20 September 1974, Volume 185, Number 4156 SCIENCE

Thermal Structure of the Lithosphere: A Petrologic Model

Temperature and depth estimates are made for rocks derived from the outer 200 kilometers of the earth.

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Various models have been used to describe the thermal structure of the earth. First, there are the historical models which trace the variation in the earth's thermal field with time. These models are critically dependent on the boundary conditions, which include the initial temperature distribution, the distribution of heat sources and sinks in space and time, mechanisms of heat transport and surface temperatures, and the heat flow. Examples, which include attempts by Lubimova (*1*), Levin and Majeva (*2*), MacDonald (*3*), Reynolds *et al.* (*4*), Birch (*5*), and Hanks and Anderson (*6*), rely on radiation and conduction as the prime thermal transport mechanisms and, with modifications to allow for historical eccentricities, essentially describe the conductive and radiative cooling history of a solid sphere.

Other models include a calculation by Clark and Ringwood (*7*) of the equilibrium thermal gradients which satisfy observation of the heat flow distribution and estimates of heat source abundance and distribution within the framework of our understanding of important phase boundary changes within the mantle. Similarly, Anderson and Sammis (*8*) have used the constraints supplied by our knowledge of the density and seismic profiles to calculate equilibrium gradients, and Tozer (*9*) has produced calculated geothermal gradients based on knowledge of the electrical conductivity distribution within the earth.

The recognition of the theory of plate tectonics as a useful way of reconstructing the geometric configuration of the continents and oceans with time has led to the postulate that convection (*10*) is the driving mechanism. Thus convective heat transfer must be an important process affecting the thermal structure of the outer few hundred kilometers. A number of thermal models have been proposed to explain the consequences of convection (*11–14*) on the thermal structure of the oceans. In each case the convective cycle is seen as a perturbed event superimposed on the long-term evolutionary or steady-state models produced above.

Sclater and Francheteau (*12*) have pointed out that the heat flow in continental and oceanic provinces is related to the age of the province; but the time scale for the thermal decay of heat flow to a steady-state heat flow varies from 10^9 to 10^8 years for the continental and oceanic regions, respectively, suggesting fundamentally different thermal histories for the continental and oceanic lithospheres. It would appear that the heat flow cycle of the continental regions fall closest to the long-term evolutionary models dominated by the processes of conduction and radiative transfer, whereas the oceans represent a convective heat transfer cycle superimposed on the longer-term cooling of the earth.

An independent method of studying the thermal structure involves the examination of rock samples believed to have been derived from the mantle. The correct assignment of temperature and pressure to suites of samples from single localities allows reconstruction of paleogeotherms related to the tectonic environment. Such independent sets of data serve as useful tests of the theoretical models posed above. In this article we examine sets of data from mantle rock suites characteristic of different tectonic environments and the implications concerning the thermal and petrologic structure of the mantle.

Method Used to Determine Paleogeotherms

High pressure–high temperature experimental studies (*15*) have shown that most ultramafic rocks found at the earth's surface have been derived from the upper mantle and may thus supply information on the nature of this region. Thus determination of the ambient pressures and temperatures at which various groups of ultramafic rocks last equilibrated in the upper mantle may be used to characterize a paleogeothermal gradient for that area. Such estimates have been made for a number of suites of ultramafic rocks from different tectonic regimes. First we estimated the temperatures by a comparison of the composition of coexisting pyroxenes from the $Mg_2Si_2O_6$-$CaMgSi_2O_6$ join at 30 kilobars (*16*). We then calculated the pressures from the Al_2O_3 content of the enstatite, using experimental data for ultramafic compositions in the system MgO-Al_2O_3-SiO_2 (*17*).

The comparison of natural composi-

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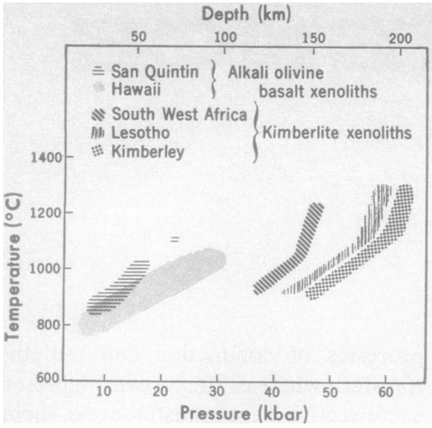

Fig. 1. Pressure-temperature plot of ultramafic xenoliths from alkali basalts and kimberlite pipes.

tions with experimental data for synthetic systems inevitably leads to uncertainties in the absolute values derived. In the case presented here the effect of the ferrosilite, jadeite, and tschermak component in the pyroxenes has not been considered, and the resulting temperature estimates are slightly higher than the true values. Important components that we ignored in using the three-component system MgO-Al_2O_3-SiO_2 are FeO, CaO, and Cr_2O_3. Theoretical calculations of the effect of CaO and FeO suggest that the pressure values are high by approximately 10 to 15 percent (*17, 18*). The effect of Cr_2O_3 has not been evaluated and must be considered as a major potential problem in ultramafic rocks with chromite spinel as the aluminous phase. Despite the uncertainties in the absolute values, it appears that the relative assignments of temperature and pressure for single suites are essentially correct (*17, 18*) and have some validity in defining geothermal gradients.

Using the above method, we estimated pressure-temperature conditions of equilibration for various groups of ultramafic rocks from different tectonic environments. A pressure-temperature plot for a particular group of ultramafic rocks from a particular tectonic setting defines a "geotherm" in that tectonic region. This geotherm relates temperature to depth beneath that tectonic region at a time just prior to the emplacement of the ultramafic rocks.

We chose four broad groups of ultramafic rocks for the present study. These are: (i) spinel lherzolite xenoliths in alkali basalts, (ii) spinel and garnet peridotite xenoliths in kimberlite from the South African shield region, (iii) high-temperature peridotite intrusions, and (iv) alpine-type peridotites.

Ultramafic Rock Suites

Spinel lherzolite xenoliths in alkali basalts. The sources of data in this group are from San Quintín, Baja California (*19*), and Hawaii (*20*). In all these cases the host basalts are Quaternary in age.

The San Quintín locality is approximately 260 kilometers south of the United States border along the west coast of the Baja Peninsula. The xenoliths are found in an alkali olivine basalt flow of Pleistocene to Recent age. The lithosphere beneath San Quintín has had a complex history, having been associated with the interaction of the North American continent with the East Pacific Rise approximately 30 $\times 10^6$ years ago and with the opening of the Gulf of California at least 4×10^6 years ago (*21*). At present, the faults and structural relations in the northern region of the gulf indicate that the present spreading center lies 200 kilometers to the southeast of San Quintín in the Delfin Basin (*22*). Although complex, it seems reasonable to assign the lithosphere in this region to a near-ridge tectonic environment for the last 30×10^6 years.

Recently (*23, 24*), considerable attention has been given to the origin of the Hawaiian volcanic chains. According to Morgan (*23*), these chains have formed by thermal plumes under fixed "hot spots," whereas Shaw and Jackson (*24*) introduce the term "melting anomaly" to emphasize that the unusual volcanic activity may be the result of a variety of processes including shear melting caused by plate motion. For our purpose, we consider the tectonic setting of the Hawaiian lithosphere as that of a mid-plate situation away from a mid-oceanic ridge.

The pressure-temperature assignments for ultramafic xenoliths from the San Quintín and Hawaii localities (Fig. 1) define approximately linear trends which reflect the ambient geothermal gradients in the lithosphere at the time of extrusion. The San Quintín lherzolites define a steeper thermal gradient than that of Hawaii.

Spinel and garnet peridotite xenoliths from kimberlites. Temperature and pressure assignments for a large number of ultramafic xenoliths from three South African kimberlite localities are shown in Fig. 1 (*25*). Suites of xenoliths from each locality define a separate trend in the pressure-temperature space. These trends have been interpreted to mark paleogeothermal gradients beneath the South African region in the Cretaceous time when the kimberlites erupted (*17, 18, 25*). In each of these three regions there is a sudden break in the slope of the interpreted paleogeothermal gradient. The break corresponds to the boundary between xenoliths with sheared textures at higher pressures and xenoliths with granular textures at lower pressures, and has been interpreted to mark the top of the low-velocity zone under the South African shield in Cretaceous time (*17, 18*).

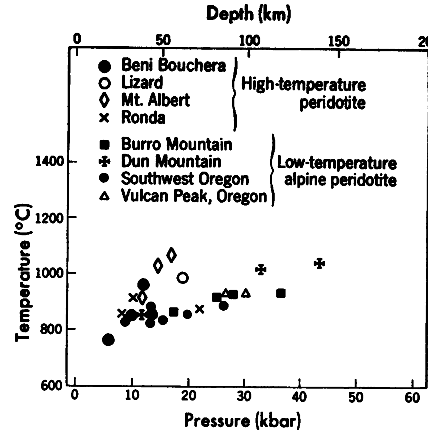

Fig. 2. Pressure-temperature plot of samples from high-temperature and alpine-type peridotite massifs.

High-temperature peridotite intrusions. Data from the Mt. Albert, Quebec, Canada (*26*); Lizard, England (*27*); Ronda, Spain (*28*); and Beni Bouchera, Morocco (*29*); peridotite intrusions are plotted in Fig. 2. All these peridotite massifs show well-developed contact metamorphic aureoles and are considered to have intruded into the country rocks at high temperatures. Figure 2 indicates that these peridotite bodies equilibrated at relatively high temperatures, conforming to field observations. Only one sample from Ronda shows anomalously low temperature. Except for this data point, all the others conform to a curvilinear trend of increasing temperature with increasing pressure.

Alpine-type peridotites. Alpine-type peridotites are one of the principal classes of ultramafic rocks characterized by their associations in the folded mountain belt. Examples of these intrusions for which appropriate chemical data are available are Burro Mountain, California (*30*); southwest Oregon (*31*); Vulcan Peak, Oregon (*32*); and Dun Mountain, New Zealand (*33*). Pressures and temperatures of equilibration of these alpine-type peridotites (Fig. 2) show an essentially linear variation of temperature with pressure.

Discussion of Pressure-Temperature Plots

The wide range of temperatures and pressures determined for the ultramafic xenoliths from volcanic pipes indicates that they are not related to the genesis of the host magma but are accidental samples torn off the walls of the volcanic pipe. Each sample thus gives the environmental conditions at its assigned depth, and the set of samples from a single locality should define the ambient geothermal gradient at the time of intrusion. On the basis of this assumption, the suites of samples from San Quintín; Hawaii; Louwrencia, South West Africa; and Lesotho and Kimberley, South Africa; define successively lower geothermal gradients (Fig. 1). The interpreted gradients for the kimberlite suites show a sudden change of slope which occurs at successively greater depth for the Louwrencia, Lesotho, and Kimberley areas. The change of slope has been interpreted to coincide with the lithosphere-asthenosphere boundary (*17, 18*), a conclusion commensurate with the sudden increase in the degree of deformation for xenoliths derived from the high-pressure side of the kink. Similarly, the kink marks the location of a potential mantle melting curve where the partial pressure of H_2O (P_{H_2O}) is less than the total pressure (P_t), or where $P_{fluid} = P_t$ when the fluid is composed of H_2O-CO_2 mixtures in which the mole fraction of H_2O is 0.25 (*34*). The apparent geothermal gradient is thus composed of two components. The first is a lithospheric component which is compatible with calculated models for the cooling of the lithosphere, and the second shows a dramatic steepening of the thermal gradient in the asthenosphere not anticipated by the calculated models (*11–14*). Boyd (*18*) has suggested that the change in slope results from frictional heating in the asthenosphere during the period of drift associated with the breakup of Gondwanaland. The asthenospheric data thus would illustrate a perturbation on the steady-state model. A comparable interpretation for a perturbed gradient suggests that the steepened gradient results from diapiric movement of mantle material within the lithosphere associated with the event of kimberlite extrusion (*35*). Any model ascribing the change of slope to a steady-state phenomenon must ascribe dramatically different thermal properties to the asthenosphere, a conclusion not supported by the general similarity of rock types found in both the lithosphere and the asthenosphere. The derived geothermal gradients are thus important for two reasons: they illustrate the steady-state cooling of the lithosphere and they point to the presence of important transient thermal processes in the asthenosphere.

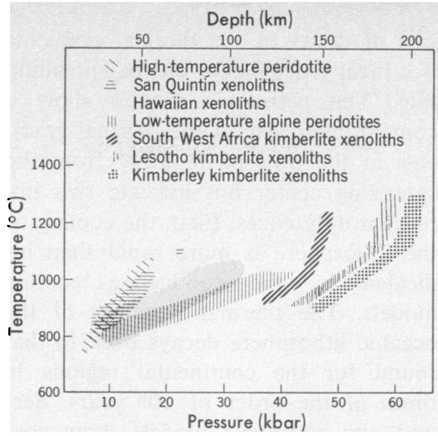

Fig. 3. Summary of pressure-temperature plots for ultramafic rock suites from different tectonic environments.

The curvilinear trend for samples from ultramafic intrusions (Fig. 2) marks the temperature-pressure path by which they diapirically migrated to the surface. The path is subject to two limiting interpretations. It may represent an adiabatic migration to the surface, or alternately it may represent complete thermal equilibrium with the environment, in which case the path also marks the local geotherm. For an adiabatic path the trajectory should be essentially isothermal. Since the high-temperature and alpine peridotites show systematic variations of temperature with depth well in excess of an adiabatic process, a nonadiabatic path is indicated, and the trajectories are assumed to be approximately parallel to the ambient geothermal gradient.

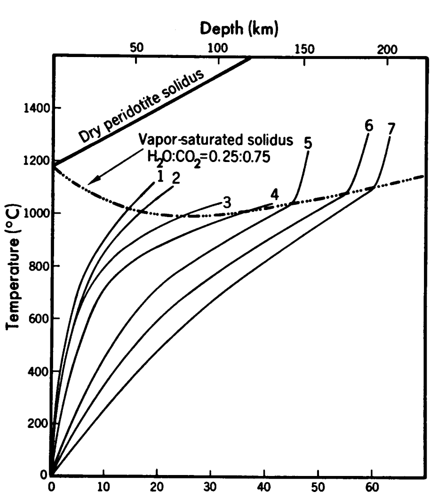

Fig. 4. Model of variations in the geothermal gradient in the lithosphere. Model geotherms: curve 1, mid-ocean ridge geotherm; curves 2 to 4, transient oceanic geotherms; curves 5 to 7, Precambrian shield geotherms.

Tectonic Setting of Rock Suites

Within the framework of plate tectonics the different ultramafic rock suites may be used to categorize the lithosphere in the following tectonic environments. The suites of xenoliths from the South African kimberlites are representative of a cross section through a continental shield (*25*). The samples in alkali basalts from the San Quintín area are close to an active spreading center, whereas comparable samples from Hawaii are characteristic of a mid-ocean region. The circum-Pacific alpine ultramafic intrusions occur in subduction zones and represent mantle material derived from the boundary regions between oceanic and continental crust. High-temperature peridotites mark the loci of hot mantle material diapirically intruding the surface and are comparable to the tectonic framework expected along spreading centers (*36*).

Reconstructed in a tectonically evolutionary model, the rock suites can be rearranged in terms of their "tectonic distance" from a spreading center. The rock suites fall in the sequence consisting of high-temperature peridotite, San Quintín, Hawaii, circum-Pacific intrusions, and South African kimberlite xenoliths with increasing tectonic distance from a spreading center. The high-temperature peridotites represent the mid-ocean ridge, the San Quintín and Hawaii samples distances successively farther from a ridge, the alpine peridotites an ocean margin, and the suites of samples from Louwrencia, Lesotho, and Kimberley successively greater distances into a continental region. Although the samples are not linearly distributed away from a single spreading center, we believe that the tectonic distance is a valid variable by which to consider the data. Thus, to a first approximation, the tectonic distance away from a spreading center may be used in the same manner as distance or spreading rate and time in the paper by Sclater and Francheteau (*12*). Figures 3 and 4 represent a compilation of the petrological data from this point of view. The derived geothermal gradients 1 through 7 (Fig. 4) are at successively greater tectonic distances from a hypothetical spreading center. The petrologic data are comparable to other thermal models (*11–14*) for the lithosphere but, as indicated above, suggest that additional thermal processes are active in the lithosphere. Since the different rock suites do not occur within a single spreading event, it has not been possible to calibrate the rate of decay. In Fig. 5 a comparable diagram shows the variation of the geothermal gradients in a hypothetical cross section through a spreading event. Possibly in preliminary calibrations the distance from Hawaii to the ridge could be used as a scale for a Pacific or fast-spreading ridge type of ocean. In contrast a scale for an Atlantic or slow-spreading center would be the distance from the mid-Atlantic ridge to the Louwrencia locality.

Comparison of Petrologic with Theoretical Thermal Models

The thermal models of Sclater and Francheteau (*12*), Verhoogen (*14*), and Forsyth and Press (*13*) all show a systematic decrease in the thermal gradients away from a spreading center. The rate of decay of the thermal gradients is critically dependent on the spreading rate. The petrologic models show a comparable decay of the thermal gradients in the lithosphere away from the spreading center but indicate two important differences. First, the cooling of the lithosphere is more rapid than indicated by the conductive cooling models. The thermal structure of the oceanic lithosphere decays back to that found for the continental regions in times of the order of 10^8 years. Second, the petrologic models show considerably more detail. Of special interest is the sudden steepening of the thermal gradients which is seen in the Louwrencia, Lesotho, and Kimberley localities at successively greater depths. As previously suggested, the change of slope is correlated with the lithosphere-asthenosphere boundary. No thermal models show this phenomenon. The steeper gradients indicate significant differences in the thermal structure or history of the asthenosphere not accounted for by the theoretical models. Either frictional heating may be responsible for the steeper gradient (*18*), or the steepened slope may result from convective movement in the asthenosphere (*35*). Alternately, but more unlikely, for a steady-state model one must ascribe significantly different thermal characteristics to the asthenosphere.

The petrologic model corresponds closely to that proposed by Sclater and Francheteau (*12*) and Forsyth and Press (*13*), in which the cycle of oceanic formation is a convective event superimposed on the long-term cooling of the earth. The continental regions correspond more closely to the longer-term 10^9-year cooling history of the earth, whereas a spreading event can be viewed as a local perturbation which decays back to the longer-term thermal evolution of the earth as a whole. The oceanic cycle decays in approximately 10^8 years.

Reconstruction of Mantle Petrology and Thermal Structure

The petrologic and thermal profiles at each locality allow the use of the tectonic distance as a means of constructing a hypothetical cross section through a spreading center into an adjoining continent (Fig. 5). The phase chemistry for the lithosphere illustrates that the suboceanic lithosphere is composed of spinel-bearing peridotites. Beneath the continents, the spinel peridotites extend to depths of approximately 125 kilometers. At greater depths there is a zone of spinel- and garnet-bearing peridotites which thickens toward the interior of the continent and finally passes into a zone of garnet-bearing peridotites. The boundaries between the different phase assemblages are diffuse and should really be considered as maxima in rock type distributions since they depend on both the thermal gradient and composition (*25*). Further, all the data would suggest that the lithosphere is essentially ultramafic in composition, with only minor proportions of mafic rocks such as pyroxenites or eclogites.

The boundary between the lithosphere and the asthenosphere, previously interpreted to correlate with the kink in the geothermal gradient (Fig. 3) for the South African localities, is comparable to a mantle melting curve in which the gas phase is composed of H_2O-CO_2 mixtures in which the mole fraction of H_2O is 0.25 (*34*). The boundary has been "mapped" beneath the continental and mid-oceanic region, and, assuming a similar mole fraction of H_2O for the whole mantle, has been extended to the

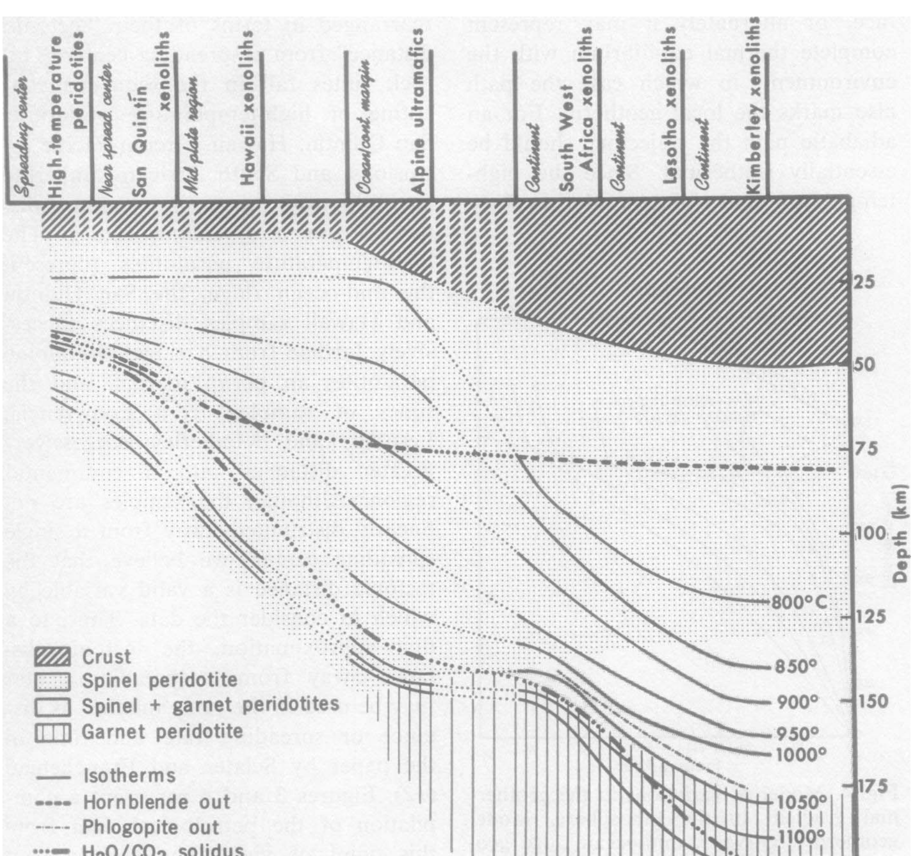

Fig. 5. Petrologic and thermal structure in a hypothetical cross section through a spreading center and the adjacent continental region.

spreading center location. It would appear that the boundary of incipient melting thus marks the top of the asthenosphere, and the systematic decrease of the thermal gradient away from a spreading center accounts for the observed thickening of the lithosphere away from a spreading center.

Dehydration reactions show possible correlations with melting phenomena in the mantle. For example, beneath a midoceanic ridge the dehydration of hornblende in a peridotite-H_2O system (*37*) corresponds to the interpolated depth of melting for a mantle with a mole fraction for H_2O of 0.25 in a H_2O-CO_2 gas phase (Fig. 5). Similarly, the dehydration of phlogopite as observed in kimberlite xenoliths (*25*) corresponds closely with the same solidus beneath the continental regions. The correspondence of mantle melting and dehydration reactions suggests a possible cause-and-effect relationship as previously proposed by Wyllie (*38*).

Conclusions

A preliminary evaluation of the thermal history of the upper mantle as determined by petrologic techniques indicates a general correspondence with theoretically derived models. The petrologic data supply direct information which may be used as an independent calibration of calculated models, serve as a base for evaluating the assumptions of the theoretical approach, and allow more careful selection of the variables describing mantle thermal properties and processes.

Like the theoretical counterpart, the petrological approach indicates that the lithosphere is dominated by two thermal regimes: first, there is a continental regime which cools at rates of the order of 10^9 years and represents the long-term cooling of the earth. Secondly, superimposed on the continental evolution is the thermal event associated with the formation of an oceanic basin, and which may be thought of as a 10^8-year convective perturbation on the continental cycle. Of special interest is petrologic evidence for a sudden steepening of the thermal gradients across the lithosphere-asthenosphere boundary not seen in the theoretical models. The unexpected change of slope points to the need for a critical reevaluation of the thermal processes and properties extant in the asthenosphere. The potential of the petrologic contribution has yet to be fully realized. For a start, this article points to an important body of independent evidence critical to our understanding of the earth's thermal history.

References and Notes

1. H. A. Lubimova, *Geophysica* **1**, 115 (1958); *Ann. Geofis.* **14**, 65 (1961); *ibid.*, p.78; in *The Earth's Mantle*, T. F. Gaskell, Ed. (Academic Press, New York, 1967), p. 232.
2. B. J. Levin and S. V. Majeva, *Ann. Geofis.* **14**, 145 (1961).
3. J. F. MacDonald, *J. Geophys. Res.* **64**, 1967 (1959); *ibid.* **66**, 2489 (1961); *ibid.* **69**, 2933 (1964).
4. R. T. Reynolds, P. E. Fricker, A. L. Summers, *ibid.* **71**, 573 (1966).
5. F. Birch, *Geol. Soc. Am. Bull.* **76**, 133 (1965).
6. T. C. Hanks and D. L. Anderson, *Phys. Earth Planet. Interiors* **2**, 19 (1969).
7. S. P. Clark, Jr., and A. E. Ringwood, *Rev. Geophys.* **2**, 35 (1964).
8. D. L. Anderson and C. Sammis, *Phys. Earth Planet. Interiors* **3**, 41 (1970).
9. D. C. Tozer, in *The Earth's Mantle*, T. F. Gaskell, Ed. (Academic Press, New York, 1967), p. 327.
10. W. M. Elsasser, *J. Geophys. Res.* **76**, 101 (1971).
11. M. G. Langseth, X. LePichon, M. Ewing, *ibid.* **71**, 5321 (1966); D. P. McKenzie, *ibid.* **72**, 6261 (1967); E. R. Oxburgh and D. L. Turcotte, *ibid.* **73**, 2643 (1968); N. H. Sleep, *ibid.* **74**, 542 (1969); J. W. Minear and M. N. Toksöz, *ibid.* **75**, 1397 (1970); Y. Bottinga and C. J. Allegre, *Tectonophysics* **18**, 1 (1973).
12. J. G. Sclater and J. Francheteau, *Geophys. J. R. Astron. Soc.* **20**, 509 (1970).
13. D. W. Forsyth and F. Press, *J. Geophys. Res.* **76**, 7963 (1971).
14. J. Verhoogen, *Geol. Soc. Am. Bull.* **84**, 515 (1973).
15. I. D. MacGregor, *J. Geophys. Res.* **73**, 3737 (1968).
16. B. T. C. Davis and F. R. Boyd, *ibid.* **71**, 3567 (1966).
17. I. D. MacGregor, *Am. Mineral.* **59**, 110 (1974); B. J. Wood and S. Banno, *Contrib. Mineral. Petrol.* **42**, 109 (1973).
18. F. R. Boyd, *Geochim. Cosmochim. Acta* **37**, 2533 (1973).
19. A. R. Basu, *Geol. Soc. Am. Abstr. Programs* **5** (No. 7), 542 (1973).
20. R. W. White, *Contrib. Mineral. Petrol.* **12**, 245 (1966); M. V. Beeson and E. D. Jackson, *Mineral. Soc. Am. Spec. Pap. 3* (1970), p. 95.
21. T. Atwater, *Geol. Soc. Am. Bull.* **81**, 3513 (1970); R. L. Larson, H. W. Menard, S. M. Smith, *Science* **161**, 781 (1968); D. G. Moore and E. C. Buffington, *ibid.*, p. 1238.
22. T. L. Heney and J. L. Bischoff, *Geol. Soc. Am. Bull.* **84**, 315 (1973).
23. W. J. Morgan, *Geol. Soc. Am. Mem. 132* (1972), p. 7.
24. H. R. Shaw and E. D. Jackson, *J. Geophys. Res.* **78**, 8634 (1973).
25. I. D. MacGregor, *Proc. Int. Kimberlite Conf.* (Cape Town, South Africa, October 1973), in press.
26. ———, thesis, Queen's University, Kingston, Ontario (1962).
27. D. H. Green, *J. Petrol.* **5**, 134 (1964).
28. J. S. Dickey, Jr., *Mineral. Soc. Am. Spec. Pap. 3* (1970), p. 33.
29. J. Kornprobst, *Contrib. Mineral. Petrol.* **23**, 283 (1969).
30. R. A. Loney, G. R. Himmelberg, R. A. Coleman, *J. Petrol.* **12** (part 2), 245 (1971).
31. L. G. Medaris, *Geol. Soc. Am. Bull.* **83**, 41 (1972).
32. G. R. Himmelberg and R. A. Loney, *ibid.* **84**, 1585 (1973).
33. G. A. Challis, *J. Petrol.* **6**, 322 (1965).
34. A. L. Boettcher, B. O. Mysen, F. J. Modreski, *Proc. Int. Kimberlite Conf.* (Cape Town, South Africa, October 1973), in press.
35. O. L. Anderson and P. C. Perkins, *Proc. Int. Kimberlite Conf.* (Cape Town, South Africa, October 1973), in press; H. W. Green and Y. Gueguen, *Nature (Lond.)* **249**, 617 (1974).
36. W. E. Bonini, T. P. Loomis, J. D. Robertson, *J. Geophys. Res.* **78**, 1372 (1973).
37. I. Kushiro, *Carnegie Inst. Wash. Year Book* **68**, 245 (1970).
38. P. J. Wyllie, *The Dynamic Earth* (Wiley, New York, 1971).

39. Financial support for this research was provided by the University of California, Davis, faculty research grant D-525 to I.D.M. and a Penrose grant from the Geological Society of America to A.R.B.