MOTION OF THE PACIFIC PLATE AND FORMATION OF MARGINAL BASINS: ASYMMETRIC FLOW INDUCTION

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We have examined agencies inducing flow in the sub-Pacific mantle. The capable agencies are buoyancy forces (causing convection), and passage of the tidal bulge. Bulge passage induces rotational flow dimensionally identical to that induced by buoyancy, and interacting with it. The stress imposed by bulge passage is additive across the width of the sub-Pacific flow cell, in the fashion that the buoyancy stress is additive in thickness. Each stress builds thus to values in excess of 10^8 dyn/cm², overcoming the flow resistance provided by mantle viscosity. The tidal flow contribution favors the development of the west limb of convection cells, because these represent rotational flow in the tidal direction. East limbs are inhibited.

Under these forces the Pacific plate has developed as the cool, rigid surface of mantle convection surfacing in the eastern Pacific, and foundering at the western margin. The west, co-tidal limb of the East Pacific Rise now extends clear across the Pacific. The counter-tidal east limb is dwindling through its replacement by the extending west limb of the Atlantic spreading center, bearing the Americas plates.

The unbroken line of gravity highs at the western Pacific margin marks excess material accumulating where the west limb of the East Pacific Rise encounters the less mobile Asian continental mass. Material added to the asthenosphere causes disruption of the Asia margin, secondary sea-floor spreading, and the formation of arc-bounded basins. At the termination of the east, counter-tidal limb (overlain by the Cocos and Nazca plates) this process is not operative. Marginal basins, if they form, are destroyed along the Andean orogenic front by the westward-advancing Americas plates.

1. *The Problem*

The Pacific lithosphere plate, 7,000 km wide but only one hundredth as thick, seems to move intact towards eastern Asia and Australasia. What causes this motion?

The only agency which seems capable of thus rafting the plate without deformation is convection in the mantle. Observation shows that convection is in fact associated with the motion of the Pacific plate. Material is seen to reach surface at the crest of the East Pacific Rise, and to return to the planetary interior in zones of subduction at the west Pacific margin.

However buoyancy convection possesses a cell structure which seems to preclude the development of a cell spanning the width of the Pacific Ocean; even if convection embraces the entire mantle, the pattern should be that of smaller cells, breaking up the Pacific plate into several. Further difficulties exist. Theoretical and numerical models of convection (see for instance MCKENZIE *et al.*, 1974) indicate that under unforced thermal convection upflow limbs should be the site of positive gravity anomalies, downflow limbs of negative. The gravity geoid (Fig. 1) shows that the observed anomalies in no way

agree with the models. The Pacific-Antarctic and East Pacific Rise is marked randomly by highs and by lows. At best a secondary relationship (SCLATER *et al.*, 1975) can be distinguished between gravity anomalies and the world rift system. On the contrary, the subduction zones marking downflow at the west Pacific margin are the site uniquely of gravity highs, the opposite of the relation to be expected under thermal convection. Deep seismicity, marking downflow, occurs world-wide only in the context of gravity highs.

To obviate such difficulties it has been suggested that the lithosphere plates move on small-scale convection in the asthenosphere. As has however been pointed out, such models cannot account for the net flow of material. The drift of plates represents unidirectional material flow at variance with that imparted by a random thermal process.

With respect to buoyancy forces, it has been pointed out (FORSYTH and UYEDA, 1975) that the forces acting on the down-going slab are greater than any others. It has been suggested that cooled, dense, foundering lithosphere (ELSASSER, 1966) draws plates after it. McKENZIE (1969) however has pointed out that not all plates are attached to sinking slabs. It seems implausible that such a frail object as the Pacific plate can be pulled or pushed from either end without disruption.

The basic problem is one of energy supply and the stress required to overcome resistance of the mantle to flow. It has frequently been concluded (see for example McKENZIE, 1972) that thermal convection, only, can supply energy sufficient to drive plates. Nevertheless, two non-thermal energy sources are of sufficient magnitude. These are: the creation of buoyancy by differentiation of mantle material (hence the release of gravitational energy); and the dissipation of Earth rotational energy by tides. Of these, differentiation buoyancy has earlier been identified (RINGWOOD, 1973). It may well play a part in driving convection, but explains no more effectively than does thermal buoyancy the geometry of plate motion. Both types of buoyancy create forces having limited lateral extent, resulting in the well-recognized limitations on cell size.

Fig. 1. Gravity geoid, after GAPOSCHKIN (1974), and incidence of deep ($h > 300$ km) hypocenters, recorded by WWSSN 1961-1967 (BARAZANGI and DORMAN, 1969). Reference ellipsoid, $f=1/299.67$; terms to 18th degree and order. Small black circles indicate axis of sea floor spreading; circles do not represent individual epicenters. To eliminate uncertain events, magnitude of deep hypocenters is not less than 5. Contour interval, 10 mGal.

Accordingly, in what follows we have examined the effect of tidal bulge passage on mantle in convection, and this specifically in regard to the Pacific plate. A companion paper (BOSTROM, 1978a) considers the flow processes responsible for the orogenesis at the eastern Pacific margin.

2. Observed Pacific Plate Motion

The seismicity of the Pacific region played a formative part in the concept of plate tectonics. MCKENZIE and PARKER (1967) noted the magnetic pattern and seismic characteristics of the sea floor spreading and consumption centers of the North Pacific, and concluded that "these observations are explained if the sea floor spreads as a rigid plate and interacts with other plates in seismically active regions which also show recent tectonic activity." The westward or northwestward motion of the Pacific plate apparent at that time from focal mechanism studies has been confirmed by later observations. The latter furthermore have been interpreted to suggest that the lithosphere at the west Pacific margin plunges into the asthenosphere and lower mantle as a slab (OLIVER and ISACKS, 1967; ISACKS *et al.*, 1969).

One line of evidence is now held as to the motion of the Pacific plate relative to the mantle (rather than to other plates). This is the trace of mantle magma sources or hot spots on the lithosphere plate (WILSON, 1965, 1973). MINSTER *et al.* (1974) believe that mantle sources may be immobile relative to each other.

The combined evidence suggests that the Pacific lithosphere plate forms at a rate of about 10 cm per year at the crest of the East Pacific Rise; moves across the Pacific Ocean where it encroaches on the eastern flank of Asia and Australasia in a belt of convergence; and reenters the planetary interior in subduction zones extending from the Kermadec region, north of New Zealand, to Kamchatka.

Deep seismicity ($h > 300$ km) is confined to low and intermediate latitudes, suggesting that deep subduction is similarly restricted. As KAULA (1970) has pointed out, the gravity relationship (Fig. 1) displays that a mass deficiency drawing material towards the west Pacific margin does not exist; material is crowded into the subduction region, where it is reassimilated in the mantle, by an external force system.

3. Vorticity Induction under Tidal Bulge Passage

A connection has been sought for some time (DARWIN, 1879; EDDINGTON, 1923; SHAW, 1970; JOLY, 1926; JEFFREYS, 1926; MOORE, 1973; DEHLINGER, 1971; DANES, 1973; JORDAN, 1974) between geotectonics and the action of tides. The connection has been obscured because usually the stress imposed by the weak, secondary retarding torque has been taken to be the only tide stress inducing cumulative deformation (DARWIN, 1879; JEFFREYS, 1929; JORDAN, 1974). The stresses caused by tidal bulge formation are 10^8 times more intense, but have been thought to entirely reverse through time, thus not result in cumulative deformation.

However the tidal bulge does not form in place, but by the passage of a wave around the Earth, namely by the addition of simple shear always in one direction. Inevitably, the real Earth is not perfectly elastic. In consequence the tidal wave is not irrotational as commonly assumed, and imposes cumulative rotational strain or vorticity. The mechanism is as follows:

Fig. 2. (a) Tide-generating force in Equator plane. At a fixed point on Earth the direction rotates without reversal. (b) Displacement of particle at point P (point of arrow), in horizontal plane, at 30°S . Particle is displaced a few centimeters twice daily around ellipse, always in same (counterclockwise) direction without reversal. (c) Instantaneous stress axes i constantly lead axes of finite strain ellipsoid, f . (d) Resulting distortion of elastic rhomboidal element. In response to rotating stress axes, strain progression is via incremental simple shears. Distortion is non-cumulative. Material consists of predominant, perfectly elastic fraction and commingled minor fraction imperfectly elastic, see Fig. 3.

The bulge advances as a result of the Earth rotating in the tide-raising potential field of itself plus the Moon (Fig. 2). At a fixed point in the Earth, the tide-raising force, hence the imposed strain field, does not oscillate back and forth; it rotates, always in one direction. The elastic fraction of mantle material does not experience cumulative strain (Fig. 2(d)).

Intermingled with the elastic fraction is the anelastic fraction. In consequence of its presence (Fig. 3), the observed tidal bulge slightly lags in time the overhead passage of the Moon or Sun. The phase lag describes the anelastic bulge. The anelastic bulge is the bulge which would remain if tidal forces were instantaneously withdrawn. Its parameters are the difference in magnitude and direction of the principal axes between those of the observed and the equilibrium or elastic bulge.

The simple shear (pure shear plus rotation, cf. LOVE, 1944) in the elastic portion of the material (representing almost all of the observed displacement of a few tens of centimeters under the bodily tides) incorporates a rotation amounting to $\tan^{-1}(\epsilon_1 - \epsilon_2)$ where ϵ is the strain, i.e. about 2.9×10^{-6} degrees. The rotation embodied in the anelastic portion is even more minute; but is cumulative, rather than self-cancelling through time.

The anelastic, bodily tidal phase displacement in the real Earth is difficult to observe, due to the gross disturbing effect of the water tides. In the case of M_2 , a large, unknown part of the observed phase lag represents elastic deformation resulting from water dis-

Fig. 3. (a) Distortion of imperfectly elastic volume element. Sides of rhombus rotate through small angle $(\gamma/2)_{\text{obs}} = 1/2 (\epsilon_1 - \epsilon_2) \approx 5 \times 10^{-8}$ rad. Stress axes rotate in counterclockwise direction. The direction of principal strain axes lags overhead position of Moon and Sun by the small angle $\alpha \approx 0.5^\circ$. $(\gamma/2)_{\text{anel}}$, the anelastic rotation of particle lines in the material (dash line), is small fraction of $(\gamma/2)_{\text{obs}}$; but is cumulative. In terms of the observed magnitude of ϵ_1 , ϵ_2 and phase lag α , $(\dot{\gamma})_{\text{anel}} \approx 2.5 \times 10^{-8}$ rad/year, entailing westward displacement of the surface relative to base of mantle of 7.3 cm/year. (b) Deformation mode of continuous medium under rotational strain $(\gamma)_{\text{anel}}$. Volume elements within surface S rotate as shown against the viscous resistance of the mantle. The cumulative deformation represents rotational flow (cf. LAMB, 1932) or vorticity, $\zeta = (\partial u_2 / \partial x_r - \partial u_r / \partial x_2)$ where u is velocity. Circulation is in direction shown (counterclockwise) under bulge passage, and in the west limb of convection cells.

Table 1. Phase of diurnal solid-Earth tide (DUCARME and MELCHIOR, 1977).

Station	Location			Phase of O_1 (α_{01})	Mean square error
	Latitude	Longitude			
Brussels-Uccle	50 48 N	04 22 E		-0.20°	0.34
Helwan-Cairo	29 51 N	31 20 E		-2.05	0.75
New Delhi	28 38 N	77 11 E		-0.61	0.69
Bangkok-Latprao	13 48 N	100 36 E		-1.50	0.54
Kuala Lumpur	3 07 N	101 39 E		-13.86	1.22
Hong Kong	22 18 N	114 10 E		-4.61	0.83
Baguio	16 24 N	120 34 E		-7.40	0.77
Bandung	6 54 S	107 38 E		+11.04	1.63
Port Moresby	9 25 S	147 09 E		+0.03	0.81
Canberra	35 19 S	149 00 E		-0.96	0.61
Perth	31 59 S	116 13 E		+2.54	0.35
Hobart	42 55 S	149 19 E		-2.25	0.25
Lauder	45 02 S	169 41 E		-1.54	0.46
Suva-Vunikawai	18 03 S	178 28 E		+2.39	1.19

—, phase lag; +, phase lead. The above observations were made using gravimeters calibrated at the base station (Brussels) for not less than 4 months.

placement in the marine tides. MELCHIOR (1974) has pointed out that the remaining, anelastic portion of α observed may be measured in terms of the phase displacement of the bodily diurnal tide O_1 . In respect to O_1 , water displacement is small. The gravimetric phase lag, which in western Europe is -0.2° to -0.4° , is a fraction of the phase lag of the actual bulge. A global profile of O_1 is now partly complete (DUCARME and MELCHIOR, 1977) (Table 1). It is already apparent that α_{grav} is small, between -0.2° and -0.5° in stable regions; but that this angle is large, ranging from -14° to $+23^\circ$ in actively orogenic regions, notably the west Pacific margin. Its variation is inexplicable in terms of water displacement.

The imposed cumulative rotational strain represents the induction of rotational flow or vorticity of the form shown in Fig. 3(b). The generation of vorticity by a "collapsing deformed surface" has already been demonstrated by McKENZIE (1968, his Eq. (3.10)). In the present case the collapsing surface is represented by the westward-sweeping tidal bulge. The state of present information therefore is that the observed anelastic phase lag is sufficient, in itself, to account for the greater part of the observed motion of the Pacific plate.

4. Convection under Buoyancy and Bulge Passage

Vorticity likewise is induced by buoyancy, resulting in convection if certain conditions are met. The driving force in buoyancy convection and the forces opposing it are represented by the flow equation

$$\bar{g}\rho\phi(T-T_0)=\eta\nabla^2\bar{u}-\nabla P \quad (1)$$

in which ρ is the density, ϕ the coefficient of thermal expansion, T , T_0 the temperature at different locations, η the dynamic viscosity, $\bar{u}$ the velocity and P pressure (cf. McKENZIE *et al.*, 1974, their Eq. (10)).

In comparison,

$$\nabla\tau_\alpha+\rho(\bar{u}\cdot\nabla)\bar{u}=\eta\nabla^2\bar{u}-\nabla P \quad (2)$$

describes the forces driving and opposing flow induced by tidal bulge passage. In (2) the buoyancy stress is replaced by

$$\tau_{\alpha} = \tau_{\text{equil}} \tan \alpha,$$

in which the first R.H. term represents the equilibrium (symmetric) stresses in bulge formation. The magnitude of τ_{α} accordingly is a few tens of dyn/cm², based on the value 0.2° of the lag angle α and the value 10⁴ dyn/cm² (TAKEUCHI, 1951) of the stress differences in bulge formation.

In respect to both (1) and (2), taking the curl of both sides eliminates P ; but vorticity remains. The reason in the case of (2) is that shears are present acting always in the same sense, as the bulge stress axes constantly lead those of the finite strain ellipsoid. In both cases, the flow represents the continual induction of rotational flow and its dissipation under viscous friction.

It is not possible to solve the Rayleigh equations of convection for the real Earth due to the unknown relationship between T , P and η , far less to reach simultaneous solutions of (1) and (2). However it is possible to discern the contribution of each to the flow. The distribution of the torques due to buoyancy and to bulge passage is compared in Fig. 4. The usual means of solution of the Rayleigh equations of convection (by exchange of variables) obscures the origin of the large stresses generated in thermal convection. The primary stress between volume elements

$$\tau_{\text{buoy}} = \Delta T \phi \rho \bar{g}$$

is minute. If for instance the lateral thermal gradient in the mantle were as great as

Fig. 4. Comparison of flow under bulge passage and flow under thermal buoyancy: Buoyancy torques (b) represent difference in gravitational attraction on neighboring volume elements 1, 2, etc., hence act only where there is present a horizontal gradient in the density. Primary stress acts only in the vertical. For Rayleigh numbers specified by mantle viscosity and thermal gradient, Pacific plate would be underlain by multiple flow cells, breaking it up. In contrast, flow under bulge passage (a) is induced by torques acting uniformly, at all points clear across Pacific. The flow induced by bulge passage is of mode identical to that representing the west limb of convection cells; and opposite to that in east limbs. Under either or both forces stems, stresses sum to values $> 10^8$ dyn/cm², according with orogenesis and seismicity of west Pacific margin.

100°C/500 km, $\rho=3.5$ g/cm³ and $\phi=2\times 10^{-5}/^\circ\text{C}$, τ_{buoy} is only 1.4×10^{-7} dyn/cm². The large stresses in buoyancy convection arise because of its organized, cellular form. Once flow commences, the buoyancy of volume elements forming a cell is cooperatively through stress transmission.

For the same reason the velocity in convection varies (TOZER, 1967) as $1/3 \kappa/d R^{1/2}$ (κ thermal diffusivity), where the Rayleigh number R is a function of the fourth power of the layer thickness, d . Stress in a layer d thick of material having a viscosity ν , attains at its maximum (KNOPOFF, 1964) the value

$$\tau_{\text{max}} = (\pi/\sqrt{2}) (\rho \nu u_{\text{max}}/d).$$

Typically, for whole-mantle convection, $\nu=10^{22}$ P, u_{max} as indicated by plate motion 8 cm/year, $\rho=4.0$ g/cm³. $d=2.9\times 10^8$ cm, τ_{max} is 1.3×10^8 dyn/cm².

In the case of the tidal torques, Eq. (2), identical conditions apply. Under flow, the stress acting at a point beneath the Pacific plate is additive to that developed at similar points across the whole width of the plate, which acts as a stress transmitter in respect to the highly coherent bulge stress. The horizontal limb in the case of tidal vorticity induction is the analog of the vertical limb in buoyancy convection. The buoyancy force acts only where there is present a horizontal gradient in the density, the physical reason why it is difficult to model cells having the width of the Pacific Ocean, one third the circumference of the Earth. However tidal vorticity induction acts ubiquitously and in phase across this great distance. Based on the plate width, the stress thus developed assuming that the anelastic phase lag α is 0.2° is (plate width) times τ_α , namely several times 10^8 dyn/cm², comparable with or greater than the stress developed in buoyancy convection. The direction of the flow under vorticity induction, representing the tidal contribution to the convection (Fig. 5), is that of the observed motion of the Pacific plate.

The observed phase angle α describes dissipation in the mantle as a whole. The viscosity distribution in the mantle requires that the flow responsible for the dissipation is concentrated in the asthenosphere, in the fashion described by TAKEUCHI and SAKATA (1970). If the base of the lithosphere is the site of a layer of partial melting or lubrication as deduced by ANDERSON and SAMMIS (1970), flow must be concentrated within it. In this case, in a sense, the lithosphere slides or works its way over the more slowly flowing lower part of the asthenosphere.

Measured in terms of the dissipation as a function of α , the tidal contribution to the flow is independent of assumed values of the mantle viscosity.

The phase-derived dissipation if α_{grav} is -0.5° is (MELCHIOR, 1974) 8.5×10^{26} erg/year. The flow must accord with the flow resistance of the mantle in respect to tide stresses, next examined.

5. Tide-Effective Viscosity

Determinations of the viscosity of the mantle with respect to surface loads (O'CONNELL, 1971; CATHLES, 1971) indicate that its value is between 10^{21} and 10^{23} P. In this case the elastic quality factor Q of Earth in respect to elastic tides has a value of millions (OROWAN, 1967) and the dissipation is negligible. Q in respect to the decay of seismic waves and eigenvibrations (ANDERSON and KOVACH, 1964; SMITH, 1972) is less by a factor of thousands, namely about 650. On this basis dissipation in respect to the oscillatory part of bodily tidal deformation is 10^{26} erg/year or less. Estimates of the displacement in re-

Fig. 5. Form of the circulation contribution under tidal vorticity, shown as stream lines. The viscosity gradient is very steep. In consequence the flow must be concentrated in the asthenosphere in the manner found by TAKEUCHI and SAKATA (1970). Return flow is slow, in the thick lower mantle. In sum, as much material is displaced east as west. Note flow is under internal torques; the crust is not "towed around over a fluid layer" by an external force system, which as JEFFREYS (1929) has pointed out would catastrophically reduce the Earth's rotation. The rings describe the sense of the vorticity vector. Vorticity is induced ubiquitously and continuously, with rotation of bulge-stress axes. Circulation causing west displacement of surface is fostered in low latitudes, thus the west limbs of convection cells are fostered. With increase in latitude, due to the angle made by the vorticity vector to the Earth's surface, clockwise circulation is fostered in the north hemisphere, and vice versa. Note switch at Equator of the polarity of the transforms.

sponse to tides, assuming perfect elasticity (TAKEUCHI, 1951; TAKEUCHI and SAITO, 1962) yield Love numbers close to those observed.

The answer to the question of why the lag-determined dissipation is several times 10^{26} erg/year is twofold.

Firstly, (*supra*) under a flow regime of large cells, bulge passage imposes steady stress which is additive across cell-wide segments of the mantle. The stress thus developed, $>10^8$ dyn/cm², is comparable to or greater than that developed by buoyancy convection.

It is therefore able to account for mantle flow and ultimately for orogenesis, either in its own right or in conjunction with buoyancy forces.

Secondly, the stress-strain relationship in respect to the processes comprising "flow" in the Earth is extraordinarily non-linear. The tide stress is imposed (Fig. 6) on mantle already stressed to the point of failure by the convection. Mantle convection comprises multifarious processes of material and energy transference. These include not only high-temperature creep, but fracture, fractionation including vulcanicity, and phase transition. CARTER (1976) has shown that in respect to flow in rock materials the strain rate is related to the stress raised to the power 2 to 9. The investigations of KLEIN (1976) and MAUK and JOHNSTON (1973) suggest that seismicity and vulcanicity are non-linear in the extreme, to the extent of being triggered phenomena. With respect to viscosity the ice-loading of continents or the imposition of a stationary load do not impose stress in the mode of a bulge sweeping across the sub-oceanic asthenosphere. The latter imposes strain on a mantle already stressed to the yield point under convection; and does so in precisely the mode (rotational flow) in which it is pre-stressed. An example of the effect to be expected at the west Pacific margin is illustrated in the next section of this paper.

5.1 Energy expenditure

The energy dissipated in the Earth including the seas is known from the acceleration of the Moon (JEFFREYS, 1921; MULLER, 1976) to lie between 8 and 15×10^{26} erg/year. This is larger than the flux required to drive convection and orogenesis. The tidal dissipation has several times been attributed almost wholly to the seas. Internal dissipation as in the mantle would then be small.

HENDERSHOTT (1972) has quantified the total dissipation in terms of the observed ocean tides. Gauge-observed water tides are used to determine the deformation of the sea floor, assuming perfect elasticity. Integrated over the oceans, the work done by M_2 is found to be 9.6×10^{26} erg/year, explaining the acceleration of the Moon.

LAMBECK (1975; p. 2923) has compared Hendershott's tidal dissipation rate for M_2 with the orbitally derived measure of the dissipation. As the difference is small he concludes that little dissipation is effected in the solid Earth. However Hendershott's formu-

Fig. 6. Viscosity $\eta = \sigma / \dot{\epsilon}$ with respect to superimposed stress in pre-stressed, non-linear material. In metastable regions of the mantle, such as those (Fig. 7) in subduction, material is pre-stressed (dotted line) to point of failure in yield, fracture and compression; superimposed stress $\delta\sigma$ as by tides results in large deformation. Effective viscosity (slope of curve) is small. On the contrary, in stable regions locally stressed such as Fennoscandia by ice loading (G), effective viscosity is high.

lation by intention includes no dissipation terms, in respect either to the seas or the solid Earth. The phase displacement in the observed water tide upon which the computation is based is assumed to be due to dissipation "somewhere" outside the boundary of the computation. His formulation thus makes no statement as to whether the phase displacement in the motion of the seas and sea floor is due to dissipation in the seas or in the solid Earth (for instance, the thick sub-oceanic asthenosphere). The elastic Love number used in the computation is insignificantly different from the anelastic number appropriate if all dissipation were to take place in the solid Earth.

Extra-terrestrial data such as the acceleration of artificial satellites cannot separate the phase lag due to dissipation in the sea and underlying floor. The seas and lossy part of the asthenosphere are coincident and their spatial spectra the same. To effect the separation MILLER (1966) measured the flux of tidal energy into shallow seas, where almost all the marine dissipation takes place. The flux thus measured is between 4.4 and 5.5×10^{26} erg/year. This accounts for about half only of the astronomically-detected dissipation, but accords with the large value of the dissipation in the solid Earth indicated by the lag in O_1 . It accords also with the large value of the quality factor Q , $Q=34$, derived by Hendershott. To account for the dissipation entirely in terms of marine processes, the value of Q_{marine} (MUNK and MACDONALD, 1960, p. 218) must be about 2.6; see also NEWTON (1972, 1973, 1976).

We therefore are inclined to conclude that about half the dissipation astronomically measured, namely between 4 and 7.5×10^{26} erg/year, is available to drive or modulate flow processes in the mantle.

6. *Structure of Pacific Margin*

The margin of the Pacific basin was at one time regarded as tectonically uniform; hence the Pacific "ring of fire." Marginal structure is now seen to vary strongly from region to region. The west margin consists of minor ocean basins outlined by island arcs and underlain by Benioff zones. Benioff zone features (the tectonic trench, the Meinesz gravity strip, sub-lithosphere seismicity) are absent in the Canada/United States portion of the Pacific rim with the exception of Alaska. The Andean sector is sublain by a Benioff zone, but this region is devoid of marginal basins. As pointed out by DICKINSON (1978) and others, the Pacific basin is remarkably asymmetric. The west Pacific is the site of a continuous string of marginal basins. In the east these are absent. The Pacific plate, representing the west limb of the East Pacific Rise axis of sea floor spreading, occupies most of the ocean floor. The east limb is one fifth this size.

6.1 *Flow regime, west margin*

Seismologic evidence (ISACKS and MOLNAR, 1969) and the geologic record indicate that the west Pacific margin, the most active region of the planet, is the site of large-scale foundering of lithosphere. In terms of convection, this represents the return to the planetary interior of material which originally upwelled to form new lithosphere at the crest of the East Pacific Rise, 6,000 km to the east. Theory, laboratory and seismologic evidence (RINGWOOD, 1973; AHRENS, 1972; JOHNSTON, 1969) require that as material under surface conditions passes into regions of high pressure and temperature its minerals transform into dense phases, to the accompaniment of energy transference. In flow the process represents

Fig. 7. (a) Creation of metastable material in subduction region. As material founders minerals such as olivine change phase to dense, high-pressure form. Transition takes place in over-pressured, metastable zone. Probability of transition to dense form increases to 100% at base of metastable zone, and transition is complete. Transition surface is complicated and may be concave upward or downward dependent upon whether slope of equilibrium curve is positive or negative. (b) Regime in metastable zone. At typical foundering rate 5 cm/year, the hydrostatic pressure change ($+5.5 \times 10^{-4}$ dyn/cm²/sec) is represented by the slope of the almost horizontal line. The pressure change due to tidal gravity disturbance is about one thousand times this rate. The oscillatory curve represents the total pressure change in transition region under tides. Over a period of many cycles, transition must take place predominantly when the total pressure is rising (dashed line), rather than falling. Note that process cannot operate in stable regions, where metastable material is not continuously created by convection.

volumetric or bulk viscosity, because the released energy becomes locally unavailable and the deformation (compression) is irrecoverable.

The flow regime in this region is illustrated in Fig. 7. Mineral species such as olivine, stable near surface, experience transition into high-density forms at depth. Within the down-flowing limb, iso-phase surfaces may be convex upward or downward, dependent upon the sign of the curve describing the conditions of transition. Laboratory and seismologic evidence (AKIMOTO and FUJISAWA, 1966; RINGWOOD, 1973; JOHNSTON, 1969) indicate that phase transitions are numerous and overlapping, and extend to the lower mantle.

Putting aside at first the effect of tides, foundering causes the increase through time of both pressure and density. On the average the pressure increase due to the downward flow component, if the latter is 5 cm/year and the density 3.5 g/cm³, is 5.4×10^{-4} dyn/cm²/sec.

Upon this slow secular increase is superimposed the rapid, much larger, but oscillatory pressure variation due to the tidal disturbance in gravity. Assuming the latter typically to be 0.1 mGal, the tidal pressure variation is

$$h\rho\Delta g$$

where h is depth of the volume element. The rate is 2.6×10^{-1} dyn/cm²/sec at a depth 350 km. Locally, water-loading increases this variation by an order of magnitude (ZSCHAU, 1977). The pressure change due to the tidal gravity cycle is thus hundreds to thousands of times greater than the pressure increase in the same length of time under foundering. The total pressure actually decreases in the negative part of the tide cycle, Fig. 7(b).

Within the metastable, over-pressured material continually created under foundering, it seems inevitable that phase transition preponderantly takes place when the total pressure is increasing; and not when it is decreasing. Conversely, with respect to its bulk viscosity

the material of the mantle must respond elastically to the tidal pressure variation in regions where upflow or downflow is not present (not creating material at the metastable point or point of failure in volume). It seems no coincidence that large tidal phase lags are typical of the west Pacific region. Thus for instance the lag in the diurnal bodily tide, little affected by water displacement, is -7° at Manila (Table 1).

In regions such as the west Pacific margin, where the resistance of the material in volume to the tidal pressure variation has been overcome by foundering and metastability prevails, volumetric viscosity must be a dominant loss mechanism. Apparently, the advance in the convection must be modulated significantly by the tidal pressure variation. GRIGGS (1972) has pointed out that the energy released by the foundering lithosphere is very large, of the order 10% of the Earth's total heat efflux, i.e. about 10% of 10^{28} erg/year. Thus even a minor modulation of the advance in the convection (such for instance that 60% is accomplished during pressure-rise times versus 40% during the remaining half of any time interval) could by itself account for the entire "missing" fraction of the tidal energy dissipation.

Analogous tidal modulation, of phase decompression and partial melting, must take place in upflow regions such as the axis of the sea floor spreading system. KLEIN (1976) has demonstrated synchronization of tide stress and seismicity in the portions he has examined, namely the north segment of the East Pacific Rise and the north half of the mid-Atlantic Ridge. In upflow regions the advance in convection takes place during tidal pressure minima.

6.2 *Formation of marginal basins*

As in his discovery of mantle plumes (WILSON, 1965a), transforms, and plates (WILSON 1965b), WILSON (1973) has pointed to the significance of relative motion in the formation of marginal basins. It seems to matter profoundly, in plate convergence, which plate is more nearly stationary relative to the mantle. If for instance a continental plate is fixed and oceanic plates encroach upon it, island arcs will form, enclosing marginal basins.

This circumstance becomes more comprehensible when considered alongside the fact (NELSON and TEMPLE, 1972) that, world-wide, subduction zones facing east differ from those facing west. As pointed out by UYEDA and KANAMORI (1978), if the motion of the mantle relative to the lithosphere is eastward, the east-west asymmetry is explicable. In terms of the tidal contribution to vorticity, the subduction zones at the west Pacific margin benefit from the tidal contribution, and from induction across the trans-oceanic expanse of the Pacific plate. Not coincidentally, the downgoing slab assumes a steep angle, and is in places overridden or overturned. At the east or Andean margin, on the contrary, we are dealing with the weak, counter-tidal east limb of the East Pacific Rise. This bucks tidal vorticity induction, and is rapidly being supplanted (BOSTROM, 1978a) by the west, co-tidal limb of the mid-Atlantic spreading center. The slope of the Benioff zone is less, the volume of material foundering is less, the continental plate is overriding the oceanic for buoyancy reasons, and no marginal basins are present.

It is possible to account for these phenomena simultaneously if we allow tidal vorticity induction across the expanse of thick asthenosphere underlying the Pacific basin. By now the west limb of the East Pacific Rise is grossly extended relative to the east limb; and the larger the west limb becomes, the greater the induction. An increasing volume of material is being carried to the terminus of the west limb, the east flank of the

Asian continent. The accumulation of material at the west Pacific margin, marked by the string of geoidal highs (Fig. 1), is subject to intense negative buoyancy, and founders. Mineral phases in equilibrium under shallow conditions cannot be in equilibrium at depth. Their transition and the role of entrained volatiles is the subject of active petrologic research (FEDORCHENKO and PISKUNOV, 1978; YODER, 1978; WYLLIE, 1978). Necessarily a fraction of the material separates upward. Material is added to the asthenosphere, resulting in secondary sea floor spreading and the formation of the arc-bounded basins described by KARIG *et al.* (1973) and others.

It is to be expected, and found to be the case, that at the terminus of the weak, counter-tidal east limb of the East Pacific Rise, basin formation will itself be weak. Basins, if they form, will be overridden and incorporated (DALZIEL, 1978) in the advancing west edge of the west limb of the Atlantic spreading center, namely the Andean margin of South America.

Basin formation and tectonic activity is latitude-dependent, as is to be expected under tidal vorticity induction. The tropical west Pacific margin is the site of the greatest geoidal high on the planet, and accounts for more than 80% of seismicity, deep, intermediate and shallow. Deep seismicity ($h > 300$ km) decreases with latitude (BOSTROM, 1973) and disappears altogether towards the poles. The fragmentation of the east Asian margin and the increasing "oceanicity" of the marginal basins is likewise latitude-dependent. The floor of the Sea of Okhotsk in the north is thought largely to consist of altered and foundering continental material (KROPOTKIN, 1978; YANSHIN, 1978). The oldest rocks in Japan formed when Japan was attached to the Asian mainland, the Sea of Japan having developed and widened since. Since Oligocene times the tectonics of Japan has changed as indicated by the development of the Fossa Magna, as a result of the alteration in the subduction regime and the opening of the Philippine Sea to the south. This coincided with the destruction in the northern hemisphere (BOSTROM, 1978 a,b) of the east limb of the East Pacific Rise, now overridden by the Californias. Co-tidal vorticity now characterizes the upper mantle continuously without a break from the middle of the Atlantic, across North America and the entire width of the Pacific, to the shores of Japan.

The Pacific plate is at its widest at the Equator, at this latitude extending furthest east and furthest west. The structure of the intensely active Indonesian region has already been attributed (KROPOTKIN and SHAKVARSTOVA, 1965; KROPOTKIN, 1978) to the fragmentation of the southeast portion of the Asian continent. With increasing latitude southward, seismic activity and subduction gradually decrease. Deep seismicity, associated with the entry of material into the deep mantle, ceases at the latitude of New Zealand.

The latitude dependence of the spreading at the crest of the East Pacific Rise, the symmetry of the Pacific basin about the Equator, and its east-west asymmetry, are shown in Fig. 8.

7. *Historical Development of the Circulation*

It seems at first curious, if the circulation is marshalled by tidal bulge passage, that mantle circulation is not ubiquitously co-tidal, and has not been thus patterned since convection commenced. The reason may lie in the re-setting of the circulation pattern by a supercessive force.

It has been postulated elsewhere (BOSTROM, 1978b) that the circulation in the mantle

Fig. 8. Pacific plate, representing surface of the west limb of the East Pacific Rise (EPR; Pleistocene at crest, *black*). The east-west spreading rate, indicated by distance to Paleocene/Eocene boundary (*stippled*), is symmetrical about Equator. The spreading maximum coincides with the Equatorial maximum in vorticity induction. The grossly extended west limb of the EPR forms most of the Pacific basin. The counter-tidal east limb is being replaced by the extending west limb of the mid-Atlantic spreading center (MAR) bearing the Americas plates. The sense of the transforms reverses at the Equator, to predominantly sinistral in the south hemisphere. Departure from north-south symmetry is mainly due to the excess of continental lithosphere in the north hemisphere; the North America plate has overridden the crest of the EPR. The convergence of N America and Siberia has caused the development of the Bering Sea marginal basin. We attribute the development of the basins flanking the west Pacific margin (*hatched*) to convergence between the co-tidal west limb of the EPR and the less mobile continental material of Australasia and eastern Asia. As a result of differentiation material is added to the asthenosphere, causing the disruption of the continental margin, sea floor spreading, and the formation of back-arc basins. At the east Pacific margin, on the contrary, the west limb of the Atlantic spreading center is overriding and extinguishing the east limb of the EPR. Data: HEEZEN and FORNARI (1976). Scale identical at all latitudes, $1: 10^8 \times 1.5$.

is determined by the interplay of two factors. The first is the marshalling effect of tidal bulge passage on buoyancy convection. The second factor stems from the fact that in a convective Earth, Goldreich-Toomre wandering of the Earth about the space-fixed pole of rotation (GOLDREICH and TOOMRE, 1969) is inevitable. The manner in which the bulge of rotation migrates is by adjustment of the sea-floor spreading pattern so as to effect the necessary mass transfer (SILLING, 1978).

We therefore suppose that tidal ordering of the convection becomes dominant during

periods such as the post-Paleozoic period of minimal pole wander. The present cycle in the Pacific hemisphere was initiated in the early Mesozoic. The pattern inherited at that time was the result of the period of pole wander occurring during the Paleozoic, extending into the Mesozoic. Thus the post-Paleozoic geological record in the Pacific basin and rim (see for instance LARSON and CHASE, 1972; BOSTROM, 1977; SUZUKI *et al.*, 1978) describes gradual marshalling of the flow pattern in the mantle, culminating in the highly ordered present pattern and concentration of activity at the west margin. World-wide, hot-spot traces unequivocally describe the present predominance of westward lithosphere motion in low latitudes (BOSTROM, 1978b).

8. Conclusions

The Pacific plate is the cool, rigid surface of mantle convection surfacing in the eastern Pacific, and foundering at the west Pacific margin. The convection represents vorticity (rotational flow) induced jointly by thermal buoyancy and the passage of the M_2 tidal bulge. Thermal convection, which would by itself produce small random flow cells, is marshalled by the highly coherent bulge stress.

The west limb of the East Pacific Rise, representing co-tidal vorticity, now extends clear across the Pacific basin. The counter-tidal east limb, underlying the Nazca, Cocos and small Juan de Fuca plates, is being replaced by the extending west limb of the Atlantic spreading center, fronted by the Andean orogenic belt.

8.1 Marginal basins

Material accumulates at the west margin of the Pacific as a result of the convergence between the co-tidal, active west limb of the EPR and less mobile material which forms the east margin of Asia. The accumulation is marked by an unbroken line of gravity highs, coincident with deep seismicity. Deep-seated differentiation results in the addition of material to the asthenosphere. This causes the disruption of the continental margin, localized sea floor spreading, and the formation of arc-bounded basins.

At the east Pacific margin this mechanism is not operative, because the co-tidal west limb of the Atlantic spreading axis (bearing the Americas plates) is overriding and extinguishing the counter-tidal east limb of the EPR.

8.2 Stress, strain, and energy flux

The stress driving the convection is supplied by thermal buoyancy and the continuously rotating bulge stress axes. The elemental shear stresses contributed by each of these agencies are minute. However in large flow cells the buoyancy stress is cooperative and additive vertically through great thicknesses of the mantle, resulting in regional stresses of the order 10^8 dyn/cm², and the stress concentrations at plate edges observed in seismicity. Analogously, the bulge stress is additive horizontally across the width of plates; the larger the plate, the larger the induced stress. Across the Pacific plate the bulge stress sums to values greater than the buoyancy stress.

The resulting strain and flow in the mantle is enhanced by the grossly non-linear relation between stress and strain. The latter includes processes of fracture, fractionation and compression. Thus in a subduction region such as the west Pacific margin, continuously-renewed volumes of material exist in an over-pressured state through downflow. The tidal pressure increment, larger than that in downflow, must effect the transition to

dense phases, modulating the convection. In mantle continuously pre-stressed to the point of failure in shear or volume, the tide-effective viscosity thus is small.

The geologic phenomena suggest that thermal buoyancy and the rotating bulge stress contribute about equally to the convection. The energy flux required in toto is about 10^{27} erg/year (mean viscosity 10^{22} P; mean plate motion, 3 cm/year). Of this the phase lag in the diurnal bodily tides suggests that about 5×10^{26} erg/year is contributed by tidal dissipation. This value accords with the dissipation required to explain the lunar acceleration, but not found in the seas. Based on this reasoning, sea floor spreading is passive, the response to forced convection rate-limited by the rate at which buoyancy is created.

8.3 Data status

The magnitude and phase of the bodily tides (Table 1) varies greatly from one region to another. Besides revealing the large dissipation, the variation has revealed the inadequacy of our station coverage. It would be of value from the geotectonic point of view to extend the observations initiated by DUCARME and MELCHIOR (1977) to a much larger number of stations. To do so, it is necessary to use instruments (gravity meters) calibrated as to their individual phase delay through side by side running at a base station (DUCARME, 1975). It would also be worth using the technique used by KLEIN (1976) in the east Pacific to identify correlations between tide stress and seismicity at the west Pacific margin. Into the bargain, such correlations might eventually prove valuable in forecasting periods of enhanced seismicity.

Note Added in Press: Tidal Dissipation in the Seas and Solid Earth

The recent important paper by LAMBECK (K. LAMBECK, *Phil. Trans. Roy. Soc. Lond., A*, 287, 545-594, 1977) reviews data as to the energy expenditure in tidal action. The conclusion is reported, in the abstract, that "dissipation occurs almost exclusively in the oceans and neither the solid Earth nor the Moon are important energy sinks."

This is based upon finding that an estimate of work done on the oceans by body forces and the heaving sea floor matches the dissipation required to account for the acceleration of the Earth and Moon. The work done on the world ocean is represented as:

$$\begin{aligned} \langle \overline{W} \rangle &= \rho_w (1 + k_2) \int_S \langle U_{lm} d\xi / dt \rangle dS && \text{term 1} \\ &+ \rho h_2 \int_S \langle \xi dU_{lm} / dt \rangle dS && 2 \\ &+ \rho_w \int_S \langle \nabla_S \cdot [u D \psi] \rangle dS, && 3 \end{aligned}$$

in which: ρ_w and ρ are the density of water and sea floor; h and k are the elastic Love numbers; U is the potential of the direct tide raising potential; ξ is the geocentric sea-surface tide; u is the horizontal velocity of tidal currents; D is the ocean depth; ψ is the total tidal potential; and integration is over the ocean-covered part of the Earth.

In evaluating $\langle \overline{W} \rangle$, term 3 is discarded on the assumption that the boundary of the region of integration is impermeable. Dissipation in the solid Earth also is precluded, through the use of the elastic Love numbers.

These assumptions make it possible to reach a solution solely on the record of coastal marine tide gauges. However, *ipso facto* they preclude our reaching conclusions as to the site of dissipation. The equation contains no dissipation terms. Dissipation is assumed

to take place somewhere, but the phase of the ocean tides upon which the computation is based may be attributed to dissipation in the waters, in the solid Earth, or in both.

In point of fact investigations extending over many decades, cited by Lambeck, have failed to find dissipation in the seas capable of explaining the observed phase of M_2 . Observations not cited, such as those by DUCARME and MELCHIOR (1977) of the phase lag in bodily O_1 , suggest that a large part of the dissipation takes place in the solid Earth. They require, furthermore, that dissipation also occurs in the sub-continental third of the Earth, arbitrarily excluded in the equation by the range S of the integration.

External observations such as those of the flight of satellites, alone or in conjunction with coastal tide gauge records, cannot distinguish whether the phase of the marine tides is determined by dissipation in the seas or subjacent sea floor (asthenosphere).

The estimate of the work $\langle \dot{W} \rangle$ done on the oceans arrived at as shown tallies with the observed acceleration of the Moon. However estimates arrived at on the basis of equally unrealistic assumptions, such as that the Earth is rigid (PEKERIS and ACCAD, 1969), also tally.

The unwelcome fact remains that computations, no matter how intricate, based upon the assumption that dissipation in the Earth is nil cannot measure the dissipation in the solid Earth. Regretfully, therefore, we find it not possible to concur with the conclusion cited at the head of this note.

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