel B. Decreases in velocities are caused by large positive temperature gradients and partial melting. The SV-SH splitting is caused by melt-rich lamellae shown schematically in panel C, which are sheared by plate motion as indicated by the arrow. The strong increase in velocities below about 150 km, which has been considered enigmatic (36), is due to subadiabatic or negative temperature gradients (39-41) (panel B).

Figure 2. Decreases in longitudinal and shear velocities across the solidus in dilute (1% and 2%) $NaCl$-H_2O solutions (*1*). The temperatures are scaled to the eutectic temperature (T* = -21.1 C) as (T-T*)/|T*|. The velocities VL and VS are scaled to V_0, the corresponding values in pure ice at the lowest measured temperature (T-T*)/|T*| = -0.6. The percent of liquid melt relative to solid ice is also indicated. This provides an analog to the G discontinuity that separates the high wavespeed seismic lid from the low velocity zone (LVZ). The theory that explains these laboratory results has been confirmed at seismic frequencies (*21*). The effects of aligned melt-rich lamellae and dynamic melting accentuate the effects of small degrees of melt, particularly on the longitudinal and bulk moduli. In addition, melt migration may increase the amount of melt relative to equilibrium partial melting calculations (16).

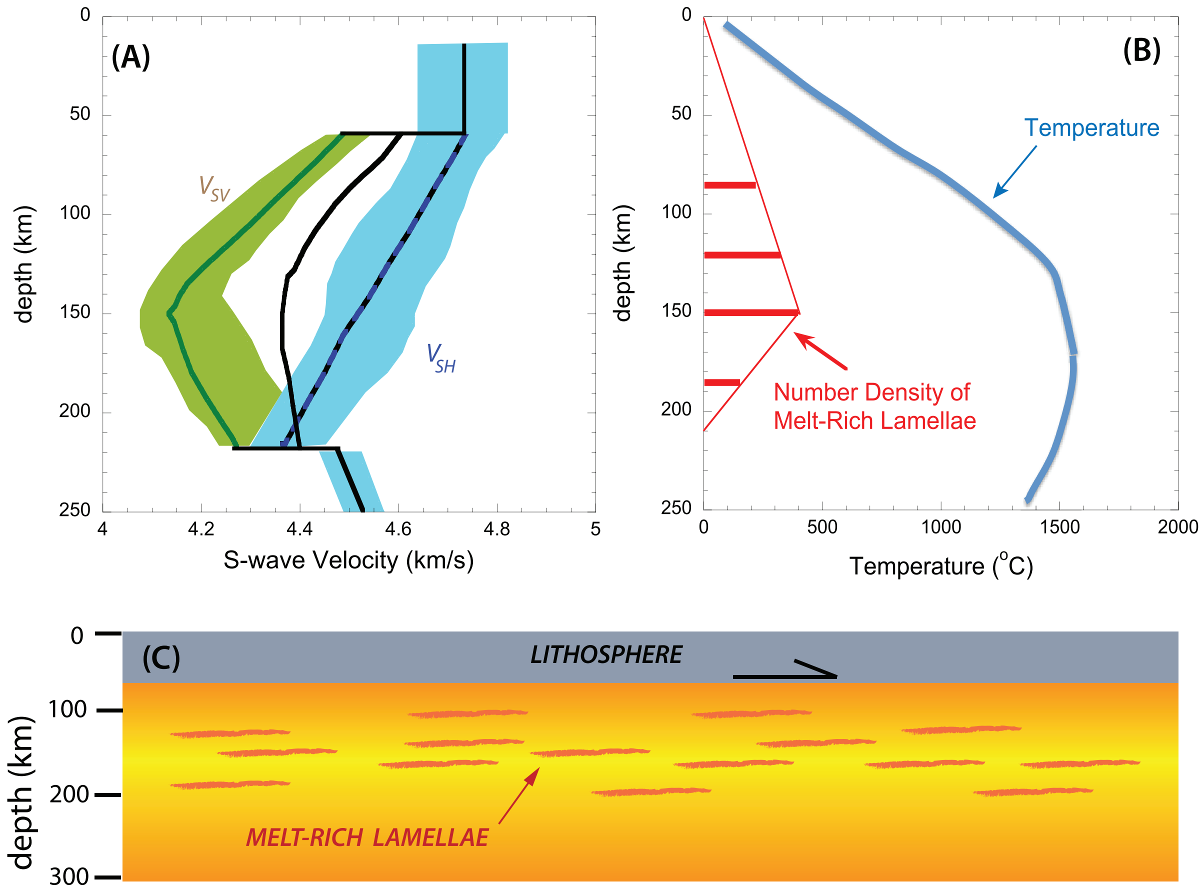
(A)
V_{SV}
V_{SH}
depth (km)
0
50
100
150
200
250
4
4.2
4.4
4.6
4.8
5
S-wave Velocity (km/s)
(B)
Temperature
Number Density of
Melt-Rich Lamellae
depth (km)
0
500
1000
1500
2000
Temperature (°C)
(C)
LITHOSPHERE
MELT-RICH LAMELLAE
depth (km)
0
100
200
300

FIG 1

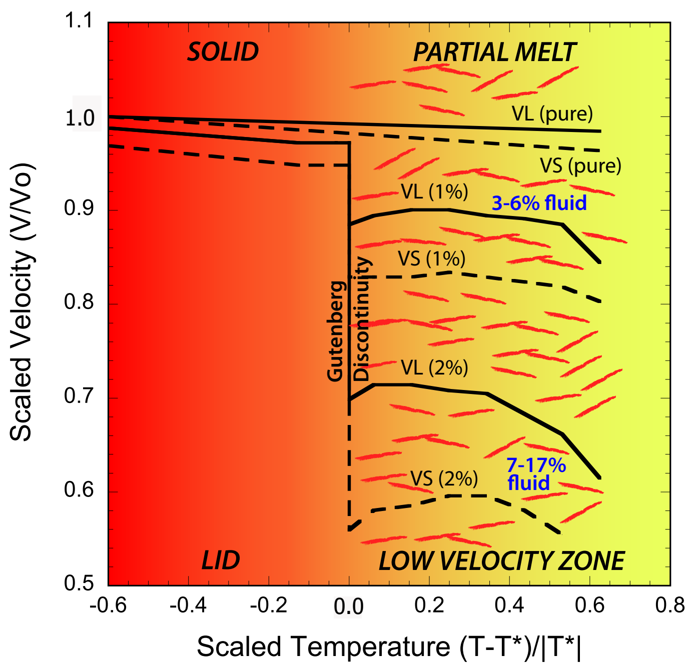
SOLID
PARTIAL MELT
VL (pure)
VS (pure)
VL (1%)
3-6% fluid
VS (1%)
Gutenberg
Discontinuity
VL (2%)
VS (2%)
7-17%
fluid
LID
LOW VELOCITY ZONE
Scaled Velocity (V/Vo)
1.1
1.0
0.9
0.8
0.7
0.6
0.5
-0.6
-0.4
-0.2
0.0
0.2
0.4
0.6
0.8
Scaled Temperature (T-T*)/|T*|

FIG 2