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Speculations on the Consequences and Causes of Plate Motions*

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Summary

Plate theory has successfully related sea floor spreading to the focal mechanisms of earthquakes and the deep structure of island arcs. It is used here to calculate the temperature distribution in the lithosphere thrust beneath island arcs, and to determine the flow and the stress elsewhere in the mantle. Comparison with observations demonstrates that earthquakes are restricted to those regions of the mantle which are colder than a definite temperature. The flow and the stress heating in the mantle can maintain the high heat flow anomaly observed behind island arcs.

Plate theory also suggests a new approach to the convection problem. The most obvious mechanism causing surface motion is the force on the plates due to the sinking lithosphere. This does not appear to be the way in which the motions are maintained. However, the input of large volumes of cold material can control convection and cause general downward movements in the mantle near island arcs. This input of cold lithosphere must cease when the island arc tries to consume a continent, since the light continental crust cannot sink through the denser mantle. Attempts to assimilate continental crust in this way can produce fold mountains, and also permit a rearrangement of convection cells.

1. Introduction

The original ideas of sea floor spreading (Hess 1962; Dietz 1961) were principally concerned with ridges, and with the creation of oceanic crust and upper mantle. They were first confirmed by the application of the Vine–Matthews hypothesis (Vine & Matthews 1963; Pitman & Heirtzler 1966; Vine 1966) to account for oceanic magnetic lineations by production of normally and reversely magnetized ocean floor along the ridge axis. Sykes (1967) used the first motions of earthquakes as independent confirmation that fracture zones on ridges are transform faults (Wilson 1965) and require non-conservation of crust.

The remarkable success of these ideas concerning sea floor creation required either expansion of the Earth or destruction of the ocean floor away from ridges. The immediate difficulty all expansion hypotheses face is the rate required. The sea floor spreading velocities are an order of magnitude greater than had been expected, and therefore require catastrophic expansion starting in the Jurassic. This suggestion seems geologically unreasonable, and therefore oceanic crust and upper mantle must be destroyed somewhere. Vening Meinesz (1962 for instance)

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had maintained that trenches were the site of such destruction for many years, but until recently there was rather little evidence in favour of this belief. In particular the negative gravity anomalies in trenches were used by Vening Meinesz to support the theory of crustal contraction, and by Worzel (1966) to support that of crustal extension. Most of the surface features on both sides of the trench suggest normal faulting (Ludwig *et al.* 1966) or slumping into the trench from the arc side (Gates & Gibson 1956; Brodie & Hatherton 1958). Perhaps two guyots, one in the Aleutian (Menard & Dietz 1951) and one in the Tonga trench (Raitt, Fisher & Mason 1955), were the best evidence that the crust under trenches was originally at a normal oceanic depth and had subsided after the formation of the guyots. Even if the subsidence is accepted, it could be caused by either extension or contraction. Also Hamilton & von Huene (1966) have now demonstrated that the guyot in the Aleutian trench is not in fact a guyot but a sea mount, since it does not possess a flat top.

The only clear evidence for crustal destruction in trenches has come from earthquake seismology. Detailed mapping of surface displacement in Alaska after the 1964 earthquake could only be explained by underthrusting of the ocean beneath the continent on an enormous scale (Plafker 1965). Focal mechanism studies of the main shock and of many aftershocks confirmed this result (Stauder & Bollinger 1966). More evidence came from a careful study of the location of intermediate and deep focus earthquakes in the Tonga area (Sykes 1966). This work demonstrated that the hook at the northern end of the Tonga Trench was matched by a corresponding feature in the intermediate and deep focus earthquake distribution. Such apparent mirroring of surface features at depths of 600 km is a remarkable result, and demonstrates that there is an intimate connection between surface and deep structures. There is now considerable evidence that this connection is the cold lithosphere moving down to great depths. Such a structure accounts for the propagation of high frequency P and S waves along the plane in this region (Oliver & Isacks 1967) and in Japan (Utsu 1967), because the oceanic lithosphere is known to have a high value of Q.

All these studies were concerned with the island arcs alone, and did not consider the problem of the conservation of surface area of a non-expanding earth. Nor did they discuss the motions of regions between ridges and trenches. Perhaps Wilson (1965) and Bullard, Everett & Smith (1965) were the first to realise the importance of rigidity of surface rocks in aseismic areas. Wilson stated the basic assumptions of plate theory, but made no further use of them. Bullard *et al.* fitted the Atlantic Continents together by a series of rotations about axes through the centre of the Earth. This procedure only succeeds because the continents have not deformed internally during their motion.

It is now clear that the major tectonic features of the Earth are produced by the relative rotation of few large aseismic plates, whose boundaries are the major seismic zones of the world. Since the seismic zones do not in general follow continental boundaries, the plates often contain both oceans and continents. For this reason 'continental drift' is a somewhat misleading name, and throughout this paper 'plate theory' is used instead. The relative motion between plates may be determined from the spreading velocities and the strike of transform faults on ridges (Morgan 1968; Le Pichon 1968) or from the focal mechanisms of earthquakes (McKenzie & Parker 1967; Isacks, Oliver & Sykes 1968). Le Pichon used the ridges to determine both the consumption rate and the direction of relative motion between the plates on either side of the trenches and island arcs, and both are in striking agreement with the seismic evidence (Brune 1968; Isacks *et al.* 1968).

The remarkable consistency of these two independent methods of determining the motions is the principal evidence in favour of plate theory, and therefore for the destruction of the lithosphere beneath island arcs and trenches. The details of this process are, however, still not fully understood. In particular it is not yet clear exactly where the huge overthrust fault intersects the surface of the earth, or how the motion takes place without disturbing the sediments in the trench (Bunce 1966; Shor 1966; Scholl, von Huene & Ridlon 1968). These difficulties are principally caused by a lack of knowledge, and do not demonstrate that plate theory is wrong. Indeed there is a similar difficulty in relating the detailed topography of ridges to the creation of oceanic crust on their axes (Atwater & Mudie 1968).

It is therefore clear from this work on plate theory that the lithosphere is consumed asymmetrically by island arcs, and it is the purpose of this paper to discuss the consequences of this destruction. Certain features of island arcs are related in a general way to the consumption of lithosphere. The northern Pacific (McKenzie & Parker 1967) demonstrates that active andesite volcanoes occur only where crust is destroyed (Fig. 1). The simplest explanation of this phenomenon is that oceanic crust is carried down into the mantle by the lithosphere. Partial melting at a depth of about 150 km then gives magmas of compositions between tholeite and andesite. Melting at different depths, and therefore pressures, may produce magmas of different compositions and hence account for Kuno's (1966) zones. This is the most obvious explanation of the observations, and appears to be in general agreement with high pressure experiments. Green & Ringwood (1966) partially melted various rocks of the calc-alkaline suite with different silica contents, and demonstrated that at 36 kb andesite had the lowest solidus and liquidus temperatures. Beneath the andesite volcanoes of island arcs the top of the sinking slab is at a depth of 100–150 km, or a pressure of 30–50 kb. Thus partial melting of the oceanic crust could produce the volcanics observed. This mechanism is simpler than that proposed by Ringwood & Green (1966), and can explain the remarkable correlation between active andesite volcanoes and consumption of the lithosphere (McKenzie & Parker 1967). Further experiments are necessary, since small concentrations of elements may have a large effect on the composition of the melt.

Other phenomena related to plate destruction are the occurrence of a trench 3 to 4 km deeper than the surrounding abyssal ocean floor, of a large negative gravity anomaly and smaller positive one over the island arc, of intermediate and deep focus earthquakes and probably also of a large heat flow anomaly above the descending slab. Of these effects the trench and gravity anomalies are probably the result of the shallow angle overthrusting of one plate by another. The explanation is consistent with what little is known about the strength and thickness of the lithosphere (McKenzie 1967).

Intermediate and deep focus earthquakes are of great interest, both because of the problem of their mechanism and also because they are a consequence of the three dimensional flow within the last 10 million years, though there are a few exceptions to this general rule (Isacks *et al.* 1968). Their focal mechanisms are not explosive, but are the usual double couple solutions generated by slip on a fault. The slip vectors demonstrate that these earthquakes are not caused by slip between the descending slab and the surrounding mantle. However, the internal fracturing of a homogeneous material does account for most focal mechanisms, at least in the Tonga–Fiji–Kermadec and Japan regions, if the axis of greatest principal stress is directed down the plane (Isacks *et al.* 1968; McKenzie 1969). This mechanism probably requires larger stress differences than those observed in shallow earthquakes.

The high heat flow anomaly behind the island arcs of the western Pacific is now well established (Vacquier *et al.* 1966), but is not easy to explain. McKenzie & Sclater (1968) attempted to do so by producing the heat by viscous shearing on the plane containing the deep earthquakes. They were unable to transmit this heat to the surface at a geologically reasonable rate. Another attempt to account for the heat flow is made below, and avoids the time scale difficulties.

The major remaining problem in plate theory is the driving mechanism. Thermal convection in some form is the only source of sufficient energy, but agreement goes no further. A recent lengthy discussion of the convection problem (McKenzie 1968a)

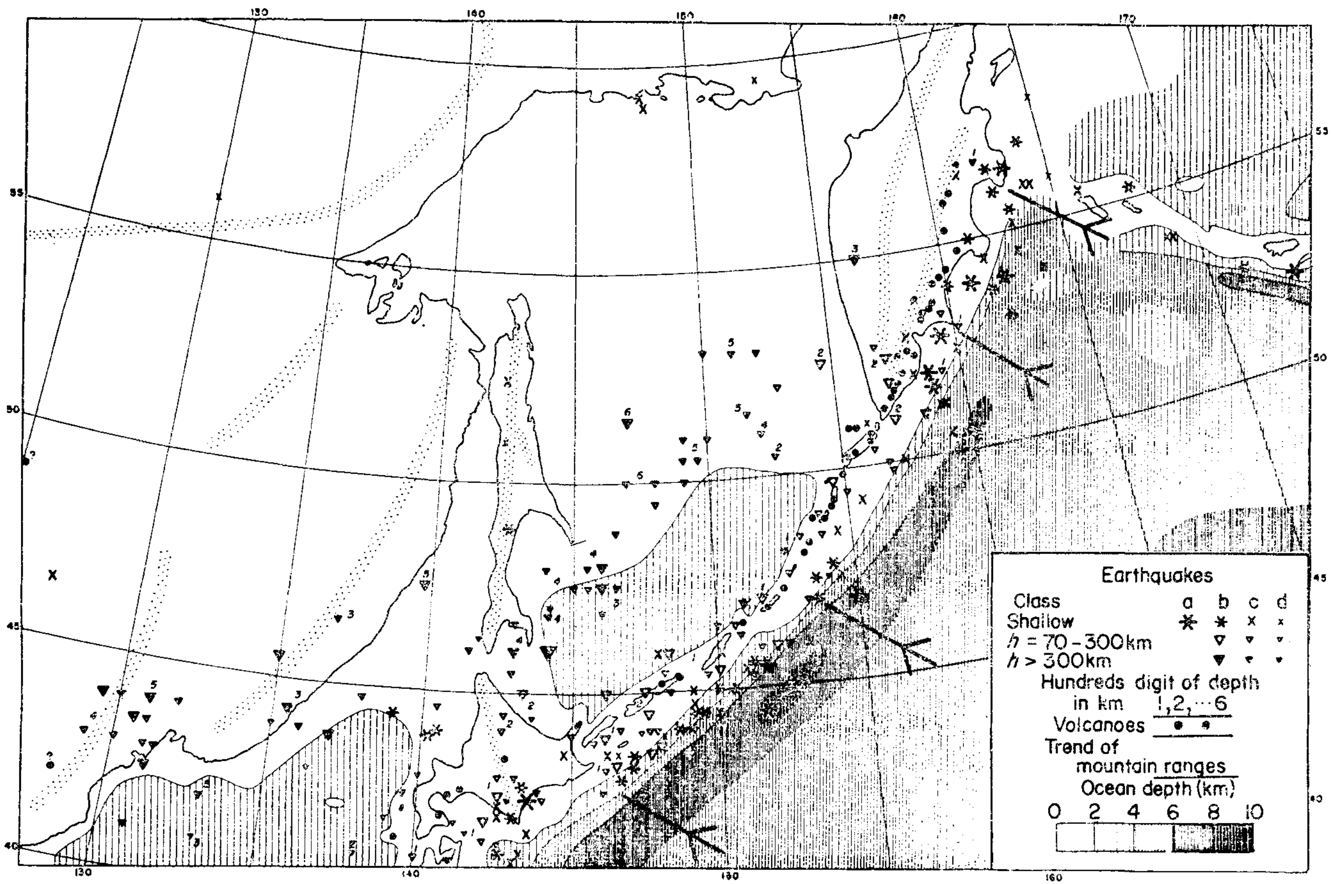


FIG. 1. The arrows show the direction of the relative motion between the Pacific and the Kamchatka–Kurile Island plate. At the west end of the Aleutian Arc the motion is strike slip, and there are no active volcanoes or intermediate focus earthquakes. Both appear where the Pacific plate is being underthrust beneath Kamchatka and the Kurile Islands. From Gutenberg & Richter (1954).

merely demonstrated that there are no obvious objections to large-scale convection. This result is scarcely surprising because almost all surface effects previously believed to be related to mantle wide convection are more easily explained by creation and destruction of plates. Elsasser (1967) has suggested that the motion of the plates themselves is not caused by viscous coupling to the mantle beneath, but that the cold slabs beneath island arcs sink and pull the rest of the plates with them. This solution of the convection problem requires the lithosphere to act as a stress guide. The argument is very appealing because it is simple, and previous attempts at solving the convection problem (see McKenzie (1968a) for a discussion of earlier work) were not. The mechanism is still a form of thermal convection because it is driven by the temperature induced density contrast between the cold slab and the hot mantle through which it sinks. Such convection is quite different from the Rayleigh–Benard problem of marginal stability because in Elsasser's problem convection of heat dominates the temperature distribution, and also because the flow is controlled by the extreme temperature dependence of the viscosity. Very little is known about convection of this type, and therefore the analysis below is rudimentary. It does, however, suggest that Elsasser's mechanism cannot maintain the motion of the major plates. This conclusion is supported by the observation that not all pairs of plates have a sinking slab attached to either of them. It is disappointing that neither observations nor theory support the idea, since the surface effects of all other types of flow are confused by the strength and rigidity of the lithosphere. Though the non-hydrostatic gravity field appears to be dominated by large-scale motions in the mantle, it cannot yet be used to determine the flow field because it is not known whether the geoid is elevated or depressed over a rising convection current (McKenzie 1968a).

The analysis of the temperature within the slab and the flow and stress fields caused by its motion require some knowledge of the mechanical and thermal properties of mantle materials. The properties of the sinking slab are least in doubt, since it must be brittle enough to produce earthquakes by fracture, and sufficiently undeformable to maintain its shape at a depth of 600 km after passing through perhaps 1000 km of mantle (Sykes 1966). Therefore for all purposes except that of generating earthquakes it behaves as a rigid plane slab. The mechanical properties of the mantle at depths of 100 km and greater, which are not part of the sinking slabs, have been the subject of fierce debate for many years. A recent collection of relevant experimental results (McKenzie 1968b) shows that pure ceramic oxides at high temperatures satisfy the viscous constitutional relationship at shearing stresses σ below $\sim 10^8$ dynes cm^{-2}. At greater stresses the creep rate $\dot{\varepsilon}$ obeys:

$$\dot{\varepsilon} = K\sigma^n \tag{1.1}$$

where n and K are constants. Typical values for n are between 3 and 7. The mantle is not pure olivine, but a complex mixture of minerals. Creep of such a solid is more rapid than that of the pure material. Throughout the analysis below the mantle is assumed to obey a viscous constitutional relationship between stress and strain rate. This assumption is a poor approximation in the region where the lithosphere bends, because the shearing stresses in the mantle outside the slab must be greater than 10^8 dynes cm^{-2}. Despite this limitation the solutions for the flow are believed to be similar to the real flow patterns, and permit a discussion of the stress distribution and viscous heating.

The thermal model for the mantle used below is considerably different from most of those commonly considered. The temperature gradient below the lithosphere is taken to be the adiabatic gradient everywhere, and all horizontal temperature variations outside the sinking slab are neglected. The lithosphere on top acts as a thermal boundary layer which can support large temperature and density gradients because of its finite strength. The effect of phase changes and of the adiabatic gradient within the mantle are contained in the analysis in Section 2 if T is defined to be the potential, rather than the ordinary, temperature. The arguments in favour of an adiabatic mantle below a mechanical and thermal boundary layer have been discussed previously (McKenzie 1967, 1968a).

The following values of parameters are used throughout these calculations:

$$\left.\begin{array}{ll} C_p = 0{\cdot}25\,\mathrm{cal\,g^{-1}\,{}^{\circ}C^{-1}} & \rho = 3\,\mathrm{g\,cm^{-3}} \\ \kappa = 0{\cdot}01\,\mathrm{cal\,cm^{-1}\,{}^{\circ}C^{-1}\,s^{-1}} & l = 50\,\mathrm{km} \\ T_1 = 800\,{}^{\circ}\mathrm{C} & \eta = 3\times10^{21}\,\mathrm{poise} \\ \alpha = 4\times10^{-5}\,{}^{\circ}\mathrm{C}^{-1} & g = 10^3\,\mathrm{cm\,s^{-2}} \end{array}\right\} \quad (1.2)$$

Only C_p, ρ and g are well determined. The thermal conductivity κ and its dependence on temperature are both uncertain, and therefore the value chosen may be wrong by perhaps a factor of two. The thickness l of the lithosphere is a poorly defined quantity, since the change in mechanical properties with depth is a gradual process, and cannot produce a sharp boundary. The value chosen is probably within 30 km of the true effective thickness. T_1 is the temperature at the base of the lithosphere, and therefore the temperature throughout the mantle in this model. Since this temperature only enters the analysis as a scaling factor, an error in the value of T_1 produces an error in the temperature, but not in the shape, of the isotherms (see also McKenzie 1967). The value of α the thermal expansion coefficient is that for pure olivine at high temperatures. Impurities are likely to increase this value, and there is some evidence (McKenzie & Sclater 1969) that the value in equation (1.2) may be too small. The viscosity η is the least well known of all these parameters. The value within the mantle is probably very variable because it depends exponentially on the temperature. The value chosen is similar to those obtained from the uplift of Fennoscandia (McConnell 1965) and Lake Bonneville (Crittenden 1963), but could be in error by an order of magnitude.

2. The thermal structure of the sinking slab

All the seismic evidence discussed in the previous section favours consumption of the lithosphere in island arcs by underthrusting. The underthrust plate then descends as a rigid slab deep into the mantle. Since the thermal conductivity of rocks is so small, the slab will retain its original temperature distributions even after it has sunk to considerable depths. The temperature structure within the slab may be obtained from the analysis used previously to discuss the heat flow anomalies associated with ridges (McKenzie 1967). This method is successful only because the slab is rigid and two dimensional. The equation governing the temperature T within the descending lithosphere is:

$$\rho C_p\left(\frac{\partial T}{\partial t}+\mathbf{v}.\nabla T\right) = \kappa\nabla^2 T+H \quad (2.1)$$

where H is the rate of radioactive heat generation $\mathrm{cm^{-3}}$ and the other quantities are defined in Section 1. Provided the spreading velocity $\mathbf{v}$ has been constant during at least the last 10 million years all transient temperature distributions will now be unimportant. It is not yet possible to test this assumption against the spreading rate obtained from the magnetic lineations because the reversal time scale at present extends back only 5 million years. Thus a world-wide pause in spreading cannot be detected if it occurs over the same period everywhere. Such an event has been suggested by Ewing & Ewing (1967) to explain sediment thicknesses on ridges. Seismic reflection records show that the thick sediments on the ridge flanks thin abruptly at anomaly 5. If the sedimentation rate has been constant, this observation requires a period of about 30 million years during which the plates did not move. Such an event is difficult to reconcile with any form of large-scale convection. Changes in convection will produce slow changes in the spreading rate over tens of millions years, and not sudden pauses: slow variations have been observed (Heirtzler *et al.* 1968) and do not occur simultaneously in different oceans. Thus it appears more likely that these observations are caused by large variations in the sedimentation rate, and not in the spreading rate. If this explanation is true, then $\partial T/\partial t$ may be neglected. H may also be ignored because of the small radioactivity of the lithosphere. Thus equation (2.1) becomes:

$$\rho C_p \mathbf{v}.\nabla T = \kappa \nabla^2 T \tag{2.2}$$

substitution of:

$$T = T_1 T', \qquad x = lx', \qquad z = lz'$$

where T_1 and l are defined in (1.2) gives:

$$\frac{\partial^2 T'}{\partial x'^2} - 2R\frac{\partial T'}{\partial x'} + \frac{\partial^2 T'}{\partial z'^2} = 0. \tag{2.3}$$

The x axis is taken to be parallel to the dip of the slab, and the z axis to be normal to the plane of the slab. The origin of the co-ordinates is chosen to be on the lower boundary (Fig. 2). The Thermal Reynolds number R is:

$$R = \frac{\rho C_p v_x l}{2\kappa} \tag{2.4}$$

R does not depend on v_y because the slab is assumed to be a two dimensional structure, and therefore $\partial T'/\partial y'$ is zero. The general solution to equation (2.3) is:

$$T' = A + Bz' + \sum_n C_n e^{\alpha_n x'} \sin k_n z' \tag{2.5}$$

if:

$$\alpha_n = R - (R^2 + k_n{}^2)^{\frac{1}{2}} \tag{2.6}$$

where A, B, C_n and k_n are constants. Since the temperature boundary conditions on both the upper and lower surfaces of the slab descending through the mantle are $T' = 1$ everywhere, for sufficiently large values of x' the temperature T' must be 1. Thus $A = 1$ and $B = 0$, and:

$$1 = 1 + \sum_n C_n e^{\alpha_n x'} \sin k_n$$

$$k_n = n\pi. \tag{2.7}$$

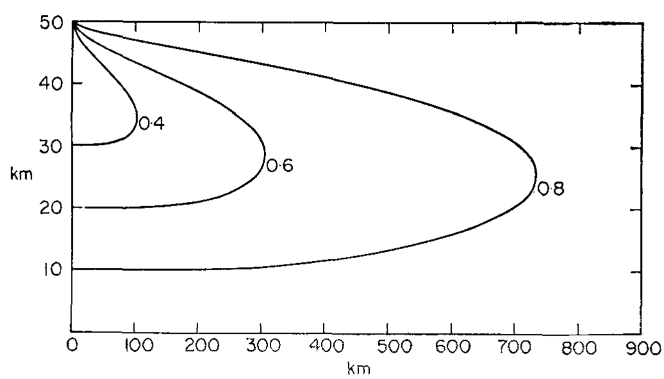


FIG. 2. Isotherms within the sinking slab, ×10 vertical exaggeration. Temperatures are in dimensionless units. If the values of the parameters in (1·2) are correct, they may be converted into °C by multiplying by 800. The ordinate is measured normal to the slab, the abscissa down its dip.

At $x' = 0$ the temperature distribution is the same as that in the lithosphere beneath the ocean (McKenzie 1967), or:

$$T' = 1 - z'. \tag{2.8}$$

Thus at $x' = 0$:

$$1 - z' = 1 + \sum_n C_n \sin n\pi z'$$

$$C_n = \frac{2(-1)^n}{n\pi}. \tag{2.9}$$

Substitution of equations (2.6), (2.7) and (2.9) into equation (2.5) gives:

$$T' = 1 + 2\sum_n \frac{(-1)^n}{n\pi} \exp\left[\left(R - (R^2 + n^2\pi^2)^{\frac{1}{2}}\right) x'\right] \sin n\pi z' \tag{2.10}$$

Isotherms from equation (2.10) for $v = 10\,\mathrm{cm\,yr^{-1}}$, or $R = 62{\cdot}5$, in Figs 2 and 3 show how they are convected deep into the mantle by the motion. An approximate expression for the temperature may be obtained by neglecting all terms in equation (2.10) except $n = 1$:

$$T' \simeq 1 - \frac{2}{\pi} \exp\left[\left(R - (R^2 + \pi^2)^{\frac{1}{2}}\right) x'\right] \sin \pi z'. \tag{2.11}$$

Since $R \gg \pi$ equation (2.11) gives:

$$T' \simeq 1 - \frac{2}{\pi} \exp\left(-\frac{\pi^2 x'}{2R}\right) \sin \pi z'. \tag{2.12}$$

The greatest depth reached by any isotherm x_M' is now easily obtained, since at this depth $\partial T'/\partial z' = 0$. Thus:

$$-2 \exp\left(-\frac{\pi^2 x_M'}{2R}\right) \cos \pi z' = 0 \tag{2.13}$$

or:

$$z' = \tfrac{1}{2}$$

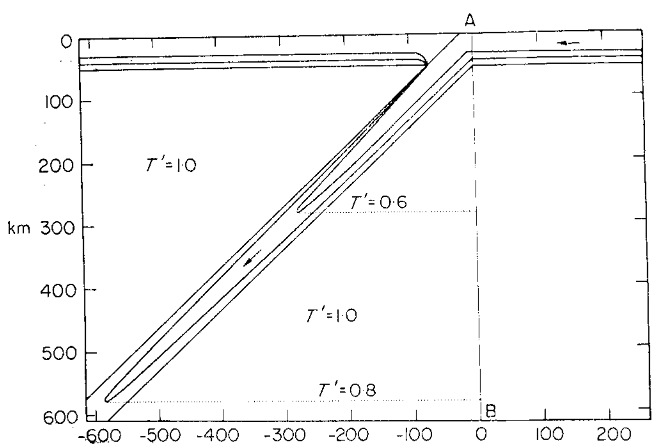


FIG. 3. Temperature structure beneath island arcs with no vertical or horizontal exaggeration. Temperatures as in Fig. 2.

since $0 < z' < 1$. Then equation (2.12) gives:

$$T' = 1 - \frac{2}{\pi} \exp\left(-\frac{\pi^2 x_M'}{2R}\right) \tag{2.14}$$

or:

$$x_M' = \frac{2R \log_e [2/\pi(1-T')]}{\pi^2}. \tag{2.15}$$

If:

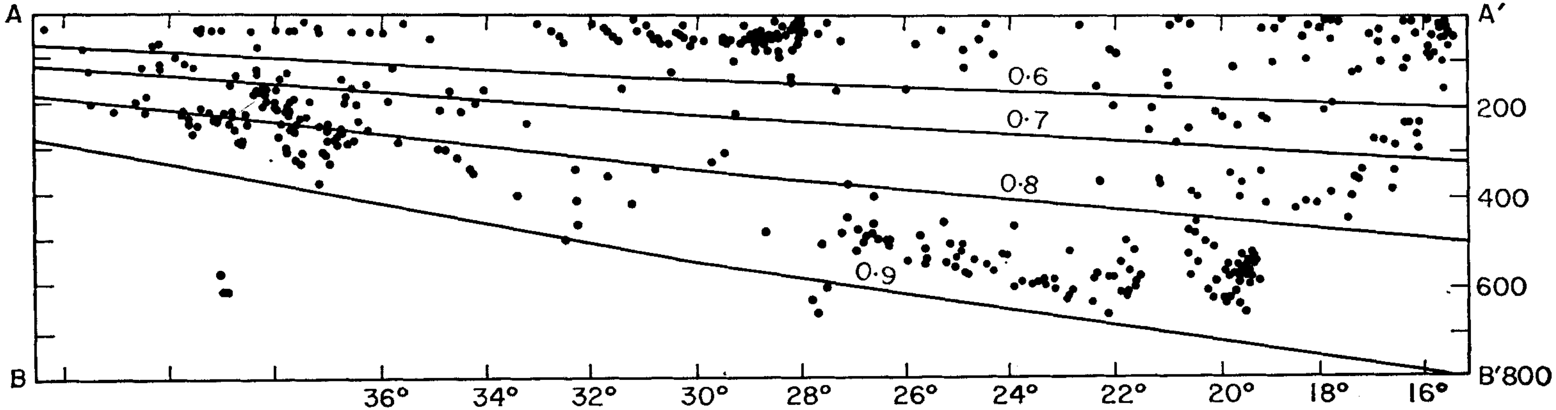
$$T' = 1 - \frac{1}{\pi} = 0{\cdot}682$$

$$x_M' = \frac{2R \log_e 2}{\pi^2}. \tag{2.16}$$

Equation (2.16) is identical to the expression obtained by McKenzie (1967) for the half width of the heat flow anomaly on ridges. Thus there is a close relation between the temperature structure in the descending slab and that in the lithosphere near spreading ridges.

A remarkable relationship is now apparent between the depth of the deepest earthquakes and the temperature distribution within the sinking lithosphere. Fig. 4 shows a projection of the hypocentres of earthquakes and the isotherms beneath the Tonga–Fiji–Kermadec–New Zealand island arc onto a vertical plane approximately parallel to the arc. Only well located earthquakes were taken from Sykes (1966), and Hamilton & Gale (1968). The depths of the deepest point on the isotherms depends on both v_x and the angle ϕ between the dipping plane and the horizontal. v_x was obtained from plate theory, using 58° S, 168° E as the Australia–Pacific pole and an angular velocity of $12{\cdot}3 \times 10^{-7}$ degrees yr^{-1}. Le Pichon's (1968) pole at 52·2° S, 169·2° E was not used because it is not consistent with Sykes' (1967) fault plane solution on the Macquarie ridge. Substitution into equation (2.4) and (2.15) then gives x, the distance measured in the dipping plane, and hence the depth $x \sin \phi$. The correspondence between the deepest earthquakes and the isotherms is striking. Though the value of 680 °C for the limiting temperature is probably inaccurate, the variation with depth of the isotherms is not. Thus Fig. 4 clearly demonstrates such a limiting temperature exists if the present distribution of earthquakes is the steady state distribution. If the earthquakes are confined to those parts of the mantle below 680 °C, then they should not occur on a plane but within a thin slab never thicker than 50 km. Fig. 3 shows the shape of the 640 °C isotherm when $v_x = 10\,\mathrm{cm\,yr^{-1}}$ and $\phi = 45°$ without any vertical exaggeration, and illustrates the great length of the plate in comparison with its thickness. If the temperature is the only controlling variable, earthquakes should occur throughout this volume. Sykes' (1966) locations are probably not sufficiently accurate to determine whether the hypocentres occur throughout the slab, or are confined to a smaller volume within it. There is, however, now some evidence (Mitronovas *et al.* 1968) that earthquakes are restricted to a layer about 25 km thick.

The temperature within the slab, equation (2.10), may also be used to determine the resultant force/unit length that the slab exerts on the lithosphere at the surface. If this calculation is to be valid the density difference between the sinking lithosphere and the surrounding mantle must be caused by a difference in temperature and not in composition. This assumption is probably justified if the sinking slab was originally oceanic lithosphere, since all oceanic crust and upper mantle has been generated from the mantle below the plates. It is not justified for continental litho-



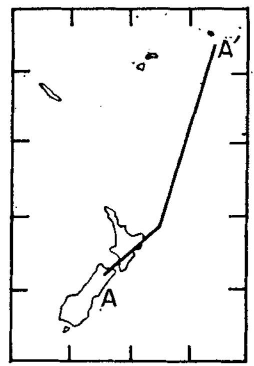


FIG. 4. Projections of foci of earthquakes and isotherms onto a vertical plane along the line A′ in the Tonga Fiji Kermadec New Zealand region (see inset). Except for six anomalous deep earthquakes, intermediate and deep focus activity ceases at about $T' = 0{\cdot}85$ or $T = 680$ °C. Accurate hypocentres only are taken from Sykes (1966), and Hatherton & Gale (1968). Numbers on the x axis refer to the position of Sykes' sections.

sphere which must be considered separately. The effect of the oceanic and continental crust on the force is considered in detail in Section 5. With this assumption the force/unit length F is given by:

$$F = \int_0^l \int_0^\infty g(\rho(T_1) - \rho(T))\, dx\, dz. \tag{2.17}$$

Substitution from equation (2.10) then gives:

$$F = 2g\alpha\rho T_1 l^2 \int_0^1 \int_0^\infty \sum_{n=1}^{\infty} \frac{(-1)^n}{n\pi} \exp\left[\left(R - (R^2 + n^2\pi^2)^{\frac{1}{2}}\right) x'\right] \sin n\pi z'\, dx'\, dz'. \tag{2.18}$$

Integration gives:

$$F = \frac{4g\alpha\rho T_1 l^2}{\pi^2} \sum_{k=0}^{\infty} \frac{1}{(2k+1)^2\left[R - (R^2 + (2k+1)^2\pi^2)^{\frac{1}{2}}\right]}. \tag{2.19}$$

Since $R \gg (2k+1)\pi$ for small k equation (2.19) may be written:

$$F \simeq \frac{8g\alpha\rho T_1 l^2 R}{\pi^4} \sum_{k=0}^{\infty} \frac{1}{(2k+1)^4}. \tag{2.20}$$

But from Jolley (1961)

$$\sum_{k=0}^{\infty} \frac{1}{(2k+1)^4} = \frac{\pi^4}{4 \times 4!}. \tag{2.21}$$

Thus equation (2.20) gives:

$$F = \frac{g\alpha\rho T_1 l^2 R}{12}. \tag{2.22}$$

This force is the total resultant force on the slab due to its thermal structure. Only the component $F \sin\phi$ acts down the dip of the slab, and can cause plate motions. The other component, $F\cos\phi$, must be balanced by the pressure distribution within the mantle outside the slab. The corresponding force/unit area f acting on the lithosphere in the direction of the island arc is:

$$f = \frac{F}{l} = \frac{g\alpha\rho T_1 l R}{12} \sin\phi. \tag{2.23}$$

This expression applies only if there is no resistance to the motion of the sinking slab. Equation (2.23) therefore gives the upper limit of f. Substitution from equation (1.2) with $v_x = 10\,\mathrm{cm\,yr^{-1}}$ gives:

$$f = 2{\cdot}5 \sin\phi \text{ kilobars.}$$

This value of f is surprisingly large, and since the numerical estimates of α and T_1 are probably too small, the true value may be even greater. Elsasser (1967) has suggested that this force maintains the observed surface motions, and this idea is discussed in detail in Section 4. Since the lithosphere is at present moving at constant velocity, the convective force f must be opposed by an equal force either on the base of the surface plates or on the sinking slab. If Elsasser's suggestion is correct, then at least in the uppermost part of the slab the axis of least principal stress must be subparallel to the dip of the plane. If this condition is not satisfied the stress system within the slab will oppose the motion and will not provide a driving force. Elsasser's idea can therefore be tested by comparing the fault plane solutions of intermediate and deep earthquakes with the predicted stress field. A necessary but not sufficient condition for his hypotheses to be true is the occurrence of tensional fault plane solutions at shallow depths in all slabs. Few focal mechanisms of intermediate earthquakes have yet been determined. Those from deep earthquakes require the slab to be in compression, with the axis of greatest principal stress approximately parallel to the dip of the slab (Isacks *et al.* 1968). Such a stress field can arise only if the deepest part of the slab is being pushed into the mantle. Thus at depths greater than about 300 km the resistance to the motion is greater than the thermal driving forces. It is not yet known whether intermediate earthquakes are produced by a stress field like that of deep earthquakes. If they are, then Elsasser's suggestion cannot be correct.

Though most intermediate and deep focus earthquakes occur within the cold sinking lithosphere, there exist also a few isolated earthquake sources which are less easily understood. Some sources, like those beneath Roumania and the Hindu Kush, produce a succession of earthquakes from a limited volume within the mantle. The most difficult sources to understand generate infrequent deep earthquakes, and perhaps the most striking of these is beneath eastern Spain. Only one earthquake, of magnitude 7 at a depth of 650 km, has ever been recorded from this source. Three deep earthquakes beneath North Island, New Zealand, in Fig. 4 are from another deep isolated source. All such special sources are beneath regions where crustal shortening has taken place in the Tertiary. It is therefore possible that the earthquakes are produced within relics of old slabs which are now no longer being regenerated by consumption of the lithosphere. The convective force, equation (2.23) generated by this density contrast is sufficient to produce earthquakes. If this explanation is correct, major changes in both the shape and relative motion between plates must have taken place in the second half of the Tertiary. Detailed geological studies of the deformation above isolated sources would therefore be of considerable interest. Such observations should provide a test of these ideas.

3. Flow within the mantle

The large plane slab which sinks beneath island arcs must set up stresses within the mantle through which it moves. Its motion will therefore govern the flow of the mantle if other forces are absent. Inside the earth there are also body forces, caused by horizontal density variations, which must strongly modify any flow produced by the sinking slab. Such thermal forces are neglected throughout this section, and all results must therefore be used with some caution. The other important assumption is that flow in the mantle outside the cold slab is governed by a viscous constitutional relationship (see Section 1). Since the stresses generated by the motion are less than 100 bars over large regions on both sides of the arc (Fig. 7), viscous flow governed by equation (3.1) probably resembles the real flow rather closely. It is much less likely that the temperature dependence of the viscosity can be neglected, as is done below. Rather important effects may well be governed by the strong dependence of viscosity on temperature. Thus several of these assumptions are unlikely to be satisfied by flow within the mantle, but without them non-linear terms dominate the equations and no analytic solutions can be obtained.

It is clear from these remarks that there is some doubt whether the resulting solutions apply to flow in the Earth. In this respect there is a considerable contrast between this section and the previous one. The motion of the slab is simple and also is rather well known from surface observations of spreading rates (Le Pichon 1968) and from the location of deep earthquakes (Sykes 1966). There is no similar method which can be used to observe the complicated three-dimensional flow within the mantle outside the slab. Even the geographic location of rising and sinking currents is in doubt. It is this extreme ignorance which makes the analysis below of at least some interest, however wrong the details may be.

Subject to these various assumptions the steady state velocity of an incompressible fluid must satisfy:

$$0 = \eta\nabla^2 \mathbf{v} + \rho\nabla U - \nabla P \tag{3.1}$$

$$0 = \nabla . \mathbf{v} \tag{3.2}$$

U is the gravitational potential and P the pressure. The curl of equation (3.1) gives:

$$\nabla^2 \boldsymbol{\omega} = 0 \tag{3.3}$$

where $\boldsymbol{\omega}(= \nabla \times \mathbf{v})$ is the vorticity. Most island arcs are approximately two-dimensional structures, and it is therefore convenient to use cylindrical co-ordinates with the z axis parallel to the arc. Fig. 5 shows a vertical section through an idealized arc, with the motions of the lithosphere and slab represented by the motion of planes. If the co-ordinate axes are fixed to ab, the lithosphere behind the arc, all boundary conditions are simple. Thus $\mathbf{v}$ may be written.

$$\mathbf{v} = \left(v_z, \frac{1}{r}\frac{\partial\psi}{\partial\theta}, -\frac{\partial\psi}{\partial r}\right) \tag{3.4}$$

where ψ is the stream function.

The resulting equations are simple if $v_z = 0$ everywhere, a condition which requires the motion between ab and bc in Fig. 5 to be normal to the arc. This condition is not satisfied by all presently active island arcs (McKenzie & Parker 1967; Le Pichon 1968), and solutions to equation (3.4) can if necessary be obtained if $v_z \neq 0$. It is, however, doubtful if these general solutions would display any features which are not possessed by the special case discussed below. Equation (3.3) then becomes:

$$\nabla^4 \psi = 0 \tag{3.5}$$

Solutions to equation (3.5) are required which satisfy $\mathbf{v} = \mathbf{a}_r \times$constant, where $\mathbf{a}_r$ is the radial unit vector, at specified values of θ (Fig. 5). Such solutions are easily obtained by substituting (Batchelor 1967).

$$\psi = r\Theta(\theta) \tag{3.6}$$

into equation (3.5) to give:

$$\frac{d^4\Theta}{d\theta^4} + 2\frac{d^2\Theta}{d\theta^2} + \Theta = 0. \tag{3.7}$$

The general solution to equation (3.7) is:

$$\Theta = A \sin\theta + B\cos\theta + C\theta\sin\theta + D\theta\cos\theta \tag{3.8}$$

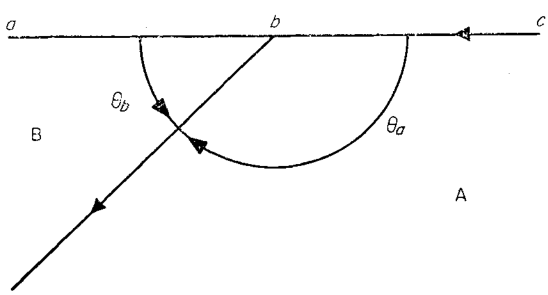


Fig. 5.

where A, B, C and D are constants which must be obtained from the boundary conditions, and are not the same on each side of the island arc. The boundary conditions within region A (Fig. 5) are:

$$\left.\begin{aligned} \mathbf{v} &= -v\mathbf{a}_r \quad \text{on} \quad \theta = 0 \\ \mathbf{v} &= v\mathbf{a}_r \quad \text{on} \quad \theta = \theta_a. \end{aligned}\right\} \tag{3.9}$$

These conditions are satisfied by:

$$\psi = \frac{-rv[(\theta_a-\theta)\sin\theta+\theta\sin(\theta_a-\theta)]}{\theta_a+\sin\theta_a}. \tag{3.10}$$

Similarly the flow within B must satisfy:

$$\left.\begin{aligned} \mathbf{v} &= 0 \quad \text{on} \quad \theta = 0 \\ \mathbf{v} &= v\mathbf{a}_r \quad \text{on} \quad \theta = \theta_b \end{aligned}\right\} \tag{3.11}$$

giving:

$$\psi = \frac{rv[(\theta_b-\theta)\sin\theta_b\sin\theta-\theta_b\theta\sin(\theta_b-\theta)]}{\theta_b^2-\sin^2\theta_b}. \tag{3.12}$$

θ_a in equation (3.10) is measured clockwise from bc in Fig. 5, with bc as zero. θ_b in equation (3.12) is measured anticlockwise from ab as zero. This sign convention simplifies the expressions and is permitted because it is possible to choose a right-handed set of co-ordinate axes in both cases.

Equations (3.9) and (3.11) are not rigorously satisfied in real island arcs because both the surface plates and the sinking slab are finite. This limitation is, however, unlikely to produce important differences in the flow close to the island arc.

The stream lines of constant ψ are easily obtained from equations (3.10) and (3.12) (Fig. 6). Throughout this discussion θ_a and θ_b are chosen to be 135° and 45° respectively. The stream lines demonstrate that the fluid is dragged down with the sinking slab, as indeed would be expected from rather simpler arguments. It is perhaps more surprising that fluid is swept upward in certain parts of region B.

The stresses which the fluid flow produces are also of considerable interest. Since the flow is two-dimensional, the stress tensor $\mathbf{S}$ may be written as:

$$\mathbf{S} = \begin{bmatrix} \sigma'_{rr} & \sigma'_{r\theta} \\ \sigma'_{r\theta} & \sigma'_{\theta\theta} \end{bmatrix} - P\mathbf{1} \tag{3.13}$$

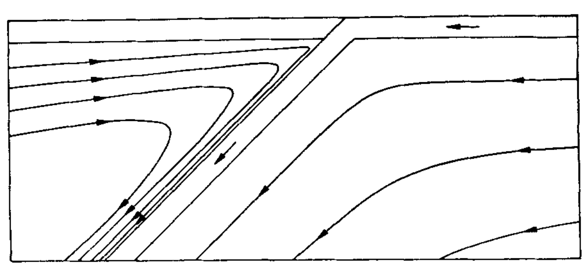


FIG. 6. Stream lines for flow within the mantle. Motion is with respect to the plate behind the island arc, and is driven by the motion of the other plate and of the sinking slab. Thermal convection outside the slab is neglected, and the lithosphere is 50 km thick.

where **I** is a 2×2 unit matrix and:

$$\left.\begin{aligned} \sigma'_{rr} &= 2\eta \frac{\partial v_r}{\partial r} \\ \sigma'_{\theta\theta} &= 2\eta \left(\frac{1}{r}\frac{\partial v_\theta}{\partial \theta} + \frac{v_r}{r}\right) \\ \sigma'_{r\theta} &= \eta \left(\frac{1}{r}\frac{\partial v_r}{\partial \theta} + \frac{\partial v_\theta}{\partial r} - \frac{v_\theta}{r}\right). \end{aligned}\right\} \tag{3.14}$$

Equations (3.4) and (3.6) then give:

$$\left.\begin{aligned} \sigma'_{rr} &= \sigma'_{\theta\theta} = 0 \\ \sigma'_{r\theta} &= \frac{\eta}{r}\left(\frac{d^2\,\Theta}{d\theta^2} + \Theta\right). \end{aligned}\right\} \tag{3.15}$$

Thus:

$$\mathbf{S} + P\mathbf{I} = \begin{bmatrix} 0 & \sigma'_{r\theta} \\ \sigma'_{r\theta} & 0 \end{bmatrix} = \mathbf{S}' \tag{3.16}$$

and forms a tensor S′ without diagonal terms. Thus the greatest shearing stress is exerted parallel to the r = const. and θ = const. surfaces, and is of magnitude $\sigma'_{r\theta}$. Substitution of equations (3.10) and (3.12) into equation (3.15) gives:

$$\sigma'_{r\theta} = \frac{2v\eta}{r(\theta_a + \sin\theta_a)}\,[\cos\theta_a + \cos(\theta_a - \theta)] \tag{3.17}$$

in A, and:

$$\sigma'_{r\theta} = \frac{2v\eta\,[\theta_b \cos(\theta_b - \theta) - \sin\theta_b \cos\theta]}{r(\theta_b{}^2 - \sin^2\theta_b)} \tag{3.18}$$

in B. Both equations (3.17) and (3.18) are singular at the origin where $r = 0$. The singularities arise because the spreading lithosphere is required by the boundary conditions (3.9) and (3.11) to bend through an angle θ_b at the origin $r = 0$. Since the radius of curvature of the lithosphere cannot in reality be zero, these singularities are not in practice possible. Thus equations (3.17) and (3.18) are probably good approximations to the stress field except perhaps within 50 km of $r = 0$ axis, and are sufficient for a qualitative discussion. Fig. 7 shows the contours of the shearing stress obtained from equations (3.17) and (3.18) with values for v and η taken from equation (1.2). The half arrows show the direction of the stresses exerted by the fluid on the rigid boundaries. The stresses in region B behind the arc are much greater than those in A. Within B there are two zones of high stress separated by a plane of zero stress. In both zones stresses exceed 100 bars and therefore the viscous constitutional relationship between stress and strain rate does not apply. The occurrence of shallow earthquakes at considerable distances inside the island arc may be related to the shallower of these zones (Hamilton & Gale 1968). Earthquakes also occur within the deeper zone, though their frequency decreases rapidly with depth (Sykes 1966). This observation agrees with the stress distribution in Fig. 7.

The stresses on both sides of the arc act in a direction which opposes the motion of the lithosphere. Thus work must be done by the mechanism which moves the plates against the viscous forces. The mechanical energy is then converted into heat by viscous dissipation within the mantle. Such stress heating is one source of energy

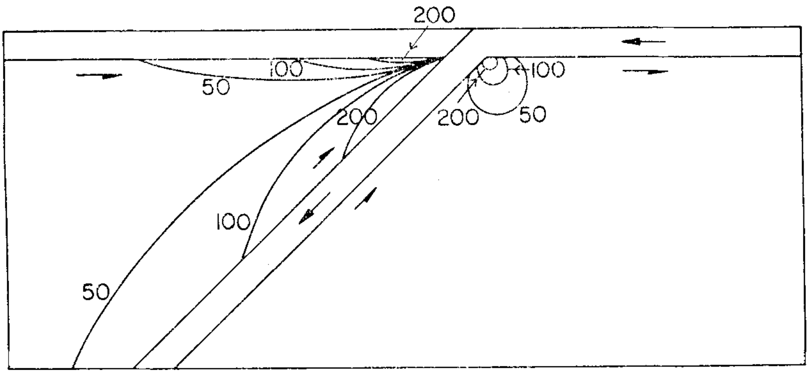


FIG. 7. Shear stresses in bars caused by the flow in Fig. 6. The half arrows show the direction of the forces exerted by the fluid on the plates and slab. The stresses on the plate behind the island arc exceed those on the plate in front. The lithosphere is 50 km thick.

which can maintain the large positive heat flow anomaly behind island arcs. McKenzie & Sclater (1968) attempted to produce the heat required from the shearing stresses between the sinking slab and the surrounding mantle. However, they were unable to transport the heat to the Earth's surface in a geologically reasonable time. This difficulty is avoided if the dissipation occurs within the shallow high stress zone in Fig. 7.

The expression for the stress heating H in a non-elastic material is:

$$H = S_{ij}\frac{\partial v_i}{\partial x_j}. \tag{3.19}$$

S_{ij} is the stress tensor in equation (3.13), and summation over repeated indices is implied. Equation (3.19) may also be written:

$$H = \tfrac{1}{2}(S'_{ij} - P\delta_{ij})\left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i}\right) \tag{3.20}$$

where $\mathbf{S}'$ is given by equation (3.16) and $\delta_{ij} = 0$ if $i \neq j$, $= 1$ if $i = j$. Equations (3.19) and (3.20) give the heat generated by friction in any material. In this problem the mantle is taken to be a viscous incompressible fluid, and therefore:

$$S'_{ij} = \eta\left(\frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i}\right) \tag{3.21}$$

and:

$$S'_{ii} = 0. \tag{3.22}$$

Equation (3.20) may be written:

$$\begin{aligned} H &= \frac{1}{2\eta}(S'_{ij} - P\delta_{ij})\,S'_{ij} \\ &= \frac{1}{2\eta}S'_{ij}S'_{ij}. \end{aligned} \tag{3.23}$$

$S'_{ij}S'_{ij}$ is an invariant of the stress tensor, and is therefore unaffected by the choice of axes. Thus equation (3.16) may be substituted directly into equation (3.23) to give:

$$H = \frac{1}{\eta}\,\sigma'^2_{r\theta}. \tag{3.24}$$

The contours of the stress heating function are therefore the same as those of the shearing stress. The values on the contours in Fig. 8 are in units of 10^{-7} erg cm^{-3} s^{-1}, and are obtained from equations (1.2), (3.17), (3.18) and (3.24). A typical value for the radioactive heat generation rate for granites is 200×10^{-7} erg cm^{-3} s^{-1}, for basalts is 20×10^{-7} erg cm^{-3} s^{-1}, and for ultrabasic rocks is 2×10^{-7} erg cm^{-3} s^{-1}. The excess heat flow behind island arcs is $\sim 1\,\mu$ cal cm^{-2} s^{-1}, and is therefore equivalent to a volume source of 80×10^{-7} erg cm^{-3} s^{-1} distributed through a depth of 50 km. This value is an order of magnitude greater than that calculated from equation (3.24). Since the viscosity is uncertain by perhaps an order of magnitude, the value of the stress heating could also be an order of magnitude different from that in Fig. 8. Thus stress heating could account for the observed high heat flow behind island arcs. Since there is considerable heat generation at shallow depths, the heat can diffuse to the surface in a geologically reasonable time. Fig. 8 shows that heat is also generated by friction between the sinking slab and the surrounding mantle. McKenzie & Sclater (1968) attempted to explain the observations by conducting this heat upwards, but were unable to do so in less than about 300 million years. No such objection applies to the suggestion above because the heat is generated by shearing at shallow depths.

Stress heating can produce the observed surface heat flow only if the viscosity in equation (1.2) is too small. If the value of η used above is too large, viscous heating is quite unable to generate the heat required. Figs 6 and 7 suggest two other possible explanations of the heat flow observations. The first depends on the shear stress on the base of the lithosphere above region B. The effective thickness of the lithosphere depends on the shearing stresses on its base; if these are large the lower part will be dragged away by the flow. Hot mantle material will be carried closer to the Earth's surface, and the heat flow will therefore be increased. Thus the temperature gradients in the thermal boundary layer at the Earth's surface must depend on the stress field.

A second effect can increase the temperature gradient at the base of the lithosphere. The fluid behind the island arc in Fig. 6 is carried towards the surface before being dragged down into the mantle by the plate motions. The flow therefore carries

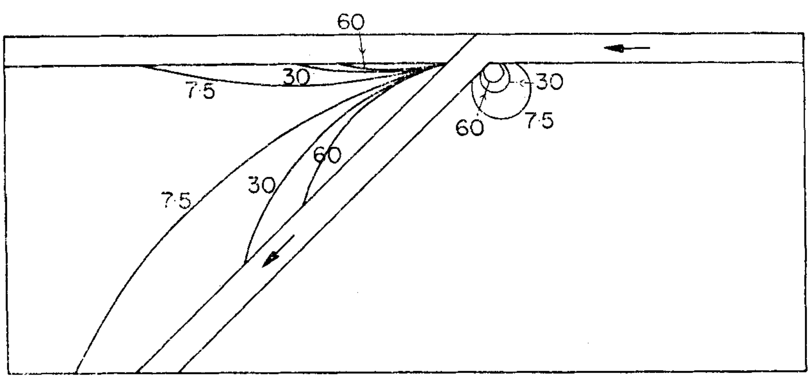


FIG. 8. Stress heating caused by viscous dissipation within the flow in Fig. 6 in units of 10^{-7} erg cm^{-3} s^{-1}. The heating within the mantle is more intense behind than in front of the arc.

hot mantle material closer to the base of the lithosphere, and hence can produce a surface heat flow anomaly. All three possible explanations may well be involved since they each can produce an increase in the surface heat flow.

All the equations above were solved without any boundary conditions on the stress or displacement normal to the lithosphere. If the displacement is taken to be zero, the normal stress must be balanced by forces within the lithosphere. The alternative boundary condition is to require the normal stress to be zero at the deformed surface. The equations then determine the surface shape. The strength of the lithosphere probably cannot support the normal stress in this problem, because of the large horizontal extent of the forces (McKenzie 1967). Therefore the deformation is calculated from the condition that the normal stress must vanish. The stress normal to any plane r = const. is:

$$\sigma_{\theta\theta} = -P + 2\eta\left(\frac{1}{r}\frac{\partial v_\theta}{\partial \theta} + \frac{v_r}{r}\right). \tag{3.25}$$

Equations (3.4) and (3.6) then give:

$$\sigma_{\theta\theta} = -P. \tag{3.26}$$

If the fluid is at rest the hydrostatic pressure P_0 at any point is:

$$P_0 = \rho g r \sin\theta. \tag{3.27}$$

The pressure P is perturbed by an amount P_1 from P_0 by the shear stresses of the flow. P_1 may be obtained from equation (3.1) by perturbation theory:

$$P_1 = -\frac{\eta}{r}\left(\frac{d^3\,\Theta}{d\theta^3} + \frac{d\Theta}{d\theta}\right). \tag{3.28}$$

If the normal stress is to vanish on the deformed surface:

$$\sigma_{\theta\theta} = 0 \tag{3.29}$$

or:

$$\rho g r \sin\theta = \frac{\eta}{r}\left(\frac{d^3\,\Theta}{d\theta^3} + \frac{d\Theta}{d\theta}\right). \tag{3.30}$$

Equation (3.30) determines $\theta = \theta(r)$ on the surface, provided the angle between the surface and the radius vector is small, and provided the fluid flow is not affected. Neither condition is satisfied close to the origin of co-ordinates. In front of the arc in region A equation (3.30) gives:

$$\tan\theta = \frac{\sin\theta_a}{\left[1 + \cos\theta_a + \dfrac{\rho g}{2\eta v} r^2(\theta_a + \sin\theta_a)\right]} \tag{3.31}$$

and behind:

$$\tan\theta = \frac{\theta_b \sin\theta_b}{\left[\theta_b \cos\theta_b - \sin\theta_b + \dfrac{\rho g}{2\eta v} r^2({\theta_b}^2 - \sin^2\theta_b)\right]}. \tag{3.32}$$

Fig. 9 shows the surface shape with a $\times 10$ vertical exaggeration. The surface inside the island arc is depressed by several kilometres, whereas outside the arc, above region A, the deformation is much less. Therefore trenches are unlikely to be maintained by flow in the mantle, but are probably supported by the strength of the lithosphere (McKenzie 1967). Depression within the arc has not been observed. If it does exist and is full of sediment it could be of considerable geological importance.

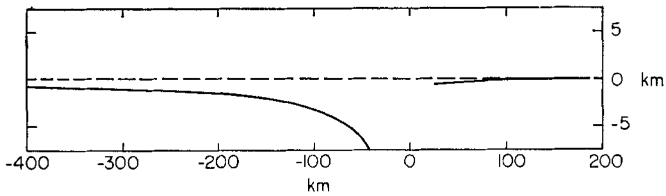


FIG. 9. Surface deformation near trenches produced by stresses due to the flow and normal to the Earth's surface. ×10 vertical exaggeration. Horizontal distances measured from the surface boundary between the plates, positive in front of the arc and negative behind (as in Fig. 3).

Though some of the results obtained in this section are probably of relevance to the flow beneath island arcs, the drastic approximations required to linearize the equations must not be forgotten.

4. The causes of plate motion

Three causes of plate motions have been suggested. The oldest theory depends on large-scale convection throughout at least the upper mantle. Viscous forces are then required to couple the plates to the moving mantle below. The history of this idea is discussed by Holmes (1965), and a modified version is compatible with all relevant observations (McKenzie 1968a). Another closely related theory has been put forward by Elsasser (1967). He suggested that the lithosphere may be considered as a stress guide, and that surface motions are maintained by the cold sinking slabs (see Section 2). At first sight it appears impossible to move the entire lithosphere by tensional forces because rocks have little strength in tension. However, this objection is not correct because materials fail in tension easily only if voids can form. Since void formation requires one of the principal stresses to be negative, this type of failure must be restricted to the rock within a few hundred metres of the surface, and does not affect the dynamics of the lithosphere. The energy source of Elsasser's mechanism is still thermal convection, but the force causing the motion is transmitted through the rigid lithosphere, and not through viscous coupling between the mantle and the plates.

A third mechanism which can maintain motions is the force exerted on the plate boundaries by neighbouring plates. In particular the motion of small plates is probably sustained by such forces. Two examples are the plate between the Juan da Fuca ridge and the Continental margin of northwestern America, and that containing the Aegean Sea in the Mediterranean. This mechanism cannot produce the world-wide movements of large plates because it has no source of energy.

It is easy to demonstrate by energetic arguments that thermal convection must provide the driving force, and therefore that vertical movements of hot and cold material must be involved. The difference between Elsasser's model for thermal convection and those of previous authors is in the importance of the transmission of stress through the upper mantle. He suggested that the cold and therefore dense slab of lithosphere sinks into the mantle beneath island arcs and pulls with it the rest of the plate to which it is attached. Buoyancy forces throughout the mantle outside the slab are neglected. Thus this model is an extreme form of convection which takes place if the viscosity is a rapidly varying function of temperature. If the flow induced within the mantle by a sinking slab affects the motion of plates other than that to which it is attached, then Elsasser's model is too simple. Fortunately his assumptions have certain consequences which can be obtained from the analysis above and are not in agreement with observations.

The force/cm exerted by the sinking lithosphere on the surface plate is given by equations (2.4) and (2.22):

$$F = \frac{g\alpha\rho^2 T_1 C_p l^3}{24\kappa} v. \tag{4.1}$$

This force is directed towards the island arc and must overcome the frictional resistance between the sinking slab and the mantle, and also that between the surface plate and the mantle. The first of these cannot be obtained from equations (3.17) and (3.18) by integration because the stress is singular at $r = 0$. The singularity is not of physical importance, and therefore the resistive force/unit length of the arc R_1 may be written:

$$R_1 = \eta v f(\theta_b). \tag{4.2}$$

R_1 acts on the sinking slab in a direction opposite to its motion. The force resisting the motion of the surface plate depends on the variation of viscosity with depth. A simple model has an upper mantle of constant viscosity to a depth d, bounded by a rigid boundary below. The resistance/unit length R_2 acting on the plate is therefore:

$$R_2 = \eta v \frac{L}{d} \tag{4.3}$$

where L is the length of the plate measured at right angles to the island arc. If the motion of the plates is driven by F in equation (4.1), then:

$$F > R_1 + R_2$$

or

$$\frac{g\alpha\rho T_1 C_p l^3}{\kappa\nu} > 24\left(\frac{L}{d} + f(\theta_b)\right). \tag{4.4}$$

The inequality (4.4) is very similar to Rayleigh's condition for convection in a fluid heated from below (see Saltzman 1962). All parameters on the left of the inequality (4.4) and also d and θ_b ($\simeq 45°$) are the same for all plates. Thus only L varies between plates. Elsasser's suggestion could be correct if the inequality is satisfied by $L = 0$. There must then be some limiting value of L, L_c, perhaps very large, at which the inequality reverses. If the inequality (4.4) is satisfied, then v is determined only by the variation of viscosity with depth. Thus this type of convection will possess some rather remarkable features. All plates with $L < L_c$ and which have a sinking slab will spread at a rate limited only by the variation of viscosity within the mantle. Large plates cannot move faster than small ones, and plates with $L > L_c$ cannot move at all.

The motion of real plates is quite different. Various small plates with sinking slabs attached are known; an example is the one between the East Pacific Rise, the Galapagos Rift and the Middle America Trench. The consumption rate along the trench may be estimated from the depth of the intermediate earthquakes beneath Middle America (Gutenberg & Richter 1954) to be about 3 cm yr^{-1}. This value is three times smaller than that of 9 cm yr^{-1} for the Western Pacific arcs (Le Pichon 1968). These arcs occur between three of the largest plates, containing India, America and the Pacific respectively. These observations suggest that the consumption rate increases with plate size. They do not demonstrate that the velocity depends only on the viscosity of the mantle, and therefore do not support Elsasser's suggestion.

It could be argued that these remarks are incorrect because they depend on the somewhat uncertain calculations of Sections 2 and 3. Though there is no reason to doubt the arguments already given, there is one other which does not depend on any analysis. If plate motions are maintained in the way Elsasser has suggested, spreading is not possible between two plates, neither of which possesses a sinking slab. The American and the African plate almost satisfy this condition. The only sinking parts of the American plate are beneath the Eastern Caribbean and the Scotia arc, both of which are small. Descending parts of the African plate are restricted to the Mediterranean, beneath the Aegean and Tyrrhenian Seas. The motion between these two plates is not small, reaching $3\,\mathrm{cm\,yr^{-1}}$ in the South Atlantic. Elsasser's hypothesis can account for this result only if stress transmitted through the mantle can cause drift. This modification is of great importance, because the equations governing such flow are those of convection in a fluid of variable viscosity. Such equations have several important non-linear terms, and must be solved by numerical methods.

Convection currents in the mantle affect plate motions in two ways. They drag the lithosphere in the direction of the flow. They also distort the surface of the Earth, and therefore gravity acts to pull the plates downhill. If the lithosphere is sufficiently thick, the gravitational force is greater than the viscous stress on the base of the plate. Under these conditions the plate motions are dominated by gravity. In oceanic regions sea level is within about 2 m of the level surface (Stommel 1966), and the structure of the plate remains the same over large regions. Thus the driving force on the plates can be obtained from the slope of the regional bathymetry, even if the bathymetry is itself maintained by convective motions within the mantle. In both the Pacific and the Indian Oceans the ocean floor slopes down towards the trenches, and therefore appears to support these arguments. Unfortunately the more detailed discussion below suggests that the viscous stress rather than gravity governs the motion of the largest plates, though in general both must be considered.

A useful model for convection in the mantle was suggested by Allan, Thompson & Weiss (1967). In this model the flow is two-dimensional, and is driven by horizontal temperature differences applied to the upper surface of a semi-infinite viscous fluid. If convection of heat is neglected, analytic expressions may be obtained for the surface deformation and the shearing stress exerted on a rigid boundary. Though the model is undoubtedly too simple, it is probably sufficient for order of magnitude calculations. Values for the gravitational and viscous forces on a plate are easily obtained from the expressions derived by McKenzie (1968a). If the surface of the fluid is maintained at a temperature T:

$$T = T_0 \cos kx \tag{4.5}$$

where T_0 is the amplitude and k the wavenumber of the applied temperature driving the flow. The difference z between the surface of the fluid and a level surface is given by:

$$z = \frac{3\alpha T_0}{4k} \cos kx. \tag{4.6}$$

The surface slope is therefore:

$$\frac{dz}{dx} = -\frac{3\alpha T_0}{4} \sin kx. \tag{4.7}$$

Provided $|dz/dx| \ll 1$ the horizontal force cm^{-3} f_x is:

$$f_x = \frac{3\alpha T_0}{4} \rho g \sin kx \tag{4.8}$$

and the effective force cm^{-2} F_x acting on the lithosphere of thickness l is:

$$F_x = \frac{3\alpha T_0}{4} g\rho l \sin kx. \tag{4.9}$$

The shearing stress σ_{xz} exerted by the flow is:

$$\sigma_{xz} = \frac{\alpha T_0}{4k} g\rho \sin kx. \tag{4.10}$$

(In the expression (3.25) given by McKenzie (1968a) the factor ρ was inadvertently omitted.)

Thus:

$$\frac{\sigma_{xz}}{F_x} = \frac{1}{3lk} = \frac{\lambda}{6\pi l}$$

where λ is the wavelength of the temperature disturbance. If $\lambda = 5000$ km, substitution from equation (1.2) gives:

$$\frac{\sigma_{xz}}{F_x} \simeq 7. \tag{4.12}$$

Thus the viscous shearing stress cannot be neglected, and may well dominate the motion of large plates.

The arguments above all suggest that the forces driving plates can be understood only if solutions to the full convection equations are obtained. Solutions to simplified models appear to give only limited insight into the general problem, and also to be internally inconsistent. Numerical solutions to the equations governing convection must be obtained for a liquid whose viscosity varies rapidly with temperature. Flow patterns within the Earth can then probably be determined by combining these solutions with the non-hydrostatic gravity field, though the relation between the two is not simple. The rising and sinking parts of convection currents are unlikely to be reflected in the surface features (McKenzie 1967). Thus there is no reason why elevations and depressions of the geoid should occur in the neighbourhood of trenches and ridges, as Runcorn (1965) has suggested.

Probably the most accurate determination of the external gravity field yet published is that of Gaposhkin (see Kaula (1966)). The corresponding geoid referred to the hydrostatic figure of the Earth (Fig. 10) shows that the non-hydrostatic equatorial bulge dominates other non-hydrostatic terms. At present there is no general agreement about how the non-hydrostatic bulge is maintained (MacDonald 1963; McKenzie 1966; Goldreich & Toomre 1969), and therefore the geoid is commonly referred to a spheroid with the observed, rather than with the hydrostatic, ellipticity. Figs 11 and 12 show such a geoid calculated from Gaposhkin's gravity field, and also the major ridges and island arcs. The ridges occur both where the geoid is elevated and where it is depressed. Therefore ridges must exist over both rising and sinking currents. This poor correlation is to be expected for the reasons discussed above. In contrast major trenches and island arcs round the Pacific occur in regions where the geoid is elevated. Unfortunately there are rather few such island arcs, and therefore this correlation need not be significant. There is, however, a physical reason why island arcs, unlike ridges, might be closely related to flow deep within the mantle. The creation of lithosphere along ridges requires a large volume of hot mantle material to be intruded along the ridge axis. This rock presumably rises adiabatically from beneath the lithosphere, then cools as the plate moves away from the ridge. Since the temperature gradient in the mantle must be close to the adiabatic gradient, plate creation along ridges requires large volumes of mantle material, but, except within the lithosphere, convects little heat. Thus plate production produces little distortion of the isotherms within the mantle beneath the lithosphere. The opposite is true beneath island arcs. No change in the temperature structure within the lithosphere can take place until it starts to descend into the mantle. Thus the motion distorts the isotherms by hundreds of kilometres throughout the upper mantle. In

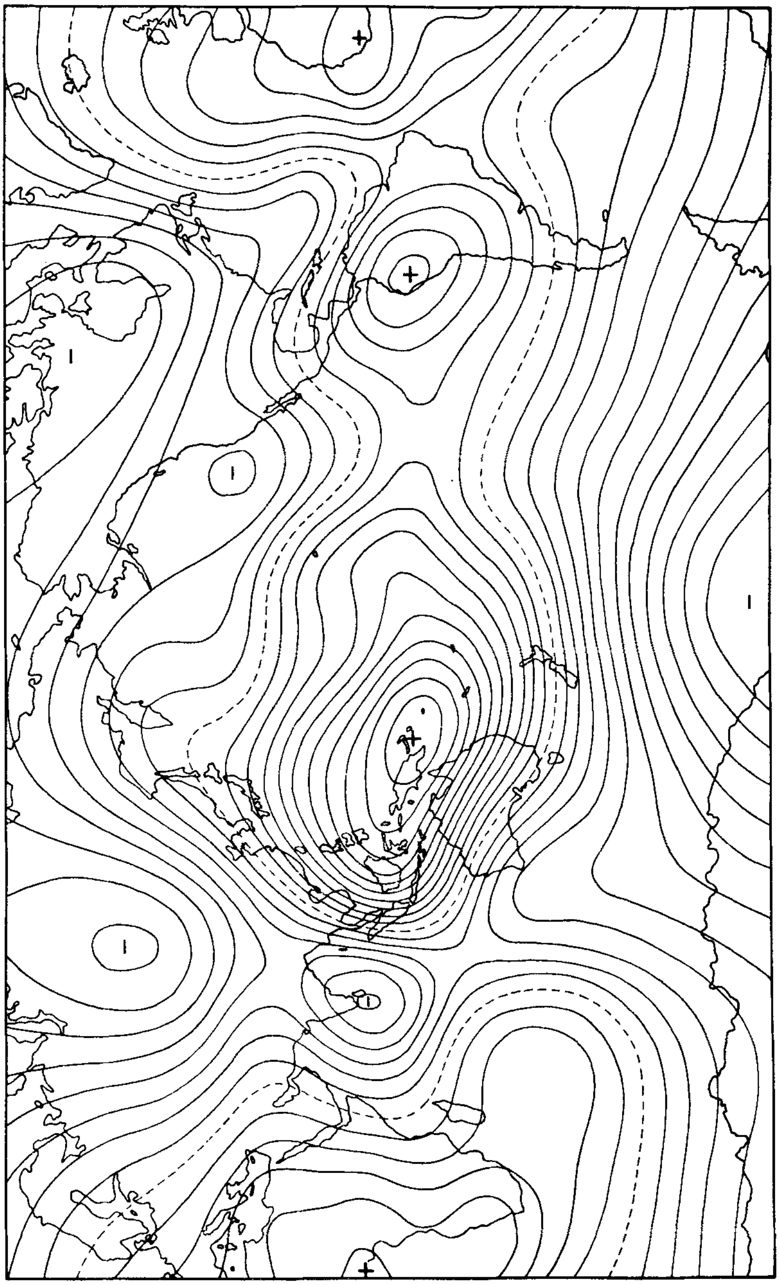


Fig. 10. Contours at intervals of 10 m of the elevation of the geoid above the hydrostatic geoid given by Jeffreys (1963). The dashed contour is of zero elevation, positive and negative signs shows elevations and depressions of the level surface. Mercator projection with the rotation axis as axis.

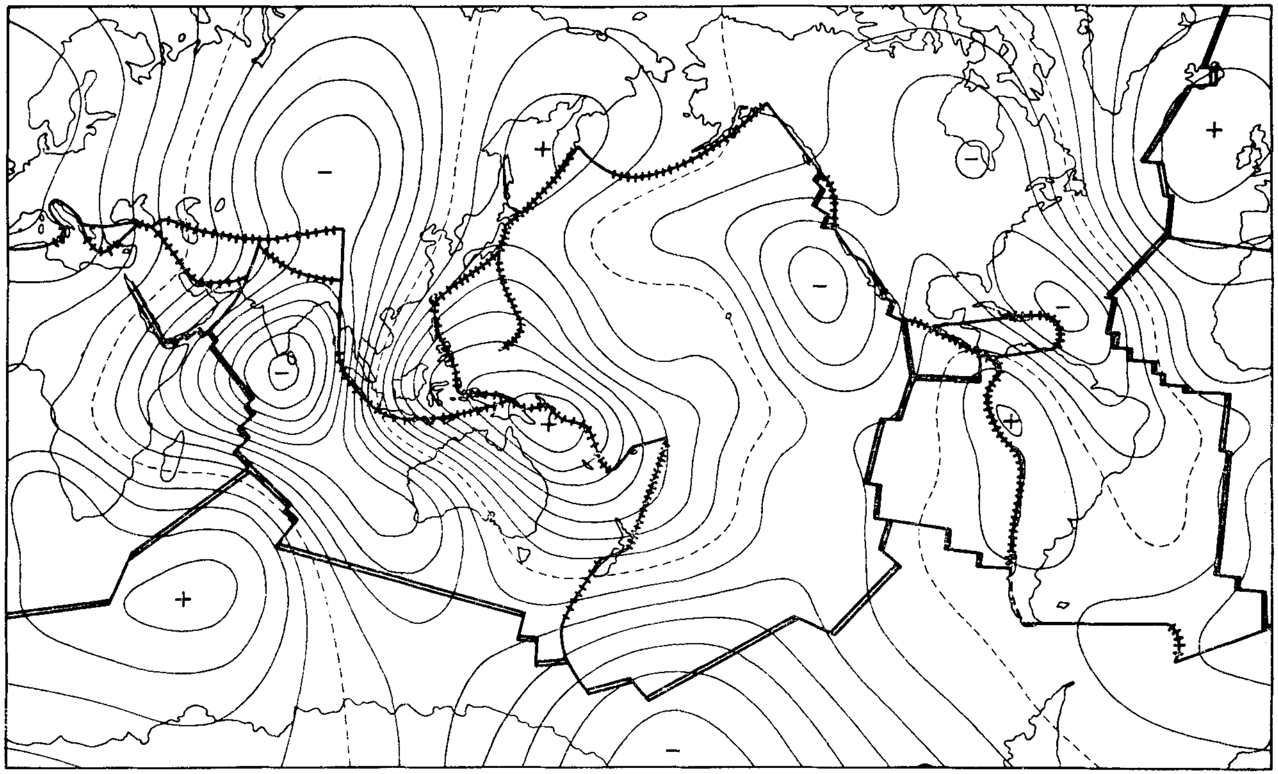


FIG. 11. Contours of the height of the non-hydrostatic geoid with respect to a geoid with the observed value of C_{20}, and not the hydrostatic value. Contours as in Fig. 10. The approximate position and nature of major plate boundaries are also shown. Major trenches all occur in regions where the geoid is above the reference surface. The opposite is not true for major ridges. Mercator projection with the rotation axis as axis. ┼┼┼┼, trench; ═══, ridge; —— transform fault.

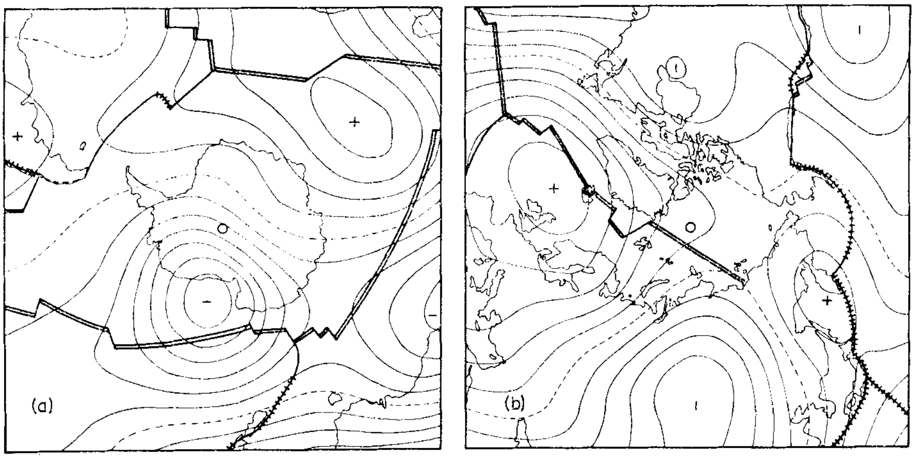


FIG. 12. The south (a) and north (b) polar regions of Fig. 11. The poles are shown as small circles. Azimuthal equidistant projections centred on the South (a) and North (b) Poles.

the absence of surface deformation the cold slab will produce an elevation in the geoid of about the observed magnitude. Thus the mantle below the lithosphere loses material beneath ridges, but loses heat beneath island arcs. Clearly this balance must be the exact opposite of that of the lithosphere. The world-wide heat loss due to plate creation on ridges (McKenzie & Sclater 1969) is therefore equal to the rate of heat loss by the mantle beneath island arcs. This mechanism of heat transfer probably accounts for more than 15 per cent of the heat lost by the mantle beneath the lithosphere. The cold descending slab produces large horizontal, as well as vertical, temperature gradients, and should therefore govern the position of the descending limb of any convection cells. Control of convection by horizontal temperature gradients maintained by the boundary conditions has been widely studied in the Earth's oceans and atmosphere. This effect is not related to flow caused by the transmission of stress through the mantle (see Section 3) though the direction of flow is the same.

There is a considerable difference between the convection hypotheses suggested here and those of Vening Meinesz (1962), Runcorn (1965) and others. For reasons already discussed, it is believed that sinking currents occur beneath island arcs because cold lithosphere is being thrust downwards into the mantle. If the line where underthrusting takes place moves, then so must the sinking current. The arc is not the consequence but the cause of the horizontal convergence beneath the lithosphere. This separation between cause and effect is of somewhat limited value, however, because the mantle and lithosphere are in thermal and mechanical contact, and therefore affect each others' motions.

The discussion in this section suggests that all detailed discussions of the convection problem so far published are too simple to apply to the Earth. However it does appear that there is now enough understanding of the mechanical and thermal properties of the crust and upper mantle for the convection problem to be posed correctly, and perhaps even solved.

5. Continental tectonics

All early theories of continental drift required strong rigid continents to smash their way through the oceans, in much the same way as an ice-breaker moves through ice flows. This is now known not to happen. The original ideas have been greatly modified (see Section 1), and in plate theory there is no difference between the motion of continents and oceans. There are, however, some striking differences between the geological and tectonic history of oceanic and continental rocks. Oceanic crust is created and destroyed rather rapidly, and the crust beneath most of the world's oceans is probably less than 200 million years old. Plate boundaries are sharp in oceanic areas, and the earthquake epicentres are restricted to a narrow belt seldom wider than the errors in location (Barazangi & Dorman 1969). Focal mechanisms of earthquakes on plate boundaries are determined by the motion of the plates alone, and are rarely complicated by the presence of small intervening plates. Intermediate and deep focus earthquakes occur only where plates are destroyed, and if the plate being consumed is oceanic there is a simple relationship between the consumption rate and the depth of the deepest earthquakes (see Section 2).

Continental crust behaves quite differently. Large areas were formed at least 1000 million years ago and have often been deformed by several orogenies. Plate boundaries within continents are rarely narrow, most commonly they are diffuse zones of seismic activity. Indeed there are few aseismic regions within the continents. Fault plane solutions for earthquakes reflect the complications and are not simply related to the motions of major plates because the boundaries consist of many small blocks in relative motion. The continental equivalents of island arcs are regions of active mountain building like Persia and the Himalaya (Le Pichon 1968). In such regions deep earthquakes are absent and intermediate sources are infrequent and localized, like that beneath Roumania or the Hindu Kush. These sources were discussed in Section 2 and could be relics of old oceanic lithosphere.

These differences are to be expected if the continental lithosphere cannot descend into the mantle, and also if it is more easily deformed than the oceanic lithosphere. Both properties are consequences of the nature and composition of continental rocks.

The buoyancy forces produced by the density contrast between the oceanic crust and the mantle were neglected in Section 2. The influence of such forces on the motion of the sinking slab is not easily determined because the composition of the oceanic crust is still uncertain, and because part of the crust will probably be converted to eclogite as it sinks. Since the density of eclogite is greater than that of peridotite, the presence of the oceanic crust could increase the downward force on the sinking slab. A simple calculation suggests that the same is not true for continental crust, which is probably sufficiently thick and light to prevent consumption. Before the lithosphere descends into the mantle the temperature within it is given by equation (2.8). Thus the maximum downward buoyancy force cm^{-2} that can be produced by the temperature difference between it and the surrounding mantle is:

$$g \int_0^l \rho\big(1-\alpha T-(1-\alpha T_1)\big)\, dz = \frac{\rho g \alpha T_1 l}{2}. \tag{5.1}$$

If the density of the crust is ρ_c, the total buoyancy force will oppose the motion of the sinking slab unless:

$$\frac{\rho \alpha T_1 l}{2} \geqslant (\rho-\rho_c)\, gd, \tag{5.2}$$

where d is the crustal thickness. If $\rho_c = 2{\cdot}7\,\mathrm{g\,cm^{-3}}$, substitution from equation (1.2) into (5.2) gives:

$$d \leqslant 4{\cdot}5\,\mathrm{km}. \tag{5.3}$$

This value is comparable to the oceanic crustal thickness, but is much less than that of 30 km often observed beneath continents. Thus oceanic lithosphere can overcome the upward buoyancy due to its less dense crust, especially if there are phase changes to eclogite as the crust sinks. The same is not true of continental lithosphere. The upper granitic part is perhaps 15 km thick and has a density of $\sim 2{\cdot}5\ \mathrm{g\,cm^{-3}}$. The lower part is probably gabbro, which has a density of $2{\cdot}8\ \mathrm{g\,cm^{-3}}$. Therefore if continental lithosphere is thrust into the mantle buoyancy forces will oppose the motion. Thus continents once formed are very difficult to destroy. If an island arc attempts to consume a continent large stresses will be generated throughout the lithosphere on both sides. Changes in the boundaries of the plates, or in their motion, probably then take place. If the trench and island arc originally had oceanic crust on both sides, the island arc is likely to flip, and instead of attempting to consume the continent it will consume the oceanic crust originally behind the arc (Fig. 13). The same cannot happen if the trench was originally between an ocean and a continent, and was consuming oceanic lithosphere (Fig. 14). Either the plate motions must then thicken the crust and the lithosphere, or the plate boundaries must change to permit consumption of oceanic lithosphere. Such crustal thickening is at present taking place in the eastern part of the Alpide belt.

If the arguments above are correct the cold descending slab is only produced by consumption of oceanic crust. It is not produced by the continental equivalents of island arcs. Thus beneath continents horizontal temperature gradients are not generated by consumption of cold lithosphere, and the position of the descending limbs of convection currents are not controlled by plate motions. Thus the whole pattern of mantle convection must depend on the motion of the continents. However, changes in the geographical position of the sinking limb will not take place as

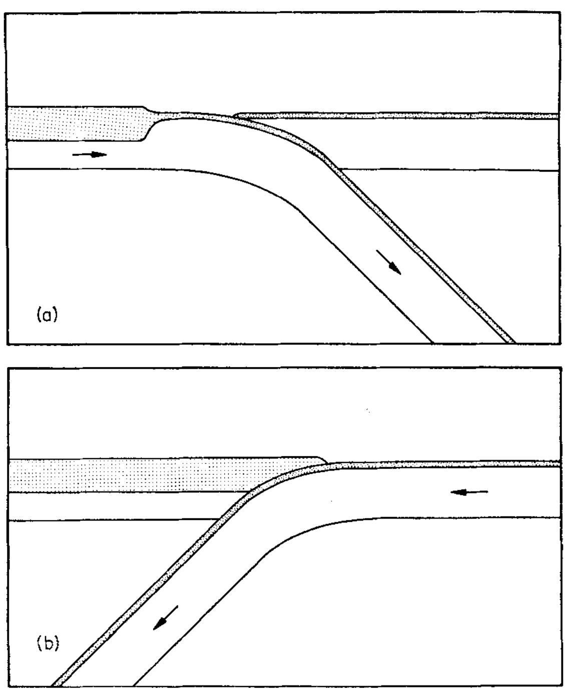


FIG. 13. A trench which originally has oceanic crust on both sides (thin stippled surface layer) may attempt to consume a continent (thick stippled surface layer). Continental crust cannot sink, and therefore the direction of overthrusting changes (b) to consume the oceanic crust originally behind the island arc.

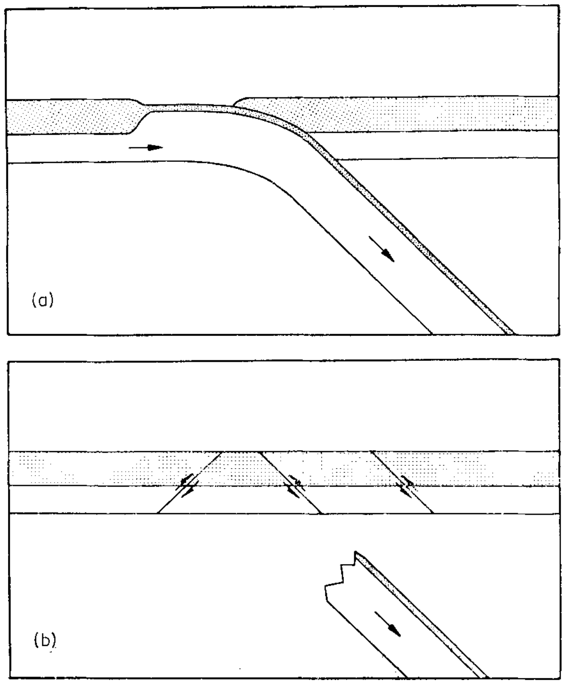


FIG. 14. If the trench in 13(b) attempts to consume a continent (a), regeneration of the sinking slab ceases (b), and mountains are built over a wide zone by overthrusting.

soon as two continents meet. The mantle below an island arc is cold and will continue to sink. The motion can only change when this material has been convected elsewhere, and when the blanketing effect of continental radioactivity penetrates to the mantle. Thus the flow will continue to move two continents together after they have collided, and therefore can produce fold mountains by crustal shortening over large areas. Examples of such mountains are the Himalaya and the Persian ranges. Mountain ranges like the Andes between continents and oceans are much less broad, presumably because most of the relative motion is taken up by consumption of the oceanic plate. Perhaps the most important consequence of the collision of two continents is that the process provides a method of changing the convection pattern within the mantle. The generation of the cold sinking slab ceases when the continents meet, and therefore the control of the sinking limb of the cell by the trench concerned will also cease. Rearrangement of convection patterns can then occur. It should therefore be possible to relate world-wide slow changes in the motion of large plates to particular collisions. However, certain apparent changes in the directions of relative motion between plates are caused by changes in plate geometry, and not in their relative motion (McKenzie & Morgan 1969). It is essential to separate these two effects, since only real changes of relative motion are connected with changes in convection patterns.

The other major difference between continental and oceanic plates is their seismicity. Throughout all continents except Antarctica some earthquakes occur which are not obviously related to the plate boundaries. The same is not true of oceanic regions. The difference in behaviour is caused by the contrast in the mechanical properties between the oceanic and continental rocks. The continental crust consists of about 15 km of granitic rocks and a similar thickness of gabbroic. The melting point of both rock types is less than about 1200 °C, and together they form more than half of the continental lithosphere. Of the upper 30 km of the oceanic plate, only perhaps 5 km consists of rocks with low melting points, the other 25 km is peridotite strongly depleted in all low melting point phases, with a melting temperature of about 1800 °C. The activation energy for high temperature creep is proportional to the melting temperature (Sherby 1962). Thus oceanic plates are more resistant to deformation than continental ones, in contrast to the early ideas of drift.

The arguments in this section suggest that the differences between the deformation of oceans and continents is entirely due to the difference in chemical composition, and not to any differences in the convection cells beneath them. Continental plates are easy to deform but difficult to consume compared with oceanic ones, and much tectonic evolution of continents must be a consequence of these differences.

6. Conclusions

Perhaps the most important result obtained in this paper is the dominance of thermal boundary conditions and time constants in convection processes. The temperature structure of the sinking lithosphere is determined by the time constant of the slab and by the spreading rate, and in turn governs the distribution of intermediate and deep focus earthquakes. The sinking slab can also control the position of the descending limbs of convection cells.

The main difference between oceanic and continental plates are due to their different composition. Continental rocks are considerably less dense than the mantle, and cannot sink through it. Any attempt to consume continental rocks must therefore produce a change in the motion or the boundaries of the plates.

The only useful simplification of the convection problem appears to be the idea of plates. Attempts to understand the driving forces by similar methods show little promise. There is therefore no obvious alternative to numerical solution of the general non-linear convection equations.

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