Melting of subducted basalt at the core-mantle boundary

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The geological materials in Earth's lowermost mantle control the characteristics and interpretation of seismic ultra–low velocity zones at the base of the core-mantle boundary. Partial melting of the bulk lower mantle is often advocated as the cause, but this does not explain the nonubiquitous character of these regional seismic features. We explored the melting properties of mid-oceanic ridge basalt (MORB), which can reach the lowermost mantle after subduction of oceanic crust. At a pressure representative of the core-mantle boundary (135 gigapascals), the onset of melting occurs at ~3800 kelvin, which is ~350 kelvin below the mantle solidus. The SiO2-rich liquid generated either remains trapped in the MORB material or solidifies after reacting with the surrounding MgO-rich mantle, remixing subducted MORB with the lowermost mantle.

Numerous seismological studies have demonstrated the complexity of the lowest 150- to 300-km-thick mantle layer situated just above the core-mantle boundary (CMB). In many areas, there is an intermittent stratification, with 1.5 to 3% velocity discontinuities, as well as lateral shear-wave anisotropy (1). These anomalies could arise from mineralogical heterogeneities (2), magma ocean crystallization (3), the descent of subducted slabs deep into the lower mantle (4), and/or chemical reactions with the outer core. In addition, ultra–low velocity zones (ULVZs), with shear-wave velocity reduction of more than 10%, have been detected in specific mantle regions (5). Their size is limited to ~40 km thickness and ~100 km across, and they are ~10% denser than the surrounding mantle. These ULVZs could be due to partial melting occurring in a steep temperature gradient when approaching a very hot outer core. Between 5 and 30% partial melting could attenuate P and S seismic wave velocities to a similar amplitude as that reported for ULVZs (6). However, it was recently shown that partial melting of the bulk lower mantle cannot produce a residue with an equilibrium partial melt that is sufficiently dense to adequately represent the ULVZ mush (7).

Partial melting of pyrolitic or chondritic mantle would be possible if the CMB temperature were higher than ~4150 K (8, 9). Although such a high temperature is not precluded, it would require a very hot core. It is more than 1000 K above the most recent determination of the mantle adiabat extrapolated to the CMB (10). Moreover, the presence of a very hot core today makes it difficult to explain how a geodynamo could have been maintained for prolonged geological periods: An even hotter core would be required in the past, or an extremely high concentration [400 to 800 parts per million (ppm)] of K in the core (11). Alternatively, partial melting in the lowermost mantle could occur below 4150 K if the mantle composition was drastically altered. Chemical differences between peridotite, pyrolite, and chondritic-type mantle would produce only minor variations, because they contain similar mineralogical assemblages. However, a local concentration of fusible (such as alkalis) or volatile (H2O or CO2) elements, together with a change in mineralogy, could have a marked effect on the solidus temperature. In particular, a small water concentration can be enough to reach the saturation limit of the bulk mantle (12). Consequently, the melting curve of hydrous pyrolite has much lower melting temperatures (13).

We investigated the melting behavior of a natural mid-oceanic ridge basalt (MORB) collected at a 2800-m depth during the Searise-1 research cruise (table S1). When a MORB eventually reaches Earth's lower mantle, it is composed of four coexisting phases: (i) Fe-rich silicate perovskite (Mg-Pv); (ii) free silica, in contrast with free (Mg, Fe)O ferropervskite (Fp) in the bulk mantle (fig. S1); (iii) Ca-bearing silicate perovskite (Ca-Pv); and (iv) enough Al, Ca, Na, and K to generate cage structures such as calcium ferrite (CF), hollandite, or an aluminous phase (14–16). Our starting material contained ~0.3 weight % H2O and traces of CO2 (fig. S9), which are common volatile concentrations for natural MORBs. At lower mantle pressures, the basaltic portion of the slab contains minerals that can carry water, such as the Al-bearing stishovite (17), δ-AlOOH (18), and phase H (19).

High pressures and temperatures were provided by a laser-heated diamond anvil cell (fig. S2 (8)). The sample behavior was followed continuously using in situ x-ray diffraction (Fig. 1 and supplementary materials). Upon heating, we first observed the appearance of a continuous diffraction ring on the charge-coupled device detector, at φinit distances typical of Mg-Pv. This

Fig. 1. Sequence of x-ray diffractions recorded with increasing temperatures from 1850 to 4050 K at 56 GPa. Diffraction peaks of the KCl pressure medium are superimposed in blue. Upon heating, we first observed the appearance of weak Mg-Pv diffraction peaks. Followed by crystallization of all other phases, including Ca-Pv and stishovite. Above 2930 K, diffraction peaks become much sharper. Also, all diffraction peaks for a given sample drastically change position on images recorded successively at the same pressure and temperature conditions. Above 4050 K, only a few diffraction peaks remain visible. The quench image contains several new peaks, showing that the lack of diffraction peaks at the highest temperature was not due to a loss of sample, but rather to complete sample melting.
ring occurred above 1800 and 2500 K, at 40 and 135 GPa, respectively. The relatively large peak profile is typical of a powder with very small grain size (Fig. 1A). At higher temperatures, Ca-Pv and silica appeared simultaneously (Fig. S3 and Table S2). The latter adopts either the structure of rutile (stishovite), CaCl₂ or α-PO₄ (seifertite), as a function of pressure (Fig. S4). Diffraction peaks of the post-perovskite phase of MgSiO₃ also appeared above 115 GPa. Diffraction features typical of CF and/or the so-called new aluminous phase remained weak because of major overlaps of their diffraction peaks with those from the other phases present in the sample (Fig. S5).

Less than 100 K above the temperature at which Ca-Pv and silica crystallized from the glass, the initially continuous diffraction rings evolved suddenly into a discontinuous juxtaposition of spots (Fig. 1B), and the peak shape sharpened drastically, showing a net discontinuity in the rate of grain growth. At slightly higher temperatures, diffraction images acquired repeatedly at the same temperature, with a time interval of 20 s (acquisition time), showed large diffraction spots in totally different positions, indicating fast grain rotation. These changes in the sample properties were associated with a flattened temperature profile, although laser power was increased continuously (Fig. S5). This effect could result from a change in laser absorption due to the structural transformation (20). Our results show that the temperature gap between rapid grain growth and fast grain rotation is less than 200 K (Fig. S8). In theory, the extensive grain rotation should take place close to the solidus temperature. The sudden change in sample behavior is most probably due to the appearance of a small amount of liquid at grain boundaries. Because the very first degree of melting can be difficult to detect, we bracketed the solidus using these two major criteria—rapid grain growth and fast grain rotation—where the solidus temperature profile should plot (Fig. S3).

The solidus increases continuously from about 2100 (±150) to 3200 (±150) K with an increase in pressure from 20 to 80 GPa (Fig. 2). At the latter pressure, which corresponds to a mantle depth of 1900 km, the MORB solidus is identical to that of a chondritic-type mantle (8). With a further increase in pressure to 135 GPa, the MORB solidus temperature increased to 3800 (±150) K, whereas that of the bulk mantle increased to 4510 (±150) K. The flattening of the MORB solidus could be linked to a change in liquid composition with pressure. Extrapolation of our solidus melting curve to ambient pressure yields a melting temperature of ~1300 K, in agreement with previous studies (21). However, we could not reproduce the rapid increase of solidus temperature at the low pressures observed in previous studies using a large-volume press (21, 22). Because analyses were performed on quenched samples, for which it is difficult to observe low degrees of partial melting, we believe that the solidus temperature was possibly overestimated in these previous studies. Our solidus curve also plots at much lower temperatures than a previous study performed up to 65 GPa, using a laser-heated diamond anvil cell (22). In that study, the melting criterion based on the change of temperature with laser power, for a MORB sample sandwiched between two Re foils, may have hampered detection of the solidus. The maximum discrepancy between the different studies is ~400 K and tends to decrease with increasing pressure above 20 GPa. At 60 GPa, the discrepancy is down to ~250 K between the previous optical (22) and our in situ measurements. To verify that the 350 K melting temperature reduction between MORB and the chondritic-type mantle was not an experimental artifact, we loaded both compositions in the same pressure chamber (Fig. S2). Using the same melting criteria as in a previous study of mantle melting (8), we observed the solidus of chondritic mantle at temperatures of 2950 (±150), 3450 (±150), and 3680 (±150) K, for pressures of 58, 85, and 107 GPa, respectively, falling within ±60 K of our previous study (8). Given the higher amount of fusible elements in MORB, one could have expected a lower solidus temperature as compared to chondritic mantle. However, the role of SiO₂ excess in MORB and Fp excess in the mantle cannot be neglected (Fig. S1).

Upon further increase of the laser power, the sample temperature eventually rose above the plateau temperature, allowing us to record diffraction patterns up to ~5000 K. We rarely observed the total disappearance of all diffraction peaks from the sample. Instead, diffraction images often contained a couple of diffraction peaks appearing at random azimuthal positions on the images (Fig. 1D). Convection in the liquid sample could have induced some crystallization at the sample/KCI interface over a very short time scale. Also, the sample temperature was probably slightly below the liquidus temperature, because we observed a clear rim of solid phase just around the melt. We define the liquidus as the temperature at which complete loss of the continuous structure of the diffraction lines was achieved. This temperature corresponds to a clear loss of the three-dimensional solid structure of the sample, with free rotation of a couple of grains in a predominantly liquid fraction.

The temperature gap between the solidus and liquidus is found to increase progressively from 1500 to 2500 K with increasing pressure from 40 to 140 GPa. This gap is much larger than that reported for the chondritic-type mantle. It results in a liquid composition at the solidus, which is very different from the bulk MORB composition (22).

Three lines of evidence point to an increase of the SiO₂ content in the liquid with increasing pressure during partial melting of MORB. First, the MORB solidus curve always plots 1000 to 1500 K below the melting curves of pure SiO₂ and MgSiO₃, as determined from shockwave experiments (21, 22, 24–26).
Ca-Pv, SiO2 ternary diagram (fig. S1). A second argument is based on the last phase(s) to disappear from the diffraction patterns at temperatures approaching those of the liquids. At pressures lower than ~70 GPa, clear peaks of SiO2 and CaSiO3 remained visible on the diffraction patterns at temperatures between the solidus and liquidus. At higher pressures, SiO2 peaks disappeared and instead those of Mg-Pv appeared (fig. S6). This indicates a sample undersaturated in SiO2, with Mg-Pv and Ca-Pv grains coexisting with the liquid at the highest pressures.

A third argument comes from mineralogical and chemical analyses of the recovered samples. After the laser was shut down, we maintained the pressure and made diffraction maps of the quenched molten region of the samples. This region was depleted in the highest pressures.

At pressures lower than ~70 GPa, clear peaks of SiO2 disappeared and instead those of MgO and Mg-Pv were observed. The loss of SiO2 was due to specific regions where MORB slabs have reached the CMB. The thicknesses of the ULVZ (~40 km) and MORB portion of the subducted slabs (~6 km (28)) are of the same order of magnitude. Mantle resistance to the slab penetration, or folding and piling-up at the CMB, could certainly induce thickening of the MORB layer. At pressures above ~70 GPa, we observed large amounts of MgO in the center of the samples. This is not likely that large-scale pockets of liquid will form. It would also explain why seismic shear waves (P) can propagate through the partially molten ULVZ (5).

Also, the decrease of Fp content in the lowermost mantle, resulting from reaction between the bulk mantle and SiO2-rich liquid, would drive mantle composition toward the perovskitic end-members, in agreement with a recent report (29). The loss of Si from the MORB would drive it toward perovskitic end-members (fig. S1). The reaction would eventually stop when the SiO2 excess in the MORB was exhausted. This would result in the disappearance of the MORB signature, except for minor and trace elements, which could remain concentrated around the ULVZ. This scenario would lead to lower mantle homogenization, in contrast to the chemical segregation generally induced by mantle partial melting. Alternatively, if the MORB proportion is high, Fp could become reactively exhausted, which would imply a lowermost mantle saturated in SiO2. The answer to these two alternative scenarios depends on how well the underlying harzburgite remains attached to the MORB layer and is entrained into the lowermost mantle. If it is sufficiently abundant, then its reaction with the liquid originating from MORB melting should produce a typical mantle composition such as pyrolite.

REFERENCES AND NOTES

Fig. 3 Phase relations in the sample recovered after partial melting at 120 GPa. The chemical maps (Mg and Si) and mineralogical maps [Mg-Pv and seifertite (Seif)] measured by SEM and x-ray diffraction, respectively, show higher SiO2 and seifertite contents and lower MgO and Mg-Pv contents in the sample region corresponding to the quenched liquid.
CORALS AND CLIMATE

Mechanisms of reef coral resistance to future climate change

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Reef corals are highly sensitive to heat, yet populations resistant to climate change have recently been identified. To determine the mechanisms of temperature tolerance, we reciprocally transplanted corals between reef sites experiencing distinct temperature regimes and tested subsequent physiological and gene expression profiles. Local acclimatization and fixed effects, such as adaptation, contributed about equally to heat tolerance and are reflected in patterns of gene expression. In less than 2 years, acclimatization achieves the same heat tolerance that we would expect from strong natural selection over many generations for these long-lived organisms. Our results show both short-term acclimatory and longer-term adaptive acquisition of climate resistance. Adding these adaptive abilities to ecosystem models is likely to slow predictions of demise for coral reef ecosystems.

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eef-building corals have experienced global declines resulting from bleaching events sparked by pulses of warm-water exposure (1–4). However, corals in naturally warm environments can have high resistance to bleaching temperatures and can survive heat exposure that would bleach conspecifics in cooler microclimates (5, 6). Similarly, recent discovery of populations of acidification-resistant corals show that physiological or evolutionary mechanisms of environmental accommodation exist (7, 8). Such populations are ideal test sites for research into the mechanisms of coral response to climate change.

Corals in adjacent backreef pools in the U.S. National Park of American Samoa on Ofu Island experience strong differences in temperature (9, 10). In the highly variable (HV) pool, temperatures often exceed the local critical bleaching temperature of 30°C, reaching 33°C during strong nighttime low tides (6). By contrast, the moderately variable (MV) pool rarely experiences temperatures above 32°C. Corals in the HV Pool have higher growth rates (9, 10), higher survivorship, and higher symbiotic photosynthetic efficiency during experimental heat stress than conspecifics from the MV pool (6). These pools provide a powerful system to test the speed and extent of coral acclimatization and adaptation to warm-water conditions in the context of future climate change.

To test corals in their native habitats for physiological resistance to heat stress, we collected branches of the tabletop coral Acropora hyacinthus [cryptic species E (7)] and exposed them to experimental bleaching conditions. A. hyacinthus is a cosmopolitan species that constitutes a large percentage of hard coral cover on Pacific reefs and shows high levels of bleaching and mortality during large-scale bleaching events (4). We chose A. hyacinthus for this study because it is a dominant reef-builder and is especially sensitive to environmental stress, making its relative ability to acclimate or adapt extremely important to the future of coral reef ecosystems as climate change proceeds. We subjected branches of corals to a prescribed ramp in water temperature of 29°C to 34°C for 3 hours, followed by an incubation for 3 hours at 34°C. These conditions mimic the natural increase in temperature observed in the HV pool during a tidal cycle. Experiments on fragments of tagged and monitored colonies showed that individuals native to the HV pool exhibit higher resistance to thermal stress, measured by retention of chlorophyll derived from photosynthetic symbionts, than corals from the MV pool (Fig. 1). The average retention of chlorophyll after experimental heat stress was 80% in HV pool corals (Fig. 1C) but only 45% in MV pool corals (Fig. 1A, t test, P < 0.00001) compared with controls.

To test for acclimatization, we transplanted coral colonies of A. hyacinthus reciprocally from their native locations in the HV and MV pools to three transplant sites within each pool. We transplanted 6 colonies from the HV pool and 12 from the MV pool. After 12, 19, and 27 months, we tested transplanted colonies for thermal resistance. For 11 separate colonies, 22 of 23 paired bleaching