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Continental Drift and the Origin of Mountains

Hot creep and creep fracture are crucial factors in the formation of continents and mountains.

E. Orowan

Geology has been eminently successful in tracing and timing upward and downward movements of the earth's surface on the basis of the sedimentary record; the ultimate causes of the movements, however, remained unexplained. The hypotheses of thermal contraction, convection, expansion, and so on could not be placed on a firm physical basis, and they did not give more than promises of an eventual understanding of the mechanism underlying geology. The situation corresponded to that of astronomy between Kepler and Newton, when the laws of planetary motion were known but their physical background was not.

In the last decades, however, there have been two important developments. First, the mechanical properties of solids have been explored and explained to a considerable extent. Second, the "oceanographic revolution" has revealed unsuspected features of the submerged part of the crust. It has disclosed the existence of ridges and guyots, the relative youth of the ocean floor, and the high oceanic heat flow, particularly along the ridges. It is clear now that, without these two developments, no amount of ingenuity could have solved the basic problems of geology; today, however, the outlines of the answer are emerging rapidly from the lifting fog.

The following account of the present situation is partly a condensed review of the relevant points of the mechanics of solids and of oceanography; partly it is a brief sketch of a recent attempt at a general synthesis of tectonophysics (*1*).

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The Oceanic Ridges

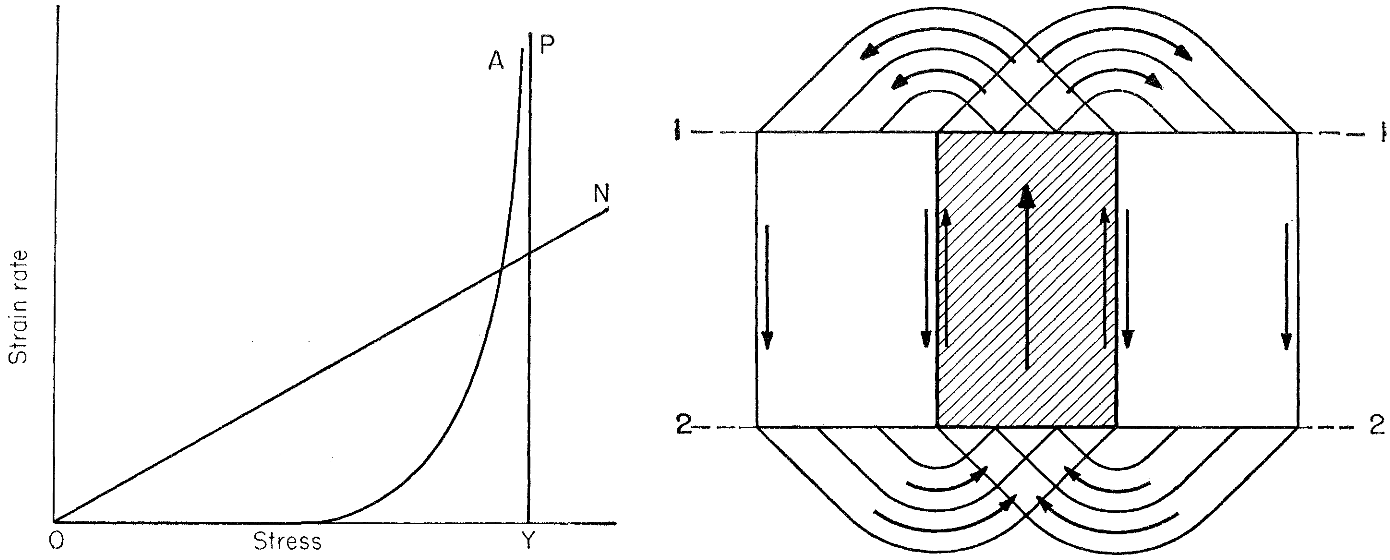
The Mid-Atlantic Ridge, the Mid-Indian-Ocean Ridge, and their seismicity (*2*) have been known for a long time, but only recently has it been recognized (*3*) that the ridges are parts of a global network extending here and there into the continents (as along the African Rift and in Nevada), and that the structure of crust and mantle under the ridges is fundamentally different from that under continental mountain systems (*4*). The latter are kept afloat by sialic roots separated from the mantle by the Mohorovičić discontinuity; the ridges are elevated by thermally light deep roots which, in their upper parts, obtain additional buoyancy from light products of thermochemical reactions, such as serpentine (*5*). A discontinuity of the Mohorovičić type under a ridge does not have the same significance that such a discontinuity has under the continents.

In view of the high heat flow in the ridges, it is plausible to see in them the sites where hot convection currents rise (*5*, *6*).

Condition of Thermal Convection in a Crystalline-Plastic Mantle

Probably all mathematical treatment of convection in the mantle was based, until quite recently, on the assumption of Newtonian viscosity—that is, of a proportionality between the shear stress and the rate of shear strain (Fig. 1, curve *ON*). Most noncrystalline materials (liquids and glasses) are Newtonian; the mantle, however, is believed to be largely crystalline. At low and moderate temperatures, crystalline materials are not viscous but plastic: apart from "transient creep," their deformation is governed by a functional relationship between the shear stress and the shear strain, not the rate of strain; the graphic expression of this relationship is the familiar stress-strain curve of the engineers. At high temperatures and not too high rates of straining, crystalline materials are viscous; their viscosity, however, is far from being Newtonian. It is represented schematically by curve *OA* in Fig. 1: the creep rate is zero, or practically zero, until the stress approaches the "creep limit" or "creep strength." When it increases further, the strain rate shoots up quasi-exponentially.

The typical mechanical behavior of crystalline matter beyond the elastic range was recognized by Andrade (*7*) in 1911–1914; it is appropriate, therefore, to call this high-temperature viscosity "Andradean." That crystalline rocks behave in the same manner was demonstrated by Griggs and others (*8*). In Fig. 1, the line *OYP* represents the "ideally plastic" (non-strain-hardening) material, on which the mathematical treatment of the deformation of plastic bodies is usually based: there is no



Figs. 1 and 2. Fig. 1 (left). Dependence of the shear stress upon the rate of shear strain in Newtonian viscosity (*ON*), in Andradean viscosity (hot creep of crystalline materials) (*OA*), and in ideal plasticity (*OYP*). Fig. 2 (right). Convection in a plastic space, due to the presence of a prism of higher temperature (shaded area).

deformation until the yield point *OY* is reached, and then any deformation, at any velocity, can be produced by the fixed stress *OY*. The Andrade curve *OA*, obviously, can be regarded as a "thermally rounded" ideal-plastic curve.

Because crust and mantle seem to be largely crystalline, the traditional treatment of convection in the mantle on the basis of the assumption of Newtonian viscosity is unrealistic; not surprisingly, it has led to difficulties which have given rise to widespread doubt about the possibility of thermal convection, except possibly in the uppermost part of the mantle. Thus, MacDonald (*9*) has recently pointed out that the present shape of the earth was its rotational equilibrium shape about 10 million years ago, and that this time lag would correspond to a Newtonian coefficient of viscosity of about 10^{26} poises; with this coefficient, however, a convection velocity of the order assumed in the theory of continental drift (1 cm/yr) would require temperature differences of about 650°C, and this would result in local variations of heat flow of more than 10 times the observed values.

Yet the alternative to the assumption of convection is very awkward. Much or most of the heat flow in the continental crust must come from the high radioactivity of granite, which is missing in the oceanic crust; nevertheless, the mean oceanic heat flow is at least equal to the mean flow in the continents (*10*). If this is not the consequence of hot convection currents rising mainly in the oceans, the oceanic mantle must contain the radioactivity missing in the crust. In this case, however, the sial of the continents, and the radioelements that, according to Goldschmidt's rule, accompany it, must have been exuded from a thin layer of the upper mantle. Since the formation of the continents in any way that could lead to a chemical difference between the continental and the oceanic mantle would also deplete a marginal belt of width comparable to the thickness of the depleted layer of the mantle, there would be a noticeable deficiency of heat flow around the continents unless the depleted layer was very thin. In addition, the exclusion of convection would practically exclude the possibility of continental drift: unless the continents moved as rigid units together with the depleted part of the mantle under them, there would be an excess of heat flow along their leading margins and a deficiency along the trailing margins. The effective exclusion of continental drift, however, is a grave difficulty of the no-convection hypothesis. The paleozoic glaciation of the southern continents, the paleomagnetic discoveries of the last decade (*11*), Carey's demonstration of the almost perfect fit of the eastern and western continental slopes of the Atlantic (*12*), and the realization that the main objections to a theory of continental movements arose from inadequate knowledge of the mechanical properties of solids (*1*) have made acceptance of the hypothesis of fixed continents amount to rejection of a simple explanation without offering any alternative.

However, the problem of convection is simply solved if the assumption of Newtonian viscosity with a constant coefficient of viscosity is replaced by the assumption of a largely crystalline mantle which behaves like any other crystalline material: it is plastic at moderate temperatures and it shows Andradean viscosity at high temperatures (in the "hot creep range"). A comparison of the curves *OA* and *OYP* in Fig. 1 shows that, as far as the stress required for deformation is concerned, an Andradean-viscous material can be regarded as approximately ideally plastic (in fact, strain hardening cannot accumulate in the hot creep range). In this case, however, an approximate convective circuit can be constructed which is so simple that one can calculate the force required to drive it without using pencil and paper (*1*). The circuit is shown in Fig. 2, in which the shaded area is the cross section of a prism of height h, of width w, and of infinite length in the direction perpendicular to the plane of the figure. If the temperature of the prism exceeds that of the surroundings by ΔT, the buoyancy force acting on it is $\rho g \alpha \cdot \Delta T \cdot hw$, where ρ is the specific gravity, α is the coefficient of thermal expansion, and g is the mean acceleration of gravity. If the prism is assumed to be rigid, its upward movement produces an indentation in the half-space above it (above the plane

1-1), and a negative indentation (plucking) in the half-space below 2-2. The slip-line solution of this case of indentation was given by Prandtl (*13*) in 1920 [see, for example, Hill (*14*)]; the slip lines and the flow pattern are indicated in Fig. 2, in which, in addition, the further simplifying assumption has been made that the two prisms flanking the hot prism are displaced as rigid bodies by the material extruded by the indentation. Prandtl has shown that the stress required for the indentation alone is $Y(1 + \pi/2)$, where Y is the yield stress in uniaxial tension or compression; the shear resistance at the sides of the prisms, of course, is $Y/2$ per unit area. It may be seen (*1*) that convection is possible if

$$Y([3 + \pi]w + 2h) = wh\rho g\alpha\Delta T \quad (1)$$

where the left-hand side of the equation is the plastic resistance to the rise of the shaded prism per unit of its length and the right-hand side is the buoyancy force. In reality, the three prisms will not move as rigid bodies, but this cannot affect Eq. 1 seriously.

Equation 1 is the convection condition under the circumstances assumed in Fig. 2; convection is possible if the yield stress is low enough to satisfy Eq. 1. If it is lower than the value demanded by Eq. 1, rapid convection sets in, and ΔT is soon reduced to the value of Eq. 1. If the layer above or below the prism is not a half-space but is of finite thickness, or if it is more fluid than the hot prism, the value of the yield stress at which convection is possible is higher than that demanded by Eq. 1.

For a numerical example, let $h = w = 1000$ km, $\Delta T = 100°$C, and $\alpha = 10^{-5}$. The critical value of the yield stress is then

$$Y = 40.3 \text{ bars.}$$

In the Newtonian treatment of the convection problem it was usual to work with the coefficient of viscosity derived from the postglacial rise in Fennoscandia. This choice of viscosity was entirely gratuitous, for the coefficient changes rapidly with temperature and pressure, and no reliable estimate of its value at different depths is possible. Fortunately, the situation is very different in the case of plasticity: the yield stress of the mantle at depths down to about 700 kilometers can be estimated from the energy released in earthquakes (*15*), independently of theoretical assumptions about its pressure and temperature dependence. The estimate is very crude at present, but its order of magnitude is hardly in question, and it can be refined by more extensive seismological observations. For the strongest deep-focus earthquakes, a stress drop of 170 bars was obtained (*15*); the weakest recorded shocks at this depth seem to have a Richter magnitude (approximately equal in this case to the "unified" magnitude) of about 6.5. If the same assumptions are made about the dimensions of the seismic fault area in the weak shocks, the shear stress drop would be about 4 times less—that is, of the order of 40 bars. This would correspond to $Y \approx 80$ bars.

Now, if earthquakes were brittle fractures there would be no relationship between the yield stress (or the Andradean creep limit) and the seismic stress drop. However, with the exception of seismic faulting in a thin veneer of the crust (perhaps down to a depth of 10 km), earthquakes are creep fractures (*1, 15*), and they take place under the shear stress responsible for creep itself. Moreover, the stress drop cannot differ much from the initial stress, because local melting should occur after a small amount of sliding (*16; 17*, p. 340), and the stress should then fall to a low value. It follows, therefore, that the yield stress (creep limit) of the mantle at the depth of about 700 kilometers, as estimated from seismic energy release, is of the same order of magnitude as the yield stress at which plastic convection can take place under plausible conditions of dimension and temperature.

Curve *OA* in Fig. 1 shows that a slight change of the stress can cause tremendous changes of the flow rate in Andradean creep. Consequently, the velocity of plastic-Andradean convection is self-adjusting: the velocity assumes the value at which ΔT is in equilibrium with the heat transport at top and bottom of the convection cell. This may have something to do with the fact that apparently different mechanisms produce almost the same mean heat flow in continental and oceanic areas.

Convection and the Mid-Ocean Ridges

If the ridges are attributed to ascending hot currents, the question must be asked how such currents can rise in the oceans, where there is only a very thin heat-insulating sedimentary blanket and where there is no highly radioactive granitic layer, when apparently the cold currents descend mainly at the continental margins. The possibility of convection due to thermal (adiabatic) instability, of course, depends on the temperature gradient; however, once the instability is present, the hot current rises where the mean temperature of a vertical column is raised either by more intense heating at the bottom or by heating and heat insulation at the top. Accordingly, it was always thought that hot convection currents would rise under the continents and that cold currents would descend under the oceans. How can a reversed situation be explained?

According to G. F. S. Hills (*18*) and Jeffreys (*17*, p. 331; *19*), Wegener's primeval continent was swept together by convection; later, the heat-insulation and radioactive heating of the protocontinent would reverse the current and lead to the disruption of the continent. What happened to the disruptive convection after this? If the disruption started, say, 500 million years ago, and the mean convection velocity was of the order of 1 or 2 centimeters per year, the current would have moved about 5,000 or 10,000 kilometers since that time. This is hardly more than a large fraction of a full rotation of the convection cell; since the convection pattern is not likely to be reversed after a fraction of a revolution, the possibility must be considered that the present mid-oceanic ridges may be maintained by hot columns which originated under the primeval continent before its disruption and which still persist because of thermal inertia, in spite of the changed conditions at the surface (*1*).

Plastic Convection Patterns

The radical difference between flow patterns in Newtonian viscous materials and Andradean-plastic materials is exemplified by the simple case of flow through a tube. In the laminar flow of a Newtonian liquid the velocity changes parabolically along the diameter; an ideally plastic material, on the other hand, would move like a plug, with the same velocity everywhere except in the boundary layer, where the shear strain is concentrated (some toothpastes show this effect fairly clearly).

Correspondingly, convection in a Newtonian mantle would give rise to a spread-out continuous distribution of flow [see, for example, Vening Meinesz (*20*)], while in a plastic-Andradean mantle the deformation tends to concentrate in relatively thin layers. The rising hot column is likely to take the form of a dike, which may be narrow relative to its depth. This would explain the typical profile of the oceanic ridges, their sharp curvatures (which could hardly occur in a Newtonian material), and perhaps also the origin of the Murray-Menard faults which intersect the ridges, particularly at points of sharp curvature (*1*). Of particular interest is the question of why so many of the ridges are so accurately centered in the middle of the ocean. At first sight it may appear that this is a consequence of the rising current's pushing apart the adjacent continents symmetrically; however, this cannot be the correct explanation. Africa is largely surrounded by mid-oceanic ridges from the west, south, and east; the Mid-Atlantic Ridge would have to push it eastward, and the Mid-Indian-Ocean Ridge, westward. If, however, the ridges owe their existence to rising hot dikes, they may have a self-centering mechanism. If they were closer to one adjacent continent than to the other, the isostatically uncompensated load they dump upon that side would be higher per unit of area than the load they place on the other side of the ocean floor; the rate of subsidence of the ocean floor would be greater where a continent is closer, and the resulting horizontal spread of the mantle underneath would tend to push the hot dike away from the closer continent (*1*).

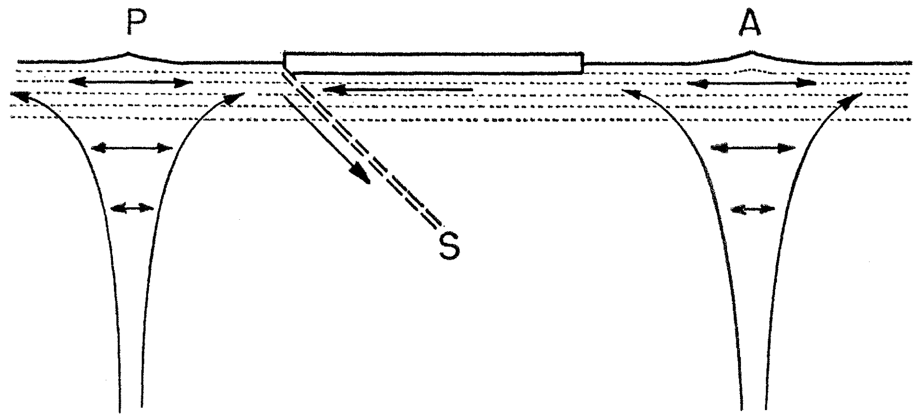


Fig. 3. Scheme of convection with rising hot dikes under the Mid-Atlantic (*A*) and East Pacific (*P*) Ridges and a sink under the circumpacific seismic shear surface (*S*). The dotted horizontal lines are shear planes of continental drift in the soft layer (the "asthenosphere").

Convection and Continental Drift

Continental drift has often been attributed to thermal convection; but convection alone need not produce any crustal displacement. If hot currents rise under the Mid-Atlantic and the Eastern Pacific Ridges, the cold current may sink down under America, which may then be held in fixed position over the sink.

However, the remarkable discovery of the circumpacific seismic shear surface, by Gutenberg and Richter (*21*) and Benioff (*22*), shows clearly that the pattern of convection must be very different from this. Most circumpacific earthquake foci lie on a surface that dips from the margin of the ocean down under the continents. The strike-slip component of seismic faulting seems to predominate; in the long run, however, the process is probably essentially a dip-slip faulting. This is indicated by the slope of the surface, which is at about 30 to 60 degrees to the horizontal (*22*); pure strike-slip faulting would take place on a vertical plane. Since strike-slip faulting requires only a fraction of the stress needed for compressive dip-slip faulting (*1*), it may dominate in the short run; the true character of the deformation, however, seems to be a compressive dip-slip faulting with a strike-slip component superposed. That the dip-slip component is probably compressive follows from two facts. First, the dip of the focal surface at shallow depths is often much less than 45 degrees; since in this case the shear failure should be of the Coulomb-Rankine frictional type, an angle of less than 45 degrees indicates compressive faulting. Second, in the Coulomb-Rankine case tensile faulting would be easier than strike-slip faulting (*1*); if the latter is nevertheless abundant, the dip-slip component must be compressive. Since a marginal seismic focal surface is present around the Pacific but not around the Atlantic [apart from the localized sink at the Gulf of Mexico (*23*) and the Caribbean], the pattern of convection seems to be of the type sketched in Fig. 3. If hot dikes rise under the Atlantic Ridge and the Eastern Pacific Ridge and the current fans out horizontally in the upper mantle, the only major sink would be, as indicated by observation, under the seismic shear surface of the Pacific coast of the Americas. If there is an upwelling in a fixed position under the Mid-Atlantic Ridge and no sink (apart from the Gulf-Caribbean one) between the Ridge and the Pacific, the western half of the Atlantic crust and the American continents must move westward. Continental drift, then, is the consequence of an active shear layer present along the Pacific margin but not along the Atlantic margin; and the crucial question is, What is the cause of this remarkable difference?

Shear strain concentrates in a narrow layer instead of being uniformly distributed, if an instability of the deformation is present. The necking of a tensile specimen is due to a geometrical instability (the decrease of the cross-sectional area at a point of lower resistance to the extension); the Lüders bands in steel are caused by a physical instability (the drop of stress from the upper to the lower yield point). Apart from the upper crust, the deformation of the earth should be of the hot-creep type, which is inherently unstable: fast creep reduces the stress required to maintain creep, and so the strain tends to concentrate in layers where, for some reason, the rate of creep was higher in the beginning. Figure 4 illustrates this process, showing torque-twist curves recorded in torsion tests with fine-grained polycrystalline alumina at 1200°F (*24*). Curve *A* represents an experiment in which the specimen broke with a brittle type of fracture after a small deformation, which nevertheless must have produced a crack or cavity responsible for the brittle-crack propagation. Curve *B*, obtained at the same temperature and the same rate of strain, reaches a maximum and the torque begins to drop before fracture; several parallel layers of visible porosity or incipient cracking developed before fracture occurred along one layer. Curve C represents an experiment made at the same temperature but with a rate of strain 25 times lower; after reaching a maximum, the torque decreased nearly to zero (this portion of the curve is not shown in Fig. 4) before fracture was complete.

Figure 5 shows schematically how the formation of shear bands follows from the character of the curve for stress plotted against strain rate. Relative to Fig. 1, the coordinate axes are interchanged; the Andrade curve is drawn with a solid line. In reality, the stress does not rise steadily with increasing strain rate: it reaches a maximum and then drops as the cohesion is gradually destroyed. This is indicated by the dashed part of the curve. Figure 5 is a crude simplification; in general, the stress begins to depend on the strain as well as on the strain rate when the maximum is approached.

There is always a weakest cross section in a specimen, in which the maximum of Fig. 5 is first reached. After this, the force drops in this section, as shown by arrow 1; in the other sections, which have not reached the maximum, the force must drop by the same amount under decreasing strain rate (arrow 2). If a given rate of compression is imposed on the specimen, the strain rate is no longer uniform: it has the value FH in a narrow shear band and FG in the rest of the specimen. FH corresponds to the shear rate in the circumpacific shear layer; FG, to that on the Atlantic side of America. Thus, the rate of underthrusting of the ocean floor is much higher on the Pacific side. Moreover, the underthrusting is seismic on that side, but aseismic on the Atlantic side, because the curve drops at H but rises at G. Where it drops, any local increase of the creep rate produces a drop of stress, culminating in local melting and a seismic shock (in the top 10 or 15 km of the crust, of course, there is no hot creep and the seismic mechanism is different; but even shallow-focus earthquakes must be caused by strain-rate concentration at depth).

In the case of polycrystalline aluminum oxide, the drop of the torque and the subsequent creep fracture are essentially consequences of the fact that the grains slide on their neighbors as rigid blocks along the viscous grain boundaries; in the course of this sliding they are lifted out of their geometrically interlocking positions, pores

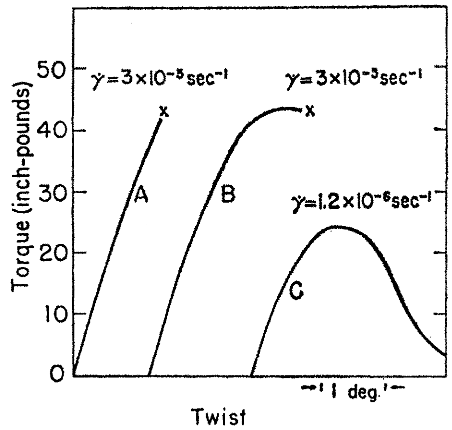


Fig. 4. Torque-twist curves for polycrystalline specimens of Al_2O_3 at 1200°F, showing brittle fracture (A) and creep fracture (B and C).

are formed, and the resistance to creep weakens.

It seems that earthquakes, except in the top veneer of the crust, are creep fractures (*15*). The instability of creep appears to be the fundamental cause of the concentration of deformation in the circumpacific shear planes; the discontinuity of the crustal structure at the continental slope is probably only a releasing factor.

In the scheme shown in Fig. 3, the lithosphere between the Mid-Atlantic Ridge and the Eastern Pacific Ridge plays the role of a compression specimen. If a shear band occurs at the Pacific coast of the Americas, the stress drops both here and everywhere else between the two ridges and may not be sufficient to form another shear band on the Atlantic coast or elsewhere. The consequence is that the lithosphere between the Mid-Atlantic Ridge and the Gutenberg-Richter-Benioff shear surface at the west coast of America moves westward as a unit, and continental drift occurs as a con-

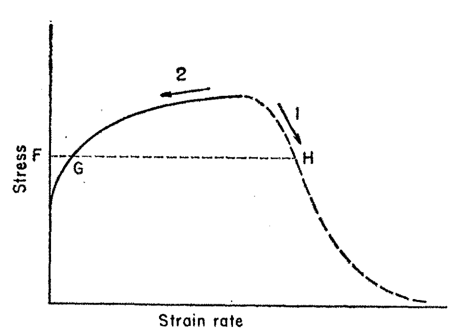


Fig. 5. Schematic curve for stress versus strain rate, illustrating the instability of hot creep. When the creep rate in the shear layer has risen to FH, it has fallen to FG elsewhere.

sequence of the physical instability of hot creep manifested in earthquakes and in the formation of the seismic shear surface.

The circumstance that the lithosphere between the two ridges seems to behave as a compression specimen would behave indicates that the continents are not firmly anchored to the mantle. The existence of a seismic shear plane on one side of continents but not on the opposite side, therefore, is additional evidence of their mobility.

If the convective circuit is of the type shown in Fig. 3, the velocity of continental drift can be calculated, in principle, by two independent methods. The height of the oceanic ridge determines the excess pressure underneath, which drives the lithosphere away from the ridge; if the effective viscosity in the soft, "asthenospheric" layer of the mantle under the lithosphere can be estimated, the order of magnitude of the drift velocity is obtained. On the other hand, the elastic energy released in earthquakes determines the fault displacement if the fault area is known; with an assumption about the ratio of dip-slip component to strike-slip component and an estimate of the annual release of seismic energy, the order of magnitude of the drift velocity of the Americas relative to the Pacific floor can be obtained. With plausible assumptions about the quantities involved, both methods give velocities of the order of 1 centimeter per year. The principles involved in the calculations are sketched briefly in the next sections.

Velocity of West-Drift Estimated from Height of Mid-Atlantic Ridge

Underneath the crest of the oceanic ridge, at the level of the abyssal plain, the pressure is higher than on the abyssal plain outside the ridge; it is higher by the weight of the ridge, minus that of the displaced sea-water. Since the density is lower under the ridge than elsewhere, the difference between the pressure under the ridge and that at the same level outside the ridge decreases with increasing depth and vanishes at the depth of isostatic compensation. It is this pressure difference that drives the upper layers of the mantle and the lithosphere away from the ridge; it is indicated by arrows in Fig. 3. It can be calculated if an assumption is made about the depth of compensation and the variation of the pressure difference with the depth.

It is widely assumed, for several reasons, that there is a soft layer (the "asthenosphere") in the upper mantle under the lithosphere; its possible origin is discussed below. In this case, the horizontal pressure exerted by the weight of the ridge causes the lithosphere to slide away from the ridge upon the soft layer. The process is somewhat similar to shallow convection of the types considered by Tuzo Wilson (*25*) and Elsasser (*26*). Although the soft layer must be highly non-Newtonian, it is of some interest to calculate the velocity of westward drift from the known wedging pressure of the oceanic ridge, with the assumption that the soft layer has an apparent viscosity of 10^{22} poises, the value calculated from the rate of the Fennoscandian postglacial uplift. The velocity obtained is of the order of 0.3 centimeter per year if the level of isostatic compensation is assumed to be 200 kilometers; if it is assumed to be 500 kilometers, the drift velocity is of the order of 2 centimeters per year (*1*).

In this estimate the resistance of the circumpacific shear plane to the displacement is disregarded; because of the instability of hot creep, this resistance must be much lower than the resistance of the lithosphere to compression elsewhere. In particular, the mean excess pressure under the Mid-Atlantic Ridge is of the order of 350 bars (*1*); on the other hand, the resistance to compression of the circumpacific shear plane, as estimated from seismic energy release (*1, 15*), is probably of the order of 40 to 200 bars. The difference between the horizontal pressure of the ridge and the resistance of the seismic shear plane, therefore, should be of the same order of magnitude as the horizontal pressure itself, and so the circumpacific resistance should not alter the order of magnitude of the estimate for the drift velocity.

Velocity of Continental Drift Estimated from Earthquakes

If the length of the seismic fault area (parallel to the movement) is much less than its width, the depth of the stress-relieved volume (perpendicular to the fault area) is of the order of magnitude of the length, and the fault displacement is of the order of the shear-strain decrement multiplied by the depth of the strain-relieved volume (*1, 27*). If, therefore, the dimensions of the fault area can be estimated or at least plausibly assumed, the seismic energy release gives the order of magnitude of the fault displacement. If annually a known fraction of the Gutenberg-Benioff circumpacific fault surface suffers seismic faulting, and if an assumption is made about the fraction of the energy release due to the compressive component of faulting, the mean annual normal movement of the Pacific against the Americas can be estimated. On the basis of plausible assumptions about the quantities involved, the velocity of movement of the Americas toward the Pacific (or vice versa) is estimated to be of the order of magnitude of 1 centimeter per year (*1*). Of greater interest than the agreement between this crude estimate and the usual assumptions about the velocity of continental drift is the fact that the uncertainties of the estimate can be reduced, and a reliable result may be obtainable by more extensive and refined seismological observations.

The Soft Layer under the Crust

In making a tentative estimate of the velocity of westward sliding of the western Atlantic and the Americas, an effective coefficient of viscosity of 10^{22} was assumed, and the yield stress was neglected. Does this not make the estimate entirely illusory? As mentioned above, MacDonald (*9*) has shown that the observed shape of the earth would correspond to a Newtonian viscosity of about 10^{26} poises.

The contrast between the apparent rapidity of isostatic adjustments and the permanence of other geological features has lead to the idea that there is a soft layer in the upper mantle (the "asthenosphere"). This idea is supported by seismological observations of low-velocity layers in the upper mantle (*28*), in view of the remarkable parallelism between the elastic moduli and the plastic yield stress of crystalline materials. In the past, the existence of such a soft layer was usually attributed to the dependence of the coefficient of viscosity upon temperature and pressure. With increasing depth, the rise of temperature would first reduce the coefficient; however, the increase of temperature levels out, probably at a depth between 400 and 800 kilometers, and then increasing pressure would raise the viscosity.

That this argument is not satisfying is seen from the fact that it demands a more or less uniform soft shell everywhere. Yet the remarkable absence of isostatic adjustments in spite of gravity anomalies at some places (for example, the "Hidden Range" in India) where orogenic processes do not seem to provide sufficient explanation indicates that the soft layer is probably very patchy. (In active orogenic belts the softening progresses to local melting; it is plausible to regard magma pockets as the ultimate result of the process responsible for the "asthenosphere.") Besides, crystals in general do not become particularly soft at the melting point: one can skate on melting ice.

A more plausible explanation of the soft layer is indicated by Fig. 6 (*24*). *A* and *B* are torsional creep-relaxation curves, showing the decay of the torque with time after the torque is rapidly applied, and then the drive of the testing machine is stopped. Both curves were obtained at 1200°C with polycrystalline specimens of Al_2O_3; the two specimens were of the same shape and had the same average grain size (42 microns). Yet at the point of intersection (where values of stress, strain, and temperature are the same for the two curves), the creep rate for specimen B was about 45 times higher than that for A. The factor responsible for this difference was apparently the impurity content: that of B was 0.7 percent; that of A, only 0.2 percent. It is a familiar fact in metallurgy that the most common cause of high creep rate at high temperatures is the presence of a relatively low-melting grain-boundary envelope due to impurities migrating into the boundaries; the effect has been beautifully demonstrated on glacier ice recently by Kuroiwa (*29*). In the earth's mantle, water, silica, and other low-melting and volatile constituents can form such envelopes. At great depth the high pressure and temperature keeps them dissolved in solid solution; in the upper mantle, however, they segregate to form glassy boundary envelopes, and this process can produce a wide range of mechanical behavior, from plasticity to Newtonian viscosity (the latter in magma pockets). Soft layers, therefore, are likely to be caused by low-melting or volatile constituents—above all, by water. It would be interesting to know whether the apparent absence of an effective soft layer under a part of India is the consequence of the loss of plasticizers and fluxes with the outpouring of the Deccan Trap, and whether the extreme nonequilibrium shape of the moon is maintained by lack of the most important silicate plasticizer, water. In other words, the moon may be hard because it is dry.

Rigidity of the Continents

A familiar objection to the Wegener theory of continental drift is that the compressive strengths of granite and basalt in conventional laboratory tests are about equal; how then can continents float as rigid rafts in the mantle covered by the oceanic crust?

The deformation resulting in drift would be composed of the extension along the ridges and the compensating compression along the Gutenberg seismic shear layers, where the instability of hot creep produces "soft" deformation bands. The resistance to these deformations is not directly related to the compressive strength of mantle rocks as measured in the laboratory, with or without confining pressure. An additional consideration is that the oceanic crust seems to consist mainly of serpentinized peridotite (*5, 30*); during the reaction

$$\text{Olivine} + H_2O \rightleftarrows \text{serpentine}$$

the increased molecular mobility should cause increased plasticity, as is well known from the physics of solids.

Origin of Mountains

The remarkably characteristic presence of serpentines in most mountain belts of the Alpine type has been emphasized by Steinmann (*31*) and particularly by Hess (*32*). The possibilities sketched above raise an interesting point: serpentine may play a central role in geotectonics and, in particular, in mountain building, similar to that of water in surface geology.

If the floor of the ocean slides under the margin of a continent (Fig. 3) downward along a Gutenberg-Benioff shear plane, the contained serpentine is decomposed around the 500°C isotherm, in the lower half of the continental crust. The crust above the shear plane, therefore, is soaked by the "rain" of water rising from depths around the 500°C isotherm, and by the light and radioactive materials it carries.

This may give rise to orogenesis, with its characteristic phenomena. The supply of water from the seismic shear belts would account for the remarkable plasticization of the rocks during active orogenesis (as manifested in the vast nappes of the Alps and elsewhere), for the magmatic and metamorphic phenomena, and for the preferential localization of mountain belts along continental margins and major rifts. It could also explain geosynclinal subsidence and orogenic uplift, perhaps a little more than qualitatively, in the following manner.

Geosynclinal Subsidence

Why do geosynclinal troughs sink, during a period of the order of a hundred million years, at an average rate estimated usually at some 0.1 to 0.2 millimeter per year?

The average volume of lava annually discharged to the surface of the earth is of the order of 1 cubic kilometer. It is difficult to estimate the magnitude of the active geosynclinal areas; however, a strip 500 kilometers wide and 20,000 kilometers long should represent the correct order of magnitude. If 1 cubic kilometer of lava is withdrawn uniformly each year from under a geosynclinal area of 10^7 square kilometers, sinking of the surface of the geosyncline occurs at the rate of

$$10^{-7}\ \text{km/yr} = 0.1\ \text{mm/yr}$$

It seems, therefore, that the outpouring of lava, insignificant as it seems at first glance, is of the right order of magnitude to account for the rate of geosynclinal subsidence. The formation of lava, as indicated above, seems to be due to hydration by the decomposition of serpentine at the Gutenberg seismic shear surface. This, then, may be the cause of geosynclinal subsidence.

Orogenic Uplift

Of two opposing schools, one attributes the uplift of mountains to the vertical rise of matter from the mantle; the other, to the buckling of the lithosphere by horizontal compression. The vertical-rise hypothesis poses the problem that, according to the evidence of the oceanic ridges, convective rise of hot masses seems to occur along the ridges and so can hardly take place in the geosynclinal belts. The horizontal-compression hypothesis, on the other hand, does not explain the circumstance that, as Smoluchowski has shown (*33*), gravity prevents the elastic buckling of a sialic crust floating on a substrate of density 3.3 if the thickness of the crust exceeds some 20 meters. Plastic buckling (*34*) is easier to explain, but the compression of the crust would have to be very rapid if the simultaneous removal of buckling by isostatic adjustment were to be overbalanced.

These difficulties disappear, and the opposing hypotheses of vertical rise and of horizontal compression turn out to represent two sides of the same process, if the assumption is made that the continued supply of water and other light and volatile materials not only plasticizes the crust under the geosyncline but also reduces its density and ultimately creates a slight density inversion between the upper crust and the layer of deserpentinization. Bold as this assumption seems, Daly made it one of the main themes of his book *The Architecture of the Earth* (*35*). He arrived at the idea from circumstantial geological evidence; lack of support from geophysics relegated it to temporary oblivion.

If continued soaking by deserpentinization leads to increasing plasticization of the crust and finally to a density inversion, the cause of vertical rise and of magmatic phenomena is no longer

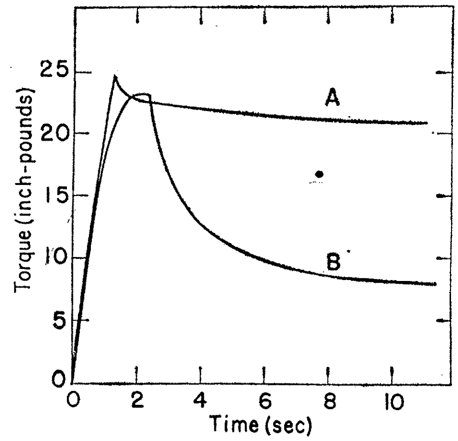


Fig. 6. Torsional creep relaxation in polycrystalline Al_2O_3 with impurity content of (*A*) 0.2 percent and (*B*) 0.7 percent.

a problem. At the same time, the Daly inversion would remove the difficulty which stood in the way of hypotheses in which buckling by horizontal compression was assumed. If magma can break through the crust, downbuckling is essentially the process considered above as the possible cause of geosynclinal subsidence: the sinking of a heavier layer into a lighter substrate in the presence of horizontal compression. The shorter wavelength of preorogenic folding may result from the greater softness and plasticity of the crust at the end of the geosynclinal period.

According to geological evidence, orogenic folding seems to precede the uplift; how subsidence, folding, and uplift can be produced by the same basic mechanism was one of the tantalizing puzzles of geology. The model of Fig. 3, however, seems to lead to this sequence. If a marginal continental shear layer develops, with a level of deserpentinization at which the water and other substances arriving from the oceanic ridge are discharged, these substances accumulate until the resulting magma breaks through the crust and geosynclinal subsidence starts, followed by folding due to compression by the flow from oceanic ridges. In the course of these processes the crust and the top of the mantle are deformed by hot creep which involves sliding of the crystal grains on their neighbors. This opens up intercrystalline pores, and more and more of the light liquid phase is soaked up by the deformed hot rock. As its density decreases, mountain roots grow and isostatic uplift takes place. During the subsequent volcanic and plutonic activity the volatiles leave the mountain roots, and a locally thickened brittle crust is left behind. If the convective supply of water, silica, and so on continues for a time, accumulation without breakthrough starts again; another uplift takes place, accompanied this time by crustal fractures (block faulting) instead of plastic deformation. A similar sequence can be observed on boiling jam: a rising bubble of steam produces uplift of the surface; a jet of steam blows out when the bubble bursts; the resulting subsidence is isostatically adjusted, and the play repeats itself.

Conclusion

There is no need to remark that the foregoing synthesis, connecting relatively few hard facts and many soft ones by bridges unavoidably built with much hypothetical mortar, will be subject to alteration and repair. However, the very possibility of a rather comprehensive synthesis indicates that the progress of the physics of solids and of oceanography may have inaugurated the end of the attractive age of "geopoetry" (5).

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