Growth, Shrinking, and Long-Term Evolution of Plates and Their Implications for the Flow Pattern in the Mantle

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Surficial manifestations of the long-term evolution of plates are examined and used to derive constraints on the pattern of flow in the mantle. Generation and consumption of plates are not balanced around single plates, so the sizes of plates and the relative positions of ridges and descending slabs change, a complex return flow thus being required. At present a net flow should occur from the sites of much subduction toward the southern hemisphere, where sea floor spreading predominates. To allow the relative motions between descending slabs, masses in the mantle should be displaced at a rate which is several times larger than the rate of generation of lithosphere. The patterns of plate motions and of the return flow change considerably over periods of 100 m.y. The formation of instabilities occurs much faster than the cycling of mantle material. Hence the flow in the mantle cannot consist of regular steady state cells in which material moves in simple circuits and which are linked to plate boundaries. The deep flow is not parallel to the relative motions of overlying plates, so it is not two dimensional. The magnitude and nonsteady state of the sublithospheric flow make it doubtful that instabilities originating within the lithosphere alone drive the flow. This fact, together with the abrupt changes of plate motions, including jumping of ridges and continental breakup, suggests that instabilities at depth, e.g., mantle diapirs, are most important. The plate tectonic regime helps generate such instabilities in the mantle by disturbing the temperature distribution there.

INTRODUCTION

The theory of plate tectonics requires considerable mass movement in a large part of the earth: plates move about on the surface, material rises from depth to generate oceanic lithosphere at ridges, and lithospheric slabs descend in subduction zones to depths reaching 700 km. To balance this overturn of lithospheric mass, a return flow should occur to at least a depth of 700 km, i.e., in about a quarter of the mass of the mantle. However, while the motions of the surficial plates are fairly well known, the pattern of the deep return flow remains obscure.

In addition to the continuing overturn of lithosphere the plate tectonic regime leads to characteristic long-term effects. Notably, the sizes and shapes of plates change continuously. In the long run these changes are related to changes in the pattern of plate motions. The purpose of this paper is to discuss such long-term features of plate evolution and to examine their implications for the pattern of mass flow in the mantle.

The discussion is based on surficial phenomena, so the conclusions do not depend on physical models of the flow, though it is accepted that the flow is some kind of thermal convection. The aim is to derive constraints on the principal features of the flow pattern rather than to offer a physical mechanism; successful physical models should satisfy these constraints.

Quantitative data on plate behavior are available only for a short part of the earth's history. However, the gross uniformity of the main traits of the earth's behavior during the Phanerozoic and much of the Precambrian, as recorded on continents, suggests that the same fundamental processes were in operation. It is assumed therefore that the processes which now drive and control the motion and evolution of plates were active during a considerable portion of the earth's history. Observations on the geologically recent past are supposed to represent much older periods, for which direct information about plate behavior is not available. It is also assumed that the size of the earth did not change during this period.

The flow in the mantle is supposed to involve only the outer 700 km or so of the earth, as is suggested by the seismic behavior of the descending slabs [*Isacks and Molnar,* 1971] and by the differences between the composition of the lower mantle and that of its shallower parts [*Anderson and Jordan,* 1970]. However, some conclusions do not depend on this assumption.

GROWTH AND SHRINKING OF PLATES

Rates of production and consumption of surface area along various plate boundaries during the last 5–10 m.y. are summarized in Tables 1 and 2. Plate motions during this period are known quite well, so the errors are probably less than 10%. These values are used to calculate the rates of change of the areas of plates (in square kilometers per year), which are given below:

Antarctican	+0.50
African (including Somali)	+0.26
American	+0.30
Eurasian	+0.06
Arabian	−0.11
Indian	−0.37
Pacific	−0.45
Cocos	−0.08
Nazca	−0.11

Clearly, sea floor spreading and subduction are not balanced locally, i.e., around single plates, so the areas and shapes of plates change significantly in geologic time.

The Antarctican plate grows fastest, being almost surrounded by fast spreading ridges. The African plate (including Somali) is also largely surrounded by ridges, but part of this growth is offset by subduction in the Mediterranean Sea region. The American plate grows along its Atlantic margin, while its Pacific margin remains virtually unchanged.

The Pacific, Cocos, and Nazca plates shrink at a total rate of about 0.7 km^2/yr. Here the overturn of oceanic lithosphere is faster than it is elsewhere, but the result is a net loss of area, so the Pacific Ocean shrinks. However, in the south the Antarc-

TABLE 1. Rates of Sea Floor Spreading Along Various Plate Boundaries

Plate Boundary	Relative Motion		Distance From Rotation Pole*		Rate of Area Generation, km²/yr	Reference
	Geographic Position	V_{max}, cm/yr	θ_1	θ_2		
			Atlantic Ocean			
Eurasian-American	68°N, 137°E	3.08	0†	72	0.136	*Pitman and Talwani* [1972]
African-American	70°N, −33°E	4.44	30	127	0.414	*Pitman and Talwani* [1972]
Total					0.550	
			Pacific Ocean			
Pacific-American	55°N, −65°E	−8.5	41	49	0.050	Author's estimate
Cocos-Nazca	5°N, −122°E	12.	20	30	0.116	*Herron* [1972]
Pacific-Nazca	58°N, −93°E	−20.	56	98	0.880	*Herron* [1972]
Pacific-Cocos	40°N, −108°E	−28.4	23	38	0.246	Calculated from *Herron* [1972]
Antarctican-Pacific	−70°N, 118°E	−12.	17	64	0.398	*Le Pichon* [1968]
Antarctican-Nazca	38°N, −110°E	9.3	81	89	0.080	
Pacific–Juan de Fuca					0.070	Estimated from *Atwater* [1970]
Total					1.840	
			Indian Ocean			
Arabian-African	36°N, 18°E	3.2	16	33	0.026	*McKenzie et al.* [1970]
Arabian-Somali	26°N, 21°E	3.9	25	37	0.026	*Laughton et al.* [1970]
Somali-Indian	16°N, 48°E	−6.9	10	48	0.140	*McKenzie and Sclater* [1971]
Indian-Antarctican	11°N, 32°E	7.1	53	116	0.470	*McKenzie and Sclater* [1971]
Antartican-African	16°N, 161°E	2.5	98	139	0.098	Calculated from *McKenzie and Sclater* [1971]
Total					0.760	
Grand total					3.150	

* The rate of generation of area between points at angular distances θ_1 and θ_2 from the rotation pole is given by $ds/dt = \int_{\theta_1}^{\theta_2} V_{max} \cdot \sin\theta \cdot R \cdot d\theta = V_{max} \cdot R \cdot (\cos\theta_1 - \cos\theta_2)$, where R is the earth's radius.

†A fair approximation.

tican plate grows at a rate of 0.24 km²/yr. The Indian plate also loses area.

Many lines of evidence show that the area of continental crust has increased during geological history [*Wilson*, 1967; *Taylor*, 1967; *Ringwood*, 1969]. To produce the area shallower than 1 km below sea level, i.e., 190 × 10⁶ km² [*Menard and Smith*, 1966] during 3500 m.y., the average rate of growth had to be 0.054 km²/yr. Compared with this value, the rate of loss of continental area along the margin of the Arabian plate and along the Himalayas (Table 2) is surprisingly large: it is almost 0.3 km²/yr, which is about 5 times the long-term average rate of continental growth. This is a transient circumstance, because it is improbable that large volumes of relatively light and buoyant continental crust can descend into the heavier mantle. Except for this feature the values of the rates of change of plate areas given in the in-text table above probably represent in a general way the geologic past.

The oldest oceanic crust is of Jurassic age, about 180 m.y. old. The possible growth of continents during this period is negligible (amounting to only 10 × 10⁶ km² when the above long-term average is used; even a value several times larger will not seriously affect the following calculation). Thus on an earth of constant size the entire oceanic area, about 310 × 10² km² [*Menard and Smith*, 1966], was generated within the last 180 m.y. In addition, the asymmetric magnetic anomalies in the Pacific Ocean show that an area equal to about three quarters of this ocean was generated during this time interval but has already been subducted [*Larson and Pitman*, 1972]. Hence during the last 180 m.y., oceanic lithosphere was generated at an average rate of about 2.4 km²/yr, which is not very different from the estimate of 3.1 km²/yr (Table 1) for the present rate. Some fluctuations could have occurred.

The shrinking of the Pacific Ocean is a long-term process; the fit of the continents at the beginning of the Mesozoic [*Bullard et al.*, 1965; *Smith and Hallam*, 1970] shows that the area of the Tethys Ocean was then about 30 × 10⁶ km². This area has been consumed, and concurrently, the Atlantic and Indian oceans were formed, covering areas (deeper than 1 km) of 76 × 10⁶ and 69 × 10⁶ km², respectively [*Menard and Smith*, 1966]. As the area of the continents hardly changed during this period (see discussion above), the area of the remaining ocean, that is, the Pacific, must have decreased by about 115 × 10⁶ km². Hence the average rate of area loss since 180 m.y. ago has been about 0.65 km²/yr, similar to the current rate of 0.7 km²/yr (see in-text table above). While the Pacific Ocean shrank, the Kula plate was entirely consumed, and so were large parts of the Phoenix and Farallon plates [*Atwater*, 1970; *Larson and Pitman*, 1972]. At the present rate of loss of area, most of the Pacific Ocean can be consumed within about 200 m.y., and the bordering continents will collide.

Other oceans disappeared in the geologic past when the bordering continents collided, a situation indicating that subduction and sea floor spreading were not locally balanced. In addition to the Tethys, one may cite the proto-North Atlantic, which disappeared early in the Paleozoic at the site of the Caledonides [*Wilson*, 1966; *Harland and Gayer*, 1972], and the Siberian Ocean, which disappeared in the Paleozoic at the site of the Uralides [*Hamilton*, 1970].

The Pacific Ocean could not have been shrinking at the present rate from before the Mesozoic. If it had been, then in the Paleozoic it would have been much larger than the total oceanic area on earth. At that time the Pacific Ocean was probably growing while other oceans were being consumed at the Caledonian and Hercynian orogenic belts. Thus the overall balance between subduction and spreading in the Pacific Ocean must have been reversed.

Neither oceanic plates nor oceans can grow indefinitely to exceed the total area of oceanic crust on earth; otherwise, much continental crust would have to be subducted, an unlikely prospect. If the Atlantic and Indian oceans continue to

TABLE 2. Rates of Subduction Along Various Plate Boundaries

Plate Boundary	Relative Motion: Geographic Position	Relative Motion: V_{max}, cm/yr	Distance From Rotation Pole: θ_1	Distance From Rotation Pole: θ_2	Rate of Area Consumption, km²/yr
		Alpine Orogenic Belt			
African-Eurasian	26°N, −37°E	3.0	14	63	0.104
Arabian-Eurasian	34°N, −10°E	5.7	38	66	0.138
Indian-Eurasian (Himalayas)	28°N, 26°E	7.2	38	64	0.155
Indian-Eurasian (Java trench)	28°N, 26°E	7.2	64	102	0.300
Total					0.697
		East Pacific			
Cocos-American	31°N, −119°E	21.6	19	41	0.256
Nazca-American	56°N, −113°E	12.	55	106	0.652
Antartican-American	−79°N, −81°E	4.4	34	19	0.033
Pacific-American	55°N, −65°E	−8.5	38	62	0.175
Juan de Fuca-American					0.075*
Total					1.191
		West Pacific			
American-Eurasian	68°N, 137°E	−3.08	0	17	0.008
Pacific-Eurasian†	69°N, −72°E	−10.5	49	109	0.660
Pacific-Indian‡	−62°N, 174°E	15	22	48	0.246
			60	70	0.151
Pacific-Indian	−62°N, 174°E	15	9	22	0.058
			48	60	0.162
Total					1.285
Grand total					3.173

All relative plate motions are from Table 1 or are calculated from it.
*Estimate.
†Includes Philippine plate.
‡Pacific plate consumed in New Guinea and Tonga-Kermadec trenches; Indian plate consumed along rest of plate boundary.

grow at the present rates, the maximum possible size will be reached in about 250 m.y. Then either sea floor spreading will cease, or new subduction zones will have to develop. The balance between generation and subduction of oceanic lithosphere will change over large areas.

The constantly changing sizes and shapes of plates clearly must be accompanied by modifications of the geometry of plate motions. In addition, when continents collide, the relative motion of the plates containing them should change or stop to avoid subduction of much continental crust.

However, motions of plates also change without apparent relation to changes of their sizes and shapes. Such events, accompanied by jumping of midoceanic ridges, occurred in the oceans [*McKenzie and Sclater*, 1971; *Herron*, 1972; *Pitman and Talwani*, 1972]. Most spectacular changes result from continental breakup, e.g., the dispersal of Laurasia and Gondwanaland. Similar events probably also occurred earlier in the earth's history: the western margin of North America, for instance, bears evidence of Precambrian events of continental splitting [*Monger et al.*, 1972; *Hoffman*, 1973].

At present the absence of local balance between sea floor spreading and subduction is well expressed not only in terms of changing areas of individual plates but also on a broader scale: more than three quarters of all sea floor spreading occurs in the southern hemisphere, or rather in the hemisphere centered on the Scotia Sea, whereas only about one third of all subduction occurs there. An area of about 0.96 km² is annually consumed in a relatively small region around Southeast Asia (Table 2). This is about 30% of the global rate, yet is not obviously related to any nearby region of sea floor spreading, since the neighboring Indian and Pacific plates are shrinking. Along the Central and Southern American trenches an area of about 0.8 km² is consumed annually (Table 2); in the neighboring South Atlantic Ocean, only half of this area is generated, while in the eastern Pacific Ocean an area twice as large is generated (Table 1). The ridges around the Antarctican and African plates do not alternate with subduction zones, so the spreading along these ridges cannot at all be related directly to subduction in neighboring regions.

The foregoing considerations show that shrinking of some plates and growth of others, coupled with continuous and episodic changes in the pattern of plate motions and interaction, are essential features of the long-term evolution of plates, expressing the absence of local balance between sea floor spreading and subduction. The quantitative relations discussed above show that 100 m.y. or a few hundred million years is a typical period for some fundamental changes in the pattern of plate motions to take place, and this may also be the lifetime of some oceanic plates.

CONSEQUENCES REGARDING FLOW IN THE MANTLE

The features of the plate evolution and motions discussed above have important consequences with regard to the nature of flow in the underlying mantle.

The non-steady state nature of plate motions implies that the underlying mass motions must also be time dependent, because (1) generation and consumption of plates are coupled with a return flow in the mantle which extends to a depth of 700 km at least and (2) the overturn of lithospheric mass relative to the mass of the mantle is significant during geologically long periods. To produce a 70-km-thick lithosphere [*Kanamori and Press*, 1970] at a rate of 2.4 km²/yr, a mass of about 5.5×10^{17} g (ρ = 3.3 g/cm^3) should rise annually from the mantle; at this 180-m.y. average rate the entire mass shallower than 700 km, which is assumed to participate in the flow, can be recycled in somewhat more than 2000 m.y., and the mass down to 1000 km can be recycled in about 3000 m.y. In the Proterozoic this rate was probably higher than it is now, because more radiogenic heat was then produced in the earth, an action which could induce more vigorous convection; hence a complete overturn could occur in 1500 m.y. This value is similar to the differentiation events recorded by lead isotopes from volcanic islands [*Oversby and Gast*, 1970], which *Morgan* [1972] interpreted as recording previous times when the source rocks of the volcanics were part of the lithosphere. Therefore the mass motions in the mantle associated with the plate tectonic regime cannot be approximated well by a steady state model over periods of a few hundred million years.

The continuing changes of the sizes and shapes of plates require that the relative positions of the midoceanic ridges and of the subduction zones must also change, so at least some must move in relation to the deeper mantle. *Elsasser* [1971] noted that as a result of the shrinking of the Pacific Ocean the subduction zones surrounding it should approach each other; therefore at least some descending slabs must sink in relation to the mantle and not merely descend parallel to themselves. Elsasser called this motion 'retrograde subduction.' Such a behavior is in line with gravitative sinking of the slabs, which are colder and denser than the surrounding mantle [*Elsasser*, 1971; *Turcotte and Schubert*, 1971]. When a descending lithospheric slab sinks in such a manner through the mantle, it displaces some material from its lower side, and an equal volume must be filled on the other side (Figure 1). Since equal volumes are displaced on the two sides of the slabs, these mass motions cannot be simply related to the return flow from the subduction zones to ridges.

The magnitude of the sublithospheric mass motions necessary to allow retrograde subduction can be estimated from the rate of shrinking of the Pacific Ocean. While this ocean shrinks at a rate of about 0.7 km^2/yr, the subduction zones which surround most of it sweep out a comparable area between themselves. The descending slabs reach to depths exceeding 500 km, so while their relative positions change, they displace from the Pacific side a mass exceeding 12×10^{17} g/yr (ρ = 3.5 g/cm^3), which is more than twice the mass that rises in all the ridges. A similar mass is displaced on the other side of the slabs. This is a crude estimate because it does not correct for absence of subduction along some parts of the Pacific Ocean or for possible differences in the depths to which various slabs extend. However, this estimate does show that the sublithospheric mass flow which accommodates retrograde subduction is several times larger than the mass flow necessary to generate oceanic lithosphere.

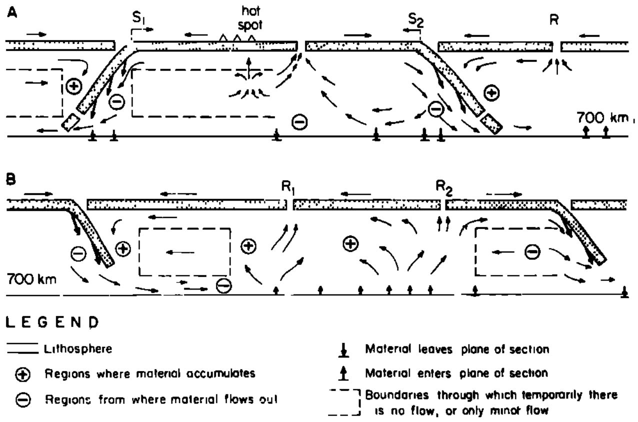

Fig. 1. Possible patterns of the master flow in the mantle (see text), compatible with the long-term behavior of plates. The flow is not two dimensional, so it is not adequately represented in cross sections. (*a*) Shrinking mantle compartment between approaching lithospheric slabs (Pacific Ocean type). Subduction zones S_1 and S_2 approach each other as a result of retrograde subduction, while ridge R recedes from S_2. The plates between S_1 and S_2 lose area. (*b*) Expanding mantle compartment (Atlantic Ocean type). Ridges R_1 and R_2 recede from each other, while the plates delimited by them grow.

Two circumstances are essential in the foregoing calculation: (1) The moving subduction zones around the shrinking Pacific Ocean sweep out areas which are not much smaller than the global rate of sea floor spreading. (2) The descending slabs reach a depth as much as 10 times deeper than the base of the lithosphere. These conditions probably obtained also in the geologic past: it may be expected that, similar to the present situation, shrinking of ancient oceans was of the same order of magnitude as plate displacement and rates of sea floor spreading. The existence of deep descending slabs in the geologic past is indicated by the record in old orogenic belts, especially by the record of andesitic volcanism and related plutonism [*Dickinson*, 1972].

Hence plate tectonics is the direct surface expression of, and is directly linked to, mass movements in the mantle that are considerably more extensive than the overturn of oceanic lithosphere. The relation between the extra flow and the retrograde descent of slabs strongly suggests that some features of subduction zones are surficial manifestations of these additional motions and are not related only to the behavior of the descending slabs.

In the absence of local balance between subduction and sea floor spreading the sublithospheric flow should allow for a net flux from the shrinking plates to the growing ones. At present this requires a net flow toward the southern hemisphere, or rather toward the South Atlantic and circum-Antarctica regions, where most plate growth occurs, and away from the shrinking northern and central Pacific Ocean, this flow not being obviously related to the generally north-south trending subduction zones bordering the Pacific. Such a globally asymmetric deep flow is perhaps related to the pear shape of the earth [*King-Hele and Cook*, 1973].

In the Cretaceous and earlier, before the growth of the Antarctican plate became important, the net deep flow was toward the spreading regions in the central and southern Atlantic Ocean and the Indian Ocean and away from the shrinking Pacific Ocean (the Tethys was also shrinking, but this was of little importance at that time). Still earlier, in the Paleozoic, most plate growth probably occurred in the Pacific Ocean while continents aggregated on the opposite hemisphere, where the Tethys, Siberian, and other oceans were shrinking or being closed. Thus global oscillations are indicated.

Hence the long-term evolution of plates indicates that the sublithospheric flow is quite complex and time dependent and the part of it that is directly related to plate evolution is considerably larger than the mass motions involved in the overturn of lithosphere. This flow accounts for the changes of the relative positions of the descending slabs and the ridges and for the absence of local balance between plate generation and consumption, and it also allows a net sublithospheric mass flux from some parts of the earth where much subduction occurs to others where plate growth predominates.

Since the surficial plates control the manner in which the heat generated in the earth is given off, the above conclusions show that the plate tectonic regime introduces irregularity into the temperature distribution in the mantle; the present average heat flow from the oceans is about 1.5 HFU. If it is assumed that the continental lithosphere (which does not participate in the mass overturn) receives a heat flux of 0.5 HFU from the underlying mantle, then the total heat flow from the convecting part of the earth is 1.8×10^{20} cal/yr. The heat flow from the oldest oceanic areas in the western Pacific, west central Atlantic, and northeastern Indian oceans is only 1.1–1.2 HFU [*Langseth and von Herzen*, 1970]. The difference between this value and the average oceanic heat flow results from cooling of the oceanic lithosphere as it becomes older; i.e., this is the heat released as a consequence of the generation of plates. Over the entire oceanic area this amounts to about 0.35×10^{20} cal/yr, or about 20% of the heat given off by the convecting part of the earth. This is a minimum estimate, because some oceanic lithosphere may not reach a steady thermal state prior to subduction, and interaction with seawater may be important at ridges. Thermal models of the lithosphere lead to higher estimates [*Sclater and Francheteau*, 1970].

As plate generation controls the disposal of a significant fraction of the thermal budget of the earth, it follows that the irregular distribution and long-term changes of sea floor spreading should influence the temperature distribution in the mantle. Complementary to this relationship is the uneven descent of cooled lithospheric material into various parts of the mantle. At present much more heat is given off by plate generation in the southern hemisphere, where most sea floor spreading occurs, than in the northern hemisphere, while most cooled lithosphere descends in a few rather limited regions. As similar irregularities existed in the past, important complexities in the temperature distribution, including lateral gradients, should have been generated in the convecting part of the mantle, provided that the plate tectonic regime operated long enough.

It is not known when this regime was first established, but the similarity of Proterozoic structures to more recent ones strongly suggests that the plate tectonic regime has been in operation during the last 2000 m.y. at least [*Salop and Scheinmann*, 1969; *Burke and Dewey*, 1973; *Bridgewater et al.*, 1973; *Hoffman*, 1973]. It was estimated above that the entire outer 700 or 1000 km of the mantle can be recycled in less than 2000 or 3000 m.y., respectively. Thus the plate tectonic regime had ample time to impose its influence on the thermal regime in the convecting part of the mantle.

PATTERN OF FLOW IN THE MANTLE

The foregoing conclusions will now be used to discuss the flow pattern. While numerous possibilities present themselves, some generalizations are possible.

Flow in the mantle was often described in terms of discrete, episodic or steady state, convective cells or rolls in which material moves in a circuit. Following the physical theory of convection and the experimental results [e.g., *Brindley*, 1967], periodic patterns were generally considered. The cells were assumed to be linked to subduction zones or to all nonconservative plate boundaries [e.g., *Hess*, 1962; *Holmes*, 1965; *Turcotte and Oxburgh*, 1972]. Models of convection cells not specifically linked to plate boundaries were also considered [e.g., *Knopoff*, 1964; *Runcorn*, 1965].

Models of a flow consisting of discrete cells, which are linked to plate boundaries, are unsatisfactory irrespective of the depth of the cells, because they predict, contrary to observation, local balance between the material which rises to build new lithosphere and the material which returns to the mantle in subduction zones. Consequently, such models do not allow changes of the sizes and shapes of plates or changes of the relative positions of subduction zones or of midoceanic ridges. To account for these features of plate evolution, cells linked to plate boundaries should have unequal sizes and irregular and continuously changing shapes, and some should be leaking and should lose material to the surrounding regions, while other cells, linked to growing plates, should expand. However, even such irregular and time-dependent cells are incompatible with ridges which do not alternate with subduction zones, like those around Africa and Antarctica. If periodic cells exist in the mantle, they are decoupled from plate boundaries, and they cannot account for the overturn of lithosphere.

Hence the overturn of lithospheric material and the long-term evolution of plates cannot be described by a flow pattern consisting of discrete circuital cells; instead, a worldwide flow must be considered. It can be discussed in terms of two components: (1) a mass flux, to be called the master flow, which is required to account for observable features of plate evolution (this flux provides the return flow from sites of plate consumption to sites of plate generation and accounts for the other mass motions required by the long-term features of plate evolution, which were discussed above) and (2) additional mass circuits underneath the lithosphere superimposed on the master flow. The master flow is the minimal flow required by the known features of plate evolution, whereas the other components of flow, of unknown magnitude (which is perhaps relatively small), do not have a net result in terms of plate evolution, though they may contribute to the mechanism which drives and modifies the motions of plates (see discussion below).

Present knowledge does not define uniquely the master flow, but the previous discussion revealed many of its features, so a simple model with the required properties can be outlined. In this model (Figure 1) the descending lithospheric slabs lose their identity at some depth, probably 600–700 km or somewhat deeper, and their mass is incorporated into a large reservior, possibly of global extent. Elsewhere material rises directly from this reservior or from shallower parts of the mantle to build new oceanic lithosphere. A rising plume is obviously formed beneath any ridge segment, but it need not extend deeper than the low-velocity zone. The latter may be replenished in a diffuse manner or by deep plumes whose position and geometry are unrelated to ridges, so migration of ridges relative to underlying mantle is possible.

Additional motions in the mantle are required to accommodate retrograde subduction or any motions through the mantle of the descending slabs not parallel to themselves, as well as the other long-term features of plate evolution. Now such motions cause the mantle compartment under the Pacific Ocean to shrink while lithospheric slabs which border it approach each other. Some material of this compartment is incorporated into plates at ridges and eventually is swept away from the Pacific Ocean as outwardly slanting slabs and may accumulate elsewhere (Figure 1). For some time, no material at all need move into this shrinking mantle compartment. Concurrently, other compartments of the mantle grow by influx of material.

It is noteworthy that all descending slabs slant away from the plates to which they are attached, and the distribution of earthquakes in Benioff zones [*Isacks and Molnar*, 1971] does not reveal any tendency of the slabs to turn backward. On the contrary, some deep detached pieces of lithosphere are displaced farther away from the subduction zones than the shallower parts of the slabs are [*Isacks and Molnar*, 1971; *Pascal et al.*, 1973]. The sublithospheric flow seems to carry these detached pieces away from their parental plates. As most descending slabs are parts of shrinking plates, they probably help evacuate material from the underlying parts of the mantle.

In this model it is possible for large volumes of the mantle to move, one relative to the others, without undergoing changes of their external shapes. For instance, large regions under the Pacific Ocean probably move toward ridges where their marginal parts rise. Groups of mantle diapirs which generate 'hot spots' or localize volcanism may be embedded in such regions and may therefore retain their relative positions for some time.

In this model, various particles follow complicated paths, quite unlike simple circuits of the cellular models, and their velocities may change considerably with time. The entire flow system is balanced only on a global scale and probably cannot be divided into a few partial mass circuits. An essential feature is that motion at depth is generally not parallel to the directions of relative motions of overlying plates. This is necessary to balance the uneven distribution of plate consumption and generation. Thus the flow is not two dimensional and is not adequately represented in planar cross sections.

Various local mass circuits may be superimposed on the master flow, such as systems of periodic cells or rolls, which may even be excited by plate motions [*Richter*, 1973] but cannot contribute to the overall overturn of lithosphere and the long-term evolution of plates. Also, the cells will occasionally be caught between approaching lithospheric slabs which delimit shrinking oceans, or alternatively, they may collide with slabs whose positions relative to the surrounding mantle change. In such events the convection cells should be modified or destroyed, so in the long run (possibly within 500 m.y.) the flow in the cells should be time dependent.

Local overturns of mass underneath the lithosphere may also be caused by ascent of buoyant masses such as diapirs or plumes [*Ramberg*, 1972*a*, *b*]. Unlike systems of periodic cells underneath the lithosphere, to which surficial features cannot readily be related, localized diapirs or plumes can be associated with various surficial manifestations. Such manifestations are 'hot spots,' i.e., loci of volcanic activity in plate interiors or loci of exceptional volcanic productivity on constructive plate margins [*Wilson*, 1973; *Morgan*, 1971, 1972], and especially, continental rifts (or aulacogenes) in which volcanism is combined with limited crustal separation and which may pass laterally into midoceanic ridges or develop into them.

It was suggested that the gravitative sinking of lithospheric slabs that have negative buoyancy, possibly supplemented by instabilities at ridges, provides the driving force behind plate motions and the overturn of lithospheric material [*Elsasser*, 1969, 1971; *Jacoby*, 1970; *Turcotte and Schubert*, 1971]. The influence of the descending slabs (but not of ridges) is expressed in the present absolute plate motions [*Morgan*, 1972]: Plates with descending slabs move faster than other plates and move toward the subduction zones. However, there is no clear correlation between the velocity of plate motion, plate size, and the size of the attached descending slab. The motions of the Antarctican and African plates, practically surrounded by ridges, are not well explained. It is unlikely, therefore, that instabilities within the lithosphere alone drive the plate motions.

Not less important are the following observations which emerge from the above discussion: (1) Motions of plates and their evolution are direct surface manifestations of a complex sublithospheric flow that is much larger than the overturn of lithospheric mass; this flow is not well explained as being driven only by instabilities within the plates. (2) Abrupt changes in the geometry and rates of plate motions are not explained by instabilities which exist all the time within the plates. Known instances cannot be related to changes in the configuration of subduction zones. In particular, there is no explanation for jumping of ridges or for changes of plate motions which were accompanied by jumping of ridges. Nor can continental breakup (e.g., dispersal of Gondwanaland) be specifically related to instabilities within plates or to the influence of preexisting or the birth of new descending slabs.

These arguments show that instabilities in the lithosphere probably are not of prime importance in driving the master flow, since they do not account for the long-term features of the plate tectonic regime. These features are best explained as resulting from instabilities in the sublithospheric mantle. It was shown above that the plate tectonic regime generated an irregular temperature distribution in the mantle, amounting to horizontal temperature gradients, which are superimposed on the gradients resulting from the heat generation within the mantle. Thus instabilities arise and should eventually lead to convection. In these circumstances the motions are not expected to have a simple, e.g., periodic, pattern.

In this context, local mass overturns, that is, hot spots or diapirs, may have a special role. Inception of these features and their activity require abrupt disturbances of the thermal and mechanical conditions in a relatively small region and a supply of heat to generate magmas. These disturbances are best explained as a result of the ascent of hot material from below the plates. Association with deep ascending masses is also strongly suggested by the cases in which continental rifting is a precursor of continental breakup and the birth of new midoceanic ridges [*Burke and Dewey*, 1973]. Such disturbances are to be expected in a mantle with a complex temperature distribution. Furthermore, the volcanics produced in such places are much richer in radioactive elements than are the normal igneous products at midoceanic ridges [*Engel and Engel*, 1970], an indication that their source materials were richer in such elements than the normal surrounding mantle material and therefore became heated and buoyant, thus producing local disturbances in the mantle. These disturbances can grow to become active diapirs within geologically brief times [*Ramberg*, 1972*a*, *b*]. However, the small areal extent of the surficial expression of many hot spots and their geologically short life span show that the underlying disturbances are small and decay within 10–100 m.y. These are mostly local features which just keep the convecting part of the mantle agitated in an irregular and constantly changing manner. Only a few disturbances, which are large enough and properly located, trigger much larger instabilities in the surrounding mantle. Such disturbances modify the patterns of flow in the mantle and of plate motions while they become integral parts of the master flow.

Occasionally, new subduction zones should form, and this too should modify considerably the master flow in the mantle, but such events are probably rare. There are no well-documented cases of transformation of Atlantic-type coasts to Pacific-type coasts with formation of new subduction zones within the last 350 m.y. (post-Devonian time), though the details of structure of many subduction zones have changed considerably within this period and many have been eliminated. The much more common mantle diapirs appear to be more important as agents modifying the master flow in the mantle.

The pattern of the flow is therefore the cumulative result mainly of the superposition of numerous mantle diapirs, though development of new subduction zones should also contribute to the secular changes of the flow pattern. Instabilities within the lithosphere, especially the negative buoyancy of descending slabs, certainly contribute to the maintenance of the flow and supplement the deep instabilities. Continued activity of the plate tectonic regime maintains an ever-changing temperature distribution in the mantle, so new instabilities develop, and these in turn maintain the complexity of the flow and its time-dependent nature.

The stress on the time-dependent nature of the flow expresses the circumstance that geologic observations are concerned with manifestations of the plate tectonic regime during periods that are comparable to, or much shorter than, overturn periods of the mass in the mantle. As is discussed above, particles of the convecting part of the mantle are incorporated into the lithosphere at intervals of the order of 10^9 yr. Since the sublithospheric mass flux is several times larger than the overturn of lithosphere, the average interval between successive times that any particle comes closest to the surface is of the order of several 10^8 yr, possibly 5×10^8 yr. Relative to the time resolution of geological methods, convection in the mantle is very slow, and time-dependent features are very conspicuous.

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