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Deep Mantle Convection Plumes and Plate Motions[1]

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Abstract Evidence shows that volcanic island chains and aseismic ridges are formed by plate motion over fixed-mantle "hot-spots" (Iceland, Hawaii, Galápagos, etc.) and new arguments link these hot-spots with the driving mechanism of continental drift. It is assumed that the hot-spots are surface expressions of deep mantle plumes roughly 150 km in diameter, rising 2 m/year, and extending to the lowest part of the mantle. The rising material spreads out in the asthenosphere, producing stresses on the plate bottoms. Order-of-magnitude estimates show these stresses are sufficiently large to influence plate motion significantly. The total upward flow in the plumes is estimated at 500 cu km/year, which would require the entire mantle to overturn once each 2 billion years.

INTRODUCTION

We may account for the main features of the Hawaiian Islands (the long linear chain, the uniform progression of ages toward the northwest, the transition from the tholeiitic main stage to more alkalic later stages of volcano growth) by assuming that the Pacific plate is moving northwestward over a fixed-mantle "hot-spot." Likewise the Greenland-Iceland and Iceland-Faeroe ridges emanating from Iceland, and the Rio Grande and Walvis ridges emanating from Tristan da Cunha and Gough Island, may be interpreted as the result of plates moving away from fixed hot-spots located on the crest of a spreading mid-ocean rise. Wilson (1963a, b; 1965a, b) advanced this hypothesis for the origin of island chains and aseismic ridges, and in a sequence of papers developed how these features may be used to determine the present motion of each plate, how the aseismic ridges are important guides in reconstructing pre-drift continental configurations, and how aseismic ridges and transform faults interrelate. Morgan (in press) has presented three additional lines of evidence supporting the concept of rigid plates moving over fixed-mantle hot-spots; this evidence is summarized in Figures 1–3.

In Figure 1, I quantify observations of the parallelism of the Pacific island chains and the continuity of the Hawaiian Islands and Emperor Seamount chains. Four sites of present volcanism are noted: (1) the Juan de Fuca Rise near Cobb Seamount, (2) Hawaii, (3) MacDonald Seamount (for a report on its discovery see Johnson, 1970), and (4) the Pacific-Nazca rise near Easter Island. The four heavy lines in Figure 1 were generated by rotating the Pacific plate backward in time over these four fixed hot-spots, first 34° about a pole at 67°N, 73°W (0–40 m.y.), then 45° about a pole at 23°N, 110°W (40–100 m.y.). The close agreement of these predicted lines with the trends of the Gulf of Alaska seamount chains, the Hawaiian-Emperor chain, the Austral-Gilbert-Marshall chain, and the Taumotu-Line chain substantiates this hypothesis. Even more exact agreement can be obtained by removing the constraint of fixed hot-spots and allowing the hot-spots to migrate at about ½ cm/year—a small fraction of the roughly 7-cm/year motion of the Pacific plate.

Figure 2 shows paleomagnetic pole positions determined from seamounts in the Pacific. The letter shown with each circle of confidence identifies the seamount group used by Francheteau *et al.* (1970); the number shows the age of the pole in millions of years. The heavy line is the Pacific polar-wandering curve predicted by the motion of the Pacific plate shown in Figure 1. (It is implicitly assumed that, at least during the past 100 m.y., the geomagnetic pole has not wandered relative to the hot-spots fixed in the lower mantle.) The paleomagnetic evidence thus verifies the plate motion predicted by the hot-spot trajectories. This test should be repeated for each of the other major plate units.

Figure 3 shows the present plate motion over the fixed hot-spots. This figure was constructed by finding the relative plate motions deduced from fault strikes and spreading rates on the

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Kenneth Deffeyes first made several of the order-of-magnitude estimates shown herein. I thank him and many others at Princeton for their contributions to this problem. This work was partly supported by the National Science Foundation and the Office of Naval Research.

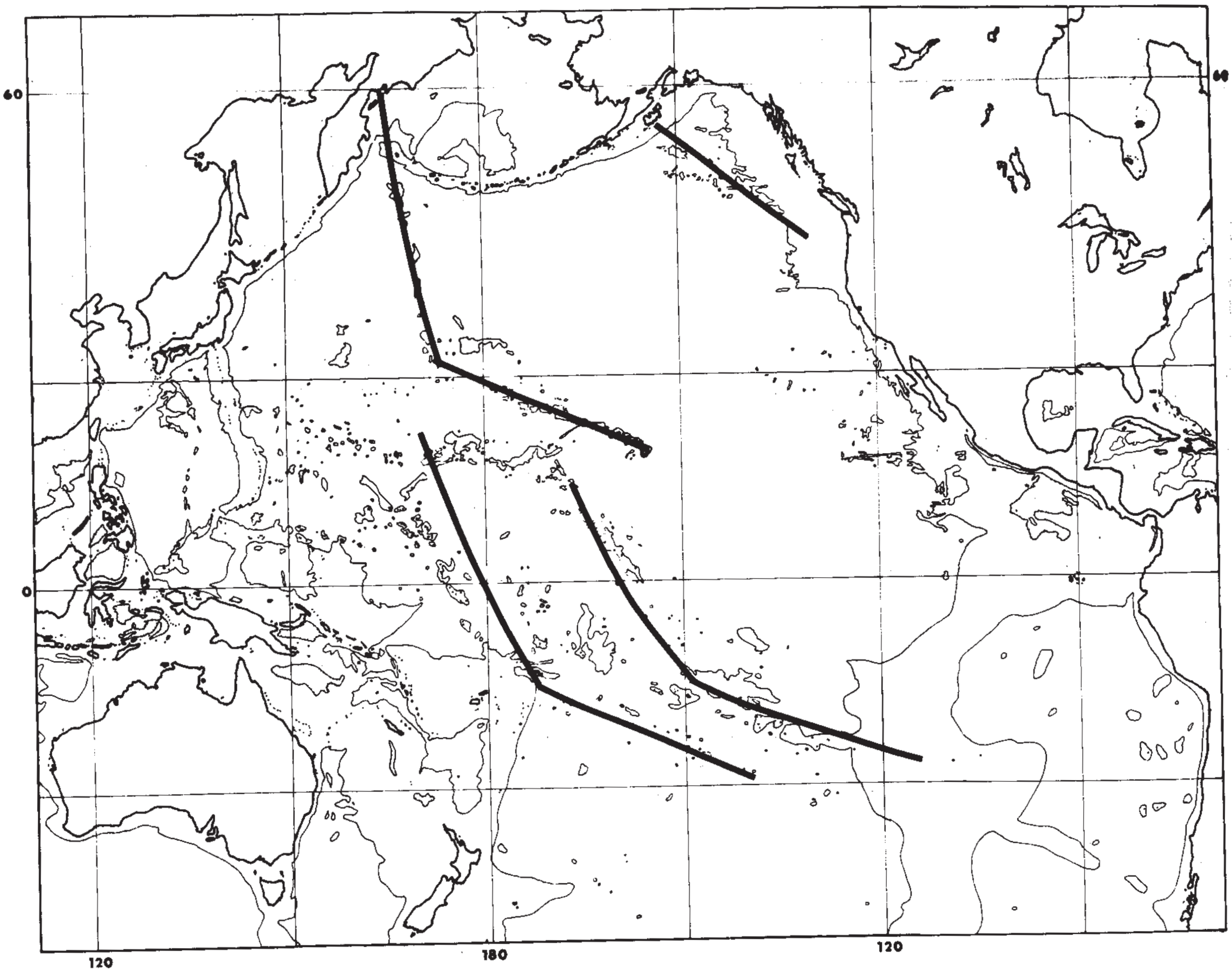

FIG. 1—Hot-spot trajectories constructed by rotating Pacific plate over four fixed hot-spots.

rise boundaries, and then adding a constant rotation to make the Pacific plate rotate properly over its hot-spots. If the synthesis tabulated by Morgan (in press) is correct, and if all the hot-spots are fixed in the mantle, then the velocity vectors shown in Figure 3 should predict accurately the trends of the island chains/aseismic ridges away from hot-spots.

Morgan (1971; in press) proposed that these hot-spots are surface manifestations of lower mantle convection which provides the motive force for continental drift. Assume that about 20 deep mantle plumes bring heat and relatively primordial material up to the asthenosphere, producing horizontal currents in the asthenosphere which flow radially away from each plume. The points of upwelling have unique petrologic and kinematic properties, but I assume there are no corresponding unique points of downwelling—the return flow is uniformly distributed throughout the mantle. The deep convection thus has a thunderhead character, whereas the shallow convection, constrained by the rigid plates at the top surface, has a roll or two-dimensional character.

Some of the consequences of the interactions of rigid plates with localized upwellings and an interpretation of the observed petrologic differences of oceanic island type basalt and oceanic ridge type basalt were presented by Morgan (1971; in press). In this paper I shall amplify the arguments supporting the claim that the hot-spots provide the driving force for continental drift. These arguments fall into three categories: (1) the observation that most hot-spots are near rise crests and evidence that hot-spots become active before continents split apart; (2) an interpretation of the gravity and topography around each hot-spot, showing that the mantle plumes generate moderately large stresses; and (3) estimates comparing the magnitude of stresses generated by plumes to the magnitude of rise and trench stresses.

Location of Hot-Spots

The primary criterion for selection of the hot-spots in Figure 3 was recent volcanic islands not associated with andesitic trench-type activity. Several volcanic islands were removed from this list based on the assumption that some volcanic activity is delayed by a magma-storage mechanism somewhere in the lithosphere. For example, the oldest rocks on Heard Island on the Kerguelen Ridge are 40 m.y. old, but there has been some activity in historic times. I assume that the entire Kerguelen Ridge is a hot-spot feature and that the recent activity on Heard Island is a delayed action of hot-spot material placed there in the lithosphere 40 m.y. ago; thus Heard is eliminated from the present hot-spot list. A more puzzling case is the Cameroon Trend, which lines up with the ridge that heads northeast from St. Helena. I have tentatively assumed that the Cameroons are a delayed action of the St. Helena hot-spot, and that the Cretaceous volcanic rocks northeast of Mt. Cameroon are the expression of the St. Helena hot-spot prior to the breakup of South America and Africa.

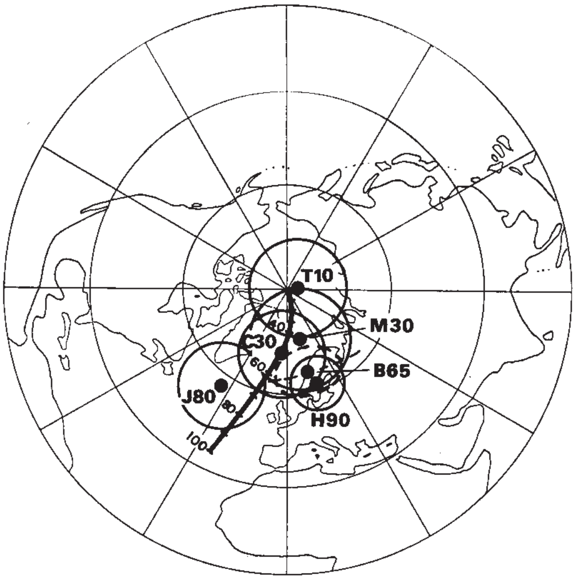

Fig. 2—Pacific paleomagnetic pole positions (adapted from Francheteau *et al.*, 1970) and polar-wander curve predicted by motion of Pacific plate shown in Figure 1.

Four oceanic centers of volcanism are not near mid-ocean rises: Hawaii and MacDonald in the Pacific plate, the Canary Islands in the African plate, and the Comores Islands in the Somalian plate. In addition, Yellowstone (and the Snake River basalts), Tibesti (in central Sahara), and Mount Kenya have characteristics suggestive of continental hot-spots. The Yellowstone, Kenya, and Comores hot-spots are near present-day breakups in the western

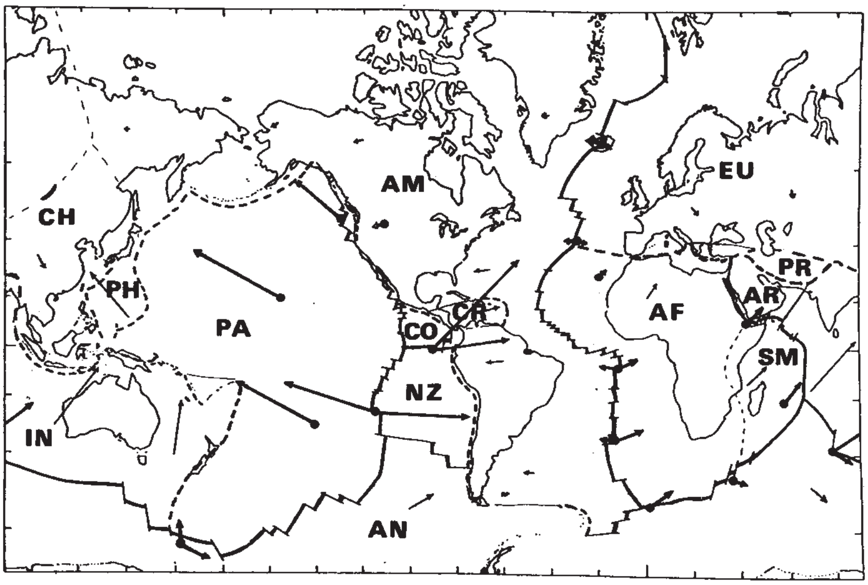

Fig. 3—Present motions of plates over hot-spots. Relative plate motions were determined from fault strikes and spreading rates on rise boundaries; with an appropriate constant rotation added, absolute motions of each plate over mantle were determined. Lengths of arrows are proportional to plate speed.

United States and East Africa, therefore only four of the locations listed are far from present spreading axes. In contrast, 14 hot-spots are near rise axes; four are in the Pacific (Juan de Fuca, Galápagos, Easter, and Balleny) and two lesser hot-spots suggested by seamount chains are near the mouth of the Gulf of California and the Eltanin fracture zone. There are four in the Indian Ocean (Amsterdam, Reunion, Afar, and Prince Edward), and six in the Atlantic (Bouvet, Tristan de Cunha-Gough, St. Helena, Ascension, Azores, and Iceland). One may argue that the lack of identifiable hot-spots on continents results from continental complexities camouflaging their presence. However, the reverse argument is no less valid; the hot-spots are mostly in the open ocean because they have pushed the crust away.

More dramatic than the location of the present hot-spots near the present rises is the evidence that the same hot-spots became active *before* the rises were formed. This evidence is best displayed in the lands bordering the Atlantic. The Jurassic volcanics in Patagonia may be regarded as the early expression of the Bouvet plume. (The even earlier Cape Vòlcanics in South Africa may be an expression of this plume. This interpretation depends on how Gondwanaland moved over this plume.) The flood basalts in the Parana basin and the ring dike complex of Southwest Africa may be due to the Tristan da Cunha plume. The White Mountain Magma Series in New Hampshire can be associated with the same hot-spot that produced the New England Seamount chain (probably the Azores plume). The Skaergaard and the Scottish Tertiary volcanic province are associated with the Iceland plume. I claim this line of plumes produced currents in the asthenosphere which led to the continental breakup creating the Atlantic. Likewise the Deccan Traps (Reunion plume) were symptomatic of the forthcoming Indian Ocean rifting, and, if my premise is accepted as proved, the Snake River basalts (Yellowstone plume) foretell a breakup of North America.

GRAVITY AND TOPOGRAPHIC HIGHS

Figure 4 shows a worldwide gravity map computed for spherical harmonics up to order 16 (Kaula, 1970). Isolated gravity highs are apparent over Iceland, Hawaii, and most of the other hot-spots (Galápagos is a notable exception). Such gravity highs are symptomatic of rising currents in the mantle—the less dense material in the rising plume produces a broad negative gravity anomaly; but the satellite passes closer to the excess mass in the elevated surface pushed up by this current, and the net gravity anomaly in the area over the rising current is positive. The mid-ocean rises are exceptionally shallow near the hot-spots; note particularly the 10^6 sq km areas surrounding the Iceland, Juan de Fuca, and Galápagos plumes. This regional high topography is another manifestation of the rising plume, and I shall now use the magnitude of the high topography and gravity to estimate roughly the size of the rising current.

The formulas following are adapted from derivations shown by Morgan (1965). A spherical ball of mass deficiency M, located a distance D

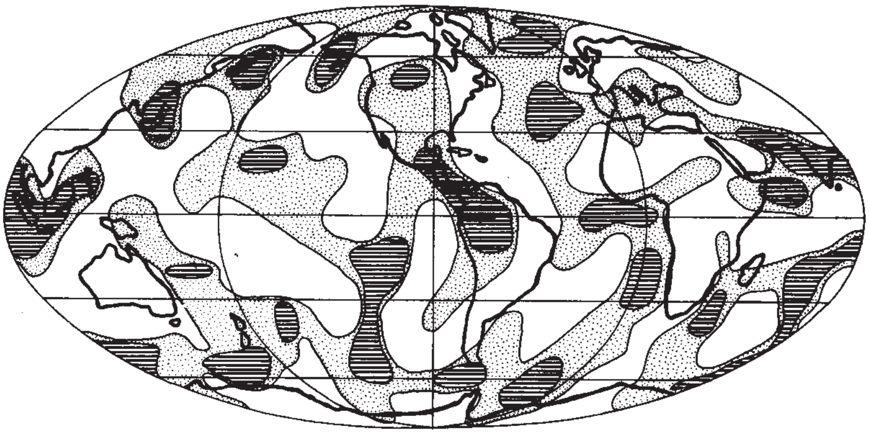

FIG. 4—Isostatic gravity map of earth constructed from spherical harmonic coefficients of degree 6 through 16. Shaded areas are regions of positive anomalies; heavier shaded areas are regions where anomalies are greater than +10 mgal. Note correlations of gravity highs with Iceland, Hawaii, and most other hot-spots (adapted from Kaula, 1970).

below the surface, is rising in a fluid of uniform viscosity. The Navier-Stokes equations were solved for a uniform half-space with rigid-plate boundary conditions for the top surface. The formulas give the normal stress σ_n, the shear stress σ_s, and the gravity anomaly δg at the rigid-plate boundary. The height the surface is pushed up is related to the normal stress by $\sigma_n = \rho g h$, and δg is the gravity anomaly produced by adding the effect of the raised surface to the gravity effect of the ball M. In these formulas, g is the gravitational field strength, G is the Newtonian Gravitational Constant, and r is the horizontal distance from the point directly above the ball.

$$\delta g = GMD\frac{(2D^2 - r^2)}{(D^2 + r^2)^{5/2}} \qquad (1)$$

$$\sigma_n = \frac{Mg}{2\pi}\,\frac{3D^3}{(D^2 + r^2)^{5/2}} \qquad (2)$$

$$\sigma_s = \frac{Mg}{2\pi}\,\frac{2r}{(D^2 + r^2)^{3/2}} \qquad (3)$$

From Figure 4, we estimate that δg over a plume is typically +20 mgal and that the gravity anomaly falls off with distance, so that it is zero at 1,000 km away. Using equation (1) with these values, I find $D = 700$ km and $M = 0.8 \times 10^{21}$ g. If from topographic maps we estimate 1 km extra height as typical of the region near a plume, then equation (2) yields a value of $M = 3 \times 10^{21}$ g. These estimates are based on a uniform viscosity model; what effects would a more complicated viscosity pattern have? The uniform viscosity case yields the result that the total mass excess of the elevated surface equals the mass deficiency of the rising ball; *i.e.*, the stresses produced by the rising ball are balanced by the stresses produced by the increased surface load. If plumes deep in the earth are rising in pipes surrounded by a very viscous mantle (much more viscous than the asthenosphere), then much of the stress of the rising material will be distributed throughout the more rigid mantle and will not produce a local elevated surface. Thus, a larger mass deficiency may be present than that estimated from the uniform viscosity formulas. We therefore estimate $M = 10^{21} - 10^{22}$ g as a typical mass deficiency of a single plume.

Using plume dimensions to be discussed later, the density of a plume can be calculated from its total mass deficiency ($M = 3 \times 10^{21}$ g). Assuming a cylindrical shape 150 km in diameter and assuming that only the top 1,000 km of the cylinder contributes to the M estimated by the surface gravity and topography, we find $\delta\rho = -0.2$ g/cu cm, or about a 5 percent density deficiency. This density change could be produced by a migration of "400–600"-km phase change boundary. This magnitude density deficiency is ideal; if the mass and dimensions yielded a density difference 10 times larger or 10 times smaller, the result would be respectively unreasonable or uninteresting.

Formula (3) may be used to estimate the shear stresses acting on the plate bottom. On the assumption that $M = 3 \times 10^{21}$ g and $D = 700$ km, then at $r = 500$ km the shear stress is 80 bars. How the shear stress falls off with distance away from the plume is very sensitive to the exact viscosity pattern; however I shall use 100 bars as a rough estimate of the shear stress on a plate near a plume.

STRESSES AT RISES AND TRENCHES

If the stresses produced by plume currents were clearly larger than the push of a rise or the pull of a trench, then the problem of finding the stresses acting on the plates would be greatly simplified. I have constructed the following mathematical model with three sources of stress: (1) stresses on plate bottoms falling off as $1/r$ away from each hot-spot, (2) a drag stress on the bottom of each plate proportional to the plate's velocity over the lower mantle and (3) stresses generated by plate-to-plate interactions of rises, trenches, and faults. The last category would have a moderately complicated set of equations predicting the stress generated by a specified closing rate at a trench or slip rate at a fault, *etc.*, but it would have Newton's Third Law of action and reaction as a simplifying feature. With this model the torques on plates with the present boundary locations can be determined, and the direction and rate of motion of each plate predicted. The parameters specifying plume size and plate-to-plate interactions then could be accurately found by adjusting them until the present observed plate motions were predicted. If the plate-to-plate interactions were smaller than the stresses produced by plumes, very elementary assumptions could be made about rises and trenches, as small errors in this specification will not be important. If the push of rises and pull of trenches are stronger than the plume-generated stresses (as appears to be the case), the equations relating stress and strain rate at boundaries must be known accurately. Thus I consider the evidence relating to the magnitude of stresses at rises and trenches, and in particular reexamine the argument that the symmetry of rises shows that the rises exert no push on the plates.

A simple calculation places an upper limit on the amount of stress that can be generated by a spreading rise. Equate the work done pushing the plates apart (the total force on a plate times the rate of displacement) with the gravitational energy available in the light material rising into the spreading area (the buoyant force times the rate of upward movement). This calculation is performed most easily in a triangular geometry with a wedge-shaped unit rising and pushing two lithospheric plates apart. The calculation is independent of the shape of the triangle and the spreading rate; it depends only on the thickness of the lithosphere and the density deficiency of the material entering at the bottom compared to the average density of the lithosphere. If it is assumed that $L = 70$ km and $\delta\rho/\rho = 3$ percent, the horizontal compressive stress in the lithosphere would be 300 bars. Any of the gravitational energy of the rising wedge that is dissipated in viscous flow will not be available to do the work required to push the plates apart; thus 300 bars is an upper limit. A 1 percent density deficiency may more accurately describe the material entering the region below the rise; in this case the horizontal compressive stress averaged over the thickness of the lithosphere would be less than 100 bars.

Focal mechanism studies of earthquakes along rise axes have shown that the lithosphere at a rise is under tension in the direction of spreading. Wyss (1970b) has concluded that this tension has a magnitude of roughly 200 bars; this result is based on studies of the seismic moments of rise earthquakes showing an "apparent stress release" of 20 bars, combined with an estimate of 10 percent for the seismic efficiency of stress release. This tension of 200 bars can be interpreted several ways. In one interpretation, we assume that some distant forces are pulling the plates apart and that the asthenosphere is rising passively to fill the void that would be created by plate separation. The lithosphere is very thin beneath the rise crest, and this "necking" of the lithosphere acts as a stress concentrator. That is, the tensional stress may be 200 bars in a region 5 km thick at the rise crest, but only 50 bars spread out over the entire 70 km thick lithosphere at some distance away from the rise. The amount of stress concentration in the thin lithosphere depends on how the stress load is distributed between the cool, strong lithosphere and the hotter, weaker asthenosphere flowing into the broad "gap" between the plates. In an alternate interpretation, the rising asthenosphere is pushing the plates apart, causing horizontal compression in the plates except in the small thin section of lithosphere at the rise which resists the separation. The average compression in the plates thus will be reduced by a factor which depends on the effective viscosity and thickness of the lithosphere at the rise crest. The dissipation in this thin section of lithosphere may form the major part of the viscous dissipation mentioned in the preceding paragraph. Therefore, the stresses at rises may be compressive or tensile, but in any event have a magnitude less than a few hundred bars.

Why are the mid-ocean rises "mid-ocean", and why is the seafloor magnetic pattern symmetrical about the rise crest? It would be easy to imagine that a rise creates new sea floor on one side only, analogous to the one-sided consumption of crust in a trench system, and yet new sea floor is created in equal amounts on the two sides of a rise. As a consequence, rise crests cannot be fixed with respect to the mantle; they must migrate over the mantle to maintain their position midway between continents. An example of such rise migration is seen for the rise boundaries that enclose Africa on three sides. As the Mid-Atlantic Rise spreads symmetrically, there is ever more sea floor between the rise crest and the African coastline. With a similar increase in the distance from Africa to the crest of the Mid-Indian Rise, the distance from the crest of the Mid-Atlantic Rise to the crest of the Mid-Indian Rise must be increasing. Thus both rises cannot be fixed with respect to the mantle—one or both must be migrating over the mantle.

The Mid-Atlantic Rise apparently is fixed to the mantle, as all the Atlantic plumes are near the present crest. Thus it is the Mid-Indian Rise that is migrating east at a rate faster than the African plate is moving northeastward. The Mid-Indian Rise has migrated over the Reunion plume—this plume was once in the Indian plate (Deccan Traps, Laccadive-Maldive island chain), but is now on the African side of the rise. Similarly, the growth of the Afar Triangle on the southwest may be regarded as the plume staying fixed while the Red Sea and Aden rifts migrate northeastward.

It has been argued that the symmetrical spreading and the migrating rise crest indicate

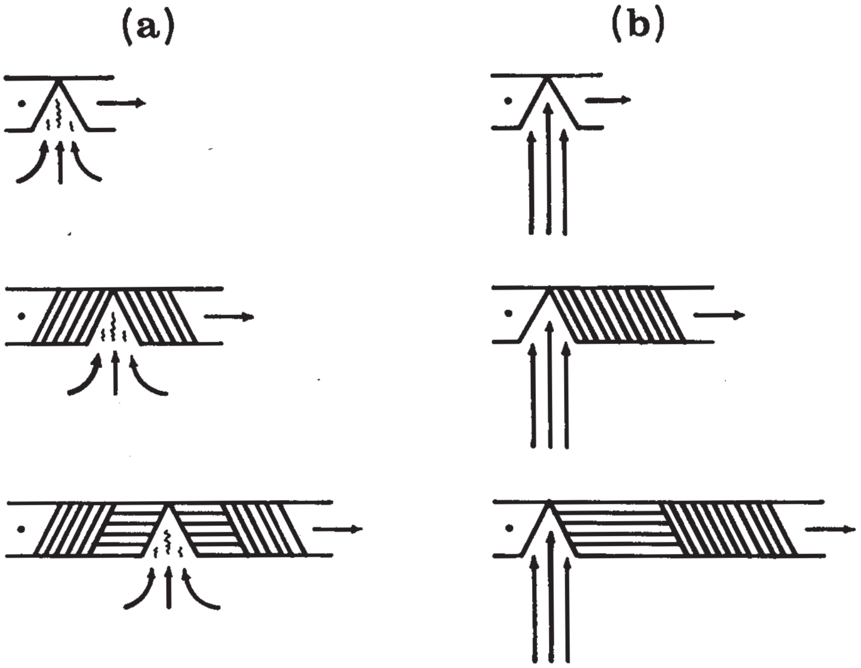

FIG. 5—In these symmetrical and asymmetric models of sea-floor spreading, left lithospheric plate is constrained to be fixed and right plate to move away at constant velocity. **(a)** If asthenosphere near rise fills gap made as plates move apart, symmetrical sea-floor spreading results. **(b)** If location of rising current is influenced strongly by conditions near bottom of asthenosphere, one-sided sea-floor spreading results.

that the rises do not drive the plates. If the two sides of a rise are moving away from a fixed rise crest with equal velocities, there is little problem; in contrast in Figure 5 I postulate a case in which the left plate has zero velocity over the mantle and the right plate is moving away with constant velocity. I do not inquire why the plates have these velocities—there may be a trench nearby on the right and another rise off left, or whatever is needed to produce the motions depicted in Figure 5. As the plates move apart at the rise crest, material from the asthenosphere rises to fill the void that would otherwise develop. The exact center of the most recently injected "dike" is hotter than any other part of the lithosphere, and because strength is extremely temperature dependent, this is where the plates will tear apart and another dike be inserted. Thus if the temperature pattern about the rise crest is symmetrical, the temperature dependence of strength will assure a symmetrical pattern of sea-floor spreading.

Any arguments about symmetrical or asymmetric spreading must thus concentrate on those factors which will make a symmetrical or asymmetric temperature pattern within a rise. It was concluded that the important factor is to have a soft asthenosphere below the plates, and that material flowing into the "gap" should be drawn from very shallow depths to avoid any kind of coupling with conditions at the bottom of the asthenosphere. It appears that passive pulling of asthenosphere into the gap or active driving of the asthenosphere upward into the gap is not related to the question of symmetry. The tensional or compressive nature of rises is related to whether or not there is a density inversion between the lithosphere and asthenosphere; the symmetrical spreading is related to the laws of heat conduction and the temperature dependence of strength. Thus it is thought that earlier conclusions of the writer (Morgan, 1971) and of Elsasser (1969) relating symmetry to passiveness are in error. A two-dimensional numerical rise model with viscosity and density varying with temperature could aid in answering this question.

A related problem concerns the existence of fracture zones. Objectors to sea-floor spreading have used the pattern of transform faults as an argument against spreading; namely, it is inconceivable that convection currents beneath the surface could have the numerous offsets of the surface pattern. The notion of a crustal plate removes the objection, as the deeper flow may have a smoother, more fluid pattern and only the "rigid" lithosphere need be broken into the irregular pattern observed at the surface. How-

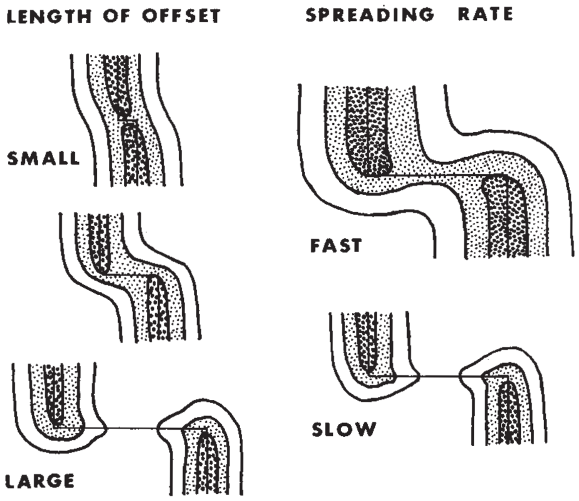

FIG. 6—Stability of fracture zone is influenced by length of offset and spreading rate. Contours are, schematically, thickness of lithosphere or, alternatively, depth to particular isotherm. If offset is very small, obliquely spreading rise may develop and transform fault would no longer exist. Length of offset of minimum stable transform fault would be dependent on spreading rate as with high spreading rates broader region of "thin lithosphere" would enclose rise (from Vogt *et al.*, 1969).

ever, the causative factors behind the maintenance of many short transform faults and the tendency for the rise to form segments perpendicular to the transform faults remain unknown.

Vogt *et al.* (1969) have noted that the minimum observed offset of transform faults is less in the Atlantic than the Pacific. They have devised the model shown in Figure 6 to account for this difference. The contours in this figure schematically represent either isotherms at a given depth (*e.g.*, 10 km), with the hottest temperature at the rise crest, or alternatively they represent the depth to a given isotherm (*e.g.*, 1,000°C). We may say arbitrarily that all material cooler than 1,000°C is lithosphere and any hotter material is asthenosphere; in this case the contours in Figure 6 show the thickness of the lithosphere. These contours are patterned after heat flow models of rises (see McKenzie, 1967) in which it was shown that the distance between the isotherm contours is directly proportional to the spreading rate of the rise. The model in Figure 6 also assumes that heat is generated on the transform faults by frictional heating. Vogt *et al.* (1969) used this diagram to illustrate that, if the offset of a transform fault is "‘sufficiently small" (where "sufficiently small" is directly related to spreading rate), then the regions of thin lithosphere may merge and an oblique spreading ridge replaces the two short perpendicular segments and the transform fault. Figure 6 also indicates that an offset greater than a certain minimum length will be stable and will not be replaced by an obliquely spreading rise.

The flow pattern chosen by nature must minimize the total dissipation rate. Therefore, the friction per unit length along a transform fault must be markedly less than the tensile resistance of a rise segment, otherwise an oblique rise would be observed more commonly. Further, I conclude that there is more tensile resistance for an oblique rise than a perpendicular rise, or else the dissipation rate of "perpendicular rise and fracture zone" would not be less than "oblique rise." The larger resistance of the oblique rise must be due to increased heat loss through the longer sides resulting in a higher effective viscosity for the oblique rise. Several models to relate these factors were constructed by the writer to attempt to relate the tensile stress to the dissipation rate. However, it was realized that a constant compressive stress produced by buoyant material would have no effect on the model. The buoyant terms are determined by the rate of the upflow, which is the same in the oblique and normal cases; thus the solution of this problem could indicate much about the mechanisms at rise crests but cannot be used to place a limit on the overall tensile or compressive nature of the rises.

I conclude that we do not really know whether rises are characterized by compressive stresses aiding the plate motion or by tensile stresses acting as a brake to plate motion. Inasmuch as tensile stresses at rises would be most favorable to the plume driven model, it seems conservative to expect all rises to generate compressive stresses of about 100–300 bars.

On the question of the tensile (actively pulling down) or compressive (resisting being pulled down) character of trenches, the earthquake mechanism solutions of Isacks *et al.*, (1969) show that earthquakes at 100–200 km depth indicate tensile stress along the sinking slab; the slab above is being pulled down by a density excess at or below this depth. Deeper earthquakes, from 300–700 km, show compressive stresses along the slab; the deeper part is resisting being driven deeper into the mantle by stresses generated above. From earthquake seismic moments (Wyss, 1970a), from surface gravity anomalies (Morgan, 1965), and from calculations based on assumed temperature profiles in the sinking slab (McKenzie, 1969), the stress produced by the sinking slab has been estimated at from a few hundred bars to a thousand bars. The motion of the sinking slab is resisted by the material on the slab boundaries, particularly on the boundary of the highest seismic activity, where the two lithospheric plates rub together. Wyss (1970a) found that 200-bar stresses are indicated by the shallow earthquakes at trenches, the same magnitude of stress as found at rises. There are many questions about the magnitude of the stresses at trenches. How much of the pull of the sinking slab is cancelled by the shallow friction of the two lithospheric plates? Do deep trenches pull with more tension than intermediate depth trenches, or does the deeper slab push into a very resistant media which tend to reduce the pull of the slab? Do areas of continental underthrusting, as in the Zagros and Himalayas, exert any pull at all, or are all stresses in these regions compressive, resisting the closing motions?

I add one new point to the discussions of trenches. Figure 3 (and more precisely if not more accurately, Table 1 of Morgan, in press) shows the absolute rate of each plate over the mantle, as determined by relative spreading rates plus the trajectories of hot-spots. The Nazca plate is moving eastward toward the Peru-Chile trench at about 7 cm/year, whereas the South American plate is moving westward toward this trench at about 1.5 cm/year. When we examine the velocities of plates at other trench systems, we find the following to be a general rule: both plates move toward the trench with the underthrusting plate moving about four times faster than the overriding plate. (This claim is made with great reservations, because, to establish it a much more accurate determination of absolute plate motions is needed.) However, note the special situation of the Philippine Sea. It is moving westward toward Asia and thus is not moving toward the Marianas trench. Also, the Indian plate does not have a component of motion toward the Tonga trench. It thus appears that the areas of anomalous spreading behind trenches (Karig, 1970) can be identified if the absolute motion of the plates is known. Moberly (in press) and Elsasser (1971) have given theoretical explanations of this phenomenon; they suppose the lithospheric slabs are not exclusively sliding obliquely into the mantle but also have a vertical sinking component. The sliding and sinking cause the trench axis to migrate "seaward," and the overthrusting plate must migrate toward the trench or a gap will open behind the trench.

This motion of both plates toward a trench appears to clinch the argument in favor of tensile stresses at trenches. In the following discussion I shall assume this horizontal tensile stress has a magnitude of a few hundred bars. I shall neglect the stresses produced by plates rubbing together at great faults, partly because such stresses must surely be reactive and not drive plates, and partly because the model of Vogt *et al.* (1969) on transform faults shows the dissipative stresses on faults must be much less than on rises.

ESTIMATES OF PLUME MAGNITUDES

Estimates of stresses, heat flow, and lead isotope data are now used to estimate the sizes of plumes. A self-consistent set of relations is found if each plume is 150 km in diameter, with an average upward velocity of 2 m/year; 20 such plumes would bring up a total volume of 500 cu km/year. Other values used in this model are a lithosphere 70 km thick and an asthenosphere 200 km thick, with an average viscosity of 3×10^{21} poise. No exactitude should be placed on any of these numbers, as it is the overall effects of the parameters and not the precise value of any one which is significant in the ensuing discussion. My purpose in the use of this model is to show that plumes can provide the stresses needed to move the surface plates and have important implications in the interpretation of heat flow and age of the mantle, and at the same time do not violate any of the "known" values of the earth.

Using equation (3) I found that the shear stress 500 km away from a plume was about 100 bars. I obtain this same magnitude of stress in the plume and thin asthenosphere model specified above. The average velocity $\bar{v}_a$ of the asthenosphere at a distance R from the plume is related to the upward velocity of the plume v_p by $2\pi R D \bar{v}_a = \pi d^2 v_p/4$, where D and d are the thickness of the asthenosphere and diameter of the plume respectively. Using the model values given above, I calculate that the average asthenosphere velocity 500 km from the plume is 5 cm/year. The asthenosphere flow is channeled between the rigid upper plate and the lower mantle (assumed to be slightly more viscous), and the velocity profile in this channel takes the well-known parabolic shape. (If the upper plate is moving, a linear velocity profile will be superposed on this pattern.) The stress at the top boundary (σ) is related to the average velocity in the asthenosphere ($\bar{v}_a$) by $\sigma = 6\eta\bar{v}_a/D$, giving $\sigma = 150$ bars at 500 km radius.

Integration of the effect of this stress on the bottom of a plate determines the total force one plume can exert on a plate. I assume that a rise axis passes right over a plume and integrate the component of force directed away from the rise. I choose as limits of my integration 75 km (the radial velocity is zero directly above a plume and increases to a maximum value a plume radius away) and 1,500 km (roughly half the distance between plumes). This integration yields $F = 1 \times 10^{24}$ dynes for the total force exerted on a plate.

I then compare this magnitude with the stress created by a plate moving over the asthenosphere. With the properties of the asthenosphere given above, a plate moving 3 cm/year over the mantle creates a shear stress of 15 bars. If we assume the plate is a square about 5,000 km on a side, then the total drag on a plate is $F = 4 \times 10^{24}$ dynes. Another comparison is with the stress generated by plate-to-plate interactions. I assume a stress of 100 bars on a plate 70 km thick along a boundary 10,000 km long; the total force acting on the plate is $F = 7 \times 10^{24}$ dynes. The plume generated stress is smaller than the other two, but it could be increased by changing the values of the viscosity and thickness of the asthenosphere, or by increasing the flow up a pipe. My conclusion is that, given the uncertainty in these factors, all three mechanisms (plume, drag, and plate-to-plate) should be considered in a model of plate motion.

The key estimate is the volume of flow up the pipes. There is a clear-cut way to obtain a lower limit on the rate if the plumes are driving the plates. From roughly 40,000 km of rise axis with an average (half) spreading rate of 3 cm/year, I can determine how much new crust is generated each year. An accurate summing of the spreading rates along each rise yields 2.5 sq km/year for this rate (Deffeyes, 1970, p. 214). If this is multiplied by the thickness of the lithosphere, it is apparent that lithosphere is being generated at a rate of 170 cu km/year, and of course destroyed at an equal rate at the trenches. Suppose I had concluded that the total volume brought up by the plumes was only 10 cu km/year, or some other equally small number. Then asthenosphere currents of total flux 10 cu km/year would be spreading horizontally away from several points on or near rise crests, whereas 170 cu km/year would be flowing toward the rises as a counterflow to the lithosphere motion. The net stresses on plate bottoms would be such as to close up the plates. Thus the total volume emanating from the plumes must be several times larger than 170 cu km/year. The value of 500 cu km/year specified in the model is three times this; one may like a larger multiple, but this value does satisfy the lower bound and fits the lead isotope criteria to be discussed subsequently. In this light, even small plume flow aids the shallow convection in a way not noted in the preceding paragraph, where I found that a plate spreading at 3 cm/year created a viscous drag of 15 bars. If I include in this calculation the assumption that the 200-km-thick asthenosphere must have a net flow to counter the mass transport of the lithosphere, then the stress on the bottom of the plate is 50 bars instead of 15 bars.

These numbers have interesting consequences for the interpretation of heat flow data. Suppose 500 cu km/year (ϕ) brought up by the plumes is on the average 300°C (ΔT) hotter than the nonplume mantle at the same depth. Then using $C_p = 0.25$ cal/g°C, I find that the total upward heat available from the plumes is $Q = \rho\, C_p\, \Delta T \phi = 1.5 \times 10^{20}$ ergs/sec. This number is half the total heat flow of the earth. Here is a mechanism for concentrating all the deep mantle's heat production into predominantly oceanic regions. The correct interpretation of this may show why the oceanic and continental heat flow averages are so nearly equal. There is another surprise in these numbers: the return flow of the plumes, involving the slow sinking of the entire mantle, is at a rate of 0.1 cm/year and downward convection at 0.1 cm/year can dominate over conduction or radiation mechanisms of heat transport. For example, if I assume a temperature (T) of 1,500°K at the base of the asthenosphere, then the downward flux of heat by convection is $q = \rho\, C_p\, T\, v = 3.7$ μcal/sq cm/sec. Thus, heat could leave the lower mantle only at the plumes. How much heat recycles compared with the amount of heat lost at the upper surface is a measure of the efficiency of the heat engine—thus the earth could be regarded as being a moderately efficient heat engine.

The isotopic composition of lead from Tristan da Cunha and St. Helena has been discussed by Oversby and Gast (1970) together with earlier results from Ascension and Gough. They showed that the lead data cannot be interpreted with a simple one-stage growth model; *i.e.*, a mixing of lead and uranium isotopes 4.5 b.y. ago, when the mantle was formed, with no separation or mixing since (except perhaps in the last few million years as the rocks were brought to the surface). Instead, they found a two-stage growth history with lead events at 4.5 b.y. and 1.8 b.y. (plus possible changes in the last few million years). That is, they interpreted their data to show that there was an homogenization and then separation of the lead and uranium isotopes 1.8 b.y. ago in the material that now makes up these islands, and that no further mixing or separation occurred until very recently.

I incorporate this observation into the plume model as follows—I assume that the rocks which now make up these islands were last near the earth's surface 1.8 b.y. ago; *i.e.*, that 1.8 b.y. is the cycle time required for a particle to sink slowly in the mantle and then to rise in a plume back to the asthenosphere (or in this case, for part of the plume to reach the surface). The rate of upwelling of 500 cu km/year fits these data; the total volume of the mantle, 1×10^{12} cu km, divided by this rate gives 2 b.y. If the lead isotope data are interpreted in this manner, it makes two restrictions on the plume model. First, this is evidence that the entire mantle is involved in the overturn—that the plumes extend all the way down to the core-mantle boundary. Second, the rate of upwelling may be estimated most accurately by knowing the period of the mantle overturn—a very straightforward estimate if lead isotope data from other plumes far from the South Atlantic also show the 1.8 b.y. lead event.

In conclusion, the mid-ocean position of most of the plumes and the land evidence of plume activity prior to continental breakup suggest that the plumes produce the stresses which drive the plates apart. An order of magnitude estimate shows that stresses produced by plume currents are comparable to other stresses. The model implies that the entire mantle overturns once each 2 b.y., a conclusion which would require a new interpretation of heat flow and chemical evolution problems.

REFERENCES CITED

Deffeyes, K. S., 1970, The axial valley: a steady-state feature of the terrain, *in* H. Johnson and B. L. Smith, eds., Megatectonics of continents and oceans: New Brunswick, New Jersey, Rutgers Univ. Press, p. 194–222.

Elsasser, W. M., 1969, Convection and stress propagation in the upper mantle, *in* W. K. Runcorn, ed., The application of modern physics to the earth and planetary interiors: London-New York, Wiley-Interscience, p. 223–246.

——— 1971, Sea-floor spreading as thermal convection: Jour. Geophys. Research, v. 76, p. 1101–1112.

Francheteau, J., C. G. A. Harrison, J. G. Sclater, and M. L. Richards, 1970, Magnetization of Pacific seamounts, a preliminary polar curve for the northeastern Pacific: Jour. Geophys. Research, v. 75, p. 2035–2061.

Isacks, B., L. R. Sykes, and J. Oliver, 1969, Focal mechanisms of deep and shallow earthquakes in the Tonga-Kermadec region and the tectonics of island arcs: Geol. Soc. America Bull., v. 80, p. 1443–1470.

Johnson, R. J., 1970, Active submarine volcanism in the Austral Islands: Science, v. 167, p. 977–979.

Karig, D. E., 1970, Ridges and basins of the Tonga-Kermadec island-arc system: Jour. Geophys. Research, v. 75, p. 239–254.

Kaula, W. M., 1970, Earth's gravity field: relation to global tectonics: Science, v. 169, p. 982–985.

McKenzie, D. P., 1967, Some remarks on heat flow and gravity anomalies: Jour. Geophys. Research, v. 72, p. 6261–6273.

——— 1969, Speculations on the consequences and causes of plate motions: Royal Astron. Soc. Geophys. Jour., v. 18, p. 1–32.

Moberly, R., in press, Origin of lithosphere behind island arcs, with reference to the western Pacific, *in* A volume to honor H. H. Hess: Geol. Soc. America Mem. 132.

Morgan, W. J., 1965, Gravity anomalies and convection currents: Jour. Geophys. Research, v. 70, p. 6175–6204.

——— 1971, Convection plumes in the lower mantle: Nature, v. 230, no. 5288, p. 42–43.

——— in press, Plate motions and deep mantle convection, *in* A volume to honor H. H. Hess: Geol. Soc. America Mem. 132.

Oversby, V. M., and P. W. Gast, 1970, Isotopic composition of lead from oceanic islands: Jour. Geophys. Research, v. 75 p. 2097–2114.

Vogt, P. R., O. E. Avery, E. D. Schneider, C. N. Anderson, and D. R. Bracy, 1969, Discontinuities in sea-floor spreading, *in* The world rift system: Tectonophysics, v. 8, p. 285–317.

Wilson, J. T., 1963a, Continental drift: Sci. American, v. 208, p. 86–100.

——— 1963b, Hypothesis of earth's behavior: Nature, v. 198, p. 925–929.

——— 1965a, Evidence from ocean islands suggesting movement in the earth, *in* Symposium on continental drift: Royal Soc. London Philos. Trans., ser. A, v. 258, p. 145–167.

——— 1965b, Submarine fracture zones, aseismic ridges and the International Council of Scientific Unions line; proposed western margin of the East Pacific ridge: Nature, v. 207, p. 907–911.

Wyss, M., 1970a, Apparent stresses of earthquakes on ridges compared to apparent stresses of earthquakes in trenches: Royal Astron. Soc. Geophys. Jour., v. 19, p. 479–484.

——— 1970b, Stress estimates for South American shallow and deep earthquakes: Jour. Geophys. Research, v. 75, p. 1529–1544.