:HN VERHOOGEN *Department of Geology and Geophysics, University of California, Berkeley, California 94720*

Possible Temperatures in the Oceanic Upper Mantle and the Formation of Magma

Note: This paper is dedicated to Aaron and Elizabeth Waters on the occasion of Dr. Waters' retirement.

ABSTRACT

Horizontal heat transfer, either by plates that cool as they move away from their source at a ridge axis or by convection currents, invalidates all temperature distributions calculated on the assumption of purely radial outward heat transfer. Temperature gradients in the upper mantle in oceanic regions are estimated on the assumptions that (1) the top of the low-velocity zone corresponds to the "wet" solidus of peridotite containing a very small amount of water, and (2) that the seismic discontinuity near 400-km depth corresponds to the olivine-β spinel inversion. The gradients are very low, suggesting that convection is the dominant mode of heat transfer throughout the upper mantle. Possible temperature profiles are accordingly drawn by joining a conduction solution in the lithosphere to an adiabatic (or near-adiabatic) curve below the lithosphere. The temperature profile at any point depends on the local age of the plate; it is impossible to devise a representative average oceanic geotherm that could be used, when superposed on a petrological phase diagram, to predict the depth of formation of magmas or to account for their compositional variations.

INTRODUCTION

To explain the nature of magmas, the petrologist usually superposes on a phase diagram appropriate to the assumed composition of the upper mantle a theoretical curve (geotherm) representing a calculated distribution of temperature with depth in the earth. It seems to be generally accepted that, as suggested some time ago (Verhoogen, 1954), melting begins when and where the temperature of a convectively rising mass intersects the solidus; the rising mass is assumed to start from a point on a geotherm (*see*, for instance, Green and Ringwood, 1967, Fig. 12; or Wyllie, 1971, Fig. 8-22). A mass starting at point S (Fig. 1) rises and cools adiabatically, following the path SM. Melting begins at M and continues as temperature drops from M to M_1 (the slope of MM_1 is steeper than that of SM by the effect of the latent heat of melting). At some point M_1, the liquid separates from the mush and rises to the surface along the liquid adiabatic curve M_1M_2. The degree of fractional melting that occurs between M and M_1 is commonly estimated to vary between a few percent, as for olivine nephelinites, and perhaps as much as 30 or 40 percent, as for olivine-rich tholeiites and picrites. The composition of the melt is determined by conditions prevailing at M_1 which are themselves determined by conditions at S (that is, by the initial choice of a geotherm).

GEOTHERMS

The first problem is then to determine a suitable geotherm appropriate to the particular region where a certain magma forms. The widely used geotherms of Clark and Ringwood (1964) and Ringwood (1966) now seem to be obsolete, for the following reasons:

1. Geotherms calculated from the law of heat conduction are sensitive to the assumed value of the thermal conductivity, particularly in the temperature range where radiative heat transport, which is proportional to T^3, may be important. Very slight changes in the parameters lead to large changes in temperature. It now seems that radiative heat transfer in the mantle is less effective than Clark and Ringwood took it to be (Pitt and Tozer, 1970; Shankland, 1970).

2. Geotherms for oceanic and shield areas differ by 200° to 300° at depths of 100 to

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200 km. Such lateral variations in temperature are almost certainly sufficient to induce convection, which is not considered in the calculations.

3. The basic datum for these calculations is the surface heat flow, and a steady state is assumed in which the heat escaping at point A on the surface is equalled by the heat generated beneath that point. Plate theory contradicts this assumption. Much of the high heat flow near oceanic ridges is now accounted for by cooling of the lithosphere formed at the ridge axis (McKenzie, 1967; Turcotte and Oxburgh, 1969; Sclater and Francheteau, 1970). Thus, heat escaping through the sea floor at A may have come up at the ridge axis B, several thousand kilometers away, whence it was carried laterally by plate motion. There need no longer be an exact correspondence between heat flow at A and heat generated at a point C vertically beneath A. In fact, the temperature beneath A need not increase with depth at all depths, as it must in the usual conduction models. Temperature inversions may occur; for instance, the horizontally moving upper limb of a convective cell may be hotter than the core of the cell itself. The application of conduction theory is now limited to the lithosphere, and convective solutions must be devised for the mantle beneath the lithosphere. Difficulties arise in estimating (1) the viscosity of the upper mantle and its dependence on temperature and pressure, (2) the initial temperature of the lithosphere where it forms at a ridge axis, and (3) thermal conditions at the base of the lithosphere. Turcotte and Oxburgh (1969) have worked out a possible geotherm applicable to a point of the oceanic floor, 1,000 km from the ridge, on a plate moving at 4.4 cm/yr; the local value of the heat flow would be 2.3×10^{-6} cal/cm^2 sec. In this solution, which justifiably ignores heat generation by radioactivity in the plate itself, it is assumed that at about 50 km depth, where the temperature is close to 1,200°C, the temperature profile passes down into a profile (essentially a 1,400°C adiabatic profile) established within the horizontal flow beneath the lithosphere. A temperature profile drawn at a point farther away from the ridge axis would be somewhat different. At 4,000 km from the ridge, for instance, the surface heat flow would be about half what it is at 1,000 km, and a temperature of 1,200°C would be reached only at a depth of 100 km.

As local variations in temperature profiles are also to be expected in continents (as, for instance, between the western and central regions of North America), it is no longer possible to talk of "oceanic" or "shield" or any other kind of generalized geotherm that ignores horizontal transport of heat and convection. But convective features of the mantle remain essentially unknown. It is not known, for example, whether the whole mantle, or only the upper few hundred kilometers of it, participates in the motion. The rheological properties of the mantle are uncertain, and much doubt remains as to whether it is permissible to assume that it has anything like a Newtonian viscosity independent of stress or strain rate. Most studies of thermal convection have considered only plane layers of liquid heated from below; the geophysical problem of a spherical body with internal heat generation has hardly been touched.

Perhaps the only statement that can be made with some assurance of general validity concerns the average temperature gradient in a convecting layer of the earth: if the layer is thick, the average gradient in it will not much exceed the adiabatic (=isentropic) gradient. For convection to start, the Rayleigh dimensionless number R characterizing the system must exceed a value which depends somewhat on boundary conditions but is generally of the

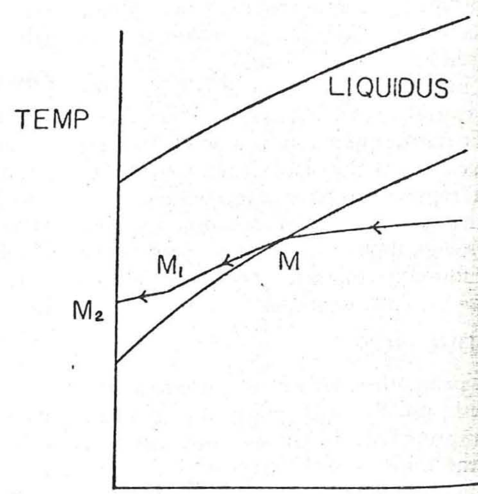

Figure 1. Mechanism for magma production. Matter starting at S rises convectively and cools by adiabatic expansion. Melting begins at M. At M_1 the liquid segregates and rises to the surface, which it reaches at temperature M_2.

r of 10^3. The Rayleigh number is:

$$R = g\alpha\beta d^4/\nu K$$

re g is the acceleration of gravity, α is the ficient of thermal expansion, d is the thick- of the convecting layer, K is the thermal sivity, ν the kinematic viscosity, and β is temperature gradient minus the adiabatic ient. With $g = 10^3$ cm sec^{-2}, $\alpha \sim 2 \times$ /deg, $\nu \sim 10^{21}$ cm^2 sec^{-1}, $K \sim 10^{-2}$ cm^2 , the Rayleigh criterion $R > 10^3$ is satisfied $^4 > 5 \times 10^{23}$ deg cm^4. If d is 10^7 cm, $\beta >$ 10^{-5}, or 5^0/km; but if $d = 2 \times 10^7$ cm, required gradient is only 0.3^0/km. The batic gradient $dT/dz = g\alpha T/c_p$ (where z pth, counted from the surface and c_p is the aric specific heat) is directly proportional he temperature T and is therefore higher re the temperature is higher. Depending the values chosen for α and for T, the batic gradient in the upper mantle might from, say 0.2 to 0.5^0/km. To this we must in phase transition zones, a term cor- onding to the entropy of the polymorphic sformation itself. Thus the gradient, aged over a thick ($\sim$200 km) section of a vecting mantle, is not likely to exceed 1^0/km may be about half that much, particularly re the temperature is low, as in a descending ent. It follows that any layer of the upper tle in which the temperature gradient is nd to be less than a few degrees per kilom- is likely to be convecting.

MPERATURES IN THE UPPER NTLE

ut how do we determine the actual gradient he mantle? Some attempts have been based he distribution of seismic velocities V_P and n the low-velocity zone of the upper mantle. e increasing pressure usually increases h V_P and V_S, while temperature has the osite effect, a decrease in velocity with easing depth implies a minimum gradient ch Anderson and Sammis (1970) estimate etween 15 and 32 deg/km for V_P and 8 to eg km for V_S. One difficulty here is that the e temperature gradient does not seem to unt for the observed distribution of both nd V_S, although, as noted by Birch (1970), velocity distribution in the low-velocity originally proposed by Gutenberg can be unted for by the temperature effect.

he properties of the low-velocity zone are largely explained in terms of incipient melting, as suggested by Anderson and Sammis (1970). It seems that the presence of very small amounts (about 1 percent) of interstitial melt could account for the marked decrease in V_S, slight decrease in V_P, and high attenuation characteristic of the zone. If this hypothesis is accepted, it follows that the temperature of incipient melting must be reached at the depth where V_S begins to decrease discontinuously. This depth is also commonly taken to mark the base of the lithosphere.

The temperature at which peridotite begins to melt (solidus) depends, of course, on its composition and on pressure; it depends also on the amount of water and on the presence of hydrous minerals and their stability field. The solidus for dry peridotite starts at about 1,100° and its slope is about 10°/kb; the slope probably increases as the mineralogy changes, with increasing pressure, from plagioclase to spinel to garnet peridotite (Wyllie, 1971, Fig. 6-10). An excess of water present as vapor lowers the dry solidus to about 1,000°C at 10 kb or more (Kushiro and others, 1968). If water is present in hornblende or phlogopite, its partial pressure is controlled by the stability of these minerals. The wet solidus initially drops below the dry solidus, but then starts to rise with a slope comparable to that for the dry system; a broad minimum occurs near 1,000°C at about 30 kb (100 km). Thus partial melting in the wet system, which marks the bottom of the lithosphere, might occur at a depth less than 50 km at 1,000 km or so from a fast-spreading oceanic ridge, and at a greater depth at a greater distance from that ridge. In shield areas, where the low-velocity zone is not well marked and lies deeper, the lithosphere may have a thickness of 150 km or more. Such large variations in lithospheric thickness must undoubtedly have an effect on the pattern of flow in the underlying mantle, with currents constrained to flow under or around the down- ward protrusions of the rigid plates; this aspect of plate theory may require further study.

A noteworthy feature of the low-velocity layer is its finite thickness, perhaps about 100 to 150 km in oceanic areas. Dorman's (1969) model for the Pacific Ocean places the top of the low-velocity layer (for V_S) at about 60 km, and the bottom at about 225 km. Its down- ward termination probably corresponds to the point where the local geotherm crosses a second time the wet solidus, on the downward side of the minimum (Anderson and Sammis, 1970).

According to Wyllie (1971, Fig. 6-18) the wet solidus at a depth of 200 km ($\sim$65 kb) is at about 1,155°C. These figures are difficult to reconcile with the higher temperatures calculated by Turcotte and Oxburgh for an oceanic plate and discussed above. Perhaps the lower termination of the low-velocity zone is not caused by the disappearance of partial melting, or perhaps there is a temperature inversion in the upper mantle, or perhaps the wet solidus for the mantle rises more rapidly to join the dry solidus, implying that there is even less than 0.1 percent of water in the upper mantle, or that its activity is reduced by dilution with CO_2.

OLIVINE TRANSITION

Another thermometer for the mantle may be provided by the occurrence near 350- to 400-km depth of a narrow zone in which seismic velocities and density rapidly increase downward. This zone has been interpreted to correspond to the phase transition from olivine to the β-phase (distorted spinel). Ringwood and Major (1970) have presented an experimental phase diagram according to which the transformation of an olivine $(Mg_{0.89}Fe_{0.11})_2SiO_4$ would begin at 109 kb and be completed at 118 kb (mean 114 kb), at 1,000°C. The temperature dependence of the transition pressure is not exactly known; Ringwood and Major take it to be 30 bars/deg and calculate that if the transition occurs at a mean depth of 400 km (132 kb), the temperature at that depth must be 1,600°C. The depth of the transition zone, which is not precisely known, may be slightly less than 400 km and locally variable. A variation in depth of about 9 km corresponds to a temperature variation of about 100°; thus if the transition occurs at 373 km, the temperature there is about 1,300°C. Graham (1970), using different arguments, estimates it to be 1,450° ± 120°C.

Return now to an oceanic plate not too far from a ridge axis where the temperature, as we have seen, might be some 1,200° at 50 km depth; the average gradient between that depth and the 370-km transition might be as low as 100°/320 km = 0.31 deg/km. Or consider a point farther away from the ridge axis where the temperature might be 1,000° at 50 km; if the temperature at 370 km is 1,570° (Graham's upper limit), the average gradient would be 1.8 deg/km. It is not argued that such gradients actually exist, but it seems permissible to argue that actual gradients, which vary locally, will generally fall between these limits. Their very low value seems to imply clearly that (1) the upper mantle is indeed convecting, and (2) the convecting layer is thicker than 100 km. Convection is thus probably not restricted to the low-velocity zone.

MAGMA FORMATION

What can magmas that appear at the Earth's surface tell us about thermal conditions in the upper mantle? We restrict our inquiry to basaltic magmas of oceanic areas, as thermal conditions in continental regions or under island arcs are even less clear.

Note first the compositional variety of basalts. As far as this author can discover, there is no consensus as to the mechanism by which this compositional variety is achieved; there is disagreement, for instance, as to whether tholeiitic magmas rise as such from the mantle, or whether they are products of fractionation of a parental picritic liquid. There is general agreement, though, that all basaltic magmas can form, in one way or another, by partial melting of peridotite of appropriate composition, with perhaps subsequent fractionation of the liquid so formed, at pressures of some 30 kb or less (that is, in the upper 100 km or so of the mantle).

The amount of water present in the mantle is important. An anhydrous pyrolite at 25 kb might begin to melt at 1,450°C, but 0.1 to 0.2 percent of H_2O will bring the solidus down to 1,100°. The amount of liquid formed at the wet solidus is very small and insensitive to temperature throughout the temperature range between the wet and dry solidus. Appreciable amounts of liquid are produced only when the temperature exceeds the dry solidus. There is some disagreement on this point, however, and Green (1970) thinks that fairly large amounts of liquid can be formed below the dry solidus, even when there is only very little water in the system. The composition of the liquid changes as the fraction of melt increases.

The fact that magmas are not observed to reach the surface at temperatures above 1,200°C must have implications with regard to conditions in the mantle. A melt rising straight from a depth of, say, 50 km would cool (by adiabatic expansion) by about 50° or less, so that its initial temperature could not be much above 1,250°. Hotter magmas forming at that depth by fractional melting would necessarily have a different composition and

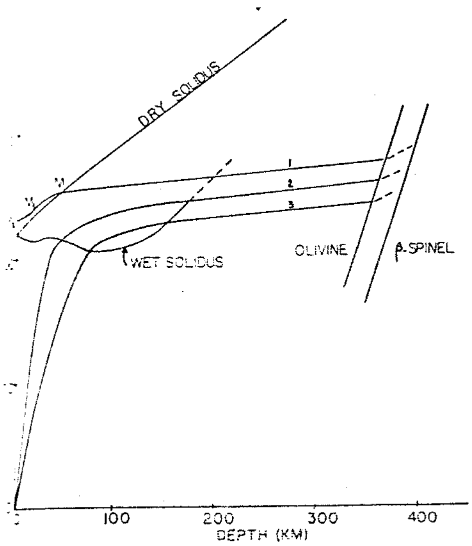

Figure 2. Possible oceanic temperature profiles. Curve 1 refers to a ridge axis, where magma segregates at a depth of about 15 to 20 km and reaches the surface at 1,150°C; the profile below M is adiabatic. M_2 is the temperature at which magma reaches the surface; M_1 is the point of segregation. Curves 2 and 3 refer to points progressively farther from the ridge axis; they represent solutions of the conduction equation in the moving lithospheric plate joined smoothly to a near-adiabatic curve below the plate (Turcotte and Oxburgh, 1969).

might be too dense to rise by buoyancy. Perhaps this is why we never see peridotitic liquids at the Earth's surface, although their absence at any exposed depth of the crust leads to the suggestion that they never form at all. The absence of magmas more basic than picrite may imply that nowhere does the temperature ever exceed 1,500°C at 100-km depth, or 1,300° at 50-km depth. The degree of fractional melting of a pyrolite mantle does not seem to ever exceed 40 percent.

To account for common magmas, most petrological schemes demand partial melting to the extent of about 20 to 30 percent. Green and Ringwood (1967) suggest 20 percent partial melting for alkali olivine basalt (segregation depth 35 to 70 km), and 20 to 25 percent partial melting for high-alumina olivine tholeiite (segregation depth 30 km). Kay and others (1970) propose, mainly for chemical reasons, that olivine tholeiites of oceanic ridges form at depths of 15 to 25 km with 30 percent partial melting. In view of this, the relatively small volume of lava that reaches the surface is noteworthy. If indeed 30 percent of the material rising at an oceanic ridge melts to form lava flows and new oceanic crust, while the residue of depleted pyrolite forms the rest of the new plate, the thicknesses of crust and plate would be in the ratio of about 30:70. The actual ratio is closer to 1:10 to 1:20, since the oceanic crust cannot consist of much more than 5 km of basalt. Thus presumably only a small fraction of the rising material undergoes partial melting to the extent of 30 percent and the bulk of the plate must consist of undepleted pyrolite which has not melted on its way up. For every 5 km of crust, there is only $5 \times 70/30 = 12$ km of depleted pyrolite and only a total of 17 km of material undergoing partial melting to the extent of 30 percent. Thus segregation of liquid magma occurs presumably no deeper than 17 km below the ridge crest.

Figure 2 shows a possible temperature profile below the axis of an oceanic ridge (curve). The profile has been drawn for magma segregating at a depth of 17 km and reaching the surface at about 1,150°C. Location of the point M is rathern arbitrary as the slope of M_1M is not known. Below M, the temperature profile is the adiabatic profile with a slope $d \ln T/dz = g\alpha/c_p = 2.5 \times 10^{-9}$ cm^{-1} (T is measured in degrees Kelvin).

The profiles marked 2 and 3 in Figure 2 are calculated following Turcotte and Oxburgh's (1969) model for a plate moving at 4.4 cm/yr. Curve 2 is for a point at 1,000 km from the ridge, where the surface heat flow is 2.3×10^{-6} cal cm^{-2} sec^{-1}. Curve 3 might represent a point on the same plate at 4,000 km from the axis where the heat flow is 1.16×10^{-6}. All calculated temperatures depend, of course, on the assumed values of heat conductivity, thermal expansion, and so on, and are therefore subject to uncertainties of 10 percent or more; such temperature profiles are drawn for illustrative purposes only.

What these curves do show is that there is no standard oceanic geotherm of general validity. Temperature at a given depth varies laterally as a function of the local age of the plate. Magma formation, also a local phenomenon, is another indicator of local disturbances; for instance, the temperature distribution beneath Hawaii is presumably quite different from what it is elsewhere in the Pacific at comparable distance from the East Pacific Rise. It would be convenient to think of a temperature profile as the sum of a local disturbance imposed on a fundamental geotherm of general validity representing some sort of equilibrium at steady-state conditions, but there are no indications that such a geotherm exists anywhere, or as to what its shape might be.

CONCLUSIONS

Plate theory requires that we now abandon the two assumptions on which geotherms are usually calculated, namely, that heat is transported mainly by conduction and vertically upward. Horizontal heat transfer in moving plates is now recognized to be important and accounts well for oceanic patterns of heat flow and the topographic profile of a spreading ridge and adjacent sea floor (Sclater and Francheteau, 1970). Convection is probably the dominant mode of heat transfer below the lithosphere, and vertical transport of heat by moving masses and rising magma may account for a large fraction of the total surface flow, even though vertical movement may be restricted, at any one time, to limited areas forming only a small fraction of the Earth's surface. Vertical temperature profiles ("geotherms") vary locally, as with the local age of a plate, or distance to a ridge axis.

As there is as yet no complete theory of convection in the mantle that takes into account internal heat generation, phase changes, and variable (temperature- and pressure-dependent) rheological properties, the temperature distribution in the mantle remains largely conjectural. The only relatively reliable way to determine that distribution is to assign certain phase changes to certain seismic discontinuities; for instance, to interpret the top of the low-velocity layer as the depth of incipient melting, and the 400-km discontinuity as the olivine-β spinel transition. Experimental data on these transitions suggest very low temperature gradients (0.3° to 1.8°/km) between the base of the lithosphere and the 400-km discontinuity, which in turn suggest convection as the dominant mechanism of heat transfer.

In Figure 2, tentative and very approximate temperature profiles are drawn for an oceanic plate at various distances from the ridge axis. The purpose of these sketches is to emphasize lateral variability; there is no single profile that can be calculated from average properties (for example, average heat flow) and taken to represent a widely applicable, average, oceanic geotherm. Nor is production of magma a normal event that occurs everywhere all the time; where magma forms, conditions are not average, nor are they representative of the whole upper mantle. It does not appear likely, therefore, that petrological variation in magmas could be accounted for by superposing on a petrological phase diagram a geotherm that is valid where magma does not form. More probably, the depth and temperature at which magmas form will be determined from petrological thermometers and barometers of the kind recently suggested by Nicholls and others (1971).

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