scattering, the time delay to the peak would be directly proportional to the epicentral distance, while in the presence of transverse scattering the delay will approach proportionality to the square of the distance that diffusion would give. The increased seismic paths proposed for future experiments should demonstrate this effect, and, if the present discussion represents the essentials of the phenomenon, one would expect the peak amplitude to be reached much sooner than if the process were one of two-dimensional diffusion.

Appendix

The result of the impact of the Apollo 13 S-IVB casing has just been reported (*7*). The signal, measured at a distance of 142 km from the impact, appears to be of the same general nature as the LM signal of Apollo 12. However, there is reported to be a clearly recognizable onset of signal at 32 seconds after impact, and this increases the confidence that the first arrival signal suspected at 23 seconds for the Apollo 12 case is also genuine. These values can be used to specify more closely some of the properties of the medium.

The main shape of the signal received is not very critically dependent on the value of the velocity gradient and is almost independent of the circumstances below the depth where a velocity of more than 40 times the surface velocity is reached, since very few rays and therefore very little of the energy reaches that depth. Nevertheless, the onset time for the first arrival of the fastest wave is dependent on these quantities. If we assume that solid rock or fully compacted material exists at a certain depth, and if we assume for it a speed of approximately 6 km sec^{-1}, increasing only slowly with depth thereafter, one can derive the initial signal arrival time as a function of this depth. Using a surface velocity of 150 m sec^{-1} and a linear velocity gradient from there down to the layer of full compaction, we derive the following relations between that depth h and the first arrival time T for the Apollo 12 LM and for the Apollo 13 S-IVB impact events.

h (km)	Apollo 12 T (sec)	Apollo 13 T (sec)
3	16.1	27.1
6	19.6	30.6
9	23.1	34.1
12	26.5	37.5

The times quoted for the first arrival signals are approximately 23 seconds for Apollo 12 and 32 seconds for Apollo 13, and this would agree in both cases with a depth for full compaction of between 6 and 9 km. This depth would be less for a lower velocity in the fully compacted material.

THOMAS GOLD
STEVEN SOTER

Center for Radiophysics and Space Research, Cornell University, Ithaca, New York 14850

References and Notes

1. G. V. Latham, M. Ewing, F. Press, G. Sutton, J. Dorman, Y. Nakamura, N. Toksöz, R. Wiggins, J. Derr, F. Duennebier, *Science* **167**, 455 (1970).
2. L. D. Leet, *Earth Waves* (Harvard Univ. Press, Cambridge, Mass., 1950), p. 83.
3. E. Orowan [in *Non-Elastic Processes in the Mantle*, D. C. Tozer, Ed. (Blackwell, London, 1967), p. 196] has stressed a similar point in relation to cracks in solids. B. I. Pandit and D. C. Tozer [in *Nature* **226**, 335 (1970)] discuss aspects of sound propagation in porous materials in a vacuum.
4. L. M. Brekhovskikh, *Waves in Layered Media* (Academic Press, New York, 1960), p. 499.
5. J. V. Evans, *Radio Sci.* **69D**, 1637 (1965).
6. E. Schreiber, O. L. Anderson, N. Soga, N. Warren, C. Scholz, *Science* **167**, 732 (1970). However an *unstressed* longitudinal velocity of 1.07 km sec^{-1} reported by these authors is an order of magnitude larger than that measured on the moon. The density of 2.2 g cm^{-3} that they associate with this figure indicates substantially more compaction than seems to be the case for the undisturbed lunar surface.
7. Apollo 13 Seismic Experiment Press Conference, 15 April 1970.
8. We thank H. Bondi for valuable discussions concerning wave propagation in inhomogeneous media. We also wish to thank L. Whitehill for suggestions relating to the calculations. Work on lunar studies at Cornell is supported by NASA grant NGL-33-010-005. S.S. is supported by a NASA traineeship.

23 April 1970; revised 22 June 1970

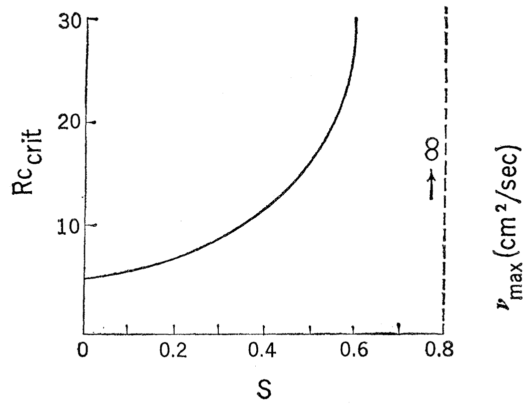
Phase Change Instability in the Mantle

Abstract. *In the presence of a temperature gradient, phase changes of the type believed to exist in the upper mantle, in which the less dense phase lies above the dense phase, may be unstable. Approximate calculations show such phase change instabilities are possible for both the 400-kilometer olivine-spinel phase transition and also for partial melting at shallower depths. The resulting flow patterns may provide a driving mechanism for the new global tectonics.*

Thermal convection within the earth's mantle has been proposed to explain continental drift (*1*). The Rayleigh number for the mantle, based on the value of viscosity inferred from postglacial uplift of Scandinavia (*2*), is several orders of magnitude larger than the critical Rayleigh number for the onset of convection (*3, 4*). Also, estimates of surface velocity and heat flux from constant property theories of thermal convection are in good agreement with observations (*5*).

Seismological (*6*) and geochemical (*7, 8*) evidence indicates that one or more phase changes occur in the upper mantle at a depth of about 400 km. The most important is likely to be the olivine-spinel phase transition. It is of interest to consider the influence of such a phase change on mantle convection. The volume and the entropy changes have the same sign for this phase transition (*7, 8*), so that heat is evolved when going from the olivine (light) to the spinel (heavy) phase and is absorbed when going from the heavy to the light. In this case the Clapeyron curve has a positive slope. Seismic evidence shows that the dense phase lies beneath the light phase; under isothermal conditions such a phase change is stable. If the fluid moves upward, the dense phase transforms to the light one and heat is absorbed. Thus the fluid is cooled and there is a downward stabilizing body force. On this basis it has been argued that the mantle phase change would act as a barrier to thermal convection (*4*). However, others (*9*) have argued qualitatively that large-scale convection could penetrate the phase change interface. A quantitative analysis of the influence of phase transformations on the stability of a fluid has recently been made (*10*). In this report we apply this theory to phase changes in the mantle.

We consider a simple model in which a two-phase fluid is confined between horizontal planes separated by the distance $2d$. The phases are assumed to be in thermodynamic equilibrium, so that the location of the phase boundary is determined by the intersection of the Clapeyron curve with the pressure-temperature curve for the fluid. In order for the light phase to lie above the heavy phase, it is necessary that the slope of the pressure-temperature curve exceed the slope (assumed positive) of the Clapeyron curve. The univariant phase boundary is initially midway between the planes. Both phases are assumed to have the same values of absolute viscosity μ, thermal conductivity k, and specific heat at constant pressure c_p (μ, k, and c_p are constants). A constant negative temperature gradient of absolute magnitude β, and a constant pressure gradient $-\rho g$ (g is the gravitational acceleration and ρ is the density) are present in this static state. The change in density $\Delta\rho$ at the phase boundary is an essential feature of the model. However, elsewhere each phase can be assumed to be an incompressible fluid of density ρ (for the olivine-spinel phase change the fractional density change $\Delta\rho/\rho$ is only about 0.1); the thermal expansion of the fluid is not considered. Since the coefficients



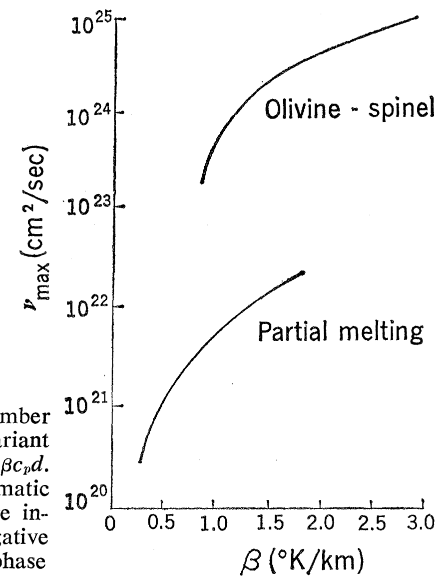


Fig. 1 (above). The critical Rayleigh number Rc_{crit} for the onset of instability of a univariant phase boundary as a function of $S = Q/\beta c_p d$. Fig. 2 (right). The maximum value of kinematic viscosity ν which will lead to a phase change instability versus the magnitude of the negative temperature gradient for the olivine-spinel phase transition and for partial melting.

of thermal expansion for both phases are taken to be zero, the Rayleigh instability is *not* present.

It is found that this basic state may be unstable to infinitesimal disturbances. The instability mechanism may be understood as follows. In a region of downward flow there is an inflow of relatively cold material from above the interface (due to the zero-order temperature gradient). Since the interface must remain on the Clapeyron curve, the lower temperature forces the interface to a region of lower hydrostatic pressure—that is, upward. With the interface displaced upward, the heavier material below the interface gives a hydrostatic pressure head tending to drive the flow downward, which leads to instability. However, the downward flow of fluid through the interface releases heat, thus tending to warm the fluid and return the phase boundary to its unperturbed location. The inflow of cold material tends to promote instability, whereas release of heat by the phase change promotes stability. The necessity of overcoming viscous dissipation in the fluid provides a further stabilization mechanism.

A quantitative analysis of this instability has been given elsewhere (*10*). The parameters determining the stability of the unperturbed state against infinitesimal perturbations are

$$Rc = \frac{gd^3\Delta\rho/\rho}{8\bar{\nu}\kappa(\rho g/\gamma\beta - 1)}$$

and

$$S = Q/\beta c_p d$$

where Q is the energy required per unit mass to change the dense into the less dense phase, T is the temperature at which the phase transition occurs, $\nu = \mu/\rho$ is the kinematic viscosity, $\kappa = k/\rho c_p$ is the thermal diffusivity, and $\gamma = Q\rho^2/T\Delta\rho$ is the slope of the Clapeyron curve in a pressure-temperature diagram evaluated at the point on the curve that represents the unperturbed state. The parameter Rc plays the role of a Rayleigh number. The parameter S is the ratio of the temperature increase due to the energy released in the phase change Q/c_p to the temperature difference between the boundaries and the interface βd. This parameter provides a quantitative measure of the stabilizing influence of the latent heat release versus the destabilizing influence of the inflow of cold material from above. The static state is unstable if $0 < S < 4/5$ and $Rc > Rc_{crit}$, where Rc_{crit} is given as a function of S in Fig. 1.

We assume that a breakdown of olivine to a spinel structure occurs in the mantle near a depth of 400 km. After approximating this phase change by a simple univariant system, we apply the results discussed above to determine whether such a phase change would be stable or unstable to small perturbations. First we will determine the value of the temperature gradient necessary for the parameter S to be sufficiently small for instability. For the olivine-spinel phase change, we take $Q = 9.8$ kcal mole^{-1} (*7*), and $c_p = 0.3$ cal g^{-1} °K^{-1}. Since the phase change occurs at a depth of 400 km, we take $d = 400$ km. If $S = 4/5$, it is necessary to have $\beta = 0.73$°K km^{-1}. Values of $\beta > 0.73$°K km^{-1} could lead to instability. Conduction gradients in the mantle are 10°K km^{-1} or higher and the adiabatic gradient in the mantle is near 0.5°K km^{-1}. Therefore it is likely that β will be sufficiently large for instability to occur.

It is also necessary that the equivalent Rayleigh number Rc must exceed its critical value if an instability is to occur. We take $\gamma = 0.0625$ kb °K^{-1}, $\rho = 4$ g cm^{-3}, $g = 10^3$ cm sec^{-2}, $T = 1800$°K, and $\kappa = 10^{-2}$ cm^2 sec^{-1}. All the quantities necessary for evaluation of the Rayleigh number can be estimated with fair accuracy except the viscosity. Therefore we find the maximum value of the viscosity which leads to instability. This maximum value of the kinematic viscosity is plotted as a function of β in Fig. 2. It is seen that the maximum viscosity is of the order of 10^{24} cm^2 sec^{-1} for reasonable values of the temperature gradient β. If the viscosity is less than this value, an instability can be expected. Uplift data suggest that, at least over some range of depths, the viscosity of the mantle is in the range $\nu = 10^{21}$ to 10^{22} cm^2 sec^{-1}, so that this requirement on the viscosity is not unreasonable.

Although the analysis presented above is only approximately valid for the mantle, the results obtained indicate that the phase change instability may occur. It is doubtful that a more accurate analysis can be made until more information is available on the mantle phase change.

A second possible application of the phase change instability is to the problem of migrating mid-oceanic ridge systems. The causes of the motions of the large lithospheric plates which are generated at mid-oceanic ridges are at present rather uncertain, and a number of alternative hypotheses must be entertained (*11*). If, however, the plates are driven by a diverging upward flow centered beneath the ridge and extending deep into the mantle for 500 km or more, it is difficult to understand how the ridges themselves are able to migrate relatively rapidly (in some cases, at many times the spreading rate) because the deep flow structure would have to migrate with them.

It is possible that the phase change instability provides a mechanism for the migration of flow beneath ridges. One of the essential features of mid-oceanic ridges is that they are the loci of intense volcanic activity and that sufficient basaltic magma is continuously extruded close to the crest for a surface layer some 4 to 5 km thick to be built up, continuously generating new oceanic crust. We therefore consider the phase change instability which might exist between the crystalline peridotite of the upper mantle and partially molten peridotite (the partial melt fraction has the composition of basalt).

If the mantle at, say, a depth of 100 km were just above its temperature of beginning of melting (as has been suggested for the seismic low-velocity zone), the introduction of some pressure perturbation (conceivably arching of the lithosphere through tectonic activity, or some kind of surface rifting) might be sufficient to initiate a self-sustaining upward flow beneath the locus of the perturbation. If fragmentation of the lithosphere ensued, sea-floor spreading could take place with ridge formation, eruption of magmas, and generation of new crust. If the ridge remained stationary or nearly so, a relatively deep flow might ultimately become established. The boundary conditions for the motion of the new plates generated at the ridge might, however, be such that the ridge itself was constrained to migrate; in that the ridge is the locus of rifting, this amounts to the migration of a pressure perturbation. In turn, this should cause the phase change instability to propagate laterally at the level of the low-velocity zone; the instability should be able to sustain the horizontal surface flows as it migrated.

This suggestion is advanced somewhat tentatively, since the theory of the instability is only approximately valid for this phase change (it is at least divariant). It would, however, imply that slowly moving or static ridges, if they could be recognized, should be characterized by deep flows, and rapidly migrating ones by shallow flows. The perturbation of mantle isotherms associated with these two kinds of flow might ultimately be distinguishable by seismic methods. It is also possible that the two kinds of flow might be characterized by slightly different compositions of magmas extruded at the surface.

Finally, we test this melting instability quantitatively. We assume one phase to contain no liquid and the other phase to contain a partial melt fraction of 5 percent. We take the partial melt fraction to be basalt. The denser (no partial melt) phase lies beneath the lighter phase, and heat is absorbed in going from the denser to the lighter phase. For the 5 percent partial melt fraction, we take $Q = 5.4$ cal g^{-1}, $c_p = 0.3$ cal g^{-1} °K^{-1}, $d = 100$ km, $\gamma = 0.118$ kb °K^{-1}, $\rho = 3$ g cm^{-3}, $g = 10^3$ cm sec^{-2}, $T = 1500$°K, and $\kappa = 10^{-2}$ cm^2 sec^{-1}. We can determine the maximum value of the kinematic viscosity which will lead to an instability as a function of β. The result is given in Fig. 2. The maximum value of the kinematic viscosity for instability is of the order of $\nu = 10^{21}$ cm^2 sec^{-1}. This value is not unreasonable, particularly since the presence of the partial melt is likely to reduce the viscosity significantly.

GERALD SCHUBERT
Department of Planetary and Space Science, University of California, Los Angeles 90024

D. L. TURCOTTE
Graduate School of Aerospace Engineering, Cornell University, Ithaca, New York 14850

E. R. OXBURGH
Department of Geology and Mineralogy, University of Oxford, Oxford, England

References and Notes

1. A. Holmes, *Trans. Geol. Soc. Glasgow* **18**, 559 (1931).
2. N. A. Haskell, *Physics New York* **6**, 265 (1935).
3. C. L. Pekeris, *Mon. Notic. Roy. Astron. Soc. Geophys. Suppl.* **3**, 343 (1935); D. C. Tozer, *Phil. Trans. Roy. Soc. Ser. A* **258**, 252 (1965).
4. L. Knopoff, *Rev. Geophys.* **2**, 89 (1964).
5. D. L. Turcotte and E. R. Oxburgh, *J. Fluid Mech.* **28**, 29 (1967); *J. Geophys. Res.* **74**, 1458 (1969); E. R. Oxburgh and D. L. Turcotte, *ibid.* **73**, 2643 (1968).
6. K. E. Bullen, *Bull. Seismol. Soc. Amer.* **30**, 235 (1940); H. Jeffreys, *Mon. Notic. Roy. Astron. Soc. Geophys. Suppl.* **4**, 498 (1939).
7. S. Akimoto and H. Fujisawa, *J. Geophys. Res.* **73**, 1467 (1968).
8. S. P. Clark and A. E. Ringwood, *Rev. Geophys.* **2**, 35 (1964); A. E. Ringwood, *Earth Planet. Sci. Lett.* **5**, 401 (1969).
9. F. A. Vening Meinesz, in *Continental Drift*, S. K. Runcorn, Ed. (Academic Press, New York, 1962), p. 145; J. Verhoogen, *Phil. Trans. Roy. Soc. Ser. A* **258**, 276 (1965).
10. F. H. Busse and G. Schubert, in preparation.
11. E. R. Oxburgh and D. L. Turcotte, in preparation.
12. Supported in part by NASA under NGL 05-007-002. One of us (G.S.) would like to thank the Alexander von Humboldt Foundation for a fellowship while visiting the Max-Planck-Institut für Physik and Astrophysik, München. We thank Dr. F. H. Busse for helpful discussions.

23 April 1970; revised 15 July 1970

Sea-Floor Spreading, Carbonate Dissolution Level, and the Nature of Horizon A

Abstract. *Evidence from leg 2 of the Deep Sea Drilling Project suggests a constant spreading rate for the floor of the North Atlantic over the past 80 million years; a major lowering of the carbonate dissolution level during the early Pliocene; and an early to middle Eocene age for horizon A.*

During leg 2 of the Deep Sea Drilling Project (DSDP), paleontological dates were obtained from sediments immediately above acoustical basement. These dates indicate that the North Atlantic Ocean has been opening at a constant rate of 1.1 cm per year during the past 80 million years, probably without any major episodes of quiescence or rapid spreading. In addition, a significant lowering in the calcium carbonate dissolution level is indicated during early Pliocene time. Pelagic sediments deposited near the compensation depth consist of calcareous ooze from early Pliocene time on, but older sediments are barren of calcareous constituents. This may result from an increase in the amount of calcium carbonate in the ocean, caused indirectly by the onset of glaciation on the Antarctic continent. The prominent reflector, horizon A, appears to be present over parts of the eastern North Atlantic, as well as in large areas of the western North Atlantic. It formed during early and middle Eocene time, probably as a result of widespread silica deposition in the North Atlantic.

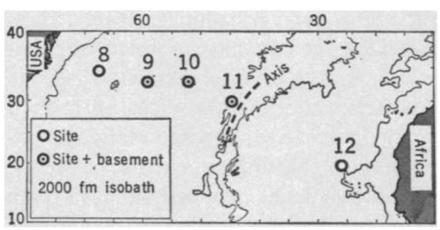


Fig. 1. Approximate location of sites drilled during leg 2 of the Deep Sea Drilling Project. [Courtesy of DSDP]

Leg 2 of the DSDP covered the 2-month period of October and November 1968; it involved crossing the North Atlantic from North America to Africa (*1*). Five sites (Nos. 8 to 12) were drilled by the *Glomar-Challenger* during this leg on a line from New York to Dakar (Fig. 1). Sites 8 through 11 are in the western half of the North Atlantic, roughly evenly spaced between the North American continent and the Mid-Atlantic Ridge; site 12 is in the eastern North Atlantic, northwest of the Cape Verde Islands.