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Convection in a Non-Newtonian Mantle, Continental Drift, and Mountain Building

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XXIV. Convection in a non-Newtonian mantle, continental drift, and mountain building

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[Plate 4]

A condition for thermal convection has been derived for a mantle of non-Newtonian viscosity, with the properties of a crystalline material in the hot creep range. The critical magnitude of the yield stress (creep limit) at which convection can take place with plausible values of the dimensions of the hot column (*ca.* 1000 km) and of the temperature difference (*ca.* 100 degC), practically coincides with the stress drop estimated from the energy release in deep focus earthquakes. Thus, convection in the mantle appears probable. It is likely to take for the form of the rise of relatively narrow hot dikes. If the oceanic ridges are ascribed to convective dikes, their characteristic profile, sharp curvatures, and the direction of the intersecting Murray–Menard wrench faults can be understood. Non-Newtonian dike convection involves a mechanism which drives the ascending current towards the central position between two continents. Bernal's view that the oceanic heat flow may not be caused by an excess of radioactivity but by the rise of convection currents in the oceanic mantle can be based on the assumption that the present hot convective dikes originated under primeval continents before their disruption.

The cohesive, gravitational, and frictional components of the strength of the continental and oceanic lithosphere are quantitatively compared, and reasons are given for the relative incompressibility of continents and for the relative frequency of strike-slip earthquakes. The coefficient of quasi-viscosity of sedimentary rocks, as manifested in the rise of salt domes, is estimated; it has the Fennoscandian magnitude.

The soft layer (asthenosphere) in the upper mantle is attributed to low melting and volatile constituents forming glassy grain boundary envelopes. Absence of isostatic adjustments, as in India or in the Moon, may be due to loss of water (in the case of India, perhaps by the outpouring of the Deccan trap).

The velocity of the westward drift of the Americas is estimated both from the gravity spreading pressure under the Mid-Atlantic Ridge, and from the energy release in circumpacific earthquakes. Both estimates give the order of magnitude of 1 cm/y.

The plasticization and fluxing of the crust on which the orogenic process is based may be the consequence of the precipitation of water from serpentine flowing continentward from the oceanic ridge where the horizontal flow is directed downward at the marginal continental shear plane. The accumulation of water and other low melting and volatile components under the shear plane (above the 500 °C isotherm) reduces the density until the density inversion assumed by Daly occurs. The density inversion reverses the gravitational stabilization of the crust against horizontal buckling and permits orogenic folding by the compressive stress exerted by the rising hot dikes. Geosynclinal subsidence may be a consequence of the outpouring of lava from under geosynclinal areas. With recent order-of-magnitude estimates of the annual outpouring of lava and of the magnitude of the active geosynclinal areas, a rate of subsidence of about 0·1 mm/y is obtained.

1. Non-Newtonian viscosity and thermal convection

Although Griggs (1939), Vening Meinesz (1948), and Jeffreys (1952) have discussed the effect of a finite yield stress ('strength') on thermal convection, all quantitative work on the subject has so far been based on the assumption of Newtonian viscosity. Yet the mantle is usually believed to be substantially crystalline, and crystalline materials show no viscosity but only plasticity with transient creep in the low temperature range, and a characteristic non-Newtonian viscosity at higher temperatures; with many materials, the gradual transition between the two temperature ranges takes place around one-half of the absolute melting temperature. The typical non-Newtonian viscosity of crystalline matter was recognized, and separated from transient creep, by Andrade in 1911–14; curve A in figure 1 shows schematically the dependence of the strain rate upon the shear stress in Andradean viscosity, while the straight line N represents Newtonian viscosity. 'Ideal' plasticity corresponds to the line OYP: there is no deformation until the yield stress OY is reached, and then

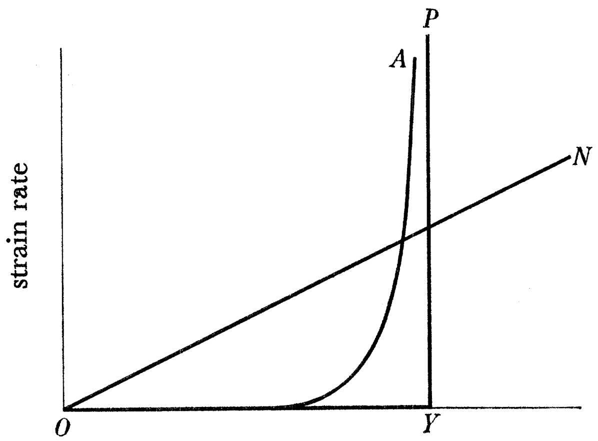

FIGURE 1. Curves of strain rate against stress for Newtonian viscosity (N), Andradean viscosity (A), and ideal plasticity (OYP).

any strain, at any strain rate, can be produced by the fixed stress OY. Andradean viscosity is closer to ideal plasticity than to Newtonian viscosity: the creep curve A can be regarded as a thermally rounded ideal-plastic line. If the stress sinks below the creep limit, the strain rate vanishes rapidly; if it rises above the limit, the flow rate increases so fast that it is often experimentally impossible to exceed the creep limit by a factor greater than 2 or 3.

It has been emphasized by Griggs and Jeffreys that, if the mantle has a finite yield stress, most conclusions about convection based on the assumption of Newtonian viscosity are quantitatively and qualitatively unfounded. In fact, the matter is worse. If the mantle had a low yield stress OY (figure 2) and a linear relation between the strain rate and the stress beyond the yield point according to curve OYB, the Newtonian treatment of the convection problem would not be unrealistic as an approximation for higher stress levels, provided that convection had already started and finite local temperature differences had been established. A material that behaves according to curve OYB is called a 'Bingham body'; crystalline materials at higher temperatures, however, are not of the Bingham but of the Andrade type. The quasi-exponential rise of the Andrade curve is an additional reason why the convection problem in the mantle cannot be treated on the basis of Newtonian viscosity.

That crystalline materials behave in laboratory experiments according to curve A in figure 1 does not prove, of course, that they cannot have a very high Newtonian viscosity below the creep limit, with a coefficient of viscosity of the geological order of magnitude.

If a rod of 10 cm length, of a Newtonian coefficient of viscosity of 10^{22} P, is subjected to a tensile stress of 100 b, it extends by about 100 Å/y; this is the change of length produced by a temperature fluctuation of about 0·01 degC. The presence of a Newtonian viscosity of the Fenno–Scandian magnitude, therefore, can hardly be proved or disproved experimentally at present. There is, however, one experimental observation giving a hint in the negative direction. At relatively high stresses the strain rate can often be expressed approximately as an exponential or a hyperbolic-sine function of the stress; the latter would involve a high Newtonian viscosity at low stresses. At stresses on the low side of the creep limit, however, most materials seem to behave approximately according to a fourth-power dependence of the strain rate upon the stress (Norton 1929; Weertman 1962).

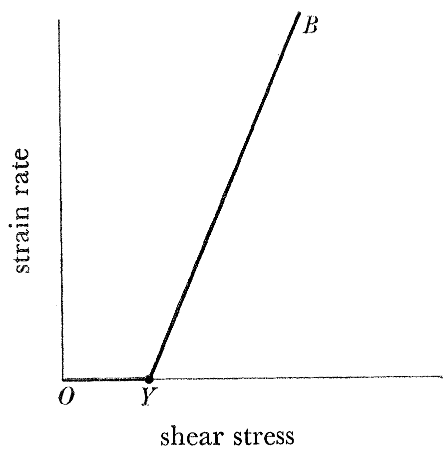

FIGURE 2. Curve of strain rate against stress for Bingham viscosity.

The question can be approached theoretically also. Viscous flow is the consequence of rearrangements within small groups of molecules, a process requiring some amorphousness of structure, i.e. a deviation from perfect crystalline order (Becker 1925). If the perturbation of the activation energy of the rearrangements by the applied stress is small compared with kT, Newtonian viscosity results; if it is not, an Andrade curve of the hyperbolic sine type is obtained, which implies Newtonian viscosity at very low stresses. However, the Andradean viscosity of crystalline matter at high temperatures, as a rule, is not of this origin (see appendix).

Since there is neither an experimental nor a theoretical basis for the assumption of Newtonian viscosity in the mantle, the problem of convection has to be approached from the opposite direction, by assuming the mantle to be plastic and of Andradean viscosity. In what follows an approximate condition of thermal convection in an ideally plastic or Andradean viscous body will be derived. Naturally, convection cannot start if the mantle is spherically symmetrical and has a finite yield stress; however, if temperature differences of sufficient magnitude and extent are present, hot masses can rise if the yield stress is low enough. As a typical order-of-magnitude example, the case of a hot column of 1000 km height, 1000 km width, and infinite horizontal length may be considered, with its temperature 100 degC above that of the surroundings. It can be seen below that such a column (i.e. a hot dike), rises if the yield stress (creep limit) in uniaxial tension does not exceed a value in the range between about 40 and 330 b, depending on the boundary conditions.

The order of magnitude of the stress drop estimated from the energy release of deep focus earthquakes (Orowan 1960) is closer to the lower than to the upper limit of this range; thermal convection in the deep mantle, therefore, appears probable.

If the mantle is crystalline-plastic, with Andradean viscosity, several difficulties of the Newtonian convection theories disappear.

If a soft (asthenospheric) layer of relatively low viscosity is assumed to exist near the top of the mantle, and if the mantle below it were Newtonian with a much higher viscosity, the lithosphere would be largely decoupled from any convection in the mantle. But if the mantle is plastic, according to curve A in figure 1 the shear stress does not depend much on the velocity of convection (provided that convection does take place). If, moreover, the asthenosphere had a more Newtonian viscosity (e.g. because of thick glassy envelopes around its crystal grains; see § 9), it would behave more according to curve N (figure 1), and the two curves could intersect. This means that both the stress and the strain rate could be continuous at the boundary of the mantle and the asthenosphere, even if the apparent viscosity of the mantle were far higher at not much lower stresses. Asthenosphere and mantle could then be mechanically strongly coupled, just as the plastic grains and the viscous grain boundaries in a polycrystalline metal or non-metal are closely coupled in the 'equicohesive temperature' range.

Another difficulty of the Newtonian convection theories has been pointed out recently by MacDonald (1963). There seems to be a lag of about 10^7 y between the rotational equilibrium shape of the Earth and its actual shape; this would demand a Newtonian coefficient of viscosity of about 10^{26} P. With this value a convection velocity of 1 cm/y would require temperature differences of about 650 degC, and the corresponding local variations of heat flow would be some 10 times greater than the observed values. This difficulty disappears if the assumption of Newtonian viscosity is dropped. In the numerical example just mentioned, a temperature difference of 100 or even 50 degC can maintain plastic convection at almost any velocity (which is determined mainly by the heat transfer at top and bottom of the colum); for a velocity of 1 cm/y the heat flow could well be of the observed order of magnitude. Expressed in other words: the condition for deep convection in a plastic-Andradean mantle, combined with plausible assumptions about temperature differences, with estimates of stress from deep focus earthquakes, and with the usual conclusions about continental drift velocities, gives the observed order of magnitude of local heat flow variations.

2. Condition of thermal convection in a plastic material

Let *abcd* in figure 3 be the rectangular cross-section of a prismatic volume in a plastic space; the height of the prism is $ad = h$, its width $ab = w$, and it is infinitely long in the direction perpendicular to the plane of the figure. If the temperature of the prism exceeds that of its surroundings by ΔT, it experiences a buoyancy force

$$F = wh\rho g\alpha\Delta T, \tag{1}$$

where α is the coefficient of thermal expansion, ρ the density, and g the acceleration of gravity.

Let Y be the yield stress in uniaxial tension or compression of the (ideally plastic, i.e. non-hardening) material. If Y is sufficiently low, the prism rises under the influence of the force F. The slip line solution of the problem is not known but, according to the theorem of plastic limit, an upper limit to the plastic resistance opposing the force F is easily obtained by considering the prism as rigid; moreover, this upper limit is likely to be close to the strict solution of the problem.

If the prism is rigid, it acts as a strip-shaped indenter on the plastic half-space above 1–1. The deformation produced by such an indenter is known (Prandtl 1920; Hill 1950); one possible set of slip lines is indicated in figure 3 and another in figure 4. In figure 3 the dashed slip lines lie in an area in which the yield condition is satisfied but the plastic deformation does not exceed the order of magnitude of the elastic deformations; in addition, it is assumed that no shear stress is transmitted in the plane 1–1. According to Hill, this slip line

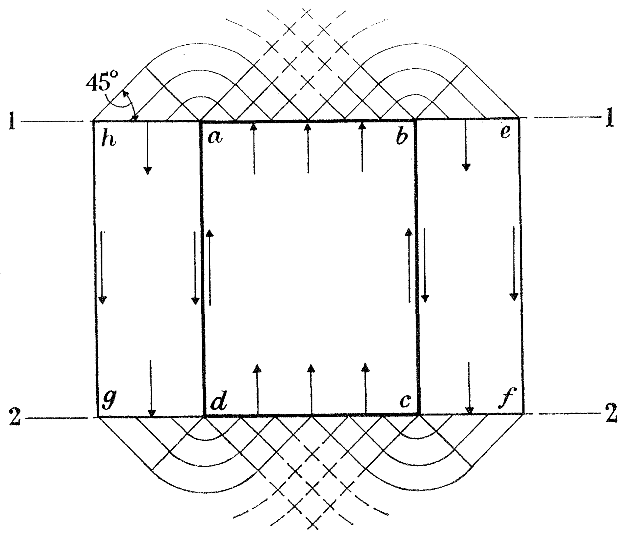

FIGURE 3. An approximate slip-line picture of convection in an ideally plastic body.

system would apply to incipient indentation. On the other hand, experiments indicate that as the indentation increases the slip line field changes towards that shown in figure 4 where, again, the dashed slip lines lie in a wedge that does not undergo plastic deformation but moves with the indenter as a rigid body. In both cases the pressure is constant over the surface ab of the indenter, and in both cases it amounts to

$$p = Y(1+\tfrac{1}{2}\pi). \tag{2}$$

The assumption of a rigid hot prism $abcd$ with high friction at its indenting surface, therefore, corresponds to the slip line solution figure 4.

The upward movement of the prism $abcd$ produces in the half-space below the horizontal plane 2–2 a negative indentation if the strength of the material, or the hydrostatic pressure, is sufficient to prevent separation along 2–2. The slip lines are identical with those in figures 3 or 4. The (tensile) stress required for the negative identation (plucking) is constant over the bottom surface of the prism, and its magnitude is again given by equation (2), independently of whether the slip lines are of the type of figure 3 or figure 4.

The penetration of the indenter extrudes material within the slip line field from the indented half-space above 1–1; similarly, as the surfaces be and ha (figure 3) move downward, the plucking of the prism at the half-space below 2–2 causes the strips cf and gd of the plane 2–2 to move downward by the same amount. Again, to obtain an approximate solution for the resistance to the movement of the prism *abcd*, the neighbouring prisms *befc* and *adgh* may be assumed to move downward as rigid bodies, against the plastic resistance $\frac{1}{2}Y$ (yield stress in shear) per unit area of their vertical faces. Finally, the same shear resistance also acts on the vertical faces *bc* and *ad* of the prism *abcd*. Per unit length of the prism (perpendicular to the plane of the figure), therefore, the force required to overcome the shear resistance at the side faces of the three prisms is $3hY$; with the identation resistance, the total plastic resistance to the upward rise of the prism *abcd* is

$$R = 2wY(1+\tfrac{1}{2}\pi)+3hY.\dagger \tag{3}$$

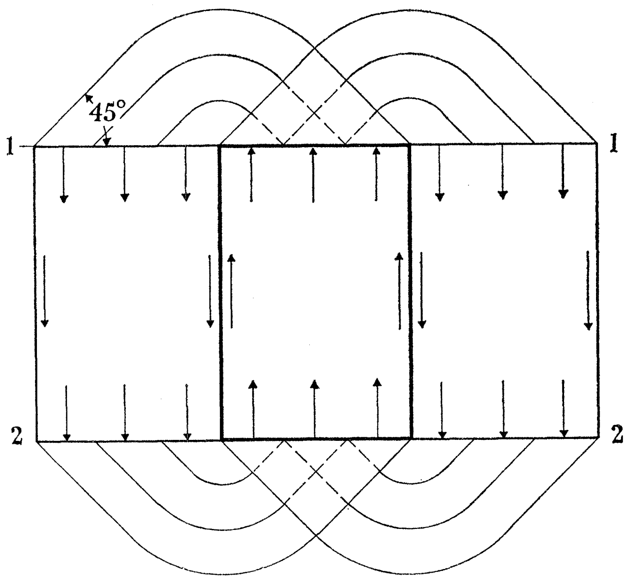

FIGURE 4. An alternative slip-line picture of convection in an ideally plastic body.

Equating this with the buoyancy force F gives the condition of convection:

$$2wY(1+\tfrac{1}{2}\pi)+3hY = wh\rho g\alpha\Delta T. \tag{4}$$

This condition applies to a hot prism embedded in a sufficiently large plastic space. The opposite case is that in which the half-spaces above 1–1 and below 2–2 have no plastic resistance (e.g. because they are liquid) while their density is the same as that of the plastic material to the right and left of the prism *abcd*. The only resistance to the rise of the prism is then that exerted on its sides, and the condition of rising is

$$hY = wh\rho g\alpha\Delta T. \tag{5}$$

For a numerical example, let it be assumed that $w = h = 1000\,\text{km}$, $\Delta T = 100\,\text{degC}$, $\rho = 3{\cdot}3$, $\alpha = 10^{-5}$ and $g = 1000$ c.g.s. Equation (4) gives then for the yield stress at which convective rise occurs

$$Y = \frac{3{\cdot}3}{5+\pi}h = 40{\cdot}5\,\text{b}.$$

If the resistance to indentation is negligible, (5) gives

$$Y = 330\,\text{b}.$$

† According to my colleague, Professor F. A. McClintock, an alternative estimate of the upper bound for the problem would give $R = 2wY(1{\cdot}5+\frac{1}{2}\pi)+2hY$.

In the case where only relatively thin or soft layers of material are present above 1–1 and below 2–2, the yield stress at which convection can take place is between these two limits, probably closer to the lower one; i.e. it is of the order of magnitude of 100 b. For smaller 'cells' and lower temperature differences the critical yield stress is proportionately higher, for larger cells and temperature differences smaller. The assumption of rigidity of the three vertically moving prisms in figure 3 leads to an estimate of the plastic resistance that is probably slightly higher than, or perhaps equal to, the exact value. In any case, (4) and (5) ought to be fairly accurate in comparison with the usual uncertainties of geophysical calculations.

Upper and lower limits of stresses and stress differences (shear stresses) in the crust and the mantle have been estimated in many different ways. The most direct and most relevant estimate is probably that of the stress drop in deep focus earthquakes derived from estimates of the energy release by means of the solution of the elastic stress distribution problem around a sheared crack due to Starr (1929). The numerical calculation requires the knowledge of the size of the seismic focal area; an estimate of this has been made by Jeffreys (1952, p. 341), based on the circumstance that the time between the first swing and the recovery in seismographic records is usually around 5 s. The greatest energy release in deep focus earthquakes is about 4×10^{22} ergs (Celebes 1934 and Philippines 1940; see Gutenberg 1956); the coresponding stress drop is about 170 b (Orowan 1960). More interesting in the present context is that deep focus shocks of Richter magnitude 6·5 have been recorded, with an energy release about 20 times less than the observed maximum, and therefore a shear stress drop of the order of 40 b. In principle, the stress drop need not have the order of magnitude of the initial stress itself; however, the heat developed during faulting should cause extensive melting (Jeffreys 1952, p. 340; Griggs & Handin 1960), and so the stress drop is unlikely to differ from the initial stress in the order of magnitude.

Since deep focus earthquakes are creep fractures, not brittle fractures (see § 11 below), the stress at which they take place should be the Andradean creep limit, i.e. the stress at which curve A in figure 1 shoots up from the stress axis. In other words, the seismogenic stress should be close to the yield stress to be used in the convection condition, (4) and (5). That the lowest stress drop in deep focus earthquakes practically coincides with the yield stress demanded by the most stringent of the convection conditions, for plausible dimensions of the hot column and a reasonable value of the temperature difference, indicates that deep-mantle convection should be possible.

3. Plastic convection patterns

Viscous flow is governed by partial differential equations of the elliptic type; plane plastic deformation, like supersonic flow, by equations of the hyperbolic type. The qualitative difference between the corresponding flow patterns appears clearly in the simple case of flow through a tube. With a liquid of Newtonian viscosity, the velocity distribution along the diameter of the tube is parabolic; with an ideally plastic material adhering to the tube, shear is confined to an infinitely thin layer at the wall and the material moves as a rigid body ('plug flow'). In the plastic case, of course, the flow can have any velocity between zero and infinity once the critical shear stress $\frac{1}{2}Y$ has been reached in the boundary layer.

Similar differences exist between convection patterns of Newtonian viscous materials and those of plastic-Andradean materials. In a Newtonian material a continuous velocity distribution arises (see, for example, Pekeris 1935). Ideal-plastic deformation, on the other hand, can be concentrated in surfaces of velocity discontinuity. A simple example is shown in figure 5 where the temperature above 1–1 and below 2–2 is T_0; between these planes it is $T = T_0 + A \sin kx$. As A increases, first an elastic stress distribution develops; then plastic deformation starts. When plastic flow is fully developed and the elastic strains are negligible in comparison, the shear is likely to be confined to the vertical planes of maximum shear stress, i.e. the planes where $\sin kx = 0$; these planes and the corresponding rigid-body displacements of the blocks between them are indicated by arrows. It will be seen below

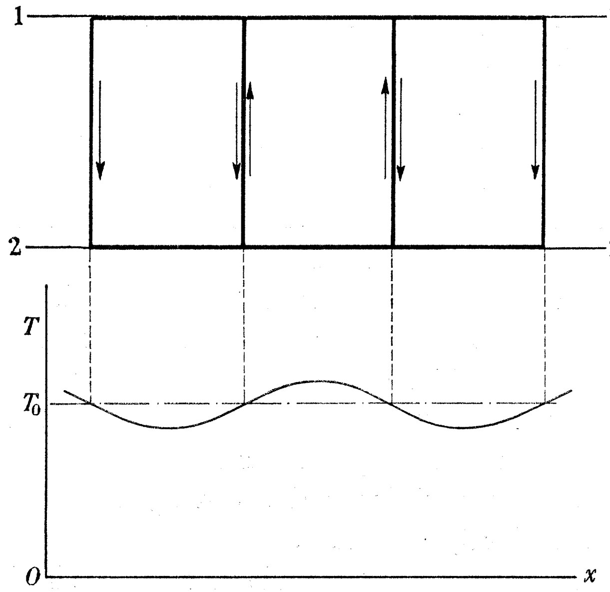

FIGURE 5. Block convection in sinusoidal temperature distribution.

that the concentration of shear in surfaces of singularity is promoted in the mantle by the mechanical-physical instability of its material, i.e. the tendency (common to all or almost all crystalline materials at high temperatures) to develop creep fracture. This seems to be the main cause of the earthquakes; that much or most of the deformation of the upper mantle is concentrated in thin layers of shear is indicated by the fact (see § 12) that the mean annual horizontal displacement in circumpacific dip-slip earthquakes seems to be of the order of magnitude of the velocity usually ascribed to convection currents and also to continental drift.

That the convection pattern typical of plastic-Andradean convection may play a dominating role in geotectonic processes is suggested by the following circumstances. Until the discovery of the intense oceanic heat flow by Revelle & Maxwell (1952) and Bullard (1954), it was usually assumed that the mantle was hotter under the continents, and therefore convection currents would rise there and descend under the oceans. After 1952 the oceanic heat flow was first attributed to higher radioactivity of the oceanic mantle which would produce a rising hot column; that the oceanic ridges were produced by the rise of convection currents was suggested by Hess (1960, 1962) and by Dietz (1961). As mentioned, an alternative view (MacDonald 1963) denied the possibility of deep convection because of the high coefficient of viscosity demanded by the observed shape of the Earth if Newtonian behaviour of the mantle was assumed; the high heat flow was attributed to large radioactive heat development in the oceanic mantle and its conductive-radiative transfer.

Both hypotheses encountered serious difficulties. Deep convection in a Newtonian mantle could hardly produce ridges of the observed sharp profile; it should give broad uplifts. Deep convection could not give the sharp curvatures of the ridges (e.g. in the equatorial Atlantic). If the assumption is made that the current descends in the central parts of the continents, the remarkably well centred position of the Atlantic ridge cannot be understood: it ought to be influenced then by the geography of the entire continent, not only its adjoining coastline. If, on the other hand, shallow convection is assumed, the ridges cannot be mere thermal upwellings: a ridge of 3 km height above the sea floor would require a temperature difference of about 1000 degC in order to be carried by a hot column of 200 km depth. If the buoyancy is attributed to a phase transformation or a chemical reaction underneath the ridge, it is difficult to see how this could be reversed or the lighter material otherwise removed a few hundred kilometres away from the crest. In addition, the assumption that the excess radioactivity of the oceanic mantle is concentrated in the top few hundred kilometres could not be reconciled so far with the possibility of continental displacements: unless the continents drag with them the mantle underneath, their highly radioactive crust would overlap the highly radioactive oceanic mantle and horizontal displacements would lead to very high heat flow at the leading side and a low heat flow at the trailing side of the continent. If, on the other hand, continental displacements are discounted, the whole complex of striking phenomena so well accounted for by the theory of continental drift remains unexplained; this assumption would demand too heavy sacrifices to be considered seriously at present.

In view of this situation it seems unavoidable to face a radically different possibility (Hess 1960, 1962; Bernal 1961). From the high oceanic heat flow it does not follow that there must be more radioactivity in the oceanic mantle: the heat flow could be the consequence of rising hot convective columns. But if the oceanic mantle does not have more radioactivity than the continental mantle, the rising column should be formed under the continent; let it be assumed, then, that the oceanic ridges are the upwellings of *convection currents which originated under a primeval continent before this was disrupted by the temperature rise to which it gave rise in the mantle.* There is no difficulty in the assumption of the survival of such 'paleo-currents': if the disruption occurred 300 or 500 My ago and the velocity of the current was of the order of 1 cm/y, the 'convection cell' has not made more than $\frac{1}{4}$ or $\frac{1}{2}$ revolution so far, and it could hardly have disappeared or reversed its direction. Before its disruption, the continental crust may effectively constrain, by its frictional strength, deformation in a plastic mantle, and thus hinder convection; after its disruption by the thermal updoming, the crustal constraint disappears, the hot column becomes hotter from the material rising from the bottom, and the convection accelerates, although (or rather because) the continental cover has disappeared over the rising column. Figure 6 illustrates a corresponding pattern of convection. *R* is the ridge in the ocean opened up by the disruption; arrows underneath indicate the hot column. The ridge thrown up by the hot column spreads out horizontally. *A* is one of the continental fragments and *P* an ocean on the opposite side of *A*;

if A and P behave like America and the Pacific, a shear layer S may arise between them, and the mantle below S is deflected vertically downwards, becoming part of the descending column. On the other hand, a shear plane S' may also arise between R and an adjacent continental fragment E; this may be the case along a narrow front between the Atlantic and western Europe, as suggested by the deep-focus earthquakes in Spain.

Convection in a plastic mantle avoids, in addition, other difficulties connected with the assumption of Newtonian viscosity. Among these are the problems raised by the rapid increase of the coefficient of viscosity with depth, by the sharp profile of the oceanic ridges, by their sharp curvatures, and their accurate centring between continents. In what follows, it should be briefly indicated how these problems may find solutions by the assumption that the mantle is plastic-Andradean, not Newtonian.

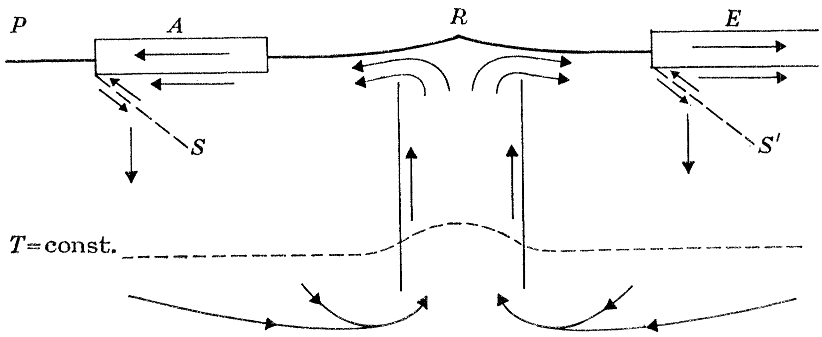

FIGURE 6. A scheme of convection due to the rise of a hot dike in an ocean.

(i) *Depth of convection.* Since the material is not Newtonian, the rapid increase of the viscosity with the pressure in materials with complex molecules need not cause difficulties. Little is known about the combined influence of pressure and temperature on dislocation movement by slip or by climb; but the remarkably small stress drop in deep-focus earthquakes shows that the yield stress (more accurately, the creep limit) can hardly be prohibitively high. The stress drop or the fracture stress in a Newtonian material would be unrelated to its viscosity; in a plastic material under hot creep conditions, however, creep limit and creep fracture stress are the same: once creep progresses, fracture follows usually after a more or less definite critical value of the creep strain. If the mantle is viscous and earthquakes (except very shallow ones) are creep fractures, the stresses revealed by them should give the order of magnitude of the creep stress.

(ii) *Profiles of the oceanic ridges.* When the continent is disrupted and free convection starts, material from the deep mantle rises upwards and isotherms of the kind shown in figure 6 develop. The bulge of the isotherm rises and widens, at first slowly; when the mean excess temperature of a vertical column and its width satisfy the appropriate convection condition such as (4) or (5), a hot dike begins to rise between two relatively narrow shear layers, somewhat like a rigid body. The top of the dike pushes up the floor of the ocean which slumps apart to form a relatively narrow and sharp-crested ridge. Under the crest the pressure is higher than at the same level away from the crest. Consequently, the ridge spreads and a rift may arise at its top where the hydrostatic pressure is not high enough to prevent tensile fracture.

(iii) *Curvature of the ridge*. The axis of a cylindrical convection cell in a Newtonian viscous material cannot have a radius of curvature less than approximately the diameter of the cell. In a plastic-Andradean mantle, however, the rising column can be a narrow dike and its axis may have a radius of curvature of the order of its thickness even if its height is close to the thickness of the mantle.

(iv) *Centring of the ridge in the ocean*. A possibility for explaining this striking phenomenon is indicated in figure 7. Over the dike the ridge is isostatically over-compensated, as is necessary for convection to occur. Where the ridge flows over the colder parts of the mantle outside the dike (where the ridge is shaded in figure 7), however, it is under-compensated and gradually it sinks into the mantle. Suppose now that the ridge, as in figure 7, is not in the centre of the ocean. Its mean uncompensated weight per unit area is then higher on

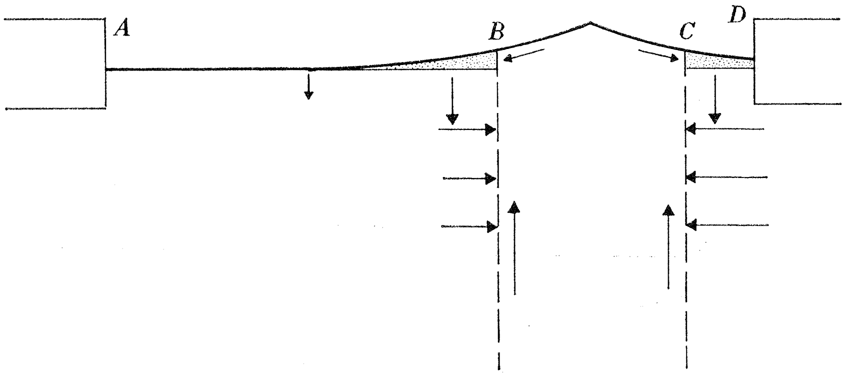

FIGURE 7. The mean pressure of the ridge outside the hot dike is higher where the continent is closer.

that side of the ridge which is closer to a continent. The higher pressure causes the ocean floor to subside faster on this side than over the area closer to the remote continent. Unless the sinking parts of the mantle behave as rigid slabs (according to the simplifying assumption made in § 2), they will spread out horizontally and the ridge is carried towards the centre of the ocean.

Whether this mechanism is supported by gravity surveys remains to be seen. The gravity at any point over the ocean is increased by the uncompensated mass of the ridge outside the hot dike but reduced by the mass deficiency of the hot column which exceeds the ridge mass over it. The net result may be a rather small anomaly, as in the case of Newtonian convection investigated by Pekeris (1935).

4. THE MURRAY–MENARD EAST–WEST FAULTS

The oceanic ridges are intersected by the more or less parallel wrench faults discovered in the Northern Pacific by Menard (1955);† in the Atlantic they are particularly numerous in the equatorial region (Heezen 1963), where the ridge deviates strongly from its dominating north–south direction. Both north of the Easter Islands, and in the Atlantic, the preponderant direction of the ridges is north to south, and the intersecting faults run approximately east to west. Since wrench faults are shear fractures, they should lie along

† One of the faults was previously observed by H. W. Murray.

planes of maximum shear strain; on the other hand, the theory of continental drift assumes that the Atlantic has been formed by a tensile principal strain of east–west direction which would mean (at least if the strain is not large) a plane of maximum shear strain at 45° to the east–west direction. How could then the east–west direction of the Menard faults arise?

The existence of the circumpacific seismic shear band discovered by Gutenberg and Richter (*S* in figure 6) indicates that the horizontal pressure produced by the upwelling at the oceanic ridges gives rise to shear fracture in a layer dipping from the continental margin under the continent. At the surface the fracture is of the Coulomb–Rankine frictional type; in the mantle it is probably a 'hot creep fracture', familiar in metallurgy. Creep fracture is preceded by the development of an instability of creep: with increasing strain or rate of strain, the material weakens and the stress drops; the physical background of this phenomenon will be discussed in § 11. If, for some reason, the shear rate increases in the Gutenberg–Richter band along a segment of the continental margin, the resistance of this segment to underthrusting by the oceanic crust drops. The crustal strip in front of the weakened segment accelerates towards the continent, and shear stresses develop between its flanks and the neighbouring parts of the oceanic crust; these may lead to shear fracture between the strip and its neighbourhood. The self-accelerating instability of hot creep can manifest itself in the upwelling along the ridges also: where the upwelling accelerates, it wedges apart the adjacent crustal strips and shear fractures may arise between these and the neighbouring crust. The shear faults will not be, in general, perpendicular to the accelerating segment of the ridge or of the continental margin, because the flow pattern is influenced by the neighbourhood and the remote parts of the ridge or margin also. In the case of the Atlantic Ridge, the flow is approximately an east–west extension even where the ridge is far from its general north–south direction; the faults developing by local acceleration of creep, therefore, should also be approximately of east–west direction in the Atlantic.

5. Continental drift as a consequence of creep instability

In figure 6 the convective upwelling under the Atlantic ridge *R* causes a horizontally diverging flow. If the conditions assumed on the left hand side of the figure are realistic, the westward flow remains horizontal and *never dips downward*: it carries America (*A*) westward while the Pacific crust and upper mantle dip under the Gutenberg–Richter seismic shear band, as indicated by arrows. This downward flow can be joined to the eastward flow in the deep mantle under America (branched arrows); the resulting convection cell, however, is fundamentally different from the types considered in the past. It has a discontinuity of flow at the Gutenberg–Richter seismic band; the convection may be called an *open cell* type. In figure 6, the open cell convection arising from the instability of hot creep in the mantle and the consequent formation of the Gutenberg–Richter band is the direct cause of continental drift. Ordinary closed-cell convection would draw a continent over its sink and keep it there in fixed position, unless the sink itself migrated for some reason.

Whether this mechanism can account for continental drift of the magnitude assumed on geological and palaeomagnetic evidence can be tested quantitatively. It will be seen in § 12 that the velocity of the relative movement between America and the Pacific can be estimated from the mean annual circumpacific seismic energy release; with what seem plausible estimates of the quantities involved, the order of magnitude of the velocity is about 1 cm/y.

On the other hand, the order of magnitude of the velocity of westdrift can also be estimated from the height of the Atlantic ridge which supplies the pressure driving America to the west; if the apparent viscosity of the upper few hundred kilometres of the mantle is of the Fenno–Scandian magnitude, again a drift velocity of the order of 1 cm/y is obtained (cf. § 10).

6. The strength of the lithosphere

Wegener assumed, without giving any physical reason, that the continents were hard plates floating in a relatively soft mantle. This assumption has been one of the main targets of attack against the theory of continental drift: the compressive strength of basalt, as measured in laboratory tests, is not lower than that of granite, and it was widely believed that the strength of the crust was equal, or at least close to, the compressive strength of the rock of which it consists. It should be pointed out in the present section that in reality the strength of a continental plate, or of the lithosphere in general, has little to do with the strength of its rocks as measured under common laboratory conditions. If the rocks behave in a brittle manner (as most rocks do at small depths), the strength of the lithosphere is predominantly of gravitational and frictional origin. At greater depths, where ordinary brittle fracture does not occur, the resistance to deformation is essentially the sum of the yield stress (creep limit) and of the gravitational strength.

It is perhaps not superfluous to remark that the term 'strength' is used with a confusing promiscuity in the literature, to mean both resistance to fracture and resistance to deformation. In general, these two quantities are fundamentally different and often unrelated. It would be desirable to use the term 'strength' only for the resistance to fracture, and to call the resistance to plastic deformation 'yield stress' (in Andradean creep, the quantity corresponding to the yield stress is the creep limit). For brevity, however, the resistance of the lithosphere or of the crust to non-elastic deformation (including fracture) will be called its strength, while the proper use of the term will be retained when properties of materials rather than bodies or structures are concerned.

In what follows, first the resistance to deformation of the topmost part of the lithosphere should be analysed, with the assumption that plastic deformation and creep are insignificant. In this case the resistance arises from three sources:

(A) *The fracture stress of the rock*

According to Griffith, the fracture stress of a brittle body is determined mainly by the largest crack present (often of submicroscopic size). A compressive strength of the order of 1000 or 2000 b, such as of granite or basalt, usually implies the presence of Griffith cracks between 10^{-4} and 10^{-3} cm in length, provided that there is no significant friction between the crack walls. Since the upper part of the crust contains joint planes and other cracks several centimetres or metres in length, its strength should be less by a factor of 10^4 to 10^6 if no friction is present (if there is a high pressure causing high friction between the crack walls, the resistance to stresses is of frictional nature; this case will be considered under (*C*) below). Consequently, the fracture stress of the rock as determined in ordinary laboratory experiments is likely to give a negligible contribution to the resistance of the uppermost parts of the lithosphere.

(B) *The gravitational strength of a floating plate*

If a continental plate floating in the mantle is compressed, its centre of gravity rises and the centre of gravity of the displaced volume of the mantle sinks. Both have the effect of increasing the gravitational energy; since the additional energy has to be supplied by the work of the compressive stress, the horizontal compressive strength of the continent has a gravitational component. Its magnitude for uniaxial compression can be calculated in the following manner.

If the thickness of the continent is t and its specific gravity ρ_c while the specific gravity of the mantle is ρ_m, the submerged depth z of the continent is given by

$$\rho_c t = \rho_m z. \tag{6}$$

If the continent suffers a small horizontal uniaxial compressive strain $\mathrm{d}\epsilon$, its thickness increases by $t\,\mathrm{d}\epsilon$ and the submergence by $(\rho_c/\rho_m)\,t\,\mathrm{d}\epsilon$. Its centre of gravity rises by

$$\tfrac{1}{2}t\,\mathrm{d}\epsilon - \frac{\rho_c}{\rho_m}t\,\mathrm{d}\epsilon = \left(\frac{1}{2}-\frac{\rho_c}{\rho_m}\right)t\,\mathrm{d}\epsilon,$$

while that of the displaced volume in the mantle sinks by

$$\frac{1}{2}\frac{\rho_c}{\rho_m}t\,\mathrm{d}\epsilon.$$

The change of gravitational energy per square centimetre of the continental area, therefore, is

$$\rho_c g t\left(t\left(\frac{1}{2}-\frac{\rho_c}{\rho_m}\right)\mathrm{d}\epsilon + \frac{1}{2}\frac{\rho_c}{\rho_m}t\,\mathrm{d}\epsilon\right) = t\sigma\,\mathrm{d}\epsilon, \tag{7}$$

where σ is the horizontal compressive stress necessary to supply the additional gravitational energy; this stress may be assumed to act uniformly in the continental plate. The gravitational contribution to the compressive strength of the continental plate is then

$$\sigma = \tfrac{1}{2}\rho_c g t(1-\rho_c/\rho_m). \tag{8}$$

With $\rho_c = 2{\cdot}8\ \mathrm{g/cm^3}$, $\rho_m = 3{\cdot}3\ \mathrm{g/cm^3}$, and $t = 33$ km,

$$\sigma \approx 700\ \mathrm{b}$$

for the continent. If it is assumed that the oceanic crust is chemically different from the mantle, and that its specific gravity and thickness have the values given by Worzel & Shurbet (1955), the gravitational component of its horizontal compressive strength is

$$\sigma \approx 130\ \mathrm{b}.$$

On the other hand, if the assumption is made that the oceanic crust arises from the mantle by a phase transformation (Dietz 1961), its gravitational compressive strength is zero, at least for sufficiently slow deformations.

A gravitational strength difference of nearly 600 b between the continental and the oceanic crust means that, if the other components of their strengths, yield stresses, and viscosities happened to be equal, the continents would be practically incompressible by oceanic crustal pressure, as it was assumed by Wegener.

The preceding calculations refer to the uniaxial compression of the crust; it is easily seen, however, that the gravitational strength is the same for biaxial compression. If the crust spreads transversely by an amount equal to its compression, the gravitational energy does not change; since this is the case of shear in a vertical plane in horizontal direction, the gravitational strength for horizontal shear (strike-slip faulting) is zero.

Naturally, the gravitational component of the tensile strength of the continental plate is negative and numerically equal to the compressive strength: a continental plate has a gravitational tendency to spread (Gutenberg 1927; Weertman 1962).

(C) *The frictional strength of the lithosphere*

If the lithosphere contains sufficiently large cracks (such as joint planes), the cohesive component of its strength is insignificant, and its resistance to stresses (apart from the gravitational component mentioned above) is due to friction between its parts. This can be calculated according to the rules of soil mechanics from Coulomb's law (see, for example, Terzaghi 1943 or Anderson 1951) down to the depth beyond which dry friction can no longer be assumed to act. According to Coulomb's law the shear resistance along a plane is proportional to the normal pressure in it:

$$\tau = \mu\sigma = \sigma \tan \phi, \tag{9}$$

where μ is the coefficient of friction and ϕ the angle of friction. In figure 8, OF is the line of frictional failure representing (9). Let $OZ = p$ be the geostatic pressure at a depth z; failure under a superposed horizontal uniaxial compressive stress $ZC = \sigma_c$ occurs when the Rankine–Mohr circle of which ZC is a diameter contacts the frictional line OF at a point P. From the figure,

$$(p+\tfrac{1}{2}\sigma_c) \sin \phi = \tfrac{1}{2}\sigma_c, \tag{10}$$

from which

$$\sigma_c = p\,\frac{2 \sin \phi}{1-\sin \phi}. \tag{11}$$

If a horizontal uniaxial tensile stress is applied, it produces frictional failure when its magnitude $ZT = \sigma_t$ is a diameter of a circle in contact at Q with OF; the frictional tensile strength σ_t satisfies the relation

$$(p-\tfrac{1}{2}\sigma_t) \sin \phi = \tfrac{1}{2}\sigma_t, \tag{12}$$

from which

$$\sigma_t = p\,\frac{2 \sin \phi}{1+\sin \phi}. \tag{13}$$

Finally, if a shear stress σ_s equivalent to a horizontal compressive stress σ_s and a numerically equal perpendicular horizontal tensile stress is superposed to the geostatic pressure, failure occurs if

$$\sigma_s = p/\sin \phi. \tag{14}$$

In this case a strike-slip fault with a vertical plane is produced.

Since $p = \rho g z$, the mean frictional strength of a crustal layer of depth d is obtained from (11), (13) and (14) be replacing p by the mean geostatic pressure $\bar{p} = \frac{1}{2}p$. With $\rho \simeq 2{\cdot}7$ for the top 10 km of the crust and $\mu \approx 1$, the frictional compressive, tensile, and shear strengths of the 10 km layer are $\sigma_c \approx 6500$ b, $\sigma_t \approx 1100$ b, and $\sigma_s \approx 1900$ b. Where the top of the crust consists of granite or basalt in unplasticized state (see § 13), plastic or viscous deformation is unlikely to be significant at a compressive stress of 6500 b (superposed to the smaller geostatic pressure). In such areas—e.g. the continental shields—the compressive strength of the crust should be higher than the sum of the frictional strength (*ca.* 6500 b) of the top 10 km, and the gravitational compressive strength (*ca.* 700 b) of the crust; the compressive strength of the crustal rocks being far lower than the frictional strength everywhere except in a surface layer of a few metres thickness, it has no significant influence upon the strength of the lithosphere. The compressive strength of the continental shields, therefore, is likely to be above 10 000 b, even if the compressive strength (fracture stress) of the rock itself is disregarded as an insignificant quantity. How does this compare with the compressive strength estimated on the same basis for the oceanic lithosphere?

The oceanic crust does not possess the gravitational compressive strength of the continents. The important question, however, is whether the oceanic lithosphere (i.e. the layer in which frictional fracture and shear dominates over creep) is significantly thinner than the continental lithosphere. If the isotherms are more crowded under the oceans, the layer in which

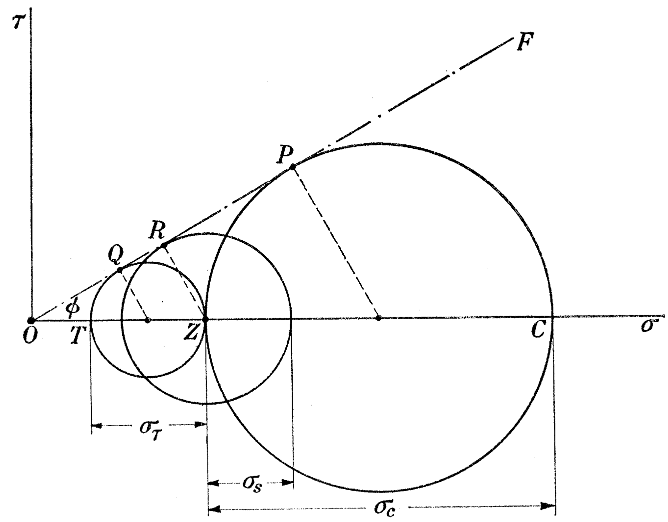

FIGURE 8. Frictional failure under compression, tension and shear.

the temperature is below a critical values is thinner; however, more important than the temperature may be the water content in reducing the resistance to creep and with it the thickness of the lithosphere in which creep is less important. It seems, in fact, that the oceanic crust and upper mantle contain more water than their continental counterparts. Geological evidence shows (Hess 1960, 1962; Dietz 1961) that the ocean floor is relatively young: it consists of fresh mantle material streaming from the upwelling at the ridges towards the continents where it is dehydrated and freed from light and volatile constituents in the course of orogenic processes. Apart from basalt flows covering it, the oceanic crust seems to consist of about 70% serpentinized peridotite (Hess 1960, 1962); but there is no reason to assume that most of the water rising from the mantle at the ridges is contained in the thin serpentinized layer. If a layer of at least 30 to 50 km thick under the ocean floors contains more water than the orogenically dehydrated layer under the continents, the silicate-plasticizing and fluxing effect of water should reduce the thickness of the oceanic lithosphere below that of the continental lithosphere. That in fact the continental lithosphere is particularly strong in compression is shown by the fact that compressive failures seem to be confined to the oceanic crust (e.g. island arcs), oceanic margins, and orogenic belts watered by the decomposition of serpentine at the 500 °C isotherm as it dips under a Gutenberg–Richter shear band.

Evidently, the floor of the ocean need not behave as a viscous liquid in order to make continental drift possible. It is sufficient if it undergoes the plastic deformation required for sliding under the Gutenberg–Richter band at the continental margin or along the island arcs.

7. Relative frequency of strike-slip and dip-slip earthquakes

In view of the preceding section it appears understandable why strike-slip faulting, and earthquakes with large strike-slip components, are so frequent in comparison with dip-slip faulting. If most of the hot convective dikes rise under the oceans, the continents must be areas of prevalent compression, except where hot dikes rise under them, as along the African rift or the Nevada seismic belt. However, compressive dip-slip faulting, according to the preceding section, probably requires more than 6 times the compressive stress that, in combination with an equal perpendicular tensile stress, can cause strike-slip faulting. In addition, compressive dip-slip faulting has to overcome the not insignificant gravitational compressive strength of the continental crust, while strike-slip faulting is not resisted by a gravitational strength.

At first sight, the argument involving the frictional compressive strength seems to apply only to shallow-focus earthquakes. However, earthquakes of great and intermediate focal depth are hot creep fractures (cf. § 11); creep fracture by grain boundary sliding in a polycrystalline material involves lifting the grains on one side of the shear plane out of engagement with the grains on the opposite side, and this requires less shear stress if the normal pressure is less. Creep fracture by grain boundary sliding, therefore, should resemble frictional failure, although the friction arises mainly by the geometrical interlocking of the grains. For this reason, a strong preference for strike-slip earthquakes ought to be expected in deep focus earthquakes also. The preponderance of strike-slip components in seismic faulting, therefore, does not contradict the view (Benioff 1955; Stille 1955; Dietz 1961) that compressive dip-slip faulting is the fundamental element of circumpacific seismicity. If the Coulomb–Rankine layer of the crust (the continental lithosphere) is so much easier to shear than to compress (cf. § 6 (C)), strike-slip faulting will predominate at the surface whenever the velocity of the underthrusting oceanic crust is not perpendicular to the continental margin.

8. Quasi-viscosity of the upper crust

There seems to be no reason to attribute a viscosity to the cold uppermost layers of the crust; in fact, in Jamieson's theory of the uplift of Fenno–Scandia by postglacial recovery the viscous deformation is usually assumed to take place in the asthenosphere. Surprisingly, however, the spectacular phenomenon of the diapiric rise of salt domes seems to indicate a quasi-viscous behaviour of the sedimentary strata in the uppermost parts of the crust. Since salt is a plastic crystal of very low yield stress, it would rise either very fast or not at all if the surrounding sedimentary rocks had a purely plastic or frictional (Coulomb-type) behaviour. In reality, there are strong reasons to assume that salt domes in the regions of the Persian Gulf and Louisiana have risen slowly by several kilometres, perhaps since the late Cretaceous. It is interesting to calculate the coefficient of viscosity with which the salt dome would rise at the observed rate in a Newtonian medium.

For an order-of-magnitude calculation, let the salt dome be considered a sphere. According to Stokes's formula, the velocity of its rise would be then

$$v = \tfrac{2}{9} r^2 \Delta\rho g/\eta, \tag{16}$$

where η is the coefficient of viscosity, $\Delta\rho$ the difference between the specific gravities of the surrounding rock, and the salt dome, and r the radius of the dome. If the velocity of rise is assumed to amount to, say, 5 km in 100 My

$$v = \frac{5\times10^5}{3{\cdot}14\times10^{15}} \simeq 1{\cdot}6\times10^{-10}\,\text{cm/s}.$$

With $r = 1$ km and $\Delta\rho \simeq 0{\cdot}4$ the coefficient of viscosity is

$$\eta = \frac{2}{9}\,\frac{10^{10}\times0{\cdot}4\times10^3}{1{\cdot}6\times10^{-10}} \approx 5\times10^{21} \tag{17}$$

in remarkable agreement with the value derived from the Fenno–Scandian uplift.

Since sedimentary rocks are not likely to have a genuine viscosity, perhaps the most plausible interpretation is that their quasi-viscous behaviour results from numerous small tectonic deformations by external causes, to which a preferential component has been added by the upward pressure of the salt dome. The process, therefore, may be similar to the slow sliding of objects placed on a slightly sloping board if this is vibrated.

9. The soft layer in the upper mantle

To explain the possibility of isostatic adjustments at the Fenno–Scandian rate, Barrell (1915) assumed the existence of a soft layer in the upper mantle, to which he gave the name 'asthenosphere'. Literally, this means 'no-strength-shell'; correctly, it ought to be 'low-hardness layer'. In what follows, it will be called the 'soft layer'. That it is not a spherical shell is seen from the fact that in some orogenically apparently inactive areas considerable anomalies of gravity do not cause anything like the rate of isostatic adjustment that would correspond to them on the ground of the Fenno–Scandian magnitude of viscosity; consequently, the soft layer must be absent or ineffective.

The physical cause of the soft layer was usually seen in the circumstance that the viscosity decreases with the temperature and increases with the pressure. Since the curve of temperature against depth should level out at a depth between 500 and 1000 km while the rise of pressure continues, there should be a minimum of the coefficient of viscosity.

This argument is inadequate for several reasons. First, the existence of persistent anomalies of gravity (e.g. at the 'hidden range' in India) shows that the soft layer can not have a general physical cause. Secondly, the mantle is substantially crystalline, and the hardness of crystals does not in general decrease steeply as the melting point is approached: one can skate on melting ice.

A more plausible reason for the existence of a soft layer is suggested by the curves in figure 9. Curves A and B represent relaxation experiments in torsion tests on two different types of polycrystalline Al_2O_3 specimens, both with an average grain size of 42 μm. The specimens were rods of square cross-section (edge length $\frac{1}{4}$ in.) and a gauge length of $\frac{3}{4}$ in. The torque was applied as fast as possible and then the drive of the testing machine stopped and the decline of the torque recorded. Where the two curves intersect, the creep rate of specimen B was about 45 times higher than that of A at the same torque and twist; yet the only difference between the two specimens was that B contained 0·7 % impurities and A 0·2 % (the major, and most effective, impurity was silicon). According to X-ray photographs, there was no perceptible deformation inside the grains; practically the entire creep was due to grain boundary sliding, and the faster creep of material B must have been caused by a grain boundary envelope of low-melting material (probably mainly an aluminum silicate). This corresponds to general metallurgical experience: high creep rates at high temperatures are usually caused by impurities producing low-melting grain boundary envelopes. In temperate glaciers the main factor accelerating flow is probably the depression of the melting point at the grain boundary by traces of ammonium nitrate formed by atmospheric electric discharges.

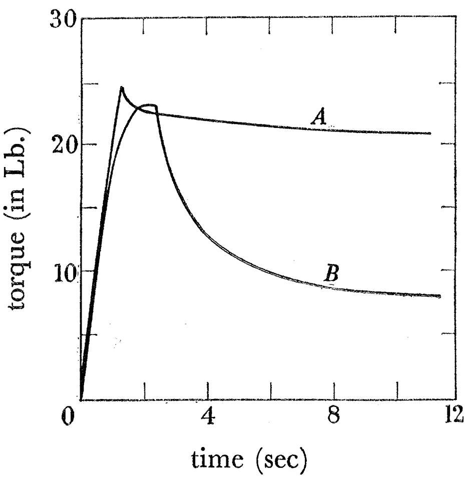

FIGURE 9. Torsional creep relaxation curves of polycrystalline alumina of higher (A) and lower (B) purity (Keith & Orowan 1965).

At greater depths the low-melting and volatile constituents of the mantle should be in a dissolved state in solid solution within the crystal grains. With diminishing depth and pressure, however, they must be expected to segregate and form low melting glassy grain boundary envelopes. Close to the surface these are rigid; below the cold surface layer, however, the grain boundary envelopes form layers of relatively low viscosity. It is natural, therefore, to expect the presence of a soft layer in the upper mantle, between the cold upper part of crust and mantle and the deep mantle where the low-melting components are dissolved in the grains.

That such a compound structure of the mantle is likely to be a decisive cause of its mechanical properties and its tectonic behaviour has been emphasized recently by Bernal (1961). An interesting hint in this direction is given by the chondritic meteorites. Besides crystalline parts similar to those found in the earth's crust, such meteorites contain frequently blobs of glass. What is more important, however, is the large number of spherolitic chondrules. A spherolite arises from a molten glass if it is cooled down slowly enough to permit crystallization, but not so slowly that the spherolites have time to recrystallize into non-spherolitic crystals in order to reduce their surface energy. Although the origin of spherolites is not known in detail, they seem to originate from the asthenospheric layer of the parent body. The frequency of chondrites as compared with achondrites may be due to the circumstance that the deep mantle of the parent body, not having thick glassy grain boundary envelopes, disintegrated into smaller fragments, most of which would evaporate before reaching the surface of the earth. If chondrites represent the asthenosphere of the parent body, which may be the result of considerable gravity differentiation of light and radioactive constituents, their composition may not represent the average composition of the disintegrated body.

The most common and most effective grain boundary 'fluxes' should be water and silica; the melting point of basalt can be depressed by 500 degC by a few parts per cent of water. If the cause of the soft layer is a low melting grain boundary envelope, there is no difficulty in understanding why considerable gravity anomalies can exist in some areas, such as India, without geosynclinal-orogenic processes providing sufficient explanation, and without significant isostatic adjustments. The absence of uplift or subsidence in such places may be the consequence of a missing soft layer due to the loss of low-melting and volatile constituents. In the case of India, for instance, it would be difficult not to connect the puzzle of the 'hidden range' with the tremendous loss of fluxing constituents which occurred during the escape of the Deccan Trap. Runcorn (1962) has emphasized that the persistence of the deviation of the Moon from its equilibrium state would demand an extremely high creep resistance of its material, unless the deviation is attributed to intense convection currents. The Moon may be so highly creep resistant because it has lost its water and with it its asthenosphere.

10. Estimation of continental drift velocities from the height of oceanic ridges

If the ridges are carried by hot dikes, their 'roots' are much deeper than the sialic roots of continental mountains; for simplicity let it be assumed that their effective depth is D and the density is constant down to this depth. If the height of the ridge over the ocean floor level is H and its specific gravity ρ, the pressure at the foot of the ridge exceeds that at the ocean floor outside the ridge by $H(\rho-1)g$, where the specific gravity of the sea water is taken as 1. At a depth z below ocean floor level the excess pressure under the crest of the ridge is

$$\Delta p = (1-z/D)\,H(\rho-1)\,g. \tag{18}$$

The mean excess pressure in the hot dike is

$$\overline{\Delta p} = m(1-z/D)\,H(\rho-1)\,g, \tag{18a}$$

where m is between $\frac{1}{2}$ and 1.

Since the mantle is plastic, and the excess pressure at its top is several times higher than its yield stress or creep limit, the pressure of the ridge acting on the dike underneath forces apart the adjacent parts of the mantle like a penetrating wedge. In the absence of gravity wedge penetration would raise the adjacent parts of the indented half-space (Hill, Lee & Tupper 1947; see also Hill 1950); in approximate isostatic equilibrium, however, the material sinks back and flows apart horizontally almost as soon as it is displaced by the dike pressure. Consider now the case in which the upwelling under the Mid-Atlantic Ridge causes simple shear in horizontal shear planes outside the hot dike; suppose that the shear deformation extends to the Pacific coast of America, over a distance, say, of $L = 6000$ km. The shearing force acting upon a horizontal plane at depth z is, per cm length of the dike,

$$F(z) = \int_0^z \overline{\Delta p}\,\mathrm{d}z = z\left(1-\frac{z}{2D}\right) H(\rho-1)\,g. \tag{19}$$

Although the mantle is not Newtonian, let it be assumed now that the Fenno–Scandian coefficient of viscosity $\eta = 10^{22}$ has the meaning of an average apparent coefficient of viscosity for the shear stresses present. The rate of shear flow at the depth z is then $F(z)/\eta L$, and the velocity at the top of the mantle is

$$v = \frac{mH(\rho-1)\,g}{\eta L}\int_0^D z\left(1-\frac{z}{2D}\right)\mathrm{d}z = \frac{mH(\rho-1)\,g}{3\eta L}D^2 \tag{20}$$

$$= 2\cdot5\times10^{-23}D^2\,\mathrm{cm/s} \simeq 8\cdot10^{-16}D^2\,\mathrm{cm/y}, \tag{20a}$$

if $m = \frac{2}{3}$ and $\rho-1 = 2\cdot3$. With $D = 200$ km, the velocity of westward drift of America would be about 0·3 cm/y; with $D = 500$ km, about 2 cm/y if there was no resistance to the movement at the Pacific coast. Suppose now that there is a resistance, and that it can be estimated as the stress required for compressing a slab of depth D and length D with the velocity $\frac{1}{2}v$ if the apparent coefficient of viscosity is 10^{22} c.g.s. From (20a), the compressive strain rate is

$$v/2D = 1\cdot25\times10^{-23}D$$

and the required compressive stress is

$$\sigma = 3\eta v/2D = 0\cdot375D\,\mathrm{dyn/cm^2}, \tag{21}$$

where 3η is the Trouton coefficient of viscous traction. With $D = 200$ km, this gives 7·5 b; with $D = 500$ km, about 19 b. Since the mean excess pressure of the ridge is

$$P = H(\rho-1)\,g/2 = 345\,\mathrm{b},$$

the compressive resistance does not influence the order of magnitude.

At first sight, this may seem to invalidate the basis of the preceding calculation of shear displacements over large distances; if the stress required for compressing a slab of length D and depth D is so small, why should shear occur over a length of many thousands of kilometres, requiring much higher wedging pressures in the mantle? The answer, of course, is that compression is possible only if the compressed slab can sink back into isostatic equilibrium. This is not possible where the top of the mantle is too warm from the proximity of the rising Mid-Atlantic Dike; nor is it possible over the continental areas of the Americas, for reasons discussed in § 6. But it does take place, according to seismological experience, at the west coast of the Americas. The presence of the seismic shear plane, if it involves thrust faulting, means simply the compression of a slab of length equal to the horizontal projection of the shear plane. Why the compression is not uniform but concentrated in a shear layer will be discussed in § 11 below.

The assumption made in figure 6 and in the above calculation is that the convective rise of the hot dike induces horizontal flow in a relatively thin surface layer. This is close to the idea of shallow convection suggested by Elsasser (1963) and in a different form by Wilson (1963).

The use of the Fenno–Scandian coefficient of viscosity, with the implicit assumption of Newtonian viscosity, introduces considerable uncertainty into the preceding calculation

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which could be removed only by more accurate knowledge of the mechanical properties of the upper mantle. This uncertainty, however, does not affect the main point, which is that the height of the Mid-Atlantic Ridge is a pressure indicator for the hot dike underneath, and that from the height, with drastic simplifications but no additional hypotheses, the order of magnitude of the continental drift velocities can be obtained in reasonable agreement with geological and palaeomagnetic estimates.

The wedging apart of the mantle by the gravity pressure of the hot and light dike is identical in principle with the gravity spreading effect investigated by Weertman (1962) as a cause of continental drift. Weertman showed that a difference of specific gravity between the mantle underneath oceans on either side of a continent causes the ocean with the lighter mantle to expand and the continent to move towards the other ocean. In the present calculation the cause of the lower mean density is seen in the presence of a hot dike. There is, of course, an oceanic ridge on the Pacific side of the Americas also; however, this ridge is somewhat shorter than the Mid-Atlantic Ridge, and its effect in reducing the mean density of the larger Pacific mantle is less. An additional point in the present calculation is the assumption that the height of the ridge is an indicator of the gravity spreading pressure.

The first use of the gravity spreading pressure for explaining continental drift was made by Gutenberg (1927) who considered the flowing apart of continental blocks as a possible cause of their disruption. Because of the thinness of the continental plates, their gravity spreading tendency could not have moved apart the fragments by considerable distances; this difficulty, however, was removed by Weertman in the manner mentioned.

11. Discontinuities of plastic creep: earthquakes

Ideal plastic deformation may develop discontinuities of displacement along surfaces of singularity; in the hot creep range, however, only high gradients of strain in relatively narrow shear layers can be present. If the Andradean creep curve *OA* in figure 1 gave a full description of the behaviour of the material, discontinuities in time could not occur. Attempts have been made to explain the intermittent nature of orogenesis by the plastic behaviour of the mantle; Griggs (1939) constructed a model which produced periodic 'convective overturns' separated by intervals of rest, but the periodicity was a consequence of mechanical inertia which is insignificant in mantle convection (a review of such attempts is given by Scheidegger 1958). It is clear now that plasticity alone cannot cause a discontinuity of deformation or convection in time.

That nevertheless the deformation of the mantle shows discontinuities in time is proved by the occurrence of earthquakes. Until 1956 (see Orowan 1960) earthquakes were usually regarded as fractures which take place when the stress reaches some critical value and are followed by sliding on the fracture surfaces and a stress drop; however, it was pointed out by Jeffreys already in 1936 that this classical theory of earthquakes was untenable. The geostatic pressure on the focal areas of deep focus earthquakes can be more than 200 kb; since the coefficient of friction cannot be considerably less than 1, frictional sliding after fracture would require shear stresses of the same magnitude. Such high stresses cannot be present in the mantle; but even if they were, a seismic shock could arise only if the fracture stress was higher than the stress required for sliding after fracture. von Kármán's (1911) experiments in which marble cylinders were crushed to a compacted powder under a superposed hydrostatic pressure illustrate the point that fracture under confining pressure, in general, takes place without stress drop, often indistinguishably from continuous plastic deformation.

The phenomenon of creep fracture is illustrated by laboratory experiments in figure 10 and figure 11, plate 4. Figure 10 shows torque-twist curves with the polycrystalline Al_2O_3 specimens mentioned in § 9 (material type B: 0·7 % impurity). Curves *A* and *B* in figure 10 represent torsion tests with two specimens at 1200 °C, at the same rate of twisting. In curve *A*, the experiment ended suddenly with fracture of a brittle type, although some non-elastic deformation was present. In experiment *B*, however, the torque-twist curve reached a maximum and started to decline when fracture occurred. Figure 11 shows this specimen

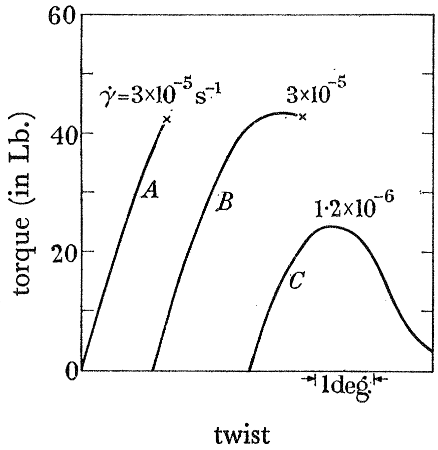

FIGURE 10. Torque–twist curves of polycrystalline alumina at 1200 °C (Keith & Orowan 1964).

after fracture. In addition to the crack that resulted in fracture, there is another which came quite close to causing fracture, and there are a number of cracks in a preparatory stage seen as white lines more or less parallel to the surface of fracture; some of these do not appear in the reproduced photograph. The lines are white because of the pores opened up between grains in the layers of shear strain concentration. Curve *C* shows an experiment with the same type of specimen at the same temperature, but $\frac{1}{25}$ of the rate of twisting. After reaching a lower maximum, the torque dropped continuously nearly to zero before fracture was completed.

The creep fracture mechanism of earthquakes explains several of their properties. Fore- and aftershocks could not happen if the earthquake was a brittle fracture, even if it took place in a viscous or plastic material; the first fracture would stop only when the stress had dropped below the magnitude needed for crack propagation, and the next shock would have to be preceded by another period of stress accumulation. Even in a visco-elastic material (Benioff 1951) progressive strain release could only produce a sequence of shocks of diminishing intensity. Benioff (1962) suggested that shallow earthquakes could develop fore- and aftershock sequences by the acceleration of creep underneath the fault in the top veneer which would give rise to repeated frictional faulting in the veneer. Below the veneer of frictional faulting, however, creep fracture seems at present the only way of explaining fore- and aftershocks (Orowan 1960); it gives rise to accelerated creep in the newly loaded

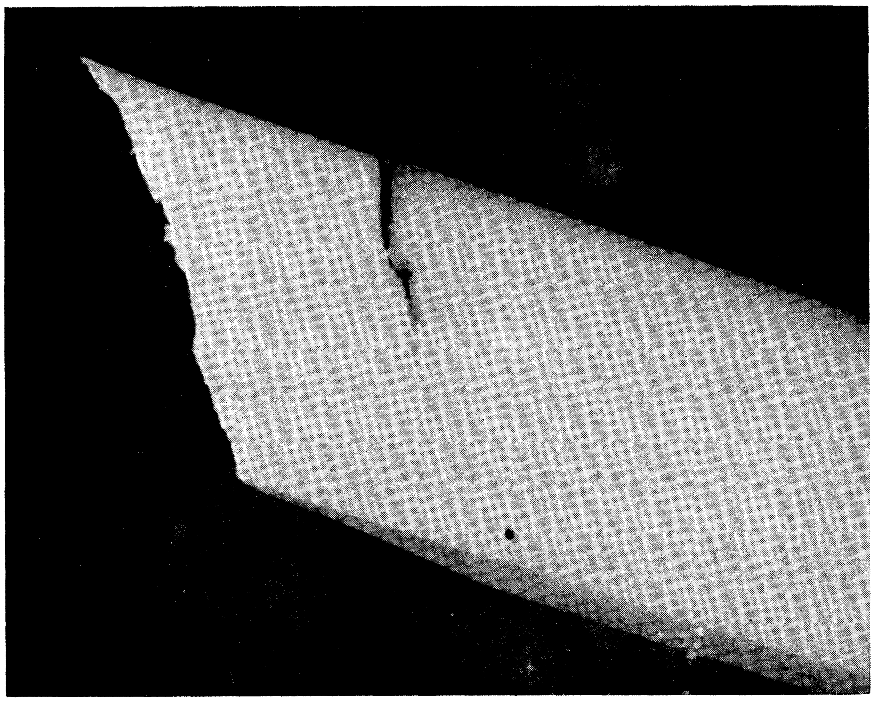

FIGURE 11. Parallel layers of complete and incipient creep fracture in a polycrystalline alumina torsion specimen broken at 1200 °C (Keith & Orowan 1964).

volumes of stress concentration, and repeated creep fractures may follow until the stress is sufficiently depleted.

It has often been discussed how the stress needed for an earthquake can accumulate in a viscous mantle; the difficulty was particularly striking in the case of deep-focus earthquakes which have a lower energy (and therefore stress) release, although both the coefficient of viscosity and the brittle fracture stress would be expected to be higher at greater depth. The answer is that creep fracture is not counteracted but produced by creep; once the stress is high enough to start creep, fracture takes place when the creep strain has reached a critical value. In other words, creep and fracture are not alternative and competitive processes, but two stages of the same process. The drop of the torque-twist curves *B* and *C* in figure 10 shows that hot creep involves a physical instability; just as the stress drop at the upper yield point of low carbon steel gives rise to Lüders bands, the stress drop in curves *B* and *C* produces the concentration of strain in narrow zones seen in figure 11. If such a strain concentration occurs, creep accelerates in the shear zones until local melting takes place (Jeffreys 1952; Griggs & Handin 1960) and the stresses in the neighbourhood are relieved.

Finally, the remarkable fact has to be explained that earthquakes do not seem to occur below a depth of about 700 km. This, of course, could be caused by the absence of sufficiently high stresses; but it is more likely that the instability of creep cannot occur below this depth. For instance, liquid-filled cavities may not be able to open up between the crystal grains if the pressure rises so high that the low-melting and volatile constituents remain dissolved in the grain.

12. Estimation of continental drift velocities from the energy release of circum-Pacific earthquakes

The discontinuity of deformation of crystalline materials in the hot range as manifested in the phenomenon of creep fracture raises the question how much of the deformation of the mantle takes place by continuous deformation preceding seismic creep fracture and how much by seismic fractures. An indication of the relative importance of the seismogenic component can be obtained thanks to the remarkable circumstance that the fault displacement in an earthquake depends only on the elastic constants and the released elastic energy per unit of width of the faulted area (perpendicular to the direction of faulting). Although some of the physical data needed for the calculation (such as the dimensions of seismic focal area) have to be represented today by crude estimates, the order of magnitude of the result can hardly change much when the estimates are replaced by measurements. It indicates that much of the deformation in the seismogenic part of the mantle takes place by seismic fractures; the drift velocities of continents calculated from the seismic energy release alone have the order of magnitude of the velocities assumed in the theory of continental drift (1 cm/y).

Let *L* be the length of the seismic focal area in the direction of slip and *W* its width at right-angle to it; *G* is the shear modulus and is Poisson's ratio. If during faulting the shear stress in the fault plane drops from τ_0 to τ, the elastic energy release is (Orowan 1960)

$$U = \tfrac{1}{4}\pi(1-2\nu)\,(WL^2/G)\,(\tau_0-\tau)^2 = \tfrac{1}{4}\pi(1-2\nu)\,W(L\gamma)^2\,G, \tag{22}$$

where

$$\gamma = (\tau_0-\tau)/G \tag{23}$$

is the seismic drop of the shear strain in the fault plane.

The strain release decreases continuously with the distance from the area of the fault: however, for a rough estimate it may be assumed that it has the value (23) within the two layers of thickness $\frac{1}{2}L$ above and below the seismic focal area, and that it vanishes outside these layers. Figure 12 shows that the shear displacement in the fault plane is

$$\delta = L\tan\beta = L\gamma. \tag{24}$$

With this, (22) becomes $$U = \tfrac{1}{4}\pi(1-2\nu)\, W\delta^2 G,$$

from which $$\delta = \frac{4U}{\pi(1-2\nu)\, GW}. \tag{25}$$

As a numerical example: with $U = 10^{23}$ ergs, $G = 10^{12}$ dyn/cm², $\nu = 0{\cdot}3$, $W = 10$ km, the shear displacement is $\delta = 565$ cm. If the entire circum-Pacific shear plane was fully but not

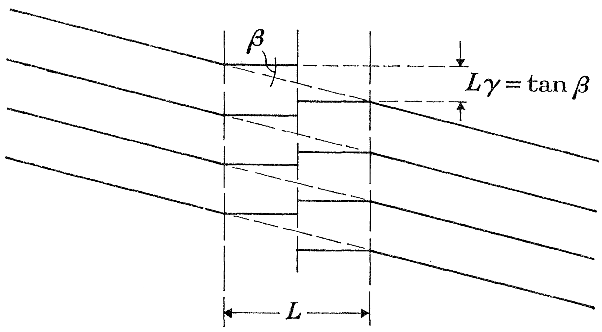

FIGURE 12. Schematic picture of strain release by local sliding.

multiply covered with such seismic faults, all of the compressive dip-slip type, the resulting penetration of the circum-Pacific continents into the Pacific would be

$$\delta\cos\alpha$$

if α is the angle of dip of the shear surface.

Suppose now that the area of the circum-Pacific shear surface is $CD/\cos\alpha$ where C is its circumferential length and D its depth. If annually n compressive-dip-slip earthquakes take place, all of the same focal area LW and the same elastic energy release U, the annual displacement of the continents into the Pacific is

$$v = \frac{nLW\cos^2\alpha}{CD}\sqrt{\frac{4U}{\pi(1-2\nu)\, GW}}. \tag{26}$$

To evaluate this expression it is not necessary to consider the entire shear surface down to the depth of 700 km. Equation (26) is valid for a part of the surface, down to an arbitrary depth D, if U means the average elastic energy release in earthquakes in this zone and n the annual number of earthquakes in it. It is of advantage to calculate only with shallow-focus earthquakes, both because their energy release can be more accurately determined and because the ratio of seismic to aseismic deformation is probably greater at smaller depths. According to Gutenberg (1957) the average annual energy release of all circum-Pacific earthquakes with focal depths between zero and 75 km is about 5×10^{24} ergs if all earthquakes down to the 'unified' magnitude $m = 6{\cdot}9$ are taken into account. The average annual number of earthquakes of such magnitudes on the whole Earth is about 20 (Gutenberg 1956); since the energy release in circum-Pacific earthquakes is about two-thirds of the global release, the number of large circum-Pacific earthquakes may be taken as 14. Of course, the majority of these earthquakes is of the strike-slip type, and some are tensile dip-slip earthquakes. On the other hand, the quantity U in the above equations does not mean the energy release in seismic waves but the release of elastic energy by faulting, of which a considerable part is lost in plastic-viscous deformation, both at the focal area and around it. In view of the uncertainties of both factors, it may be assumed that their effects approximately balance and that n can be regarded, for the present purpose, as the number of compressive earthquakes while U is identified with the average elastic energy release. Much of the global energy release comes from a few large earthquakes, some of which extend over many hundreds of kilometres; let it be assumed, then, that $W = 100$ km, $L = 10$ km, $\alpha = 45°$, $n = 14$, $U = 5 \times 10^{24}/24 = 3{\cdot}57 \times 10^{23}$ ergs. With these values, the annual velocity of the relative drift of the circum-Pacific continents against the Pacific would be

$$v = 1{\cdot}25 \text{ cm/y}.$$

Since this is the order of magnitude conventionally assumed for the velocity of continental drift, and also that calculated in § 10 from the wedging pressure of the Atlantic Ridge, it seems that a substantial part of the deformation of the upper mantle takes place in the form of seismic shocks.

13. Conclusions about the orogenic process

The main characteristics of mountain building are: (*a*) geosynclinal subsidence; (*b*) volcanism and plutonism; (*c*) plasticization and metamorphism of the geosynclinal tract; (*d*) orogenic compression of the plasticized crust; and finally, (*c*) hardening of the crust and local uplift and subsidence by block-faulting.

If hot convective dikes rise in the oceans, the origin of the compressive forces is understandable; the crucial question concerns the cause of geosynclinal subsidence and of the plasticization and metamorphism of the geosynclinal trough. Since the formation of magma, in all probability, is merely an intensification of the process of plasticization, it does not seem to present a separate problem.

The most powerful plasticizer of silicate rocks is water, particularly in connexion with free silica; the question is how water comes into the orogenic belts in large quantities? An attractive possibility of accounting for this is offered by the circumstance mentioned above that the oceanic ridges contain large quantities of serpentine. If it is assumed that a hot convective dike rises beneath the ridge and causes a horizontal flow of the oceanic crust and upper mantle towards the adjacent continents, the question of the origin of the orogenic plasticizer may have a simple answer. When the flow arrives at the seismic shear plane under the continental margin, it is deflected downwards. At a depth roughly equal to half of the thickness of the continental crust the decomposition temperature of serpentine is reached; around this depth the water in the serpentine is released and rises upwards while the remaining olivine sinks down into the mantle. Combined with the rocks and sediments of the continental margin, and heated by the highly radioactive sedimentary blanket, the water can cause the plasticization and the melting of magma on which mountain building depends.

The next important question is how the compressive stress can produce orogenic folding in the plasticized geosynclinal tract. Smoluchowski (1909) has proved that elastic buckling is impossible in a plate of the density of the continental crust and the 'strength' of granite floating in a substrate of the density of the mantle if the thickness of the plate exceeds some 20 m. The case of plastic buckling has been investigated by Vening Meinesz (1955); obviously, it would require extremely high velocities of compression to prevent the concurrent elimination of buckling by isostatic adjustment. An apparently bold, but on closer consideration very natural, solution of the problem has been suggested by Daly (1938). Daly assumed that, in the course of the orogenic process, the density of the plasticized and fluxed strata of the crust and mantle finally decreases below the density of the layers above them. This would give rise both to plutonic and volcanic outpourings of lava, and to a progressive collapse of the heavier upper layer into the soft and lighter layer beneath (Daly's 'major magmatic stoping'). If Daly's hypothesis, so strongly supported by volcanic and plutonic phenomena, is accepted, the difficulty of orogenic folding disappears; in fact, gravity promotes the compressive buckling of the upper crust if this floats in a hydrostatically unstable equilibrium on a soft layer of slightly lower specific gravity.

The last problem to be treated is that of geosynclinal subsidence, for a full century one of the most refractory puzzles of geology. It may have a solution of surprising simplicity. According to the usual estimates, the present annual discharge V of lava to the surface of the Earth is of the order of magnitude of 1 km^3. It is difficult to estimate the active geosynclinal area A of the earth; its order of magnitude cannot differ much from that of a strip of, say, 500 km width and 20 000 km length. If annually $V = 1\,km^3$ of lava is drained uniformly from under a geosynclinal area $A = 10^7\,km^2$, the annual subsidence of the geosynclinal trough is

$$S = V/A = 10^{-7}\,\mathrm{km} = 0{\cdot}1\,\mathrm{mm}. \quad (27)$$

This is in full agreement with the average rate of subsidence of geosynclines as derived from geological observations; its order of magnitude is given usually as 0·1 to 0·2 mm y.

Summed up briefly, the scheme of orogenesis suggested by the earlier sections is as follows:

Hot convective dikes rise in the mantle under the oceanic ridges which flow apart towards the continents. Because of the high frictional and gravitational compressive strength of the continent (§ 6), the flow plunges under the continental margin or it pushes the continent ahead—in which case the floor of the ocean on the other side of the continent plunges under its margin. Because of the instability of hot creep, the compression at the leading edge of the continent takes place by the formation of a shear zone, as in figure 12; this is connected with a 'thixotropic' yield stress drop, as seen in figure 11, so that the process cannot take place at opposite margins of the continent at the same time. The flow from the ridge transports serpentine towards the continent; its water content 'rains out' around the 500 °C isotherm under the continental shear plane. It plasticizes and fluxes the crust and finally produces in it a density inversion such as was assumed by Daly. This leads to the upward discharge of lava with the corresponding geosynclinal subsidence, and also to the reversal of the stabilizing effect of gravity upon crustal buckling and, in consequence, to orogenic folding.

This paper owes much to Professor Walter M. Elsasser, Princeton University, and to Professor S. Keith Runcorn, University of Newcastle upon Tyne, for extensive exchanges of ideas and for the stimulating effect of the Discussion on Continental Drift both before, during, and after the London meeting.

Appendix. The question of latent Newtonian viscosity in an Andradean material at low stress

Becker's theory of plastic flow attributed this to thermal stress fluctuations superposed to the applied stress; his expression for the flow rate (creep rate) was

$$d\gamma/dt = C\,e^{-A(\tau)/kT}, \tag{28}$$

where $A(\tau)$ is the activation energy of the elementary slip process, depending on the applied stress τ. Since the activation energy does not vanish with vanishing τ, the expression would give a finite creep rate for zero stress. To remedy this defect, (28) has to be replaced by

$$d\gamma/dt = C(e^{-A(\tau)/kT} - e^{-A(-\tau)/kT}), \tag{29}$$

where the second term on the right-hand side represents the reverse creep due to thermal fluctuations opposite to the applied stress (Orowan 1935). If the modification of the activation energy by the applied stress is a linear perturbation $\beta\tau$ (β being a constant of the order of magnitude of the molecular volume), the creep rate becomes

$$d\gamma/dt \simeq C\,e^{-A/kT}\,2\sinh\beta\tau/kT \tag{30}$$

(Eyring 1936; Orowan 1938; Glasstone, Laidler & Eyring 1941). If the perturbation of the activation energy $\beta\tau \ll kT$, the hyperbolic sine converges towards $\beta\tau/kT$, and Newtonian viscosity results.

The hyperbolic-sine type of non-Newtonian viscosity is very similar to Andradean viscosity (OA in figure 1). If the Andrade curve were of this origin, the material would have Newtonian viscosity at stresses sufficiently far below the creep limit. In fact, the corresponding Newtonian coefficient of viscosity could be calculated from any two points of the curve, even if the creep rate in the Newtonian stress range was too low to be measured. If the calculation is carried out, e.g., with the creep curves for lead wires given in Andrade's original publication, viscosities between 10^{-4} and 10^{-6} of the Fenno-Scandian value are obtained. The question, then, is whether the behaviour of a crystalline material in the hot creep range is of the hyperbolic-sine type; if it is, may not the mantle show approximately Newtonian behaviour at the stresses present? Experience accumulated in creep testing indicates that the answer to both questions is in the negative.

The strain rate versus stress curve in hot creep has been approximated by several empirical formulae, such as

$$\dot{\gamma} = C\tau^n \quad \text{(Norton 1929)}, \tag{31}$$

$$\dot{\gamma} = C(e^{a\tau} - 1) \quad \text{(Soderberg 1936)}, \tag{32}$$

$$\dot{\gamma} = C\sinh a\tau \quad \text{(Nadai 1938)}. \tag{33}$$

While the exponential and the hyperbolic-sine formulae are satisfactory at high stresses, it has been found that in general the hot creep behaviour of crystalline materials (metallic and non-metallic) at stresses substantially below the creep limit corresponds to the power formula (31), with the exponent n usually close to 4. Thus, there is no experimental indication of Newtonian viscosity at those stresses.

Equally significant is the fact that the physical cause of Andradean viscosity is usually not the high value of the perturbation of the activation energy relative to kT. Zener & Kê have determined the grain boundary viscosity of polycrystalline metals from torsional damping (cf. Zener 1948); the values obtained were of a much higher order of magnitude than the viscosity that would have been measured in creep tests under constant stress. Obviously, then, the creep limit was a consequence of geometrical interlocking of the grains which is much less effective under alternating than under a constant stress. In the case of single crystals the answer is simpler still: viscous creep, which requires the presence of deviations from perfect crystallinity, cannot occur until the yield stress has been reached and the production of lattice injuries has started.

Finally, the stresses in the mantle of the Earth cannot be assumed to be small compared with the creep limit. In the Newtonian low-stress range creep fracture could not take place; but the existence of earthquakes proves that it is a common occurrence in a substantial part of the mantle.

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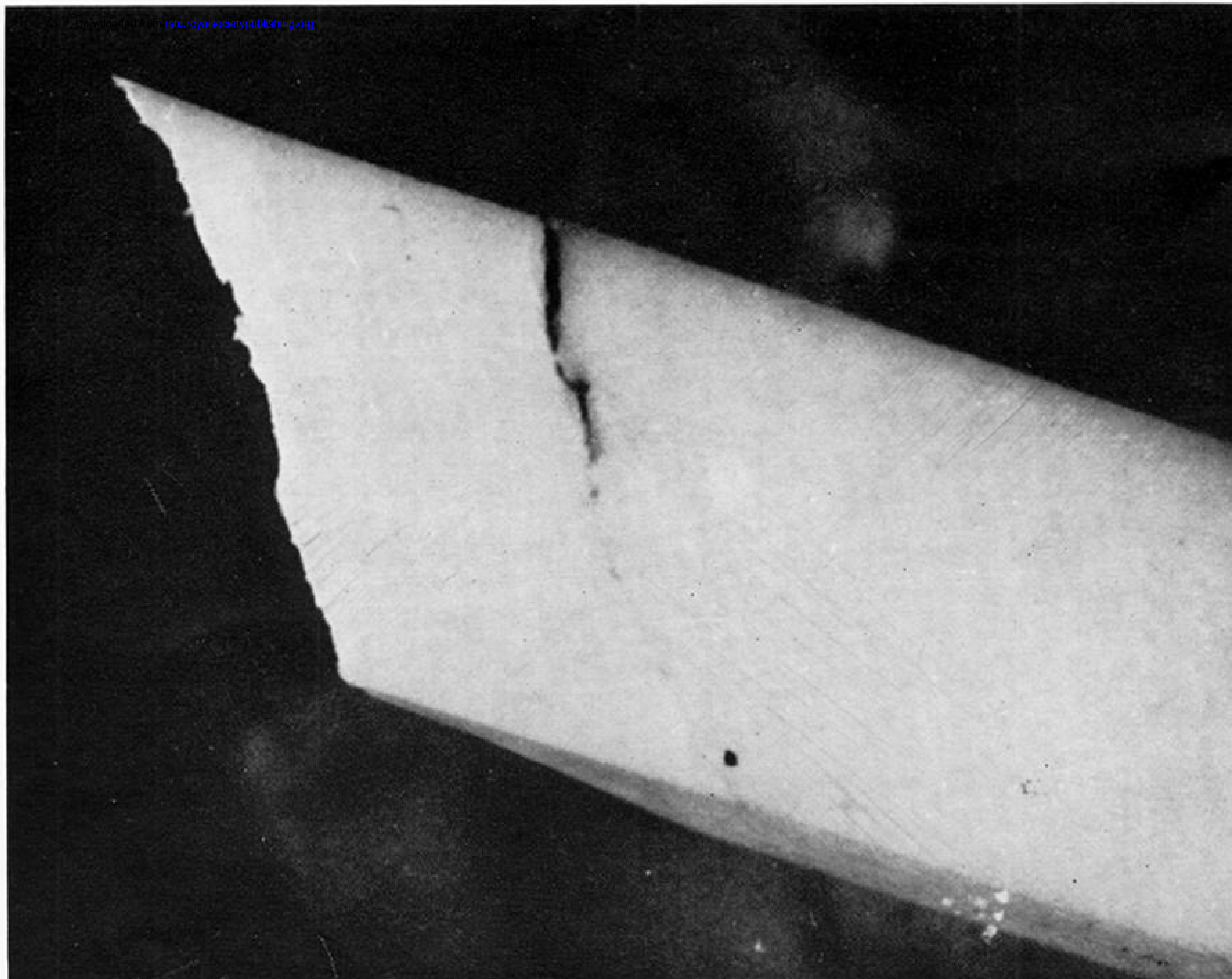

FIGURE 11. Parallel layers of complete and incipient creep fracture in a polycrystalline alumina torsion specimen broken at 1200 °C (Keith & Orowan 1964).

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