'RESS AT RANGELY

'ault Plane at Stresses over Fracture in Intact Rock*

ıge, deg	
α_2	$\alpha_2 - \alpha_1$
85	41
84	38
81	33
78	27
76	22
72	16
70	11

ipal stress is taken to be lane when derived from This angle could be in 40° where sliding on an :herwise intact rock pro- If the nodal plane corre- is known, σ_1 should be normal to the fault plane irection. In this case the l not be in error by more

Ve are grateful to our col- :, R. de la Cruz, and J. [aimson for discussion and of this paper was author- the U.S. Geological Survey. ed here was supported by h Projects Agency of the e under ARPA orders 1469

Plate Tectonics, the Analogy with Glacier Flow, and Isostasy

F. C. FRANK

Department of Physics, University of Bristol
Bristol, England

A diagrammatic method is presented for calculating resultant horizontal forces on parts of the lithosphere attributable to gravity. These forces give rise to stresses in the lithosphere mainly dependent on changes of large-scale topographic height. Spreading pressures of the ocean floors, sliding away from mid-ocean ridges, approximately balance the spreading pressures of continental mountain masses, in isostatic balance for vertical forces. Shear stresses transmitted to plates from below (by 'conveyor belt' mantle convection) cannot be significantly larger than the gravitational sliding stresses deducible from surface topography and may be small or absent.

In the discussion period following a recent colloquium by D. H. Matthews I expressed the view that the motion of the ocean floors away from mid-ocean ridges should be considered downhill sliding under the force of gravity like that of a glacier and that, as for a glacier, 'downhill' is to be interpreted with reference to the form of the upper surface (in this case, the ocean floor, descending from about 2 km below sea level at the ridge to about 5 km below sea level before reaching the continental shelf, or 10 km below in a trench). As for a glacier, the form of the lower surface is of secondary importance, and this situation is fortunate, because we know comparatively little about the form of the lower surface. Objection was taken to this view on the grounds that it would imply a particular pattern of gravitational anomalies in association with ocean ridges, whereas no strong correlation of gravitational anomalies with ocean ridges is observed. This paper clarifies this issue and shows, among other things, that even a fully isostatic situation, free from gravitational anomaly, is not incompatible with downhill sliding, according to the glacier analogy. I prefer, incidentally, to make the analogy to a glacier rather than to an ice sheet, because the ocean floor, like the glacier, is principally fed with new material in a localized mountaintop region, from which it flows unidirectionally and can flow uniformly, whereas the ice sheet, if persisting in a steady state, is fed with new material and therefore is undergoing extension (often two dimensionally) over a large part of its area; thus the ocean floor is more analogous to a very wide glacier rather than to an ice sheet. (An illness has both delayed the submission of this paper for publication by about a year and prevented a thorough search for previous work. I am indebted to referees for directing my attention to the work of *Benioff* [1949], *Weertman* [1962, 1963], *Temple* [1968], *Pollock* [1969], and *Hales* [1969]. I believe that the method of treating the problem given here is original. I am advised that what I have referred to as pressure-compensated stresses are most appropriately called Benioff stresses.)

Let us start with an extremely simplified model, Figure 1*a*, with a flat earth and the flat oceanic plate ABCD, the ocean floor AB sloping down from the ridge at A and the plate sliding on its lower surface CD and being renewed by the solidification of melt fluids injected from below into the opening gap AC. An arbitrary vertical plane BD passes through the plate, where it approaches either the seismic zone commencing with an oceanic trench or a continent receding from the ridge. We calculate forces acting on the plate, which is assumed to be rigid, by first considering that the forces acting on all its surfaces are hydrostatic. This assumption is clearly correct on AB (the drag

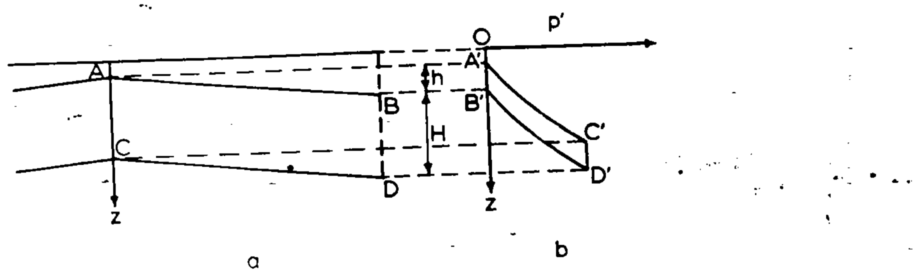

Fig. 1. (*a*) Idealized model of the downhill sliding of the oceanic floor away from the ridge A. (*b*) Pressure gradient in the plate.

ıe to ocean currents is negligible), and, as will ɜ shown later, it is a reasonable one to make ith respect to the horizontal forces on AC, here new melt fluid is injected. The resultant ɔrce thus calculated must be balanced by the onhydrostatic forces acting on BD and CD. Ve take the pressure everywhere to be given y the vertical overburden pressure

$$p(z) = \int_0^z g\rho d\zeta \qquad (1)$$

vhere g is the gravitational acceleration, ρ is he density at depth ζ, which is an integration ariable, and z is the integration limit, which is neasured downward from the surface of the sea, and the integration path is vertical. The rue pressure at a point can deviate from this value by amounts of the order of magnitude of the yield stress of the material by 'arching' or where there is too rapid horizontal change of density, but the mean value of the compressive vertical stress $\langle\sigma_{zz}\rangle$ over any horizontal plane of dimensions that are large as compared with depth must necessarily equal the mean value $\langle p\rangle$ of pressure thus calculated, so that this assumption is well-justified where the vertical distribution of density changes slowly with horizontal position, as in our case. This assumption ensures that vertical forces on the plate (gravity included) are totally balanced except for possible vertical forces on its end. The horizontal force F_H (per unit length normal to the plane of the figure) calculated with these hydrostatic boundary conditions is

$$F_H = \int_{\mathrm{ACD}} p\,dz - \int_{\mathrm{ABD}} p\,dz \qquad (2)$$

the integrals being taken with respect to depth z but along the left and right bounding surfaces ACD and ABD of the plate. Since both integrals have the same range in z, we can also introduce $p_w = \int_0^z g\rho_w d\zeta$, define $p' = p - p_w$, and write

$$F_H = \int_{\mathrm{ACD}} p'dz - \int_{\mathrm{ABD}} p'dz \qquad (3)$$

where ρ_w is the density of sea water (with fictive existence to any depth). Thus the force F_H is given by the area A′C′D′B′ in Figure 1*b*. The slope dp'/dz of either line A′C′ or line B′D′ is proportional to the density of the plate material (less that of water) at the corresponding depth. Taking densities as being uniform makes A′C′ and B′D′ parallel straight lines. Taking densities as increasing with depth, according to the same law at either end of the plate, makes them similar curves, and in both cases we have

$$F_H = (\langle g\rho\rangle - \langle g\rho_w\rangle)hH \qquad (4)$$

where g is the gravitational acceleration, ρ is the density of the plate material, the angle brackets signify an average, h is the height of the ridge A above B, and H is the thickness of the plate. With little error, g can be treated as a constant, and so can ρ_w, which is a fictive quantity for most of the depth concerned.

If there is no shear traction on CD, there is a mean horizontal nonhydrostatic compressive stress on BD:

$$\langle -\sigma_{xx}\rangle = (\langle g\rho\rangle - \langle g\rho_w\rangle)h \qquad (5)$$

Insertion of values $g = 10$ m sec^{-2}, $(\langle\rho\rangle - \langle\rho_w\rangle) = 2.3 \times 10^3$ kg m^{-3}, and $h = 3$ km gives a stress of 6.9×10^7 N m^{-2} = 0.69 kb. A zero nonhydrostatic stress on BD would imply a mean shear traction on CD, smaller in the ratio of H/L, where L is the horizontal length CD, and directed toward the ridge. A shear traction from below directed away from the ridge would

act to increase the comprehensive
BD, with amplification by a factor
stress deduced from free sliding or
to be of a reasonable order of mag
about 10 times greater than the ty
stress release in a large earthquak

Conditions at AC, where the pl
to admit melt fluid from below, w
sively solidifies to renew the plat
little more thought. To conceive of
filled with fluid in hydrostatic equ
not make a satisfactory model. Its
less than ρ, that of the plate mate
on solidification, so that the press
in it is less than the gradient of th
pressure on either side. If its p
suffices to reach the top of the rid
of the plate will be squeezed tigh
lower down, and separation will n
the fluid pressure at the bottom of t
for that matter, the mean pressure
the depth of the slot, suffices to op
the fluid will emerge with a large
sure (the consequence of which will
the ridge to greater height). A mor
model is that the fluid flow is resiste
drag in a narrow gap (or, more re
many narrow channels in a porous
medium, perhaps of substantial widt
despite the lower density, the pressu
in the fluid remains substantially e
overburden pressure gradient on
Then the horizontal forces on the
plate correspond to the 'hydrostatic
condition proposed. The viscous dr
flow makes an upward force (per
of ridge) on the edge of each pla
$\frac{1}{2}gV(\rho - \rho_e)$, if g is taken as cons
is the volume of fluid in the gap

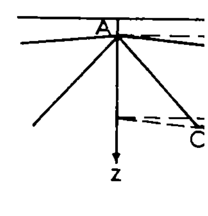

Fig. 2. (*a*) Idealized sliding mod
(*b*) Press

act to increase the comprehensive stress across BD, with amplification by a factor of L/H. The stress deduced from free sliding on CD appears to be of a reasonable order of magnitude, being about 10 times greater than the typical average stress release in a large earthquake.

Conditions at AC, where the plates separate to admit melt fluid from below, which progressively solidifies to renew the plates, deserve a little more thought. To conceive of AC as a slot filled with fluid in hydrostatic equilibrium does not make a satisfactory model. Its density ρ_e is less than ρ, that of the plate material it forms on solidification, so that the pressure gradient in it is less than the gradient of the overburden pressure on either side. If its pressure just suffices to reach the top of the ridge, the edges of the plate will be squeezed tightly together lower down, and separation will not occur; if the fluid pressure at the bottom of the plate, or, for that matter, the mean pressure of fluid over the depth of the slot, suffices to open the gap, the fluid will emerge with a large excess pressure (the consequence of which will be to build the ridge to greater height). A more acceptable model is that the fluid flow is resisted by viscous drag in a narrow gap (or, more reasonably, in many narrow channels in a porous and plastic medium, perhaps of substantial width), so that, despite the lower density, the pressure gradient in the fluid remains substantially equal to the overburden pressure gradient on either side. Then the horizontal forces on the edge of the plate correspond to the 'hydrostatic' boundary condition proposed. The viscous drag of fluid flow makes an upward force (per unit length of ridge) on the edge of each plate equal to $\frac{1}{2}gV(\rho - \rho_e)$, if g is taken as constant and V is the volume of fluid in the gap (per unit length of ridge). This force is of the right sign to maintain the tilted attitude of the plate, but it would be quite inadequate in magnitude if there were a 'true' hydrostatic boundary condition on CD, i.e., one with the material below CD behaving as an inviscid fluid. In that case, the fluid density being ρ, the water-compensated pressure p on the base, minus its mean, is $g(\rho - \rho_w)x \sin \alpha$, where x is distance from the midpoint of CD and α is the angle of tilt of the plate. If the moment of these forces about the point $x = 0$ were balanced by that of the upward force at AC, we should have

$$\tfrac{1}{4}gV(\rho - \rho_e)L \cos \alpha = (1/12)g(\rho - \rho_w)L^3 \sin \alpha$$

and thus

$$V = \tfrac{1}{3}[(\rho - \rho_w)/(\rho - \rho_e)]Lh$$

about 42,000 km^3 of fluid per kilometer of ridge, if $(\rho - \rho_w) \simeq 2.1$ Mg m^{-3}, $(\rho - \rho_e) \simeq 0.1$ Mg m^{-3}, $L = 2000$ km, and $h = 3$ km. This volume of fluid is unacceptably large. An escape from the difficulty is provided as before by supposing that the water-compensated pressure gradient parallel to CD in material just below is practically annulled by viscous drag arising from its downhill creep on more rigid material beneath. This assumption implies that the plastic zone below CD does not reach to great depths comparable to the 1000-km order of magnitude of the length of a plate. On the other hand, of course, if there are plastic motions at such depths, driven by density gradients there, there may be other mechanisms by which these motions transmit the requisite tilting torque to the plate.

Figure 2 shows the effect of modifying the

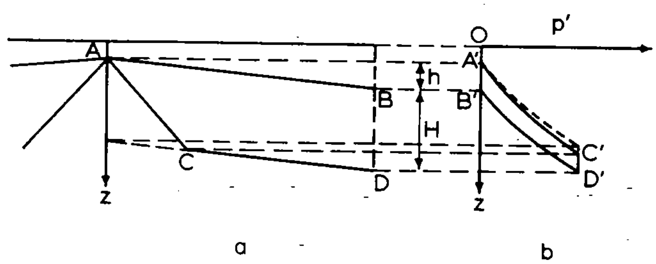

Fig. 2. (*a*) Idealized sliding model from Figure 1 with the modified accretion surface AC. (*b*) Pressure gradient in the modified model.

orm of the bottom surface of the plate. We iave the same sliding surface CD, but we have sloping accretion surface AC. The effect in 'igure 2b is to lower the point C′, reducing the rea A′C′D′B′. This change is not large. In the xtreme of bringing C into coincidence with D nd C′ into coincidence with D′, the horizontal orce F_H and the stresses deduced from it would e halved. The most reasonable assumption is robably a curved lower boundary to the plate, ɔ which the sloping lines AC and CD make a rst approximation, with perhaps a resulting eduction of 10% from the force and stress alues calculated according to Figure 1. Of ourse, it is an idealization to treat this boundry as a sharply defined surface at all: it should ather be a zone of considerable thickness in 'hich both plastic yield and plate accretion by ielt solidification occur.

We have so far taken the density distribution ı the plate to be the same at AC as at BD. Iowever, since at AC it is freshly made from ielt, it should be hotter and may still contain finite proportion of melt fluid (which is not ecessarily incompatible with rigid behavior), ıd its density should therefore be less. The orresponding modification of diagrams is shown . Figure 3 (in which, for simplicity, we revert › a vertical boundary AC, as in Figure 1, ıther than the somewhat more realistic sloping ne AC of Figure 2). No longer is CD parallel › AB, the length AC contracting thermally to D during the progress of the plate. We assume ıat the column height changes inversely as the ensity, on the grounds that the upper surface ›ols at an early stage, becoming most rigid and eventing contraction in area. The integration agram, Figure 3b, is very nearly similar to igure 2b, except that the line C′D′ is no longer ertical. The water-compensated overburden pressure at C is less than that at D by an amount $\langle g\rho_w\rangle H\langle\rho-\rho'\rangle/\langle\rho\rangle$, where ρ' is the mean density of a column of plate at AC and ρ is the mean density of such a column at BD. Thus a relatively trivial reduction is made in F_H, the more important reduction coming from the decrease of slope dp'/dz in A′C′ in proportion to the decrease in density from $\langle\rho\rangle$ to $\langle\rho'\rangle$. As a result we have, very nearly,

$$F_H = \langle g(\rho-\rho_w)\rangle Hh \left[1 - \frac{\langle\rho-\rho'\rangle}{2\langle\rho\rangle}\left(\frac{H}{h} + \frac{\langle\rho_w\rangle}{\langle\rho-\rho_w\rangle}\right)\right] \qquad (6)$$

If $\langle\rho-\rho'\rangle/\langle\rho\rangle = 1\%$ and $H/h = 20$, the first correction term makes a reduction of 10%, and the second is negligible (less than ¼%); the height of C above D is 20% less than h.

We now turn to the subject of isostasy; first, we consider it in relation to the more familiar example of an isostatic continent, since consideration of the horizontal forces is often omitted from presentations of this subject. Figure 4a shows a continent, of uniform density ρ_1, supported isostatically by a material of greater density ρ_2, covered by a layer of thickness w (5 km, say) of water of density ρ_w. Pressures are assumed everywhere equal to overburden pressures, constant across boundaries and constant on horizontal planes in the water and the medium of density ρ_2. It follows, for the parts of the continent that are not water immersed, that

$$(h + d)\rho_1 = d\rho_2 \qquad (7)$$

and therefore

$$(h + d)/h = \rho_2/(\rho_2 - \rho_1) \qquad (8)$$

where h is the height and d the depth of the upper and lower continental surfaces measured

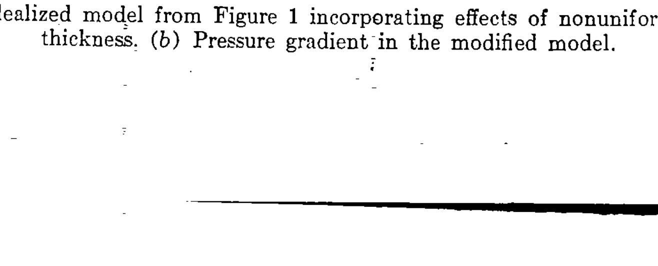

Fig. 3. (a) Idealized model from Figure 1 incorporating effects of nonuniform density and thickness. (b) Pressure gradient in the modified model.

from an effective flotation level $w\rho_w/\rho_2$ above the ocean floor, $(\rho_w/\rho_2)]$, say 3.3 km, below sea water-immersed margins of the have instead

$$(h' + d')(\rho_1 - \rho_w) = d'(\rho_2$$

h' and d' being measured from th The difference of (9) from (8) m correction that is of no further i ·us. These conditions being satis forces are balanced on every ve of the continental mass. Furthermo integrated mass of matter per uni every external point is the same, given that the vertical distribution is sufficiently slow in horizontal comparison with the total depth in is density variation (it should be co course, that the vertical scale of F in the previous figures, is greatly e it follows that g has the same value on an external surface at constant h sea level, after correction for the r the asphericity of the earth.

That is, there is no gravitation under 'free-air' reduction or the s static' reductions to uniform le attempt in their various ways to meas ture from this condition and succeed ment with each other, when the horiz ation of the vertical density distribut enough. The Bouguer reduction give anomalies in mountainous areas an anomalies in the ocean under the sam conditions.

As was done before, we calculate zontal force F_H on the part of the lying to the right of the vertical boun which passes through its highest poi the initial presumption of hydrostatic all its boundaries, thus deducing the force on boundaries AC and CD, w partures from this condition can occu the system is not entirely water imm is more convenient to use (2) here in (3). The integration diagram, Figure simple, being essentially the triangle straight lines A′C′ and E′C′ with slopes proportional to ρ_1 and ρ_2, respectively, small correction from the triangle B′T which the slope of B′D′ is proportional from an effective flotation level E, which lies $w\rho_w/\rho_2$ above the ocean floor, i.e., $w[1 - (\rho_w/\rho_2)]$, say 3.3 km, below sea level. For the water-immersed margins of the continent we have instead

$$(h' + d')(\rho_1 - \rho_w) = d'(\rho_2 - \rho_w) \qquad (9)$$

h' and d' being measured from the ocean floor. The difference of (9) from (8) makes a trivial correction that is of no further importance to us. These conditions being satisfied, vertical forces are balanced on every vertical column of the continental mass. Furthermore, since the integrated mass of matter per unit area below every external point is the same, then, if it is given that the vertical distribution of densities is sufficiently slow in horizontal variation in comparison with the total depth in which there is density variation (it should be considered, of course, that the vertical scale of Figure 4a, as in the previous figures, is greatly exaggerated), it follows that g has the same value everywhere on an external surface at constant height above sea level, after correction for the rotation and the asphericity of the earth.

That is, there is no gravitational anomaly under 'free-air' reduction or the several 'isostatic' reductions to uniform level, which attempt in their various ways to measure departure from this condition and succeed, in agreement with each other, when the horizontal variation of the vertical density distribution is low enough. The Bouguer reduction gives negative anomalies in mountainous areas and positive anomalies in the ocean under the same isostatic conditions.

As was done before, we calculate the horizontal force F_H on the part of the continent lying to the right of the vertical boundary AC, which passes through its highest point, under the initial presumption of hydrostatic stress at all its boundaries, thus deducing the resultant force on boundaries AC and CD, where departures from this condition can occur. Since the system is not entirely water immersed, it is more convenient to use (2) here instead of (3). The integration diagram, Figure 4b, is simple, being essentially the triangle between straight lines A'C' and E'C' with slopes dp'/dz proportional to ρ_1 and ρ_2, respectively, with a small correction from the triangle B'D'E', in which the slope of B'D' is proportional to ρ_w.

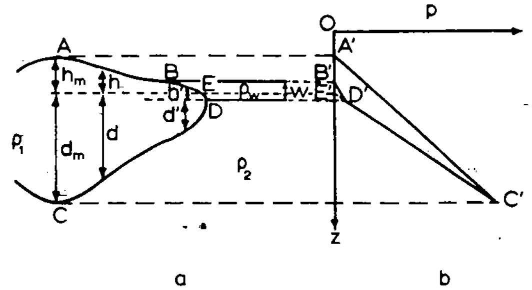

Fig. 4. (a) Model of the continent of uniform density in isostatic equilibrium. (b) Pressure gradient in the continent.

The result is

$$F_H = \tfrac{1}{2}g\rho_1 h_m(h_m + d_m) - \tfrac{1}{2}g\rho_w w^2[1 - (\rho_w/\rho_2)] \qquad (10)$$

$$= \tfrac{1}{2}g\rho_1 h_m^2 \rho_2/(\rho_2 - \rho_1) - \tfrac{1}{2}g\rho_w w^2[1 - (\rho_w/\rho_2)] \qquad (11)$$

where h_m and d_m are the greatest height and the greatest depth of the continent, respectively, above the effective flotation level, which is about 3.3 km below sea level.

With ρ_1 and ρ_2 as 2.8 and 3.1 Mg m^{-3}, respectively, the second term is about 1% and is now neglected. Thus $(h_m + d_m)/h_m = \rho_2/(\rho_2 - \rho_1)$ and is approximately 10. For mountain ranges rising 2, 3.5, and 5 km above sea level the corresponding maximum continental thicknesses are 55, 70, and 85 km, respectively. The horizontal spreading forces F_H are 3.9, 6.5, and 9.6 TN m^{-1}, sufficient to make a stress of 10^8 N m^{-2} = 1 kb on slabs of thickness 39, 65, and 96 km, respectively. If local peaks are disregarded, 2 km represents the height above sea level of the mountain ranges in most continents, 3.5–4 km may represent the Andes, and 5 km the Himalaya and Tibetan Plateau (continuing to the Pamirs at 4 km).

This force can be sustained by tensile stress across AC (of mean magnitude $\tfrac{1}{2}g\rho_1 h_m$, equal to 0.74, 0.95, and 1.16 kb in the three cases considered), by shear stress on the quasi-horizontal part of CD representing the lower surface of the continental mass, or by compressive stress at the edge of the continent. A fourth possibility, that the continent is not isostatic, is not acceptable except as a moderate correction. To annul the spreading force, the continental

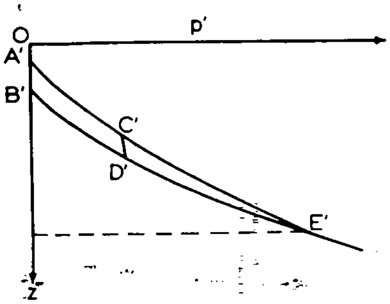

Fig. 5. Pressure gradient in the crustal plate required to produce zero gravity anomaly at depth.

roots would have to reach down to an additional depth $[d(d + h)]^{1/2}$ below the depth required for isostasy, another 52 km under a mountainous region rising to 2 km above sea level, and would produce a negative free-air gravitational anomaly of $2\pi G\rho_1[d(d + h)]^{1/2}$ (where G is the gravitational constant) amounting in the same case to 6.5×10^{-3} m sec^{-2} = 650 mgal (and a still larger negative Bouguer anomaly), which is far larger than anything observed. I prefer to reject the first of these possibilities, that the continent holds together by its own tensile strength, on the assumption that the crust of the earth is 'wet' at all depths below the water table, pore fluids being mobile, if there is enough time, at pressures generally close to the overburden pressure. It thus follows that, on the large scale and in the long term, little or no tensile deviation from lithostatic stress can persist. It is interesting to observe that the horizontal spreading pressure of the continents, except that arising from the Tibetan plateau, is just balanced by that of the oceanic plates if these plates are presumed to have thicknesses in the range 50–100 km, which are reasonable depths for cooling from the surface in a time of the order of 10^8 years to have produced rigidity. The plate thickness of 140 km required to balance the Tibetan plateau seems excessive, but here our simple two-dimensional analysis is probably at fault. We have no obligation to assume 'undertows' of mantle movement dragging the lower surfaces of oceanic plates and continents. Such drags, if present, should generally be oppositely directed for the oceanic plate and the adjacent continent. In the terms of the resultant force produced they can be smaller than or similar in magnitude to the forces we have calculated, but they can hardly be permitted to be larger by an order of magnitude without generating intolerably large stresses.

We come now to isostasy in connection with the oceanic ridges. For zero free-air anomaly with low horizontal variation of the vertical distribution of density, we require the integrated density below any point at the surface of the sea to be the same. When g is taken as a constant (a good approximation), this situation implies the equality of the integral (1) that we used to calculate pressure, which must now be evaluated down to some 'compensation depth' below which truly horizontal stratification of density is presumed to exist. Thus the requirement for zero free-air anomaly is that the pressure-versus-depth lines A′C′ and B′D′ in, e.g., Figures 1*b*, 2*b*, and 3*b*, continued to greater depth ultimately merge, as is shown in Figure 5. In Figure 3*b*, where we presumed a mean density 1% smaller at AC than at BD, the two curves had already approached 20% closer at a depth of about 80 km. Continuation of the same density difference to a greater depth would cause them to come together (and cross) at a depth of 400 km, but continuation of the same fractional density difference, with increasing density, would cause them to come together at a lesser depth. Allowance of 1% may also be ungenerous for the density difference. With a 5% density difference, doubtless an overgenerous one, we produce the situation shown in Figure 6, with a horizontal plate bottom that can itself be taken as a level of compensation. The essential point, however, is as revealed in Figure 5, that the ultimate merging of lines A′C′ and B′D′ to produce zero gravity anomaly (or their crossing to reverse its sign) in no way conflicts with having a finite area A′C′D′B′ that measures the downhill sliding force. In simplest terms, if we see a broad hill with no free-air gravitational anomaly we are entitled to infer that there is relatively light matter beneath it: but this fact should make no change in the direction in which a landslide will go (namely, downhill) nor any appreciable change in its force. And sea level is an adequate reference surface with which to define downhill; if g increases by 0.05 m sec^{-2} (=500 mgal) from A to B as it does from the equator to the poles ('anomalies' are smaller by an order of magni-

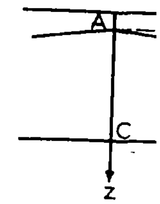

Fig. 6. (*a*) Idealized m

tude), two level surfaces separ
at A are 25 meters closer togeth

Figure 6 represents an impro
A more likely situation is one so
to Figure 5. In this case the mat
plate has its own horizontal sprea
measured by the area C′D′E′, w
similar order of magnitude as A′
measures the force on the plate,
or lesser. There are various ways
force can be sustained. The great
volume on which it acts may b
move at speeds comparable to
plate sliding. If some of it sprea
influence of this force, the larger f
resisting it may be on its lower
likelihood, only a small part of this
becomes transmitted to the plate.
CD (as in Figures 1, 2, and 3)
zone of relatively high plasticity;
CD are to a large extent decoupl
tions above.

An objection can be made to the
slope of ocean floor represented in
2*a*, and 3*a*, but this objection is
moved. The required form of the
maintenance of a steady state is
each point follows the line in its
general rigid surface permitting su
a surface of revolution, and an acc

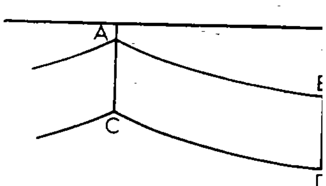

Fig. 7. Model of a plate similar
1*a*, 2*a*, and 3*a*, modified with curved
lower boundaries.

lated, but they can hardly
arger by an order of mag-
ierating intolerably large

isostasy in connection with
For zero free-air anomaly
variation of the vertical
sity, we require the inte-
y any point at the surface
same. When g is taken as a
proximation), this situation
of the integral (1) that we
essure, which must now be
some 'compensation depth'
horizontal stratification of
to exist. Thus the require-
ir anomaly is that the pres-
ies A′C′ and B′D′ in, e.g.,
l 3b, continued to greater
rge, as is shown in Figure 5.
we presumed a mean den-
AC than at BD, the two
approached 20% closer at a
n. Continuation of the same
a greater depth would cause
her (and cross) at a depth
tinuation of the same frac-
rence, with increasing den-
hem to come together at a
nce of 1% may also be un-
nsity difference. With a 5%
doubtless an overgenerous
e situation shown in Figure
plate bottom that can itself
of compensation. The essen-
is as revealed in Figure 5,
merging of lines A′C′ and
o gravity anomaly (or their
its sign) in no way conflicts
e area A′C′D′B′ that meas-
sliding force. In simplest
broad hill with no free-air
aly we are entitled to infer
ely light matter beneath it:
ld make no change in the
a landslide will go (namely,
appreciable change in its
el is an adequate reference
to define downhill; if g in-
sec^{-2} (=500 mgal) from A
om the equator to the poles
ialler by an order of magni-

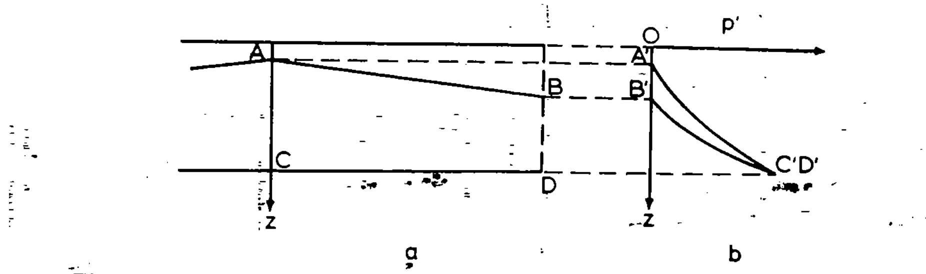

Fig. 6. (a) Idealized model of the compensated crustal plate with a horizontal base. (b) Pressure gradient.

tude), two level surfaces separated by 5 km at A are 25 meters closer together at B.

Figure 6 represents an improbable extreme. A more likely situation is one somewhat similar to Figure 5. In this case the material below the plate has its own horizontal spreading force F_H, measured by the area C′D′E′, which can be of similar order of magnitude as A′C′D′B′, which measures the force on the plate, either greater or lesser. There are various ways in which this force can be sustained. The greater part of the volume on which it acts may be too rigid to move at speeds comparable to those of the plate sliding. If some of it spreads under the influence of this force, the larger frictional drag resisting it may be on its lower boundary. In likelihood, only a small part of this force, if any, becomes transmitted to the plate. The boundary CD (as in Figures 1, 2, and 3) represents a zone of relatively high plasticity; motions below CD are to a large extent decoupled from motions above.

An objection can be made to the flat uniform slope of ocean floor represented in Figures 1a, 2a, and 3a, but this objection is readily removed. The required form of the line AB for maintenance of a steady state is one in which each point follows the line in its motion. The general rigid surface permitting such motion is a surface of revolution, and an acceptable form for the flat earth model, providing a ridge rising more steeply to the enter, is a cylinder with an elevated axis about which it rotates, as shown in Figure 7, which is a revised version of Figure 1a. No change at all is required in Figure 1b.

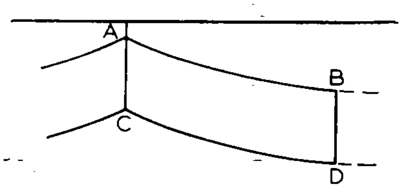

Fig. 7. Model of a plate similar to Figures 1a, 2a, and 3a, modified with curved upper and lower boundaries.

The corresponding modified form for the plate surface on a spherical earth is adequately represented by an ellipsoid having revolutional symmetry about its own axis of rotation, which is offset from the center of the earth (Figure 8). To level out to the horizontal at a depth δ below the ridge at an angular distance θ from it, the axis must be offset from the center of the earth by

$$p = \delta/(1 - \cos \theta) \qquad (12)$$

and the radius R' must exceed that of the earth taken to the ridge R by

$$R' - R = p - \delta \qquad (13)$$

these distances being 30 and 26 km, respectively, for $\delta = 4$ km and $\theta = 30°$.

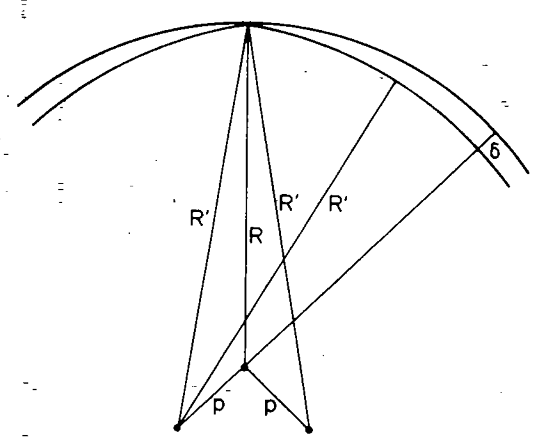

Fig. 8. Geometric model of the plate surface on a spherical earth.

In the conventional theory of the motion of spherical plates on a spherical earth, the velocity at a point $\mathbf{r}$ of the ith plate, $\mathbf{r}$ being the position vector from the center of the earth, is

$$\mathbf{v}_i = \boldsymbol{\omega}_i \times \mathbf{r} \qquad (14)$$

where $\boldsymbol{\omega}_i$ is the rotation vector for the ith plate, and the relative motion at a point where two plates i and j are adjacent is

$$\mathbf{v}_{ij} = \mathbf{v}_j - \mathbf{v}_i = \boldsymbol{\omega}_{ij} \times \mathbf{r} \qquad (15)$$

where $\boldsymbol{\omega}_{ij} = \boldsymbol{\omega}_j - \boldsymbol{\omega}_i$. (It is, in fact, only the relative velocities $\mathbf{v}_{ij}$ and relative rotation vectors $\boldsymbol{\omega}_{ij}$ that can be determined.) The modification made when the rotation vector $\boldsymbol{\omega}_i$ is offset from the center of the earth by $\mathbf{p}_i$ is to make

$$\mathbf{v}_i = \boldsymbol{\omega}_i \times (\mathbf{r} - p_i) = \boldsymbol{\omega}_i \times \mathbf{r} - \boldsymbol{\omega}_i \times \mathbf{p}_i \qquad (16)$$

$$\mathbf{v}_{ij} = \boldsymbol{\omega}_{ij} \times \mathbf{r} + \boldsymbol{\omega}_i \times \mathbf{p}_i - \boldsymbol{\omega}_j \times \mathbf{p}_j \qquad (17)$$

thus adding to the result (15) a small correction term independent of $\mathbf{r}$.

Having thus satisfied the kinematic requirements for motion on a spherical earth, we can take over the flat-earth calculations of forces unchanged. Lateral dimensions did not enter into these calculations except by the requirement of minimal variation in these dimensions. All integrations were taken with respect to z, which can simply be interpreted as the direction in which gravitational potential ϕ decreases (at rate g), replacing the elementary interval of integration gdz by $d\phi$. Horizontal forces are then reinterpreted to be torques about the rotation axis.

The glaciological analogy for plate tectonics has thus been successfully defended. Moreover, glaciology can be regarded as that branch of tectonics in which, because the motions are most rapid, the basic principles are most vividly illustrated. The essential feature of both plate tectonics and glaciology is thermal mobilization of matter via a fluid phase to produce a state of high gravitational potential energy and the action of the consequent mechanical processes to reduce this energy, thus maintaining an approximately steady state of motion.

Some Shock Effects in Piledriver Site

I. Y. Borg

Lawrence Livermore Labora
Livermore, California 94550

Postshot exploration
granodiorite indicates t
which extends 277 mete
and is asymmetric. The
point rocks have been s
at distances to the shot
diorite Hugoniot elastic
correlated with measure
temperatures of 30°C, a
quartz, planar lamellas a
grains subjected to a pr
⟨110⟩ is evident in rock t
kinking in biotite is asso
contains kink bands. At
noted in either the orth
there was a noticeable lo
in distant fractures withi
No diaplectic glass was
hydrous phases, biotite a
permeation of hot gases a

The Piledriver event, a nucle
granodiorite at the Nevada Test
an opportunity to study shock
occurred adjacent to a highly
tunnel drift system and was ac
an extensive re-entry and postsh
gram. Close-in stress gages, inst
uring particle velocities and acc
in situ cameras recording displa
ated satisfactorily at the time
propagation and provided an u
plete set of data, which was u
preshot predictions. Under these
the specific behavior of the rock
the explosion (vaporization, melti
and fracture) can be associated w
pressures with a minimum of unce

Postshot exploration resulted in
of three 3-inch cores, one of whic
the lower portions of the cavity
the explosion. This paper summari