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Intra-Plate Volcanism [and Discussion]

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Intra-plate volcanism

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[Plate 1]

A significant fraction of the volcanism on the Earth's surface cannot be associated with plate margin processes. The mantle plume hypothesis is one explanation for this volcanism. Convective plumes beneath rigid plates could be the result of turbulent thermal convection within the upper mantle. Although the hypothesis of nearly fixed mantle plumes is reasonably successful in explaining the direction and velocity of migration of intra-plate volcanism, there are a number of difficulties. It is particularly difficult to explain continued volcanism over extended linear distances. An alternative hypothesis for intra-plate volcanism is that magmas flow to the surface through lithospheric fractures. In this case intra-plate volcanism would be associated with tensional tectonics. Intra-plate volcanism and seismicity in Africa and the western United States are discussed in terms of these hypotheses.

INTRODUCTION

Most of the Earth's volcanic activity is directly associated with plate margins. Continuous volcanism on the oceanic ridge system forms the basaltic oceanic crust. As the surface plates diverge (spread) at an oceanic ridge, solid mantle must convectively rise to form new oceanic lithosphere. The volcanism at ocean ridges can be attributed to pressure-release melting in the rising mantle (Oxburgh 1965; Oxburgh & Turcotte 1968*a*). While much of the oceanic crust is of near-uniform thickness and chemical composition there are anomalous regions, e.g. Iceland, the Azores and the Galapagos Islands, where the oceanic crust is anomalously thick: it is important to decide whether these anomalous parts of the ridge system can be explained in terms of normal ocean ridge processes or whether additional mechanisms, such as deep mantle plumes, are required.

The volcanism associated with subduction zones is related to the descending lithospheric plates. Since pressure-release melting cannot be operative in a descending flow an alternative mechanism for heating and melting is required. This is probably frictional heating on the fault zone between the descending and overlying plates (McKenzie & Sclater 1968; Oxburgh & Turcotte 1968*b*). The frictional heating of the descending oceanic crust may produce magmas directly or the heating may drive off magmas and/or water which rise and partially melt the overlying mantle.

In some cases volcanic activity creates new oceanic crust in marginal basins behind subduction zones (Karig 1974). The similarity of the resulting basaltic crust to normal oceanic crust indicates that the basic processes are similar. Again pressure-release melting due to ascending convection of mantle rock is probably occurring.

Our purpose here is to examine intra-plate volcanism which cannot be associated with plate margin processes. A prime example is the Hawaiian Islands where extensive volcanism has created a chain of volcanic islands and seamounts near the centre of one of the largest

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plates. There are a number of other volcanic chains and isolated islands which are also examples of intra-plate volcanism, e.g. Easter Island and the Cook-Austral Islands.

Intra-plate volcanism also occurs on the continents. It is particularly extensive in Africa. In some cases it is difficult to tell whether continental volcanism is truly intra-plate or is associated with the development of a plate margin, e.g. the East African Rift system regarded by some as part of the worldwide ocean ridge system while by others as typical of intra-plate volcanism. The Tibesti volcanics of northern Africa are an example of continental volcanism not associated with a plate margin. There is also extensive volcanism in the western United States. This region could be considered as associated with some kind of sub-continental extension of the East Pacific Rise, or as intra-plate volcanism.

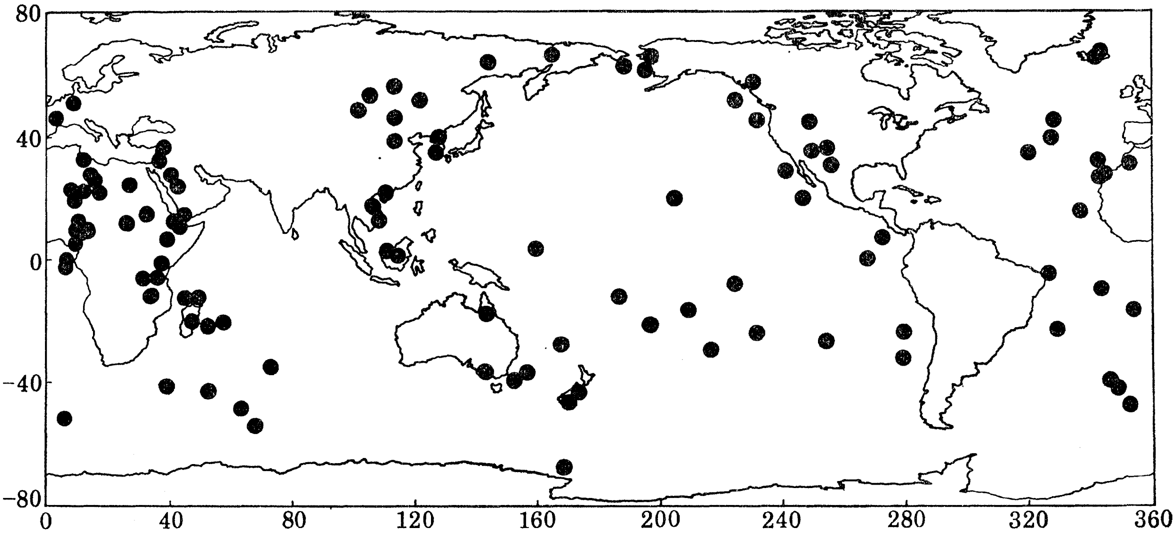

FIGURE 1. Distribution of hot spots (after Burke & Wilson 1976).

MANTLE PLUMES AND HOT SPOTS

The simplest explanation for intra-plate volcanism is to assume that mantle convection is occurring beneath lithospheric plates without fracturing them to create new plate boundaries. Ascending convection can lead to pressure-release melting and the magmas penetrate the plates leading to volcanism. This is the mantle plume hypothesis.

An early approach of this kind was that of Wilson (1963), who proposed that the Hawaiian Islands were the result of a melting anomaly in the upper mantle; the motion of the lithosphere over the mantle caused the island chain to propagate. Morgan (1971, 1972*a*, *b*) further hypothesized that deep mantle plumes were responsible for areas of anomalous volcanism both intra-plate and on ridges. In his original paper Morgan (1971) considered sixteen plumes and suggested that they were stationary with respect to each other.

However, it was soon recognized that the number of areas of intra-plate and ridge volcanism which would qualify as plume tracks was much larger than sixteen. These areas of anomalous volcanism are often referred to as 'hot spots'. The world-wide distribution of hot spots proposed by Burke & Wilson (1976) is shown in figure 1. One hundred and six hot spots are shown; not included are nine in Antarctica. Clearly hot spots are not uniformly distributed over the

Earth's surface. There are numerous hot spots in Africa, none in South America, and relatively few in North America, Europe, and Asia. A classification of many of the hot spots has been given by Wilson (1973). Many, but not all, of the hot spots shown in figure 1 are also associated with local crustal upwarping.

One of the assumptions of the deep mantle plume hypothesis is that hot spots move very little with respect to each other. A number of studies have been carried out to determine whether this is in fact the case. Clague & Jarrard (1973), Molnar & Atwater (1973), Winterer (1973), and Jackson (1976) conclude that the trends of several Pacific island chains can be explained if the plumes migrate with respect to each other at a few centimetres per year. Studies of relative hot spot motions in the Atlantic (Burke, Kidd & Wilson 1973; Molnar & Francheteau 1975) also show relative velocities of several centimetres per year. However, since the plate velocities in the Atlantic are also of the order of several centimetres per year, the relative migration of the hot spots is much greater in the Atlantic. In particular the motion of the Iceland hot spot does not appear to be consistent with the motion of the hot spots in the South Atlantic. A comprehensive world-wide study of relative hot spot motions by Minster, Jordan, Molnar & Haines (1974) confirms these conclusions.

Propagation of Pacific volcanic chains

The trends of the Pacific volcanic chains gave rise to the original fixed mantle plume hypothesis (Morgan 1972*b*; Wilson 1973), and the Pacific plate provides a good test for the hypothesis since the plate is moving relatively rapidly over the mantle. Selected volcanic chains are illustrated in figure 2.

Probably the most comprehensive study of absolute plate motions on a world-wide basis was carried out by Minster *et al.* (1974). These authors inverted the trends of twenty linear volcanic chains and aseismic ridges by use of the fixed hot spot hypothesis and minimized hot spot migration. The resultant pole they obtained for the absolute (with respect to a fixed mantle) rotation of the Pacific plate was at 67.3° N and 59.4° W. The angular velocity of the plate was 0.83°/Ma. The predicted directions of propagation of the Pacific island chains are shown in figure 2. Although the general trends of the volcanic chains are in the predicted directions, significant deviations are seen. This is particularly true for the Marquesas chain where the deviation is about 30°.

In addition to predicting a direction of propagation for volcanic islands, such a model also predicts a rate of propagation which may be tested by the dating of volcanic rocks from islands and seamounts. The data have been summarized by Jackson (1976) and the ages for the island chains in figure 2 are given in figure 3. The age of the volcanic rocks on each island or seamount is plotted against the distance of that island or seamount from the currently active volcanic island at the end of the chain. Also included in figure 3 is the speed of propagation of the chain predicted by the rotation pole and angular velocity given by Minster *et al.* (1974).

The propagation of the Hawaiian–Emperor chain shown in figure 3*a* is in good agreement with the predicted rate of 8.9 cm/a. This is even true beyond the change in the trend of the chain about 40 Ma B.P.

Several other chains exhibit similarities to this chain both in trend and age progression. However, there are also striking misfits, e.g. the Cook–Austral chain. The mantle plume hypothesis provides an attractive means of explaining these Pacific volcanic chains, although without extension and/or modification there are many features that remain unexplained – particularly synchronous volcanism along extended lengths of a chain.

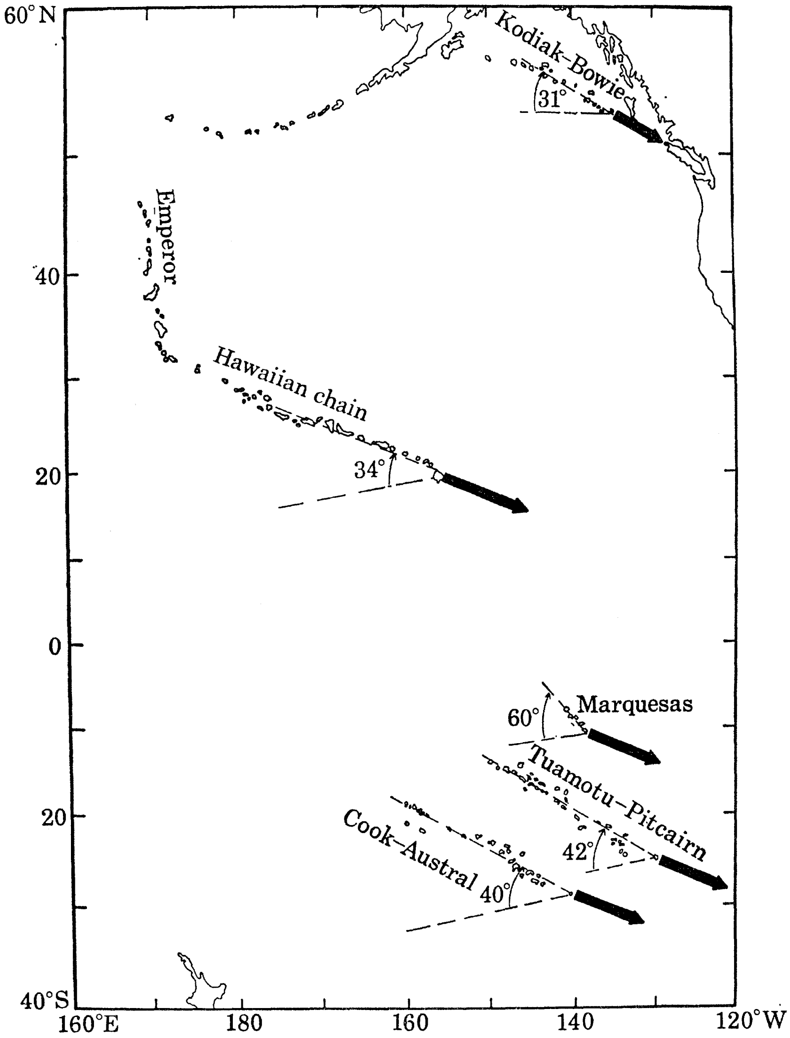

Figure 2. Selected linear volcanic chains in the Pacific. The arrows show the directions of propagation about the absolute rotational pole given by Minster *et al.* (1974). The angles shown are between the island chains and the direction of sea floor spreading at that point on the plate.

Plumes and mantle convection

Plate tectonics can be explained directly in terms of mantle convection. The lithospheric plates are the thermal boundary layers of mantle convection cells (Turcotte & Oxburgh 1967). The hot mantle behaves like a fluid on geological time scales; radioactive heating causes the mantle to become gravitationally unstable. The governing non-dimensional parameters are the Rayleigh number ($Ra = gh^5 \alpha H / c_p \kappa^2 \upsilon$, where g is gravitational acceleration, h depth of convection, α coefficient of thermal expansion, H radioactive heat generation per unit mass, c_p specific heat at constant pressure, υ kinematic viscosity, κ thermal diffusivity) and the

Prandtl number ($Pr = \upsilon/\kappa$). In the mantle the Prandtl number is very large so that the limit of infinite Prandtl number is appropriate.

Experiments on thermal convection in a fluid layer heated from below with a constant viscosity and high Prandtl number have been carried out by Somerscales & Dropkin (1966). These experiments were carried out between two rigid boundaries so that the critical Rayleigh number for the onset of thermal convection was 1708. A vertical shadowgraph of the convection pattern at $Ra = 3040$ is shown in figure 4*a*. The convection cells were primarily in the form of laminar two-dimensional rolls. A boundary-layer analysis of this type of convection pattern has been given by Turcotte (1967). Thin thermal boundary layers develop on the horizontal boundaries of the fluid layer. These thermal boundary layers become gravitationally unstable and form two-dimensional thermal plumes. The gravitational body forces on the thermal plumes drive the viscous core flow.

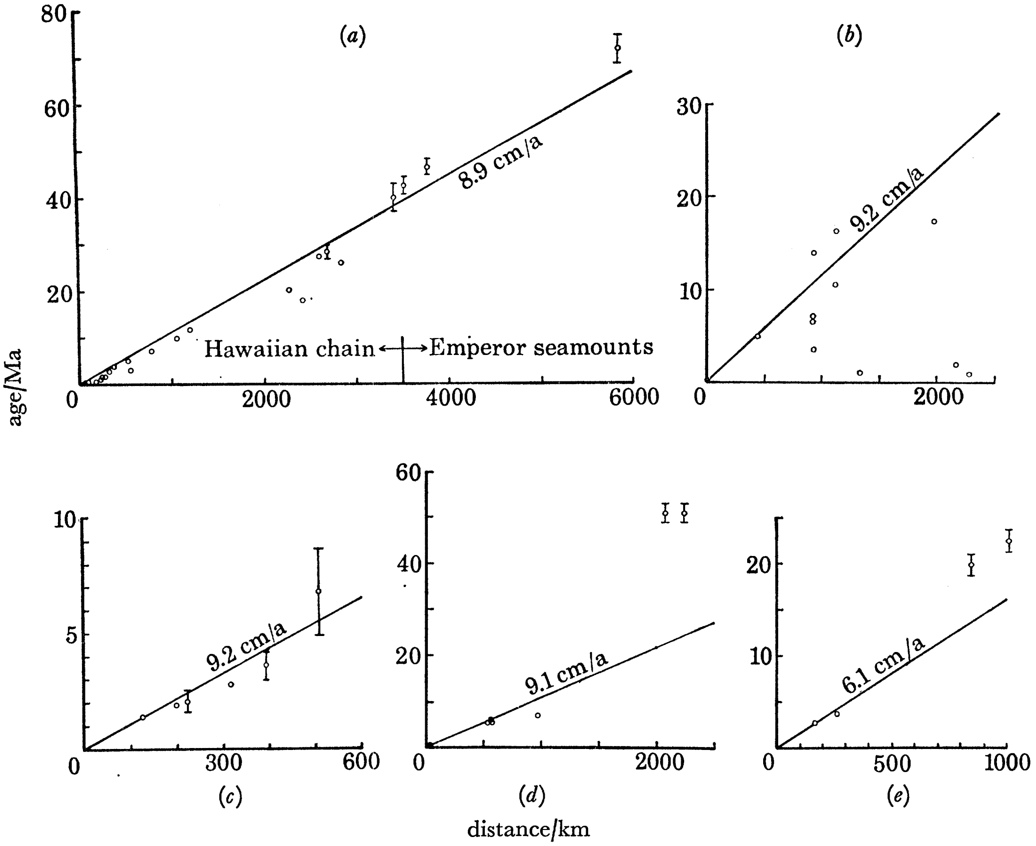

FIGURE 3. Ages of volcanic rocks from selected linear volcanic chains in the Pacific as a function of the distance from the currently active volcanic island at the end of the chain: (*a*) Hawaii–Emperor; (*b*) Cook–Austral; (*c*) Marquesas; (*d*) Pitcairn–Tuamotu; (*e*) Kodiak–Bowie. The solid line is the rate of propagation of the chain predicted by Minster *et al.* (1974).

This type of two-dimensional thermal convection can be directly associated with mantle convection (Turcotte & Oxburgh 1967). The thermal boundary layer adjacent to the upper boundary of the fluid layer is directly analogous to the thickening oceanic lithosphere. The descending two-dimensional thermal plume is directly analogous to the descending lithosphere at an oceanic trench.

However, laboratory experiments at higher Rayleigh numbers show that the pattern of two-dimensional rolls breaks down into a complex three-dimensional convection pattern. At a Rayleigh number of about 22000 the two-dimensional convection becomes unstable to transverse disturbances and laminar three-dimensional convection develops (Krishnamurti 1970*a*). A top view of the convection pattern at $Ra = 43300$ is shown in figure 4*b*. At Rayleigh numbers greater than about 55000 the convective flow becomes unsteady (turbulent). This transition as a function of Prandtl number has been studied by Krishnamurti (1970*b*). A top view of turbulent thermal convection at $Ra = 131000$ is shown in figure 4*c*.

Clearly the laboratory results described above cannot be applied directly to the interior of the Earth's mantle. It is doubtful that such effects as internal heating, viscous dissipation (Turcotte, Hsui, Torrance & Schubert 1974) and non-Newtonian viscosity (Parmentier, Turcotte & Torrance 1976) will significantly affect the qualitative behaviour described above. However, the mantle viscosity is expected to have an exponential dependence on temperature and pressure. This dependence strongly affects two-dimensional thermal convection (Torrance & Turcotte 1971*a, b*) and would be expected to influence significantly the transitions to three-dimensional and unsteady flows. It should also be emphasized that the Rayleigh number for the Earth's mantle can be only crudely estimated. There are considerable uncertainties in the mantle viscosity and the depth of convection. Taking $\upsilon = 2.6 \times 10^{17}\ \mathrm{m^2/s}$ based on post-glacial rebound studies (Cathles 1975), $h = 700$ km upper mantle convection, $\kappa = 10^{-6}\ \mathrm{m^2/s}$, $c_p = 10^3 \mathrm{j/(kg\ K)}$, $H = 8 \times 10^{-12} \mathrm{j/(kg\ s)}$, $g = 10\ \mathrm{m/s^2}$, and $\alpha = 3 \times 10^{-5}\ \mathrm{K^{-1}}$, we find that $Ra = 1.5 \times 10^6$. This is clearly within the turbulent régime based on the laboratory experiments and unsteady thermal convection beneath rigid lithospheric plates could possibly contribute to intra-plate volcanism.

It has been shown theoretically by Richter (1973) and experimentally by Richter & Parsons (1975) that a moving upper boundary stabilizes thermal convection in the form of longitudinal rolls. The axes of the longitudinal rolls are aligned with the direction of motion of the upper boundary. These authors suggest that there are two scales of mantle convection. The primary scale of convection is associated with the lithospheric plates. The secondary scale consists of longitudinal rolls within the asthenosphere and mesosphere.

There are significant differences between the analysis and experiments involving a moving upper boundary and the two-scale convection problem within the Earth's mantle. The thickening lithosphere incorporates the cooling boundary layer thereby removing the unstable cooling rock from the flow field. The mechanical strength of the lithosphere inhibits the gravitational instability of the upper thermal boundary layer.

Whether a regular secondary convection pattern occurs in the mantle must be tested by observations. The trends of the island chains of the Pacific do not appear to support a longitudinal roll pattern since they are not equally spaced and do not have a systematic relation to the ocean ridge system. Marsh & Marsh (1976) have examined the free-air gravity anomalies between degrees 12 and 22 and conclude that there are NE–SW trending anomalies across the Pacific. These authors argue that these anomalies support the hypothesis of two scales of convection. While these patterns are interesting they are close to the noise level of the data and are not obviously restricted to oceanic regions.

The original hypothesis of deep mantle plumes proposed by Morgan (1972*a*) envisioned narrowly confined ($r \approx 100$ km), cylindrical, thermally convected flows. It was pointed out by Tozer (1973) that the concept of narrowly confined flows was not applicable to high

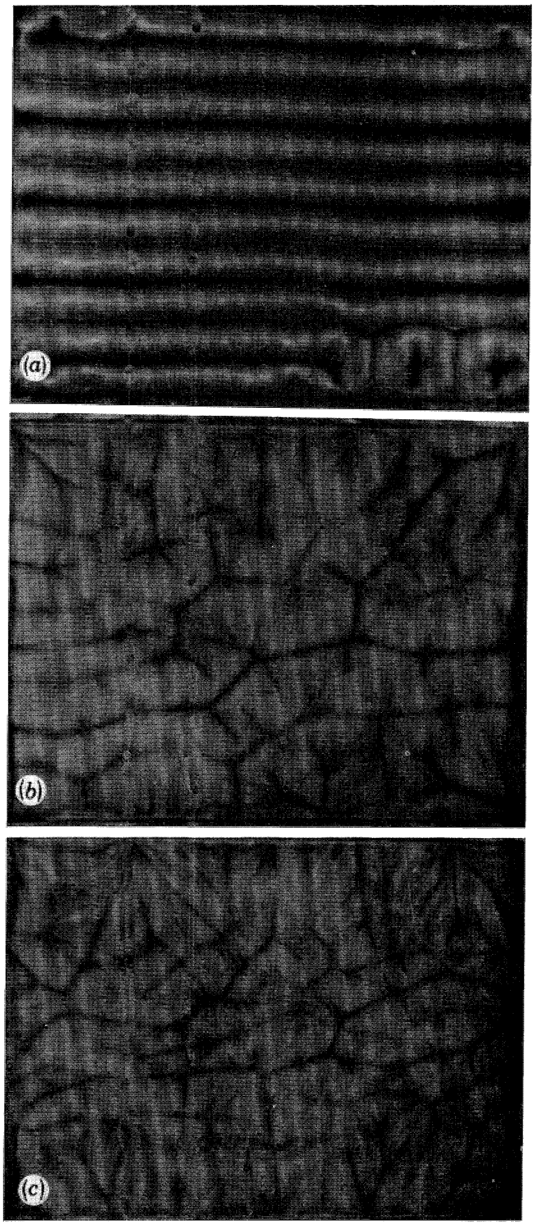

FIGURE 4. Planform shadowgraphs of thermal convection in a high Prandtl number fluid heated from below and cooled from above with rigid horizontal boundaries. (*a*) $Ra = 3040$; steady, two-dimensional; (*b*) $Ra = 43\,300$, steady, three-dimensional; (*c*) $Ra = 131\,000$, unsteady (turbulent).

Prandtl number fluids. Thermal convection at high Prandtl numbers may result in narrowly confined thermal boundary layers or thermal plumes because of the slow diffusion of heat, but high viscosity leads to a rapid diffusion of vorticity and a diffuse velocity structure.

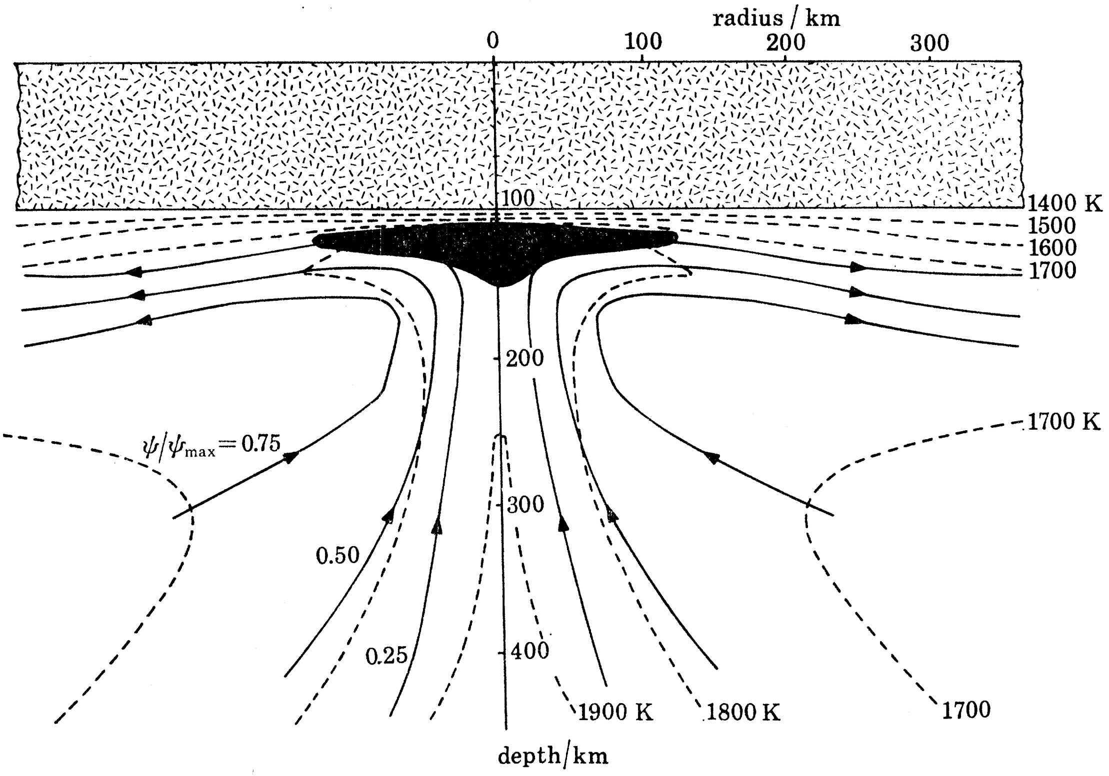

FIGURE 5. Structure of a shallow mantle plume impinging on a lithospheric plate (Parmentier *et al.* 1975); flow lines solid; isotherms dashed; partial melt zone dark.

With an exponentially temperature-dependent viscosity it might be expected that the high temperature region would also have high flow velocities because of the reduction in viscosity. Two-dimensional plumes of this type have been studied by Yuen & Schubert (1976). They find that the flow is primarily within the thermal plume if the temperature difference is sufficiently large. Numerical calculations for the structure of cylindrical mantle plumes have been carried out by Parmentier, Turcotte & Torrance (1975) using an exponentially temperature and depth dependent viscosity. These authors found that relatively narrow mantle plumes could be obtained if the convective flow is confined to the upper mantle and if the heating is primarily from below (from the lower mantle). An illustration of this type of shallow mantle plume obtained from the numerical calculations is given in figure 5. Axisymmetric shallow mantle plumes of this type could be associated with unsteady (turbulent) thermal convection in the upper mantle.

Deffeyes (1972) suggested that deep mantle plumes could be driven by phase changes. A phase change is a local driving force which might drive a local flow; however, it would not induce a narrowly confined deep mantle plume.

An alternative to thermal plumes is chemical (or compositional) plumes. Anderson (1976) has proposed that surface hot spots result from compositional plumes which rise from the lower mantle. Certainly the diapiric upwelling of lighter material from the lower mantle can result in relatively narrow 'plumes' impinging on the lithosphere. However, the instability mechanisms which lead to compositional plumes have not been clearly delineated. There is the general depth constraint on all types of plume that if they originate at great depth, the amount of melting that should result from adiabatic decompression during the ascent of the plume is much greater than that which is deduced from the petrology of surface volcanics. If anything, the volcanic suites from oceanic islands represent lower degrees of partial melting than ridge basalts.

It was shown by Gruntfest (1963) that the heat produced by viscous dissipation in a fluid with a strongly temperature dependent viscosity could lead to an instability. Anderson & Perkins (1974) have suggested that this type of thermal runaway could be responsible for magma generation in the asthenosphere and the type of volcanism observed in the western United States. The highly idealized analyses of this problem carried out to date are not applicable to the mantle. Numerical calculations of thermal convection with viscous dissipation show that the dissipation damps rather than accelerates the flow (Turcotte *et al.* 1974).

Shaw & Jackson (1973) suggest that dissipative heating and the consequent partial melting in the asthenosphere produce dense residues that sink in the mantle. The sinking residues create a 'gravitational anchor' which fixes the melting anomaly to the mantle. The descending residuum induces further flow and shear heating and melting. There is, however, strong evidence that the residues are less dense than the starting material (O'Hara 1975; Oxburgh & Turcotte 1976; Boyd & McCallister 1976; Oxburgh & Parmentier 1978 (this volume, pp. 415–429)).

LITHOSPHERIC FRACTURES

Although some fraction of intra-plate volcanism may be associated with convective processes, it is difficult to explain all intra-plate volcanism in this way. It is recognized that most intra-plate volcanism is associated with tensional tectonics, e.g. the East African Rift, the Rhine Graben and the Basin and Range Province in the western United States.

An alternative hypothesis for intra-plate volcanism is that magmas from the asthenosphere penetrate the lithosphere where tensional fractures occur. Betz & Hess (1941) proposed that the Hawaiian Islands were the result of a propagating fracture. Jackson & Wright (1970) and Green (1971) discussed the petrology of the Hawaiian Islands in terms of magmas rising through a cracked lithosphere. Turcotte & Oxburgh (1973, 1976) discussed the geometry of lithospheric fractures and the sources of the required stresses.

In order for tensional failures of the lithosphere to occur, elastic stresses must build up over long periods of time. The presence of an outer swell due to the bending of the lithosphere under the load of the Emperor seamounts (Watts & Cochran 1974) indicates that elastic stresses are not relaxed for at least 40 Ma. The failure of the lithosphere under tension is not likely to resemble the tensional failure of a rock in the laboratory. We have previously suggested (Turcotte & Oxburgh 1973) that the lithosphere will fail plastically according to the failure criteria given by Bijlaard (1935). A thin, malleable plate under tension will fail plastically at an angle $\theta = \frac{1}{2} \arccos \frac{1}{3} = 35° \, 16'$ as illustrated in figure 6.

The trend of the Hawaiian chain lies at an angle of 34° with respect to the direction of sea floor spreading defined locally by the magnetic anomalies and fracture zones. If this angle is associated with the plastic failure angle θ, the conclusion is that the tensional stress causing the plastic failure of the Pacific plate in the vicinity of the Hawaiian Islands is nearly perpendicular to the direction of sea floor spreading. Stresses with this orientation and large enough to cause failure of the lithosphere could result from thermal contraction of the oceanic plate (Turcotte 1974*a*).

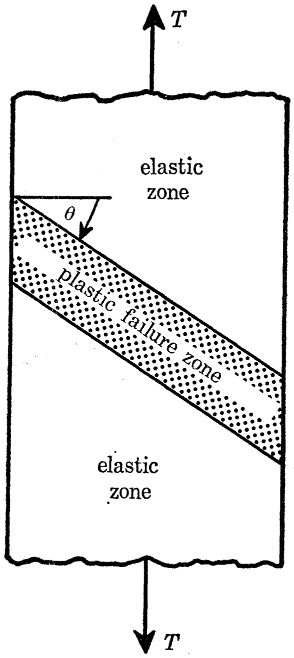

Figure 6. The plastic failure of a thin plate under a tensional load.

The angles between the trends of the other island chains of the Pacific and the directions of sea floor spreading are shown in figure 3. With the exception of the Marquesas the agreement with the plastic failure angle, 35°, is reasonably good. The discrepancies are very similar to those associated with the fixed plume hypothesis. Such deviations are perhaps not surprising insofar as there must be significant contributions to the stress fields of plates both from membrane effects associated with the ellipticity of the Earth (Turcotte 1974*b*) and stresses due to the driving mechanisms of plate tectonics (Forsyth & Uyeda 1975).

The rate of propagation of the tensional fractures is not predicted directly by the hypothesis. If thermal stresses cause the tensional failures then it might be expected that the failure would propagate in the direction of sea floor spreading at about the sea floor spreading velocity. Taking the sea floor spreading velocity to be 6 cm/a the propagation velocity would be $6/\cos 35° = 7.3$ cm/a.

An essential part of the lithospheric fracture hypothesis is that the fracture allows an essentially artesian flow of magma from the asthenosphere to the surface. Normally lithosphere behaves as an impermeable cap rock but under a tensional failure the hydrostatic head drives the magma to the surface. The physical processes by which magmas are transported through the lithosphere are poorly understood. However, Weertman (1971) argued that magma in the asthenosphere can nucleate a crack in the lithosphere provided that the lithosphere is in tension. The magma then follows the propagating fracture to the surface. Anderson & Grew (1977) suggest that magma migrates through a lithospheric plate in tension due to stress corrosion.

AFRICA

As shown in figure 1, Africa is the site of extensive intra-plate volcanism. Burke & Wilson (1972) argued that Africa was stationary with respect to the mantle and that mantle plumes therefore could produce fixed volcanic centres on the continent. However, the world-wide studies of Minster *et al.* (1974) concluded that Africa is moving towards the northeast at about 2 cm/a.

The intra-plate volcanism of Africa is associated with several different tectonic settings. Much of the volcanism is related to the East African Rift system (figure 7), which exhibits

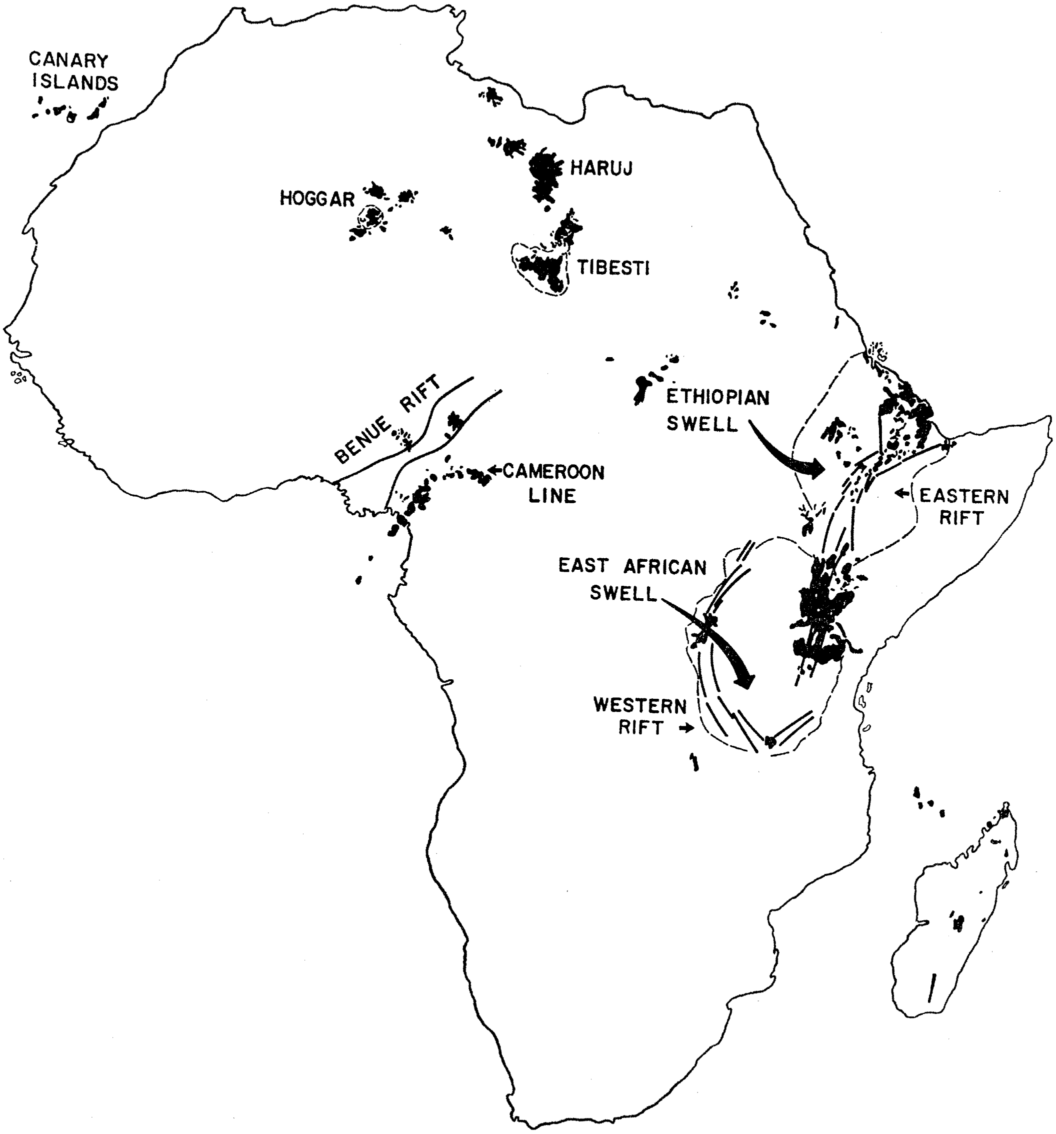

FIGURE 7. Miocene and younger (< 26 Ma B.P.) volcanics of Africa shown as dark areas.

tensional tectonics and may be propagating southward. Burke & Wilson (1976) associate the East African Rift with seven hot spots. These hot spots are defined by crustal doming which is an important aspect of the rift system. We have previously suggested (Oxburgh & Turcotte 1974) that the East African Rift is a tensional failure of the African plate due to membrane stresses. The doming and volcanism are secondary features resulting from the tensional failure.

Other areas of extensive recent volcanism in Africa are the Tibesti area in northeast Chad (figure 7) and the Hoggar to the west. In addition to volcanism, these areas are clearly domed but do not appear to be associated with any linear features. North of Tibesti the Haruj volcanics are not associated with any apparent doming.

Adjacent to the continental margin in the Gulf of Guinea is the Cameroon Line of recent volcanics. This is a linear series of volcanic centres which passes from oceanic on to continental crust. These volcanics are near the Benue Rift interpreted by Burke & Dewey (1973) as the failed arm of a triple junction. There are two similar volcanic lines on the Atlantic continental margin of South America (Fernando Noranha–Mocas and Martin Vaz–Trinidade).

Further to the north on the continental margin of Africa there is the recent volcanism of the Canary Islands. This is a group of volcanic islands similar to the Cape Verde Islands further to the south. The association of the Canary Islands with a propagating fracture has been discussed by Anguita & Herman (1975). Such volcanic centres which lie near the continental margins of the South Atlantic may be associated with large stresses caused by changes in crustal thickness (Bott & Dean 1972). The association of intra-plate volcanism with localized lithospheric fractures due to these stress concentrations is perhaps more reasonable than a fortuitous distribution of mantle plumes.

Intra-plate volcanism of the western United States

The western United States is an area of extensive volcanism. The distribution of Quaternary (< 1.5 Ma B.P.) and Pliocene (1.5–7.0 Ma B.P.) volcanics is given in figure 8. Since the San Andreas fault is recognized as a major plate boundary between the Pacific and North American plates the volcanism of this area may be classified as plate margin volcanism. However, it extends some 1500 km from the margin.

Morgan (1971) associated one of his original sixteen plumes with the extensive volcanics of the Snake River Plain. Burke & Wilson (1976) located four hot spots in the western United States (see figure 1). Even this number of hot spots cannot explain all the volcanics of the area. Scholz, Barazangi & Sbar (1971) related the volcanism and tectonics of the western United States to behind-arc spreading (Karig 1974).

The plate tectonic evolution of the western United States has been studied in detail by Atwater (1970). If the San Andreas fault is the major boundary between the Pacific and North American plates then the relative velocity across the boundary is 5.5 cm/a (Minster *et al.* 1974).

The half-spreading rate on the Juan de Fuca Ridge is 2.9 cm/a. However, the spreading direction as defined by the Blanco Fracture Zone is not parallel to the trend of the San Andreas fault. This difference led Atwater (1970) to conclude that active subduction is occurring along the Oregon–Washington coast; taking the difference in the velocity vectors the rate of subduction is about 1.5 cm/a as shown in figure 9. Active subduction explains the line of active andesitic volcanoes extending from Mt Baker in northern Washington to Mt Lassen in northern California.

Probably the most extensive recent volcanism in the western United States is associated with the Snake River Plain. A plume is postulated beneath the Yellowstone caldera at the eastern end of the plain. The trend of the eastern section of the plain is in good agreement with the direction of propagation predicted by Minster *et al.* (1974). A large number of volcanic rocks in the Snake River Plain have been dated by Armstrong, Leeman & Malde (1975). The ages of these (and other) rocks as a function of the distance from the Yellowstone caldera are given in figure 10. Also included is the rate of propagation of the volcanics predicted by Minster *et al.* (1974) on the basis of the fixed plume hypothesis. Although the age of the oldest

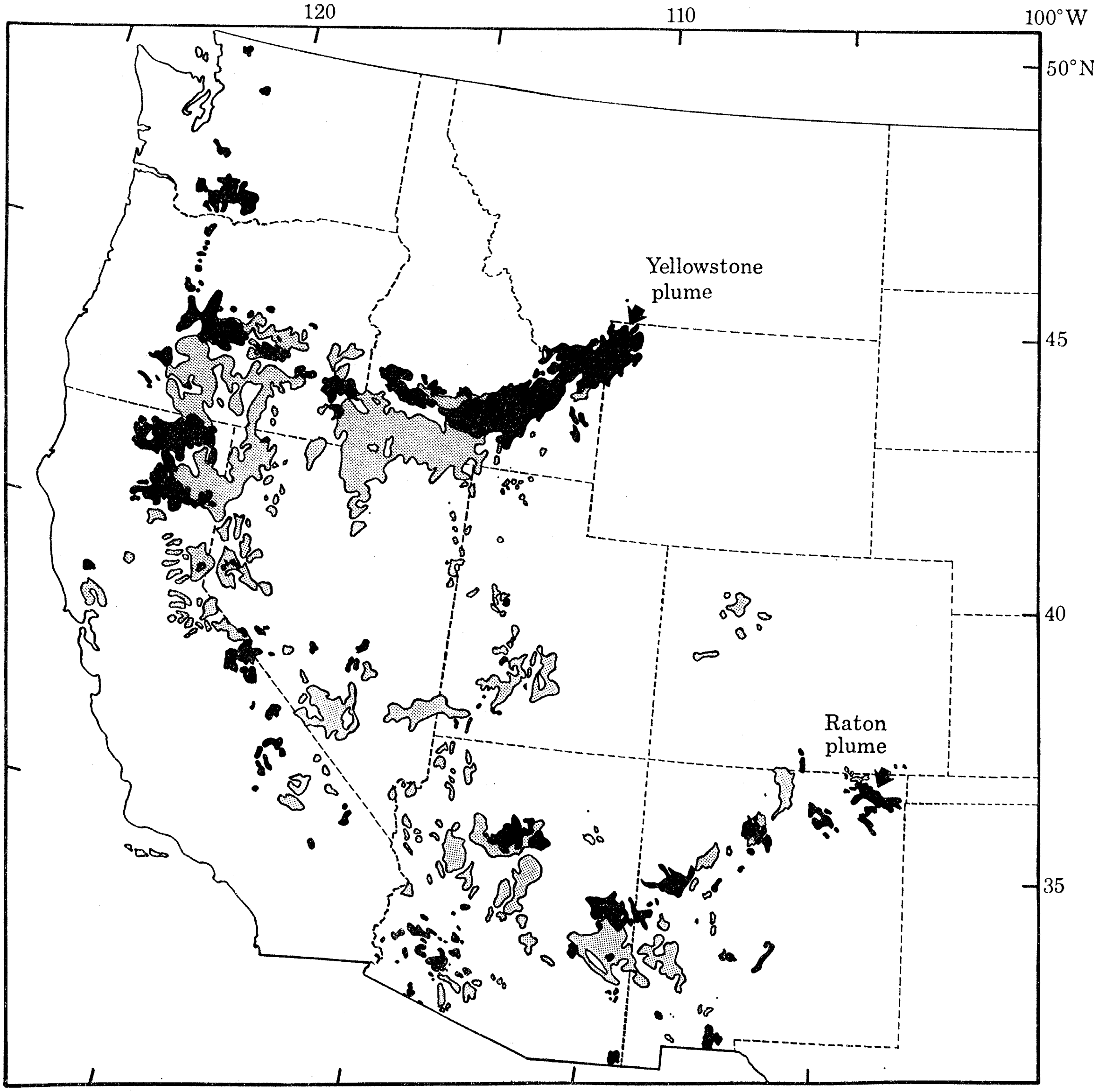

FIGURE 8. The distribution of Quaternary (< 1.5 Ma B.P.; dark areas) and Pliocene (1.5–7.0 Ma B.P.; shaded areas) in the western United States (King & Beikman 1974). Also shown are the hypothetical Yellowstone and Raton plumes (Suppe *et al.* 1975).

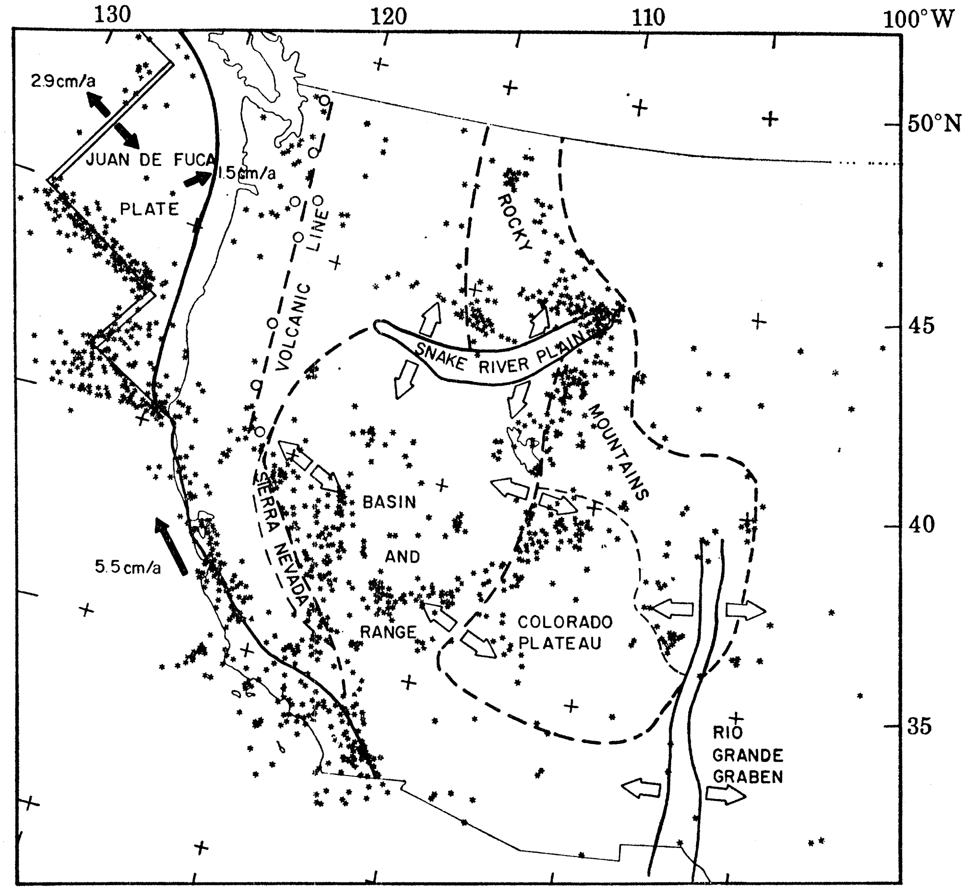

Figure 9. Distribution of seismicity in the western United States (stars). Also shown are various geological provinces. Solid arrows give directions of plate motion where velocities are known; open arrows give stress directions inferred from seismic focal mechanism studies (Smith & Sbar 1974).

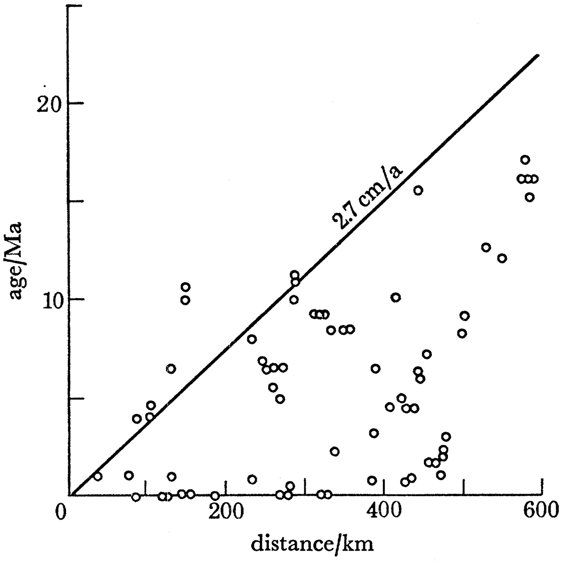

Figure 10. Ages of volcanic rocks in the Snake River Plain as a function of distance from the Yellowstone Caldera (Armstrong *et al.* 1975). The solid line is the rate of propagation predicted by Minster *et al.* (1974).

volcanics tends to increase with distance from Yellowstone the agreement with the predicted rate is rather poor. Of course, this can be attributed to the migration of the plume. It is also evident from figure 10 (as well as figure 8) that there are recent volcanics along much of the length of the Snake River Plain. It is difficult to associate extensive basaltic volcanism over a distance of 500 km directly with a plume beneath Yellowstone.

Suppe, Powell & Berry (1975) attribute much of the volcanism of the western United States to two plume tracks. In addition to the Yellowstone plume they also proposed that a plume is located near Raton, New Mexico. This is shown in figure 8. They would extend this plume track to the White Mountains of eastern Arizona and possibly as far as the Gulf of California. These authors concede, however, that these plumes cannot explain all the intra-plate volcanism of the western United States.

The distribution of seismicity in the western United States is shown in figure 9. The main area of active volcanism, seismicity, and tectonics is bounded on the east by the Rio Grande Graben, and on the north by the Snake River Plain, although there is considerable seismicity in the northern Rockies. The Basin and Range Province is an extensive zone of horst and graben tectonics, high heat flow, hot springs, and recent volcanism. Focal mechanism studies (Smith & Sbar 1974) indicate that the entire area is in tension with the directions indicated by the arrows in figure 9.

Atwater (1970) and Christiansen & Lipman (1971) attribute the tensional tectonics and the volcanism of the western United States to forces exerted at the plate boundary, with tensional failures of the lithosphere permitting the flow of magma to the surface. Smith *et al.* (1974) suggest that the Snake River volcanics have followed a pre-existing zone of weakness and that a wedge-shaped fracture is propagating eastward in response to intra-plate deformation.

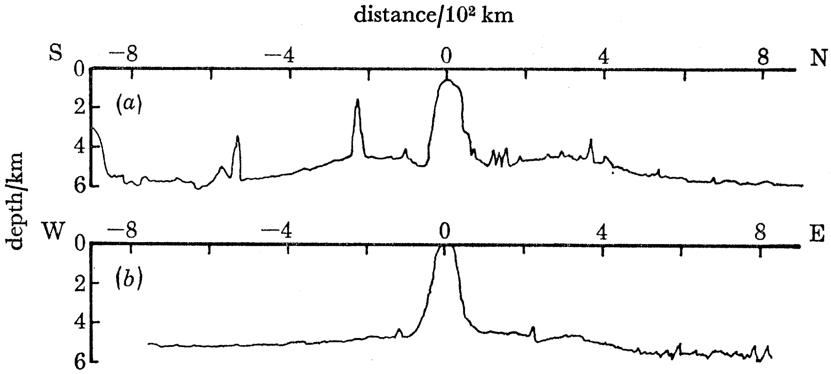

FIGURE 11. Topography (*a*) of the Hawaiian swell from Vema legs V2105 and V2004 (Watts 1976) and (*b*) of the Bermuda swell from Vema legs V2610 and V2602 (Talwani *et al.* 1974).

LITHOSPHERIC SWELLS AND BERMUDA

The Hawaiian swell is a prominent feature of the topography of the central Pacific. It is a near-circular elevation of the sea floor centred on the Hawaiian Islands. The amplitude of the swell is near 1.5 km and its radius is approximately 500 km (figure 11*a*), and there is a long-wavelength positive free-air gravity anomaly of about 15 mGal.†

The magnitude both of the topographic swell and of the gravity anomaly are such that they

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† 1 Gal = 1 cm s^{-2}.

could be associated with a mantle plume beneath Hawaii. The topography is difficult to explain by a propagating crack.

In addition, however, to the many island and seamount chains, some of which we have discussed, there are also a large number of isolated volcanic islands and seamounts. An example which has been studied in some detail is Bermuda (Reynolds & Aumento 1974). A suite of younger volcanics has an age of about 33 Ma and an older group 60–110 Ma; the sea floor beneath Bermuda has an age of about 110 Ma. Bermuda is not part of a volcanic chain and is difficult to explain by mantle plume.

It is interesting to note that, like Hawaii, Bermuda is associated with a lithospheric swell (figure 11*b*; Talwani, Hayes, Pitman & Aitken 1974). The amplitude of the swell is smaller (approximately 700 m) than the Hawaiian but the radius is similar. Since Bermuda is moving with respect to the mantle, a mantle plume that produced volcanism at Bermuda 33 Ma ago would not be centred at Bermuda today. Therefore if the Bermuda swell is not associated with a mantle plume, the Hawaiian swell may not be either.

Seismic studies

An important source of evidence for the existence of deep or shallow mantle plumes is seismic studies. Seismic studies can be used to search for velocity anomalies and/or regions of seismic attenuation in the mantle. An anomalous zone of seismic velocities and attenuation has been reported in the vicinity of the core-mantle boundary beneath Hawaii by Kanasewich, Ellis, Chapman & Gutowski (1972, 1973) and by Kanasewich & Gutowski (1975). A number of deep mantle inhomogeneities, including one beneath Hawaii but also including several others not associated with surface hot spots, were reported by Julian & Sengupta (1973). However, the interpretation of these data has been questioned by Wright (1975) and Green (1975). These authors suggest that the observed anomalies are due to upper mantle inhomogeneities beneath the receiving seismic arrays in North America. At present the seismic evidence for deep mantle plumes beneath hot spots is, at best, ambiguous. Further verification of the velocity anomaly beneath Hawaii is required. Also, deep anomalies beneath other hot spots must be found. Otherwise, the anomaly beneath Hawaii could be considered a coincidence.

Hadley, Stewart & Ebel (1976) report a velocity anomaly at depths between 200 and 400 km beneath Yellowstone using similar techniques. Ellsworth *et al.* (1975) have reported on travel delays associated with teleseismic arrival at 21 stations on Kilauea Volcano in Hawaii. They found no velocity anomalies at depths greater than 30 km. If any type of mantle convection were occurring beneath surface hot spots a zone of anomalous mantle would be expected such as occurs beneath ocean ridges and beneath behind-arc spreading centres. The failure to find anomalous mantle beneath intra-plate hot spots such as Hawaii suggests that these hot spots are not associated with special mantle features.

Seismic studies are also relevant to the propagating fracture hypothesis. Focal mechanism studies of major earthquakes can give information on the state of stress which led to the seismic rupture. On 27 June 1962, a magnitude 6.1 earthquake occurred near the Kaoiki fault system on the island of Hawaii. This earthquake has been studied by Koyanagi, Krivoy & Okamura (1966). They concluded that the earthquake was due to either right-lateral strike-slip movement on a fault trending N 25° E or left-lateral strike-slip movement on a fault trending N 65° W (figure 12). The former direction follows the trend of the island chain, the latter direction is roughly parallel to the Kaoiki fault trace. If this earthquake was due to a simple shear failure of the britle lithosphere under tension the direction of the tension would be N 15° W. This is within 2° of the direction of tension (N 13° W) given by the plastic failure hypothesis.

On 26 April 1973 a magnitude 6.2 subcrustal earthquake occurred beneath the northeast coast of the island of Hawaii. First motion studies of this earthquake reported by Koyanagi, Endo & Ward (1976) reported that the earthquake was due to either right-lateral strike-slip movement on a fault trending N 30° E or left-lateral strike-slip movement on a fault trending N 70° W. If this earthquake was due to a similar simple shear failure the direction of the tension would be N 20° W. Despite being at different depths and different locations the focal mechanisms and the direction of the inferred tensional stress for the two earthquakes agrees both with each other and with the predictions of the plastic failure hypothesis.

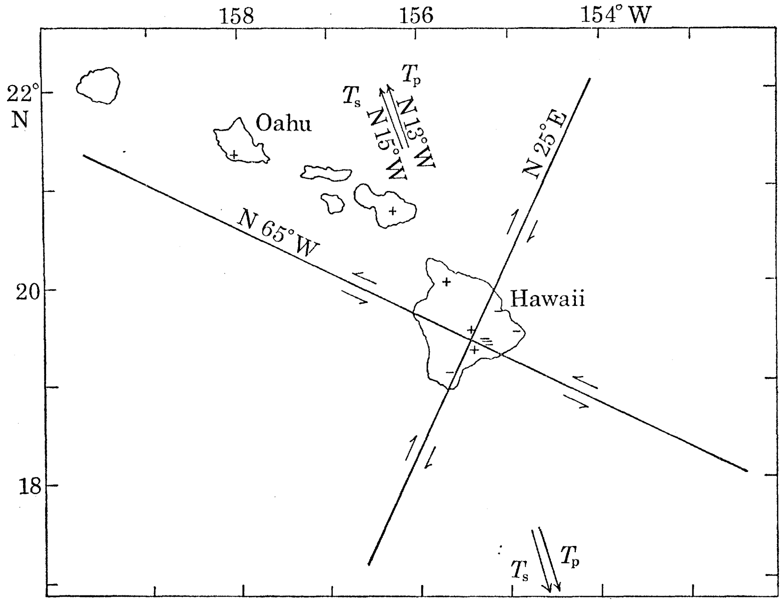

Figure 12. Focal mechanism study of the 27 June 1962 Hawaiian earthquake (after Koyanagi *et al.* 1966). The deduced direction of tensional stress is given by T_s and the direction based on the plastic failure hypothesis is given by T_p.

Studies of the focal mechanisms of earthquakes in southern Africa have been carried out by Maasha & Molnar (1972) and by Scholz, Koczynski & Hutchins (1976). These authors conclude that a tensional failure of the lithosphere is occurring in southern Africa and that the East African Rift system is propagating southward along a pre-existing zone of weakness.

CONCLUSIONS

It does not appear that either the plume hypothesis or the propagating fracture hypothesis can easily explain all aspects of intra-plate volcanism. The principal advantages of the plume hypothesis are:

1. Ascending convection in a plume explains the production of magma by pressure-release melting (but deep plumes give too much melting).

2. The fixed plume hypothesis predicts the general trends of volcanic chains and, in a number of cases, the rates of propagation.

3. Crustal doming is a natural consequence of the flow processes associated with plumes.

The principal advantages of the propagating fracture hypothesis are:

1. It explains the widespread occurrence of intra-plate volcanism, particularly the continued volcanism along volcanic lineaments.

2. It explains the common association of extensional tectonics with intra-plate volcanism.

3. It does not require anomalous asthenosphere beneath volcanic areas.

4. It provides an explanation for the migration of magmas through the lithosphere by corrosion fracturing.

With further refinement either hypothesis may be able to explain intra-plate volcanism or both hypotheses might be valid: the similarity of the products of intra-plate volcanism, however, suggests one process rather than two, or possibly an entirely new hypothesis is required.

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REFERENCES (Turcotte & Oxburgh)

Anderson, D. L. 1975 *Bull. geol. Soc. Am.* **86**, 1593–1600.

Anderson, O. L. & Grew, P. C. 1977 *Rev. Geophys. Space Phys.* (In the press.)

Anderson, O. L. & Perkins, P. C. 1974 *J. geophys. Res.* **79**, 2136–2138.

Anguita, F. & Herman, F. 1975 *Earth planet. Sci. Lett.* **27**, 11–19.

Armstrong, R. L., Leeman, W. P. & Malde, H. E. 1975 *Am. J. Sci.* **275**, 225–251.

Atwater, T. 1970 *Bull. geol. Soc. Am.* **81**, 3513–3536.

Betz, F. & Hess, H. H. 1942 *Geogr. Rev.* **32**, 99–116.

Bijlaard, R. P. 1935 *Ing. Ned.-Indië.* **11**, 135–156.

Bott, M. H. P. & Dean, D. S. 1972 *Nature, Lond.* **235**, 23–25.

Boyd, F. R. & McCallister, R. H. 1976 *Geophys. Res. Letters* **3**, 509–512.

Burke, K. & Dewey, J. F. 1973 *J. Geol.* **81**, 406–433.

Burke, K., Kidd, W. S. & Wilson, J. T. 1973 *Nature, Lond.* **245**, 133–137.

Burke, K. & Wilson, J. T. 1972 *Nature, Lond.* **239**, 387–390.

Burke, K. & Wilson, J. T. 1976 *Scient. Am.* **235**, 46–57.

Cathles, L. M. 1975 *The viscosity of the Earth's mantle.* Princeton University Press.

Christiansen, R. L. & Lipman, P. W. 1972 *Phil. Trans. R. Soc. Lond.* A **271**, 249–284.

Clague, D. A. & Jarrard, R. D. 1973 *Bull. geol. Soc. Am.* **84**, 1135–1154.

Deffeyes, K. S. 1972 *Nature, Lond.* **240**, 539–544.

Ellsworth, W. L., Cipar, J. J., Ward, P. L., Aki, K., Koyanagi, R. & Unger, J. 1975 *Abstracts of papers presented at the Interdisciplinary Symposia, XVIth General Assembly, International Union Geodesy Geophysics*, Grenoble, France, p. 109.

Forsyth, D. & Uyeda, S. 1975 *Geophys. J. R. astr. Soc.* **43**, 163–200.

Green, A. G. 1975 *J. geophys. Res.* **80**, 4028–4036.

Green, D. H. 1971 *Phil. Trans. R. Soc. Lond.* A **268**, 707–725.

Gruntfest, I. J. 1963 *Trans. Soc. Rheol.* **7**, 195–207.

Hadley, D. M., Stewart, G. S. & Ebel, J. E. 1976 *Science, N.Y.* **193**, 1237–1239.

Jackson, E. D. 1976 *The geophysics of the Pacific Ocean basin and its margin* (eds G. H. Sutton, M. H. Manghnani & R. Moberly), pp. 319–335. *Am. Geophys. Un. Monogr.* **19**.

Jackson, E. D. & Wright, T. L. 1970 *J. Petr.* **11**, 405–430.

Julian, B. R. & Sengupta, M. K. 1973 *Nature, Lond.* **242**, 443–447.

Kanasewich, E. R., Ellis, R. M., Chapman, C. H. & Gutowski, P. R. 1972 *Nature, Phys. Sci.* **239**, 99–100.

Kanasewich, E. R., Ellis, R. M., Chapman, C. H. & Gutowski, P. R. 1973 *J. geophys. Res.* **78**, 1361–1371.

Kanasewich, E. R. & Gutowski, P. R. 1975 *Earth planet. Sci. Lett.* **25**, 379–384.

Karig, D. E. 1974 *A. Rev. Earth planet. Sci.* **2**, 51–75.

King, P. B. & Beikman, H. M. 1974 *Geological map of the United States.* Washington, D.C.: U.S. Geological Survey.

Koyanagi, R. Y., Endo, E. T. & Ward, P. L. 1976 *The geophysics of the Pacific Ocean basin and its margin* (eds G. H. Sutton, M. H. Manghnani & R. Moberly), pp. 169–172. *Am. Geophys. Un. Monogr.* **19**.

Koyanagi, R. Y., Krivoy, H. L. & Okamura, A. T. 1966 *Bull. seism. Soc. Am.* **56**, 1317–1335.
Krishnamurti, R. 1970*a* *J. Fluid Mech.* **42**, 295–307.
Krishnamurti, R. 1970*b* *J. Fluid Mech.* **42**, 309–320.
Maasha, N. & Molnar, P. 1972 *J. geophys. Res.* **77**, 5731–5743.
Marsh, B. D. & Marsh, J. G. 1976 *J. geophys. Res.* **81**, 5267–5280.
McKenzie, D. P. & Sclater, J. G. 1968 *J. geophys. Res.* **73**, 3173–3179.
Minster, J. B., Jordan, T. H., Molnar, P. & Haines, E. 1974 *Geophys. J. R. astr. Soc.* **36**, 541–576.
Molnar, P. & Atwater, T. 1973 *Nature, Lond.* **246**, 288–291.
Molnar, P. & Francheteau, J. 1975 *Geophys. J. R. astr. Soc.* **43**, 763–774.
Morgan, W. J. 1971 *Nature, Lond.* **230**, 42–43.
Morgan, W. J. 1972*a* *Am. Ass. petr. Geol. Bull.* **56**, 203–213.
Morgan, W. J. 1972*b* *Studies in Earth and space sciences*, pp. 7–22. *Geol. Soc. Am. Mem.* **132**.
O'Hara, M. J. 1975 *Nature, Lond.* **253**, 708–710.
Oxburgh, E. R. 1965 *Phil. Trans. R. Soc. Lond.* A **258**, 142–144.
Oxburgh, E. R. & Turcotte, D. L. 1968*a* *J. geophys. Res.* **73**, 2643–2661.
Oxburgh, E. R. & Turcotte, D. L. 1968*b* *Nature, Lond.* **216**, 1041–1043.
Oxburgh, E. R. & Turcotte, D. L. 1974 *Earth planet. Sci. Lett.* **22**, 133–140.
Oxburgh, E. R. & Turcotte, D. L. 1976 *Tectonophysics* **32**, 107–128.
Parmentier, E. M., Turcotte, D. L. & Torrance, K. E. 1975 *J. geophys. Res.* **80**, 4417–4424.
Parmentier, E. M., Turcotte, D. L. & Torrance, K. E. 1976 *J. geophys. Res.* **81**, 1839–1846.
Reynolds, P. H. & Aumento, F. 1974 *Can. J. Earth Sci.* **11**, 1269–1273.
Richter, F. M. 1973 *J. geophys. Res.* **78**, 8735–8745.
Richter, F. M. & Parsons, B. 1975 *J. geophys. Res.* **80**, 2529–2541.
Scholz, C. H., Barazangi, M. & Sbar, M. L. 1971 *Bull. geol. Soc. Am.* **82**, 2979–2990.
Scholz, C. H., Koczynski, T. A. & Hutchins, D. G. 1976 *Geophys. J. R. astr. Soc.* **44**, 135–144.
Shaw, H. R. & Jackson, E. D. 1973 *J. geophys. Res.* **78**, 8634–8652.
Smith, R. B. & Sbar, M. L. 1974 *Bull. geol. Soc. Am.* **85**, 1205–1218.
Smith, R. B., Shuey, R. T., Freidline, R. O., Oris, R. M. & Alley, L. B. 1974 *Geology* **2**, 451–455.
Sommerscales, E. F. C. & Dropkin, D. 1966 *Int. J. Heat Mass Trans.* **9**, 1189–1204.
Suppe, J., Powell, C. & Berry, R. 1975 *Am. J. Sci. A* **275**, 397–436.
Talwani, M., Hanes, D., Pitman, W. & Aitken, T. 1974 *Underway marine geophysical data in the North Atlantic June 1961–January 1971*. Part C. *Bathymetry, geomagnetic, and gravity profiles*. Palisades, New York: Lamont-Doherty Geological Observatory.
Torrance, K. E. & Turcotte, D. L. 1971*a* *J. Fluid Mech.* **47**, 113–125.
Torrance, K. E. & Turcotte, D. L. 1971*b* *J. geophys. Res.* **76**, 1154–1161.
Tozer, D. C. 1973 *Nature, Lond.* **244**, 398–400.
Turcotte, D. L. 1967 *Int. J. Heat Mass Trans.* **10**, 1065–1074.
Turcotte, D. L. 1974*a* *J. geophys. Res.* **79**, 2573–2577.
Turcotte, D. L. 1974*b* *Geophys. J. R. astr. Soc.* **36**, 33–42.
Turcotte, D. L., Hsui, A. T., Torrance, K. E. & Schubert, G. 1974 *J. Fluid Mech.* **64**, 369–374.
Turcotte, D. L. & Oxburgh, E. R. 1967 *J. Fluid Mech.* **28**, 29–42.
Turcotte, D. L. & Oxburgh, E. R. 1973 *Nature, Lond.* **244**, 337–339.
Turcotte, D. L. & Oxburgh, E. R. 1976 *Tectonophysics* **35**, 183–199.
Watts, A. B. & Cochran, J. R. 1974 *Geophys. J. R. astr. Soc.* **38**, 119–141.
Weertman, J. 1971 *J. geophys. Res.* **76**, 1171–1183.
Wilson, J. T. 1963 *Can. J. Phys.* **41**, 863–870.
Wilson, J. T. 1973 *Tectonophysics* **19**, 149–164.
Winterer, E. L. 1973 *Am. Ass. petr. Geol.* **57**, 265–282.
Wright, C. 1975 *J. geophys. Res.* **80**, 1915–1919.
Yuen, D. A. & Schubert, G. 1976 *J. geophys. Res.* **81**, 2499–2510.

Discussion

R. N. Thompson (*Department of Geology, Imperial College, London SW7 2AZ*). I entirely share the authors' doubt that the Late Cenozoic tectonism and magmatism of the western United States is consequent upon the existence of a Wilson/Morgan mantle plume, currently situated beneath Yellowstone (Thompson 1977). Nevertheless, I am concerned that the alternative scheme proposed appears to be based upon the tectonomagmatic relations of only the last 10 Ma or so. The problem for such an approach lies in recent K–Ar age determinations of

Snake River Plain and Columbia River Plateau lavas. Armstrong *et al.* (1975) have shown that the earliest extruded rocks in the Snake River Plain (13 Ma) immediately followed the spatially contiguous youngest eruptions of the Columbia River Plateau (13.5 Ma), dated by Watkins & Baksi (1974). Wright, Grolier & Swanson (1973) have also discovered that some of the youngest Columbia River Plateau lavas are intermediate in composition between the otherwise distinctive Columbia and Snake River suites. These data persuade me to conclude that it is no longer valid to take the traditional approach of treating these two immense volcano-tectonic provinces separately.

References

Thompson, R. N. 1977 *Tectonophysics* (In the press.)
Watkins, N. D. & Baksi, A. K. 1974 *Am. J. Sci.* **274**, 148–189.
Wright, T. L., Grolier, M. J. & Swanson, D. A. 1973 *Bull. geol. Soc. Am.* **84**, 371–386.

E. R. OXBURGH. The authors are familiar with some of the geochronological work on the Columbia River Plateau Lavas and agree that there is no good reason for regarding them as chemically distinct from the Snake River Series. This relation, discussion of which was omitted in the interests of brevity, if anything strengthens the argument that the distribution of surface volcanism is fracture controlled rather than being the expression of some kind of mantle plume across which the North American plate has been moving during the last 20 Ma.

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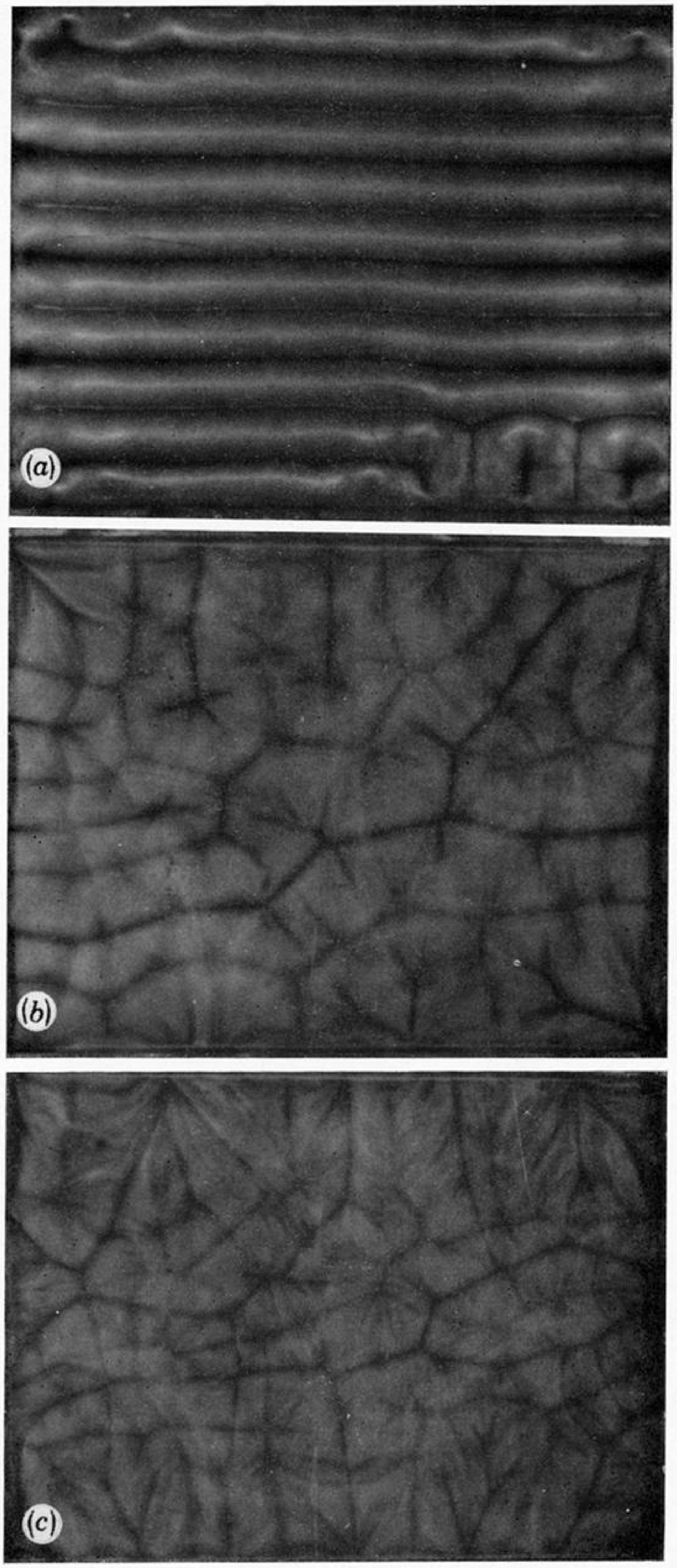

FIGURE 4. Planform shadowgraphs of thermal convection in a high Prandtl number fluid heated from below and cooled from above with rigid horizontal boundaries. (*a*) $Ra = 3040$; steady, two-dimensional; (*b*) $Ra = 43\,300$, steady, three-dimensional; (*c*) $Ra = 131\,000$, unsteady (turbulent).