A DOUBLE PARTIAL MELT ZONE IN THE MANTLE BENEATH MID-OCEAN RIDGES *

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For a lherzolite mantle with about 0.1 wt.-percent CO_2 or less, and a CO_2/H_2O mole ratio greater than about one, the mantle solidus curve in P–T space will have two important low-temperature regions, one centered at about 9 kbar (30 km depth) and another beginning at about 28 kbar (90 km depth). It is argued that the depth of generation of primary tholeiitic magmas beneath ridge crests is about 9 kbar, and that the geotherm changes from an adiabatic gradient at greater pressures to a strongly superadiabatic gradient at lesser pressures. Such a ridge geotherm would intersect the solidus at two separate depth intervals corresponding to the two low-temperature regions on the solidus. With increasing age and cooling of the lithosphere, the shallow partial melt zone would pinch out and the thickness of the deep partial melt zone would decrease. With increasing depth in a mature oceanic lithosphere, the rock types would consist of depleted harzburgite from directly beneath the crust to about 30 km depth, fertile spinel lherzolite from about 30 km to 50–60 km, and fertile garnet lherzolite from about 50–60 km to the top of the deep partial melt zone at about 90 km.

1. Introduction

Existing models for the evolution of the oceanic lithosphere show an increasing thickness with age and distance from active mid-ocean ridges. These models indicate either a gradual thickening with age, from nearly zero at ridge crests to roughly 100 km in the older parts of the oceans (Parker and Oldenburg, 1973; Leeds et al., 1974, Yoshii, 1975; Green and Liebermann, 1976), or a rapid increase in thickness to about 60 km for 10 My old lithosphere and a very gradual increase with age thereafter (Forsyth, 1975, 1977). In most of these models, the base of the lithosphere has been taken to represent the top of the seismic low velocity zone, which in turn has been assumed to be caused by the onset of partial fusion. However, the existence of a melt in the low velocity zone has recently become controversial (Gueguen and

Mercier, 1973; O'Connell and Budiansky, 1977; Shaw, 1978; Walker et al., 1978) and Anderson and Minster (1980) have pointed out that the thickness of the lithosphere depends on whether seismic, thermal, or rheological properties are being considered. It will be seen in the following discussion that if the model of Presnall et al. (1979) for the generation of mid-ocean ridge basalts is correct, it is very difficult to draw a geotherm at and for a considerable distance away from oceanic spreading ridges that does not pass through the mantle solidus in the approximate depth range of the low velocity zone. Thus, a melt is believed to exist even though this melt may be unnecessary or inadequate to account for all the properties of the low velocity zone.

The purpose of this paper is to present arguments favoring the existence of a double partial melt zone beneath actively spreading oceanic ridges. The extent to which such a melt configuration is related to measurements of lithospheric thickness based on various geophysical parameters will not be examined here,

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but if a liquid is even partly responsible for the properties of the seismic low velocity zone, verification of the model proposed here would require the modification of existing concepts of a thinning lithosphere toward ridges.

Because the mantle solidus curve is critically dependent on volatiles that may be present, it is necessary first to discuss these constituents. Then a model will be developed based on intersections of a proposed geotherm directly beneath ridges with the mantle solidus. It will be assumed that fertile mantle consists essentially of lherzolite, that is, a mixture of olivine, orthopyroxene, clinopyroxene, and one of the aluminous phases plagioclase, spinel, or garnet, depending on pressure. The relative proportions of the four phases is not critical, and if subsequent studies verify the conclusion of Anderson (1977) that a large amount of garnet must exist in the lower lithosphere, none of the general conclusions reached here would be significantly altered as long as all four phases are present.

2. Volatiles in the oceanic mantle

Delaney et al. (1978) and Muenow et al. (1979) have determined the volatile constituents of glass—vapor inclusions in olivine and plagioclase phenocrysts from mid-ocean ridge and Hawaiian tholeiites. They found 0.2–0.4 wt.-percent CO_2 and no H_2O, with a stated detectability limit for H_2O of 0.004 wt.-percent. In contrast, they found amounts of H_2O in the matrix glasses surrounding the phenocrysts of 0.1–0.7 wt.-percent, in general agreement with previous results by other methods (Hart and Nalwalk, 1970; Moore, 1970). Barring some unrecognized analytical problem, they concluded that almost all of the H_2O in the matrix glasses entered the magmas after formation of the phenocrysts. Also, they inferred that the mantle source region was nearly anhydrous, because any H_2O present in the source would be fractionated into the liquid during fusion and would be observed in the glass—vapor inclusions. As a further indication of the nearly anhydrous character of the oceanic mantle, Killingley and Muenow (1974) found that inclusions in olivine from a peridotite nodule collected at Salt Lake Crater, Hawaii, contained CO_2 but no observable H_2O. Assuming that mid-ocean ridge basalts represent approximately

10–30% fusion (Gast, 1968; Kay et al., 1970; Schilling, 1971, 1975), the above results suggest that fertile oceanic manlte contains on the order of 0.02–0.1 wt.-percent CO_2 and a negligible amount of H_2O (less than about 12 parts per million).

In contrast to the results of Delaney et al. (1978) and Muenow et al. (1979), Harris (1979) analyzed glass inclusions in olivine phenocrysts from Hawaiian tholeiites and found concentrations of H_2O and CO_2 of 0.21 and 0.05 wt.-percent, respectively. Part of the discrepancy may be due to the fact that the method of Delaney et al. and Muenow et al. determines the total volatile content of the glass and vapor phases in the inclusions, whereas the method of Harris apparently determines only the volatiles in the glass. If it assumed that (1) the analyses of Harris are correct for the glass, (2) the volume of the gas phase is about 4% of that of the glass (Delaney et al., 1978), (3) the gas bubbles consist essentially of pure CO_2, and (4) the entrapment conditions of the gas bubbles are 1 kbar—1240°C (Harris, 1979), then the CO_2/H_2 mole ratio of the glass and vapor combined would be roughly one (about 0.21% H_2O and 0.56% CO_2 by weight). Thus, the difference in CO_2 between the two laboratories can be explained but the discrepancy in H_2O remains.

The question of water in the oceanic mantle can also be addressed by direct examination of oceanic mantle rocks. Melson et al. (1967) found pargasite in peridotite mylonites at St. Paul's Rocks, but rare-earth abundance data are inconsistent with the hypothesis that these rocks represent either primitive mantle material from which basalts are derived or residua from partial fusion (Frey, 1970). Because the simplest explanation for the origin of these rocks is complex, they are unsatisfactory as an indicator of typical abundances of primary hydrous phases in the mantle. Elsewhere in the ocean basins, traces of hydrous phases that may be primary have been reported in ultramafic nodules. Jackson and Wright (1970) and Beeson and Jackson (1970) found up to 0.5% amphibole plus phlogopite in garnet pyroxenites from Oahu, Hawaii. These pyroxenites have been considered variously as high pressure magmas, crystal cumulates, parental mantle, and residua from partial fusion (Yoder and Tilley, 1962; Tilley and Yoder, 1964; Green, 1966; O'Hara, 1969; Kuno, 1969; Beeson and Jackson, 1970; Jackson and Wright,

1970). Thus, the origin of these nodules is uncertain and they may contain concentrations of volatiles that are anomalously high. Talbot et al. (1963) reported "rare" phlogopite in peridotite xenoliths from the Kerguelen archipelago, and Jackson and Wright (1970) found 0.1% phlogopite in a spinel lherzolite from Oahu, Hawaii. A mantle containing 0.1% phlogopite would contain at most about 40 parts per million water, an amount that is the same order of magnitude as the maximum indicated by the data of Muenow et al. (1979) and Delaney et al. (1978). Thus, the existence of hydrous phases in mantle nodules from oceanic regions does not contradict the data of Delaney et al. (1978) and Muenow et al. (1979), but the nodule data are also consistent with the data of Harris (1979). The only conclusion that appears clear is that hydrous phases are present in the mantle.

At pressures below about 28 kbar, dolomite would not be stable during melting of the mantle. Thus, CO_2 as well as H_2O would be totally fractionated into the melt and vapor phases during fusion. If it is assumed that the CO_2/H_2O mole ratio (>20) indicated by the data of Delaney et al. (1978) and Muenow et al. (1979) is approximately valid for the source region and the mantle generally, the mantle solidus would be very close to the pure CO_2-saturated mantle solidus (Wyllie, 1977), particularly in view of the fact that CO_2/H_2O in the vapor would be >>20 because of strong fractionation of CO_2 into the vapor relative to H_2O (Eggler and Rosenhauer, 1978). Therefore, melting in the low velocity zone would not be the result of the breakdown of amphibole (Lambert and Wyllie, 1968, 1970), but rather the result of a deep low-temperature cusp in the mantle solidus caused by a carbonation reaction (Wyllie and Huang, 1975a, 1976; Eggler, 1975, 1976). In the following discussion, the consequences of assuming that this latter mechanism is the cause of melting will be pursued. For simplicity, only the solidus for a mantle containing pure CO_2 will be considered, but it is not meant to imply that H_2O is absent from the mantle. If the CO_2/H_2O mole ratio is about one, as suggested by the data of Harris (1979), the mantle solidus would be modified somewhat but the general conclusions regarding the existence of a double partial melt zone would still hold (Wyllie, 1979).

3. Intersection of the mantle solidus with the geotherm beneath mid-ocean ridges

In the system $CaO-MgO-Al_2O_3-SiO_2$, Presnall et al. (1979) found a low temperature cusp on the simplified lherzolite solidus at 9 kbar—1300°C, where the transition between the subsolidus assemblage forsterite + enstatite + diopside + anorthite (simplified plagioclase lherzolite) and forsterite + enstatite + diopside + spinel (simplified spinel lherzolite) intersects the solidus (Fig. 1). They found that the composition of the first liquid produced at this cusp corresponds closely to the compositions of the least-fractionated mid-ocean ridge tholeiites, and concluded from this and other evidence that the production of primary tholeiites at the mantle equivalent of this cusp acts as a thermal buffer for controlling the shape of the geotherm, as shown in Fig. 1. At pressures less than that at the cusp, the slope of the geotherm shown is consistent with gradients at ridge crests calculated from heat flow models (Oxburgh and Turcotte, 1968; LePichon and Langseth, 1970; Bottinga and Allègre, 1973), and at pressures greater than the cusp, the geotherm flattens to an adiabatic gradient (0.5°C km^{-1}), consistent both with the gradient determined by Hermance and Grillot (1974) beneath Iceland, and with the concept of transport of mantle material along the ascending limb of a convection cell.

In the system $CaO-MgO-Al_2O_3-SiO_2$, the effect of the addition of CO_2 on the simplified lherzolite solidus has not been determined. However, Eggler (1975, 1976) and Wyllie and Huang (1975a, b, 1976) have shown that in the closely related system $CaO-MgO-SiO_2-CO_2$, the solidus at pressures below about 20 kbar is depressed only a small amount below the anhydrous solidus because of the low solubility of CO_2 in the melt. However, at 28 kbar a deep cusp occurs in the solidus due to the sharply increased solubility of CO_2 in the melt and the intersection of the solidus with the carbonation reaction forsterite + diopside + CO_2 = enstatite + dolomite (Fig. 1). With the addition of Al_2O_3, the same general relationships would be expected to persist (Eggler, 1978; Wendlandt and Mysen, 1978), but as the solidus curve crosses the plagioclase lherzolite to spinel lherzolite curve and the spinel lherzolite to garnet lherzolite curve, it is evident from Schreine-
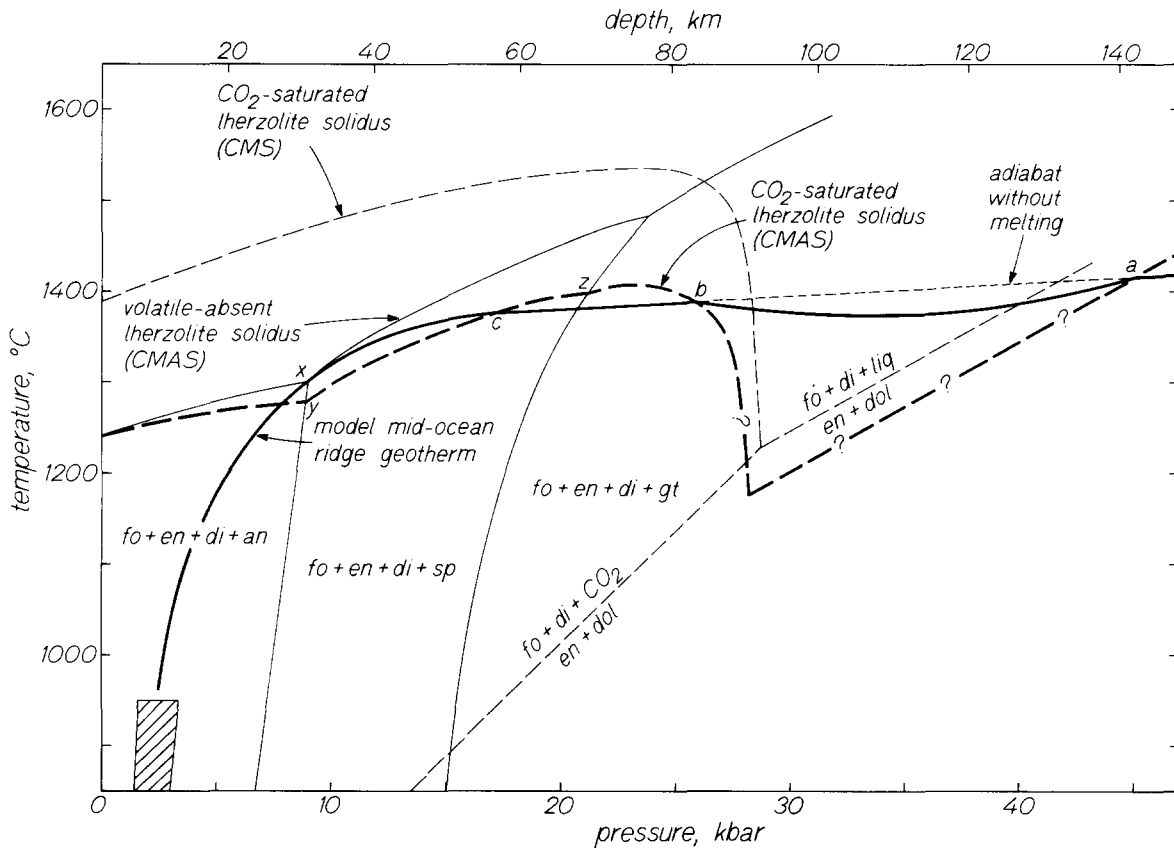
Fig. 1. Pressure—temperature diagram showing stability fields of model lherzolite phase assemblages and the relationships of solidus curves to a mid-ocean ridge geotherm. Light-dashed lines are for the system $CaO-MgO-SiO_2-CO_2$ (CMS) and are after Eggler (1978). Light solid lines are for the CO_2-free system $CaO-MgO-Al_2O_3-SiO_2$ (CMAS) and are after Kushiro and Yoder (1965, 1966), MacGregor (1965), O'Hara et al. (1971), Presnall (1976), Presnall et al. (1979) and Jenkins and Newton (1979). The spinel lherzolite to garnet lherzolite curve is an average of the various published curves and the position of the CMAS volatile-absent solidus above 1400°C is inferred. The heavy dashed line is an estimated lherzolite solidus for the system $CaO-MgO-Al_2O_3-SiO_2-CO_2$. By analogy with the system $CaO-MgO-SiO_2-CO_2$, the portion of this solidus above 28 kbar is assumed to represent vapor-absent melting in the presence of dolomite. The reaction, however, has not yet been studied. The hachured region shows the range of surface geothermal gradients published for mid-ocean ridges (Oxburgh and Turcotte, 1968, LePichon and Langseth, 1970, Bottinga and Allègre, 1973). an = anorthite, di = diopside, gt = garnet, sp = spinel, dol = dolomite, en = enstatite, fo = forsterite.

makers' rules that two additional cusps must exist (points y and z in Fig. 1).

For the model mantle system $CaO-MgO-Al_2O_3-SiO_2-CO_2$, consider the history of a parcel of lherzolite mantle material containing 0.1% CO_2 which rises adiabatically from the deep mantle in the ascending limb of a convection cell directly beneath an active mid-ocean ridge. At pressures greater than 45 kbar, the pressure—temperature path follows an adiabat appropriate for the situation in which no melting

occurs, about $0.5°C\ km^{-1}$, and CO_2 will be present as dolomite rather than vapor. When the mantle solidus is encountered at a, melting begins, dolomite quickly disappears into the melt, and the temperature drops a small amount below the adiabat that would be followed if no melting occurred, the heat for melting being supplied by the temperature drop. The amount of melting in the depth interval between a and b is not known, but for illustrative purposes, a maximum of 10% melting is assumed, which would correspond

very roughly to a temperature drop of about 30°C. As the mantle parcel approaches the solidus again at b, the melt crystallizes and the temperature rises back to the original adiabat. Between b and c, the mantle parcel is completely crystalline, with CO_2 being present as vapor. At c, melting begins again, the vapor quickly disappears, and the crystal—liquid mixture continues to melt until separation of the magma occurs at the approximate depth of the anhydrous solidus cusp at x. Arguments favoring the separation of magma from the mantle source at the approximate depth of the solidus cusp x are given in detail elsewhere (Presnall et al., 1979).

From trace element considerations, it has been shown that the amount of melting required to produce mid-ocean ridge basalts is about 10–30% (Gast, 1968; Kay, et al., 1970; Schilling, 1971, 1975). If the average amount of melting at the time of magma separation is about 20%, the temperature of the upwardly convecting mantle material would have to fall by roughly 60–70°C in order to provide the heat for melting. For this reason, the mid-ocean ridge geotherm at pressures above 9 kbar has been raised, in order to locate the adiabat appropriate for the situation in which no melting occurs about 70°C above the cusp x at which magma separates. Presnall et al. (1979) brought the adiabatic portion of the geotherm directly into the 9 kbar cusp at x, but such a geotherm would allow only a trace of melt to be generated at the cusp. Thus, the geotherm of Presnall et al. is a minimum geotherm. The average geotherm should have an adiabatic portion that is 60°–70° higher, as shown in Fig. 1. The fixed point at x on the geotherm of Presnall et al. remains, however, because the amount of fusion would have to exceed about 35% if the temperature were to rise above the volatile-free solidus at 9 kbar (Presnall et al., 1979).

Thus, a geotherm directly beneath a ridge crest would intersect the simplified mantle solidus at two pressure intervals, one from about 7–17 kbar (~25–55 km) and the other beginning at about 26 kbar (~85 km). The bottom of the deeper partially melted zone is shown at 45 kbar (~145 km) in Fig. 1, but phase relationships have not been determined at these pressures and the depth shown is only an estimate. The width of the unmelted interval from 55 to 85 km is very sensitive to the position of the geotherm relative to the solidus, and it is even possible that the geotherm could be high enough to cause a continuous, single partial melt zone. Even in this case, however, the amount of melting would be enhanced at two distinct depth intervals.

4. Changes in the constitution of the oceanic lithosphere with age

It will be assumed here that the entire mature thickness of the oceanic lithosphere, about 85 km (Asada and Shimamura, 1976), is created at mid-ocean ridges. This implies a width of mantle material rising rapidly along ridge axes of at least twice this thickness, or about 170 km. If mid-ocean ridge basalts are separated from their source region at about 30 km depth, as argued by Green and Ringwood (1967a), Kay et al. (1970), Fujii and Kushiro (1978), and Presnall et al. (1979), then mid-ocean ridge basalts would be derived only from the central portion of the rising mantle material. The margins would remain undepleted after they turned the corner to move horizontally away from the ridge and would make up the lower part of the oceanic lithosphere. Thus, that portion of the mantle shallower than 30 km depth would consist of depleted harzburgite (Kay et al., 1970; Oxburgh and Parmentier, 1977; Presnall et al., 1979). Beneath this depleted harzburgite would be a layer of fertile spinel lherzolite which in turn would be underlain at the base of the lithosphere by fertile garnet lherzolite.

The depth to the transition from depleted to fertile mantle at about 30 km would be fixed at the ridge and would therefore remain constant as the lithosphere ages. It will be noted in Fig. 1 that, with increasing age, the depth to the top of the plagioclase lherzolite to spinel lherzolite transition would decrease slightly due to falling temperatures and would coincide with the change from depleted to fertile mantle only at the ridge crest. For a mature lithosphere, the plagioclase lherzolite to spinel lherzolite transition would not be expected to correspond to a seismically observable discontinuity because the dominant rock type at this depth (<~30 km) would be depleted harzburgite devoid of significant amounts of either plagioclase or spinel. Only the bulk chemical change at ~30 km, from depleted harzburgite to fertile spinel lherzolite, would be potentially observable.

However, the decrease in V_p from depleted harzburg-ite to fertile spinel lherzolite would be only about 0.1 km s^{-1}, assuming an increase of about 6 mole-percent in fayalite content of the olivine (Kumazawa and Anderson, 1969; Chung, 1970) and the increase in density would be about 0.08 g cm^{-3} (Carter, 1970). Thus, detection of this transition by geophysical methods would be difficult.

At greater depths, falling temperature associated with increasing age of the lithosphere would cause the thickness of the garnet lherzolite layer to increase at the expense of the overlying spinel lherzolite layer. In the simplified model system (Fig. 1), the depth to the top of the garnet lherzolite layer would occur at 63 km directly beneath ridge crests and would decrease to a depth of 50–60 km in the older parts of the oceans. In agreement with the phase relationships, Hales et al. (1970) and Asada and Shimamura (1976) have observed a high velocity layer ($V_p \simeq 8.6$ km s^{-1}) at the base of the oceanic lithosphere, which they attributed to the presence of garnet lherzolite. The depth to the top of this layer was placed at 57 km in the Caribbean by Hales et al. and at 66 km in the western Pacific by Asada and Shimamura. Further-more, Asada and Shimamura noted that the depth to this seismic discontinuity decreases with increasing age, which, they pointed out, would be expected if the discontinuity is caused by the spinel to garnet lherzolite transition in a cooling lithosphere.

Anderson (1977) has noted that to produce com-pressional velocities of 8.6 km s^{-1}, a large amount of garnet (at least 26%) must be present. This require-ment led him to propose an underplating of ecologite on the base of the lithosphere. However, in the present model, a thickened high velocity layer at the base of a mature lithosphere is a natural consequence of falling temperatures, and underplating of ecologite is unnecessary. It might be argued that such an explanation is unlikely because garnet lherzolites usually contain much less than 26% garnet. However, a large amount of garnet would be possible if the lower lithosphere is fertile, as proposed here. Also, samples of this layer are rare. The island of Oahu, Hawaii, contains the only reported samples of garnet lherzolite nodules from the oceanic basins, and only one of these nodules (66SAL-1) has been described in any detail (Jackson and Wright, 1970). It contains 22% garnet, but even for this nodule, there is no

assurance that it represents mantle material un-touched by fusion processes. The garnet lherzolite nodules are found in small alkalic cones erupted on the flanks of the earlier Koolau volcano which forms the backbone of the island (Jackson and Wright, 1970), and these nodules may possibly represent partially depleted mantle material remaining after production of the main tholeiitic shield. Thus, it is possible that unmelted samples from a fertile, garnet-rich lherzolite layer at the base of the oceanic litho-sphere do not exist among nodules that have reached the Earth's surface, and this layer of fertile mantle may contain much more garnet than has previously been supposed.

Figure 2 shows a comparison between the present model and that of Forsyth (1975). For the present model, depths to the various layers at the ridge crest are taken from Fig. 1, but for older parts of the litho-sphere, depths have not been modeled quantitatively with respect to age. Thus, only the sign and relative magnitudes of the slopes for the various boundaries in Fig. 2 are significant. With increasing distance from the ridge axis, the decreasing geothermal gradient would cause the shallow partially melted region to pinch out. Because of the very steep negative slope of the solidus curve controlling the depth to the top of the deep partially melted region (Fig. 1), this depth would increase by only a small amount as the geo-thermal gradient decreases away from the ridge axis. However, the slope of the solidus curve defining the base of the deep partial melt zone is probably much smaller, so the depth to the base of this zone would decrease more rapidly with increasing age of the lithosphere.

The existence of the shallow partial melt zone, the approximate position of this zone, and the greater amount of melting compared to the deep partial melt zone are supported by a limited amount of geo-physical evidence. For the Mid-Atlantic Ridge, Solo-mon (1973) reported a low-Q zone having a width of 50–100 km and a maximum depth of 50–150 km. Solomon and Julian (1974) found that the upper boundary of the low-Q zone increases in depth with age of the lithosphere, as expected in the model pre-sented here, and the maximum width extends to an age of about 5 My at about 35 km depth. Beneath the crest of the East Pacific Rise, Kausel (1972), as reported in Forsyth (1977), found a zone of very
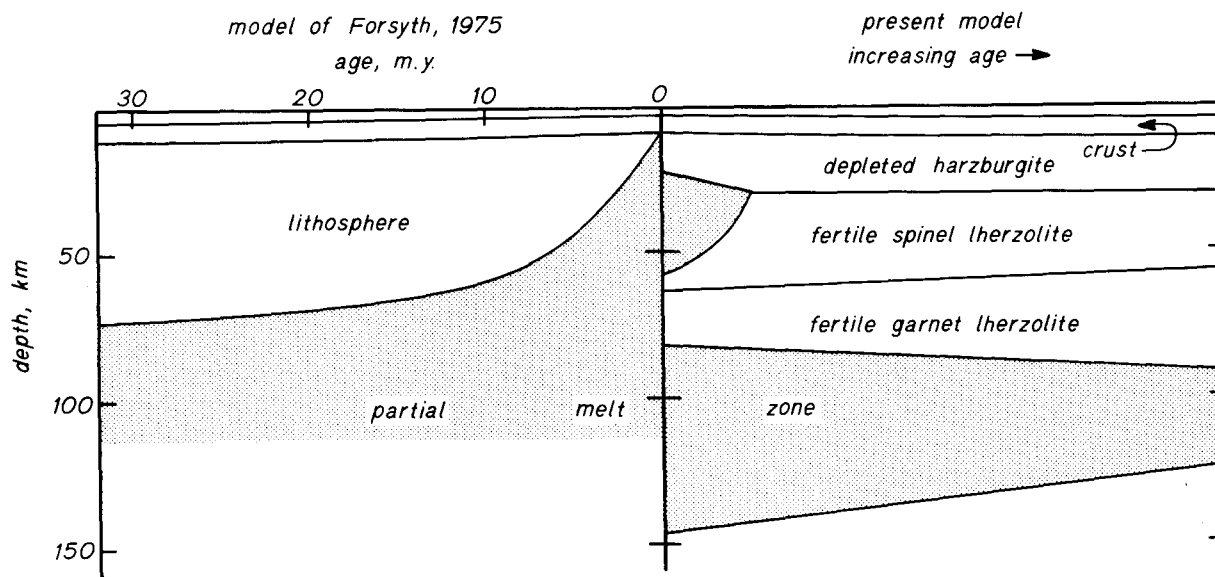
Fig. 2. Comparison of present model for the oceanic lithosphere with the model of Forsyth (1975). The bottom of the partial melt zone was not defined by Forsyth.

high seismic attenuation that was fitted best by models in which the zone extends from 20 to 60—70 km depth. He found that Q in this zone is lower than typical low velocity zone values, an observation consistent with an enhanced amount of melting.

It should be kept in mind that the present model is based on phase relationships for a simplified model system. Presnall et al. (1979) have pointed out that in the Earth's mantle, the presence of additional components would lower temperatures along the solidus curve, and the adiabatic portion of the geotherm would therefore probably be 50—100°C lower than shown in Fig. 1. Also, the phase boundaries in the model system are univariant curves, resulting in the construction of sharp boundaries between the various layers in Fig. 2. The presence of additional components would cause the phase transitions in the mantle as well as the solidus cusp to be smeared out over a depth interval. Small shifts in the positions of phase transitions would also be expected. Nevertheless, available high pressure data on complex lherzolite compositions indicate that the general features of the phase relationships in the model system can be applied to the mantle (Green and Ringwood, 1967b; Wendlandt and Mysen, 1978).

If the CO_2/H_2O mole ratio in the mantle is

approximately one, the phase relationships illustrated by Wyllie (1979, fig. 3) indicate that a high temperature ridge would still persist on the peridotite solidus at about 26—27 kbar. Depending on the exact position of the geotherm relative to the solidus, the mantle directly beneath active spreading ridges would either have two partial melt zones or a continuous melt zone with the amount of melting accentuated at two different depth intervals. Thus, the result is similar to the case of a mantle with a very high CO_2/H_2O ratio.

Anderson and Minster (1980) have proposed that the terms seismic lithosphere, thermal lithosphere and rheological lithosphere be used, depending on which properties are being discussed. Unfortunately, it is not yet clear how any of these lithospheres are related to a petrologically definable boundary. To facilitate discussions between petrologists and geophysicists, it is proposed that the term "petrological lithosphere" be used for material above the deeper of the two partial melt zones.

Acknowledgments

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