On the Forces Driving Plate Tectonics: Inferences from Absolute Plate Velocities and Intraplate Stress

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Summary

Absolute plate motions and intraplate stress both serve as tests of models for the forces acting on plate boundaries. Plate velocities relative to a presumably fixed underlying mantle are calculated from the hypothesis that no net torque is exerted on the lithosphere. Intraplate stress is calculated by solving the equilibrium equations for thin elastic shells in the membrane state of stress. The absolute velocity fields predicted from a wide assortment of physical and geological models are all very similar. While the global pattern of absolute velocity is probably close to those predicted by such models, the absolute motions do not therefore provide a strong test of the driving mechanism. Comparison of predicted intraplate stress with the long-wavelength features of the global stress field, however, as determined by in situ measurements, earthquake mechanisms, and stress-induced geological structures, does prove to be a powerful test of possible driving forces.

All absolute velocity models have several interesting properties. Lithosphere in the equatorial half of the Earth is moving significantly faster than lithosphere in the polar half. Some connection to the Earth's rotation is implied since this statement is demonstrably untrue for co-ordinate poles much different from the geographic pole. Subducted slabs are characterized by a slow horizontal translation perpendicular to strike that is independent of the plate convergence rate, confounding attempts to explain the dip angles of Benioff zones in terms of a uniform vertical sinking and a variable absolute velocity for the overthrust plate. Ridges must migrate at a wide range of velocities relative to their underlying source of new lithosphere; such rapid migration may be a necessary but is not a sufficient condition for ridge jumps.

Force models considered in velocity and stress calculations include driving forces at spreading centres and subduction zones and various parameterizations of drag at the base of the lithosphere. From the rms absolute velocities of individual plates, there is a weak indication that pull by subducted lithosphere at trenches is an important driving force and that drag may be greater beneath continental than oceanic lithosphere. The predicted intraplate deviatoric stress cannot match the well-determined stress fields in North America and Europe unless the driving force exerted at ridges is at least comparable in magnitude to other forces in the system. The mid-plate stresses are very sensitive to the nature of drag at the base of the lithosphere and thus measured stresses may ultimately provide a sensitive test of absolute plate velocities.
Introduction

Rigidity of lithospheric plates is one of the fundamental tenets of plate tectonics. From the postulate of rigid plates and Euler's theorem on the motion of a rigid body with one point fixed has come the description of relative displacements between plates in terms of simple rotations (Bullard, Everett & Smith 1965). Extension of this concept to that of the relative angular velocity of one plate to another (McKenzie & Parker 1967; Morgan 1968; LePichon 1968) has completed the kinematic description of plate tectonics and has given the theory its elegant predictive power.

The principles of rigid body mechanics may further be applied to examine the nature of the forces driving plate motions. The lithosphere as a whole is in mechanical equilibrium. By conservation of angular momentum, the torques exerted on the lithosphere by buoyancy forces at plate boundaries and by drag forces beneath plates must sum to zero. The same equilibrium condition must hold for each individual plate.

In this paper, we investigate two consequences of this simple statement that the lithosphere is in a state of equilibrium. If some fraction of the torque exerted on the lithosphere is produced by forces which depend on the velocity of the plates with respect to the underlying mantle, then the condition that no net torque be exerted on the lithosphere provides a framework for determining the 'absolute' velocities of the plates (Solomon & Sleep 1974). If the complete set of forces operating on the lithosphere is specified, then the equilibrium equations for displacements in an elastic lithosphere may be solved to obtain the intraplate stress.

The absolute plate velocities do not provide a particularly critical test of possible driving force models, for two reasons: (1) there are no measurements of absolute velocity independent of some geological or physical assumption; and (2) as we shall see, the absolute velocity fields predicted for a wide assortment of physical models are all very similar. To the extent that such models for absolute plate velocity have any validity, and there are weak grounds for believing that they have some, the absolute velocities of individual plates and of plate boundaries may contain information on the driving forces. We examine a number of such possible relationships below.

There does exist independent information on the state of stress in the lithosphere, both from the mechanisms of midplate earthquakes and from in situ stress measurements. Deviatoric stress in the lithosphere can be produced, however, from a number of different processes. If that portion of the stress field attributable to the plate tectonic driving forces now acting on the plates can be isolated, then the observed intraplate stresses can be compared with those calculated for a given force model. Because the predicted stress is much more sensitive to the details of the assumed forces than are the absolute velocities, such a comparison can serve as a powerful test of possible driving force models.

Possible driving forces

The forces available to drive and/or resist plate motions include buoyancy forces at ridges and subduction zones, viscous drag on the base of plates, and assorted resistance forces at transform faults, adjacent to sinking slabs, and possibly in the upwelling zone at spreading centres. The relative importance of these various forces and even the sign of some of them have been the subject of much debate.

The descending slabs of cold lithosphere at subduction zones are likely an important element of the driving mechanism (Elsasser 1969; Jacoby 1970). That these slabs exert a pull on the plates is suggested by the predominantly normal-faulting earthquake mechanisms, with least compressive stress aligned down-dip, at intermediate depths in most slabs (Isacks & Molnar 1971) and the trenchward component of absolute velocity for both plates adjacent to subduction zones, according to many current
models of the absolute motions of the plates (Morgan 1972a). Downgoing slabs are energetically capable of producing observed plate velocities (Richter 1973a), including those of plates not themselves being subducted.

Buoyancy forces at ridges can also drive plate motions (Hales 1969; Jacoby 1970; Morgan 1972a; Artyushkov 1973). A mechanism by which viscous losses at the ridge resisting plate separation may dominate the forces driving the separation, however, has also been proposed (Lachenbruch 1973).

Because the asthenosphere is viscous, drag at the base of the plates is an important force to consider, yet whether this drag acts to drive or to resist the plates is by no means completely settled. There are numerical problems if the descending slab is the only driving force for plate motions and linear viscous dissipation in a passive asthenosphere is the only resisting force (McKenzie 1969a) but that does not exclude either mechanism from contributing to the sum total of significant forces. Drag due to classical Rayleigh–Benard convection in the upper mantle is not an adequate driving force (Richter 1973a). In fact, transverse convection (Rayleigh–Benard) cells are unstable to shear at their top and would be converted to longitudinal rolls (Richter 1973b), which would not contribute a net driving force.

The magnitude of the drag force may vary from region to region. Drag may be much higher beneath continents than beneath oceans, either because of the greater viscosity of the sub-continental asthenosphere (Knopoff 1972) or because continental lithosphere is substantially thicker than oceanic lithosphere (Alexander & Sherburne 1972; Sacks & Okada 1974). There may also be a large drag force resisting the horizontal translation of sinking slabs (Talwani 1969; Tullis 1972). Finally, drag resistance may be greater on the lithosphere beneath hot spots because of depletion of the melt fraction in the asthenosphere (Lachenbruch 1973), a suggestion that bears at least a geometrical similarity to the gravitational anchor model (Shaw & Jackson 1973) for hot spots.

**Absolute plate velocities**

In this section we apply the principle of mechanical equilibrium to the problem of determining the absolute velocities of the plates. The discussion below extends our earlier work (Solomon & Sleep 1974), by including a number of new models and by examining more thoroughly the consequences of the absolute velocities obtained.

**Assumptions**

There are three major assumptions that make calculation of absolute plate motions a simple linear problem. We assume that (1) the configuration and relative motions of the Earth's plates are known; (2) no net torque is exerted on the lithosphere; and (3) asthenospheric drag resists plate motion according to a simple viscous law, so that drag on the bottom of a plate is linearly proportional to the absolute velocity \( u \) of the plate and is exerted in a direction opposite to that of \( v \). These assumptions were discussed by Solomon & Sleep (1974), but the first two merit further amplification.

The boundaries of the plates included in this study are shown in Fig. 1. Two alternative sets of relative velocity vectors are used. The preferred set (Table 1) is based on the RM1 velocities of Minster et al. (1974), with some modifications. We include a Philippine plate, and use the Eurasian–Philippine angular velocity vector \( (9 \times 10^{-7} \text{ deg/yr about 54^\circ N, 154^\circ W}) \) of Fitch (1972). We also include a Caribbean plate and use the American–Caribbean angular velocity vector \( (1.0 \times 10^{-7} \text{ deg/yr about 28^\circ S, 69^\circ W}) \) given in Solomon & Sleep (1974), except that the vector magnitude is given incorrectly in that paper. This American–Caribbean pole gives a good fit to the shape of the northern boundary of the Caribbean plate interpreted as a trans-
Fig. 1. Outline of the plates, continental areas (shaded borders), and subduction zones (heavy dashed lines) adopted for absolute velocity models. Cylindrical equidistant projection.
On the forces driving plate tectonics

Table 1

Adopted values for relative plate motions

<table>
<thead>
<tr>
<th>Plate</th>
<th>Area*</th>
<th>Continental Area*</th>
<th>Relative angular velocity† 10⁻⁷ deg/yr</th>
<th>α_x</th>
<th>α_y</th>
<th>α_z</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAC, Pacific</td>
<td>2.664</td>
<td>0.046</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
<td>0.0</td>
</tr>
<tr>
<td>AME, American</td>
<td>2.486</td>
<td>1.494</td>
<td>1.364</td>
<td>-3.611</td>
<td>5.687</td>
<td></td>
</tr>
<tr>
<td>AFR, African</td>
<td>1.931</td>
<td>0.872</td>
<td>2.406</td>
<td>-4.112</td>
<td>8.973</td>
<td></td>
</tr>
<tr>
<td>EUA, Eurasian</td>
<td>1.675</td>
<td>1.463</td>
<td>1.313</td>
<td>-3.569</td>
<td>8.267</td>
<td></td>
</tr>
<tr>
<td>IND, Indian</td>
<td>1.503</td>
<td>0.531</td>
<td>6.334</td>
<td>-0.221</td>
<td>10.890</td>
<td></td>
</tr>
<tr>
<td>ANT, Antarctic</td>
<td>1.477</td>
<td>0.442</td>
<td>0.675</td>
<td>-3.680</td>
<td>9.596</td>
<td></td>
</tr>
<tr>
<td>NAZ, Nazca</td>
<td>0.405</td>
<td>0.693</td>
<td>0.693</td>
<td>-9.001</td>
<td>13.692</td>
<td></td>
</tr>
<tr>
<td>PHL, Philippine</td>
<td>0.141</td>
<td>-5.836</td>
<td>0.750</td>
<td>-4.715</td>
<td>14.425</td>
<td></td>
</tr>
<tr>
<td>ARB, Arabian</td>
<td>0.121</td>
<td>0.104</td>
<td>4.709</td>
<td>-3.784</td>
<td>10.343</td>
<td></td>
</tr>
<tr>
<td>CAR, Caribbean</td>
<td>0.087</td>
<td>0.034</td>
<td>1.051</td>
<td>-2.794</td>
<td>6.332</td>
<td></td>
</tr>
<tr>
<td>COC, Cocos</td>
<td>0.076</td>
<td>0.0</td>
<td>-4.715</td>
<td>-14.425</td>
<td>13.332</td>
<td></td>
</tr>
</tbody>
</table>

* Earth’s radius equals 1.
† The listed angular velocity vectors are relative to PAC, using a right-handed rule for α. The Cartesian co-ordinates x, y and z are measured along the radius vectors through latitude-longitude pairs (0°, 0°), (0°, 90°), and (90°, 0°), respectively. North and east are positive.

form feature (Molnar & Sykes 1969). Morgan (1972b) gives a very similar pole. The rates of motion between the Caribbean plate and plates adjacent to it are quite uncertain (see a complete discussion in Holcombe et al. 1973); the American–Caribbean rate used here lies within values proposed and gives a left-lateral slip rate of 0.75 cm/yr along the northern Caribbean plate boundary.

We do not include either the Bering plate of Minster et al. (1974) or their division of the American plate into separate North American and South American plates. The Bering plate may be an artifact of systematic errors in the slip vectors of Aleutian earthquakes due to the laterally heterogeneous velocity structure in the subduction zone (Engdahl 1974). While Minster et al. (1974) were able to demonstrate that relative motion between North and South America is statistically resolvable, their North American–South American pole is inconsistent with the sense of relative motion along the northern and southern boundaries of the Caribbean plate. Specifically, a North American–Caribbean pole that fits the shape and left lateral transform sense of that boundary, taken together with the North American–South American pole of Minster et al., gives a Caribbean–South American pole predicting either dominantly left-lateral motion or dominantly thrust motion along the boundary between these two plates, in contradiction to geologic and seismic evidence (Molnar & Sykes 1969). Improved relative motion solutions in the future hold promise for resolving this dilemma (C. G. Chase, private communication 1974). In the meantime, we have opted for a Caribbean and a single American plate. The American–Pacific pole given in Table 1 is the mean of the North and South American–Pacific poles of Minster et al. (1974).

The second set of relative velocity vectors is that adopted by Solomon & Sleep (1974), with the corrected Caribbean–American rotation rate. These relative velocities are taken primarily from the tabulation of Chase (1972).

An alternative to assuming a set of relative velocity vectors and requiring that the lithosphere as a whole be in equilibrium is to balance the torques on individual plates and invert the relative velocities to obtain the magnitudes of various parameterized driving and resistive forces (Tullis & Chapple 1973; Forsyth & Uyeda 1975; Harper 1975). The latter approach has considerable appeal but at least one major drawback. The torques exerted on plates at plate boundaries are not single-valued functions of...
the boundary type and relative velocity, as evidenced by the fact that models which assume the contrary either do not precisely fit the observed relative velocities or require additional ad hoc assumptions. Some forces transmitted across plate boundaries are not readily parameterized but depend on the global state of stress. Normal forces across transform faults and tangential forces across subduction zones are examples of these.

An additional advantage to considering the entire lithosphere when evaluating absolute velocities is that torques exerted on adjacent plates at symmetric boundaries, such as ridges and transform faults, are equal in magnitude and opposite in sign so do not need to be included in the torque balance.

**Analysis**

In this section we present the equations for absolute plate velocities using the assumptions just described.

For each plate \( p \), let \( \omega_{po} \) be the angular velocity vector with respect to an arbitrary reference plate \( o \) (e.g. PAC in Table 1) and let \( \omega_{pm} \) be the ‘absolute’ angular velocity with respect to the underlying mantle \( m \). Since

\[
\omega_{pm} = \omega_{po} + \omega_{om},
\]

then \( \omega_{pm} \) is known for all \( p \) once \( \omega_{om} \), the absolute velocity of the reference plate, is determined. We may solve for \( \omega_{om} \) for a number of interesting physical models for the plate driving forces.

Consider first models in which the forces exerted at all plate edges are symmetric, i.e. at ridges, trenches and transform faults the torques on adjacent plates are equal in magnitude but opposite in sign. Then only drag at the bottom of the lithosphere need be considered explicitly in the torque balance. The relevant equations for \( \omega_{om} \) are (Solomon & Sleep 1974)

\[
\left( \sum_k D_k Q_k \right) \omega_{om} = -\sum_k D_k Q_k \omega_{ko}
\]

where \( D_k \) is a scalar drag coefficient and \( Q_k \) is a symmetric \( 3 \times 3 \) matrix given by

\[
Q_{ij} = \int_k (\delta_{ij} - r_i r_j) \, d\alpha
\]

where \( r_i \) and \( r_j \) are Cartesian co-ordinates of a point \( r \) on the Earth’s surface and \( \delta_{ij} \) is the Kronecker delta function. When the only drag force in a model is at the base of plates, the sums in (2) are taken over the population of plates and the integral in (3) is taken over plate area. Equation (2) is particularly simple when the drag coefficient is everywhere uniform:

\[
\omega_{om} = -\frac{3}{8\pi} \sum_p Q_p \omega_{po}.
\]

When drag beneath continental lithosphere is much greater than beneath oceanic lithosphere, the sums in (2) are taken over the population of continents and the integral in (3) is taken over continental area. If the drag is such as to resist motion of linear features, such as island arcs, rather than areal features, the sums in (2) are taken over the population of such linear elements and the integral in (3) is a line integral along each element. If the drag is such as to resist motion of point features, such as hot spots, then the sums in (2) are taken over the population of such point elements and the integral in (3) is identically unity. Equations (2) and (3) amount to minimizing the rms (root-mean-squared) absolute velocity of whatever tectonic elements are included in the force model.
Before solving equations (2) or (4), we need to evaluate the coefficients $Q_{ij}$, which for the case of drag at the base of the lithosphere are all integrals over the area of each plate or each continent. The plate boundaries in our calculations (Fig. 1) consist of 1833 points digitized from seismicity and bathymetric maps. Writing the area integrals as double integrals over co-latitude $\phi$ and longitude $\theta$, we evaluate the $\theta$ integral analytically and the $\phi$ integral numerically. Defining $S_{ij} = A_p \delta_{ij} - Q_{ij}$ where $A_p$ is the area of plate $p$, we have, for example,

$$S_{33} = \int \int \cos^2 \phi \sin \phi \, d\phi \, d\theta = \sum_i \cos^2 \phi_i \sin \phi_i \Delta \phi_i \theta_i$$

where $(\phi_i, \theta_i)$ are points on the plate boundary, $\Delta \phi_i = \phi_i - \phi_{i-1}$, and the line integral in (5) is performed clockwise about the plate. Because $\theta$ is discontinuous at 0 and $2\pi$, we must add a branch cut contribution to (5) for plates on the Greenwich meridian. A term

$$S'_{33} = \int_0^{2\pi} \int_0^\alpha \cos^2 \phi \sin \phi \, d\phi \, d\theta = \frac{2\pi}{3} (1 - \cos \alpha)$$

must be added to (5) when $\theta$ goes from $2\pi$ to 0 at $\phi = \alpha$ and must be subtracted from (5) when $\theta$ goes from 0 to $2\pi$ at $\phi = \alpha$. The matrices $S_p$ are given in Table 2.

The continental boundaries used in these calculations are also shown in Fig. 1. The boundaries enclose both land areas and continental shelves, and are specified by 3271 points obtained by digitizing the 1000-fm isobath on US Hydrographic Office bathymetric charts.

There are alternative methods for deriving equations (2) or (4). We could postulate that the absolute plate velocities are those which minimize the energy $E$ dissipated per unit time by drag at the base of the lithosphere. We may write

$$E = \sum_P \int_{\text{plate } p} D_p(r) \mathbf{\omega}_{pm} \cdot [r \times (\mathbf{\omega}_{pm} \times \mathbf{r})] \, dA$$

or equivalently,

$$E = \sum_P \int_{\text{plate } p} D_p(r) (\mathbf{\omega}_{pm} \times \mathbf{r})^2 \, dA.$$  

Table 2

<table>
<thead>
<tr>
<th>Plate</th>
<th>$S_{11}$</th>
<th>$S_{22}$</th>
<th>$S_{33}$</th>
<th>$S_{12}$</th>
<th>$S_{13}$</th>
<th>$S_{23}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>PAC</td>
<td>1.461</td>
<td>0.639</td>
<td>0.564</td>
<td>-0.399</td>
<td>-0.075</td>
<td>0.057</td>
</tr>
<tr>
<td>AME</td>
<td>0.558</td>
<td>0.880</td>
<td>1.048</td>
<td>-0.424</td>
<td>-0.211</td>
<td>-0.209</td>
</tr>
<tr>
<td>AFR</td>
<td>1.263</td>
<td>0.380</td>
<td>0.288</td>
<td>0.248</td>
<td>-0.059</td>
<td>-0.109</td>
</tr>
<tr>
<td>EUA</td>
<td>0.275</td>
<td>0.595</td>
<td>0.804</td>
<td>-0.097</td>
<td>0.161</td>
<td>0.413</td>
</tr>
<tr>
<td>IND</td>
<td>0.391</td>
<td>0.831</td>
<td>0.282</td>
<td>-0.159</td>
<td>0.238</td>
<td>-0.228</td>
</tr>
<tr>
<td>ANT</td>
<td>0.112</td>
<td>0.266</td>
<td>1.100</td>
<td>0.046</td>
<td>-0.062</td>
<td>-0.070</td>
</tr>
<tr>
<td>NAZ</td>
<td>0.012</td>
<td>0.334</td>
<td>0.059</td>
<td>0.013</td>
<td>0.003</td>
<td>0.116</td>
</tr>
<tr>
<td>PHL</td>
<td>0.061</td>
<td>0.063</td>
<td>0.016</td>
<td>-0.061</td>
<td>-0.029</td>
<td>0.029</td>
</tr>
<tr>
<td>ARB</td>
<td>0.047</td>
<td>0.053</td>
<td>0.021</td>
<td>0.049</td>
<td>0.030</td>
<td>0.031</td>
</tr>
<tr>
<td>CAR</td>
<td>0.007</td>
<td>0.075</td>
<td>0.005</td>
<td>-0.021</td>
<td>0.005</td>
<td>-0.019</td>
</tr>
<tr>
<td>COC</td>
<td>0.001</td>
<td>0.073</td>
<td>0.002</td>
<td>0.006</td>
<td>-0.001</td>
<td>-0.011</td>
</tr>
</tbody>
</table>

Earth’s radius equals 1.
If $D_p$ is uniform for all plates, then from (8) $E$ is proportional to the mean-squared absolute plate velocity. Using equation (1) and setting the derivatives of $E$ with respect to the components of $\omega_{om}$ equal to zero, one obtains equation (2).

Lliboutry (1974) has also derived equation (4) based on a simple model for convection in the Earth’s mantle.

In general, the forces exerted at plate edges will not be symmetric. In particular, the torque exerted on the subducting plate at subduction zones is likely to be greater than the torque exerted on the overthrust plate. In a force model with such an asymmetry at trenches, there will be some net torque $T$ exerted by the global set of subducting slabs that must be balanced by torque due to drag. For such a model, $\omega_{om}$ satisfies

$$\sum_p D_p Q_p \omega_{om} = T - \sum_p D_p Q_p \omega_{po} \tag{9}$$

instead of equation (2). The net torque $T$ will be given by

$$T = \sum_t \int_{\text{line}} C_t(r) \, dl \tag{10}$$

where the scalar $C_t$ is the force exerted per unit length of trench and where the line integral in (10) is taken counter-clockwise about the subducted plate. Torques associated with individual subduction zones are given by Solomon & Sleep (1974).

There are a family of solutions $\omega_{om}$ to (9) and (10) each of which follows from a different assumption about $D_p$ and $C_t$. To simplify matters, we only consider cases in which $D_p$ is equal to a constant, $D$, everywhere and $C_t$ is equal to a constant, $C$, everywhere. Then

$$\omega_{om} = \frac{3}{8\pi} \left( \frac{C}{D} \, L - \sum_p Q_p \omega_{po} \right) \tag{11}$$

where

$$L = \sum_t \int_{\text{line}} dl = \sum_t (r_{12} - r_{11}) \tag{12}$$

and where $r_{11}$ and $r_{12}$ are radius vectors to the two ends of trench $t$. Solomon & Sleep (1974) have discussed how to find an upper bound to $C/D$ by conservation of energy if forces at subduction zones and ridges act to drive plate motions and drag at the base of the lithosphere is the major resistive force. In such a case, the energy $U$ gained per unit time at subduction zones cannot exceed the energy $E$ dissipated per unit time by drag, where

$$U = \sum_t \omega_{tm} \cdot T_t = C \sum_t \omega_{tm} \cdot (r_{12} - r_{11}) \tag{13}$$

and from equation (7),

$$E = D \sum_p \omega_{pm} \cdot (Q_p \omega_{pm}). \tag{14}$$

Equating (13) and (14) and making the substitutions (1), (11) and (12) gives

$$\frac{C}{D} \left[ - \frac{3}{8\pi} L \cdot \sum_p Q_p \omega_{po} + \sum_t \omega_{om} \cdot (r_{12} - r_{11}) \right]$$

$$= \sum_p \omega_{po} \cdot (Q_p \omega_{po}) - \frac{3}{8\pi} \left( \sum_p Q_p \omega_{po} \right) \cdot \left( \sum_p Q_p \omega_{po} \right) \tag{15}$$
For the set of relative velocity vectors in Table 1 based on the results of Minster et al. (1974) and the set of 14 oceanic trenches in Fig. 1 (see also Table 6), equation (15) gives $C/D = 6.9$. For the alternative set of relative velocities based largely on the results of Chase (1972), equation (15) gives 7.2. (The latter figure was given incorrectly in Solomon & Sleep 1974.)

Absolute velocity models

Absolute plate velocities were calculated for the two alternative sets of relative velocity vectors and for a number of different assumptions about the forces acting on the plates. Descriptions of the models are given in Table 3. The absolute angular velocity of the Pacific plate for each model is given in Table 4. The absolute angular velocity models

<table>
<thead>
<tr>
<th>Model</th>
<th>Description</th>
<th>$v_{rms}$ cm/yr</th>
<th>$\Omega_x$</th>
<th>$\Omega_y$</th>
<th>$\Omega_z$</th>
</tr>
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<tbody>
<tr>
<td>A3</td>
<td>Uniform drag coefficient beneath all plates</td>
<td>4.79</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>B3</td>
<td>Drag beneath continents only</td>
<td>5.08</td>
<td>-0.31</td>
<td>1.00</td>
<td>-1.55</td>
</tr>
<tr>
<td>B4</td>
<td>Continents have 3 times more drag than oceans</td>
<td>4.85</td>
<td>-0.11</td>
<td>0.47</td>
<td>-0.65</td>
</tr>
<tr>
<td>C3</td>
<td>Drag opposing horizontal translation of slabs, oceanic subduction zones only</td>
<td>4.81</td>
<td>0.16</td>
<td>-0.20</td>
<td>0.35</td>
</tr>
<tr>
<td>C4</td>
<td>Same, but including ARBT and HIMT</td>
<td>4.80</td>
<td>0.14</td>
<td>0.02</td>
<td>0.10</td>
</tr>
<tr>
<td>D1</td>
<td>Maximum pull by slabs plus plate drag</td>
<td>4.91</td>
<td>0.15</td>
<td>1.15</td>
<td>-0.08</td>
</tr>
<tr>
<td>E2</td>
<td>Drag beneath 8 mid-plate hot spots of Morgan (1971)</td>
<td>4.86</td>
<td>0.73</td>
<td>0.05</td>
<td>-0.53</td>
</tr>
<tr>
<td>E3</td>
<td>Drag beneath 19 hot spots of Morgan (1971)</td>
<td>4.93</td>
<td>0.20</td>
<td>0.68</td>
<td>-1.04</td>
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</table>

Group 2: Models with relative velocities based on S005ALL of Chase (1972)

<table>
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<tr>
<th>Model</th>
<th>Description</th>
<th>$v_{rms}$ cm/yr</th>
<th>$\Omega_x$</th>
<th>$\Omega_y$</th>
<th>$\Omega_z$</th>
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<td>A0</td>
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<tr>
<td>B0</td>
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<td>-0.21</td>
<td>1.17</td>
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<tr>
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<td>-0.07</td>
<td>0.54</td>
<td>-0.68</td>
</tr>
<tr>
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<td>Same as C3</td>
<td>4.94</td>
<td>0.37</td>
<td>0.08</td>
<td>0.38</td>
</tr>
<tr>
<td>C2</td>
<td>Same as C4</td>
<td>4.94</td>
<td>0.49</td>
<td>0.18</td>
<td>0.14</td>
</tr>
<tr>
<td>D0</td>
<td>Same as D1</td>
<td>5.04</td>
<td>0.16</td>
<td>1.20</td>
<td>-0.08</td>
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<td>E0</td>
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<td>5.03</td>
<td>1.01</td>
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<td>-0.45</td>
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<tr>
<td>E1</td>
<td>Same as E3</td>
<td>5.09</td>
<td>0.21</td>
<td>0.93</td>
<td>-1.13</td>
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Table 4

Absolute velocity of the Pacific plate

<table>
<thead>
<tr>
<th>Model Group</th>
<th>Latitude, deg</th>
<th>Longitude, deg</th>
<th>$\omega_x$ $10^{-7}$ deg/yr</th>
<th>$\omega_y$</th>
<th>$\omega_z$</th>
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<tr>
<td>A3</td>
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<td>121.7</td>
<td>7.08</td>
<td>-1.490</td>
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<td>B3</td>
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<td>117.8</td>
<td>8.92</td>
<td>-1.800</td>
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<tr>
<td>B4</td>
<td>-65.2</td>
<td>119.2</td>
<td>7.86</td>
<td>-1.605</td>
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<tr>
<td>C3</td>
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<td>121.1</td>
<td>6.67</td>
<td>-1.333</td>
<td>2.211</td>
</tr>
<tr>
<td>C4</td>
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<td>119.1</td>
<td>6.97</td>
<td>-1.353</td>
<td>2.427</td>
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<td>D1</td>
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<td>110.6</td>
<td>7.59</td>
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<tr>
<td>E2</td>
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Group 2

<table>
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<tr>
<th>Model Group</th>
<th>Latitude, deg</th>
<th>Longitude, deg</th>
<th>$\omega_x$ $10^{-7}$ deg/yr</th>
<th>$\omega_y$</th>
<th>$\omega_z$</th>
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<td>A0</td>
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<td>B2</td>
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<td>D0</td>
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<td>105.8</td>
<td>8.57</td>
<td>-1.073</td>
<td>3.779</td>
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</table>

Others

AM1, Minster et al. (1974)

Morgan (1972b)

Lliboutry (1974)

S, Kaula (1975)

![A3. UNIFORM DRAG](https://academic.oup.com/gji/article-abstract/42/2/769/652631)

Fig. 2. Absolute velocities of the plates for force model A3. Plate boundaries as in Fig. 1. Arrows are proportional to the linear velocity of the plate at the tail of the arrow; a length equal to the distance between tick marks on either the horizontal or vertical axis corresponds to a velocity of 20 cm/yr. Velocities are generally shown at 30° intervals of latitude and longitude and at several proposed hot spots (dots): Hawaii, Macdonald, Easter, Yellowstone, Iceland, Tristan, Reunion (Morgan 1971). Cylindrical equidistant projection.
velocity of any other plate may then be found from Tables 1 and 4 by vector addition. The absolute linear velocities for one of these models are illustrated in Fig. 2.

The absolute velocity of the Pacific plate predicted by several techniques independent of those used here is also given in Table 4. The models of Morgan (1972b) and Minster et al. (1974) are based on the premise that a global set of fixed hotspots provides a reference frame for absolute plate motions. The Pacific poles shown are for models which provide a best fit to the azimuths of the linear volcanic features associated with such hot spots. The model of Lliboutry (1974) is similar to model A0 except that three of the smaller plates with more poorly-defined relative velocity vectors (PHL, ARB, CAR) are incorporated into larger plates; Solomon & Sleep (1974) gave an identical model (A2). The model of Kaula (1975) is one for which the rms transverse velocities of spreading centres and overthrust plate boundaries is a minimum.

The main conclusion to be drawn from Tables 3 and 4 is that for a wide variety of physical or geological assumptions from which the absolute plate velocities may be calculated, the predicted velocity field is constant to within an addition of a uniform angular velocity vector of magnitude 1 to 2 $\times 10^{-7}$ deg/yr. This statement is true for both sets of relative velocity vectors assumed and probably holds for more precise relative velocity solutions likely to be produced in the future. Is there any independent check on the validity of the class of absolute velocity models represented in Tables 3 and 4? Can we distinguish among the models?

The best check, though one that is incomplete, comes from palaeomagnetism. Models with a net lithosphere rotation (Table 2) about a pole different from the north pole will show an apparent 'wander' of the palaeomagnetic pole with respect to the geographic pole at successively greater times in the geologic past. This wander is resolvable from plate motions if the Earth's field has always been, on the average, an axial geocentric dipole. Both McElhinny (1973) and Jurdy & Van der Voo (1974) have shown that the wander amounts to no more than approximately 2' in the last 50 My, or an average of 0.4 $\times 10^{-7}$ deg/yr about a rotation axis in the Earth's equatorial plane. This upper bound on the average angular velocity is raised to 1.1 $\times 10^{-7}$ deg/yr if the uncertainty in the early Tertiary palaeomagnetic pole (Jurdy & Van der Voo 1974) is included. If the assumption is made that the net lithosphere rotation $\Omega$ for an instantaneous absolute velocity model must also obey

$$\left(\Omega_x^2 + \Omega_y^2\right)^{1/2} \leq 1.1 \times 10^{-7} \text{ deg/yr},$$

then several models in Table 3 (B0, D0, D1) can be rejected as inconsistent, though only marginally so, with palaeomagnetic evidence. Palaeomagnetic data can never be used to choose a unique model for absolute velocities since $\Omega_z$ is unconstrained by (16).

A second check on the validity of absolute velocity models is provided by apparent north-south shifts with time of the zone of maximum sediment deposition in the equatorial oceans. Such evidence has been used to show the progressive northward migration of the Pacific plate for the last 40 My (Winterer 1973; Clague & Jarrard 1973; van Andel 1974), in approximate agreement with palaeomagnetic results and with the PAC motions predicted in Table 4. As with palaeomagnetism, however, mapping the palaeoequator from sedimentation rates leaves $\Omega_z$ unconstrained.

It has been argued (Jordan & Minster 1974) that the strikes of linear volcanic features associated with hot spots can be used to decide among absolute motion models such as those in Tables 3 and 4. Such a criterion can be applied only if hot spots are fixed in the mantle. The hypothesis that a number of such hot spots are fixed with respect to each other cannot be rejected when plate motions over only the last 10 My are considered (Minster et al. 1974). When time periods of 50 My are considered, however, hot spots appear to show relative velocities averaging 1 to 2 cm/yr (Burke, Kidd & Wilson 1973; Molnar & Atwater 1973) and the hot spot frame may rotate with respect to the Earth's dipole magnetic field at a comparable rate (Mc-
Elhininy 1973). Thus the motivation for using hot spot traces to reject one or more of the models in Tables 3 and 4 is all but eliminated. Because the hot spot relative velocities are among the smallest in the entire kinematic picture of plate tectonics, however, it is probably reasonable to conclude that the true absolute velocities are not too different from the general class of models in Tables 3 and 4.

**Rms velocities of the plates**

Absolute velocity models such as those presented have a number of interesting consequences which we now consider in detail. The rms linear velocities of individual plates and of the lithosphere as a whole are of interest because of their relationship to the work done overcoming drag (equation (8)) and because of several proposed relationships between some form of average plate velocity and other plate parameters. The rms velocity

\[ v_{\text{rms}} = \left[ \frac{1}{A_p} \int_{V \text{plate}} v(r) \cdot v(r) dA \right]^{\frac{1}{2}} \]  

for each plate is given in Table 5 for the Group 1 absolute velocity models. The rms velocity of the entire lithosphere is given in Table 3 for all models.

Not surprisingly, \( v_{\text{rms}} \) for the lithosphere as a whole does not vary much from 5 cm/yr for the models considered. Models based on the relative velocities of Minster *et al.* (1974) have a slightly lower \( v_{\text{rms}} \) than do those based on Chase's (1972) velocities.

The variation in \( v_{\text{rms}} \) from one plate to another is much larger, from less than 1 to over 10 cm/yr (Table 5). The rms velocity of any given plate, however, varies by only 1–2 cm/yr over the suite of absolute motion models. Since many of the models were derived by assuming a driving or resisting force with a particular dependence on the plate area or on the length of certain plate boundaries, it is worth examining whether the resulting rms velocities also imply a similar dependence.

Two such possible relationships are explored in Fig. 3. The rms plate velocities from model A3 are plotted versus plate area. There is no obvious connection between these two parameters; the correlation coefficient is -0.29, but the 95 per cent confidence limits (-0.74 to 0.37) include zero. The rms plate velocities from model B3 are also plotted in Fig. 3 versus the continental area on each plate (see Table 1). Here a weak negative correlation is evident. The correlation coefficient is -0.66 and differs slightly from zero at the 95 per cent confidence level (limits are -0.89 to -0.08). Such an inverse relationship between plate velocity and continental area was noticed by Minster *et al.* (1974) for the AM1 model.

**Table 5**

<table>
<thead>
<tr>
<th>Model</th>
<th>A3</th>
<th>B3</th>
<th>B4</th>
<th>C3</th>
<th>C4</th>
<th>D1</th>
<th>E2</th>
<th>E3</th>
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<td>7.91</td>
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<td>6.66</td>
<td>7.59</td>
<td>7.34</td>
<td>8.05</td>
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<td>4.08</td>
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<td>0.99</td>
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<td>7.98</td>
<td>8.01</td>
<td>7.11</td>
<td>7.22</td>
</tr>
</tbody>
</table>
Two additional possible relationships are investigated in Fig. 4. In one, the rms plate velocities from model C3 are plotted versus the length of overthrust subduction-zone boundary along the plate perimeter. There is a weak negative correlation (coefficient = -0.51) which is not, however, significant at the 95 per cent level of confidence (limits are -0.84 to 0.12). Also in Fig. 4 is a graph of rms plate velocities from model D1 versus the length of subducting plate boundary along the plate perimeter. The correlation coefficient (0.73) is the highest for any of the relationships considered, and differs significantly from zero at 95 per cent confidence (limits are 0.21 to 0.92). The correlation coefficient is increased only negligibly if the length of the continent–continent collision zones ARBT and HIMT are included. A positive correlation between plate velocity and length of subducting plate boundary has also been noted by Jordan & Minster (1974) and by Forsyth & Uyeda (1975).

Thus even though the relative angular velocities of the plates are not quantitatively predicted by any of the force models we have considered, the internal consistency between rms velocity of the plates and the assumed force model is higher if subducting slabs provide a major driving force and drag beneath continents is a major resisting force than if other assumptions are made. That sinking slabs provide the major driving force for the fast-moving (oceanic) plates has been concluded from a direct inversion of relative angular velocities by Tullis & Chapple (1973) and Forsyth & Uyeda (1975).

Latitudinal dependence of plate motions
The suggestion has often been made that plate velocities show a preferred orientation with respect to the Earth's rotation axis. To test this idea, we determined the rms velocity and the mean value of longitudinal velocity $v_\theta$ within latitudinal bands.

Fig. 3. Rms absolute plate velocity $v_{\text{rms}}$ the area of each plate subjected to drag forces. Open symbols are $v_{\text{rms}}$ from model A3 vs the total area of each plate. Filled symbols are $v_{\text{rms}}$ from model B3 vs the continental area of each plate.
Fig. 4. Rms absolute plate velocity vs the total length of subduction zones subject to driving or resisting forces on each plate perimeter. Open symbols are $v_{\text{rms}}$ from model C3 vs the total length of overthrust plate boundary for each plate. Filled symbols are $v_{\text{rms}}$ from model D1 vs the total length of subducted plate boundary for each plate.

Fig. 5. The root-mean-squared absolute velocity $v_{\text{rms}}$ and the mean longitudinal absolute velocity $v_{\theta}$ of the lithosphere within latitudinal bands centred on equator. The half-width of the equatorial band is the abscissa. All velocities are for model A3. The quantity $v_{\theta}$, though plotted as a positive quantity, is actually negative (westward) for all bandwidths.
centred on the equator. To keep things simple an absolute velocity model having no net rotation of the lithosphere (A3) was used, and all lithosphere was weighted equally. The rms velocity calculated for equatorial bands of half-width 30° or less is about 1 cm/yr higher than the rms velocity for the Earth (Fig. 5). The mean longitudinal velocity, or net westward drift, is smaller and more concentrated near the equator.

To examine whether the higher velocities in the half of the Earth near the present equator are the result of chance or are related somehow to the rotation axis, we repeated the above calculation for various arbitrary choices of co-ordinate poles. For each new pole, we determined $v_{\text{rms}}$ within a latitudinal band of half-width 30° centred on the equator to this pole. The results are shown in Fig. 6. Quite clearly, the present geographic pole is near the pole with maximum equatorial rms velocity and velocities for other poles are much lower. The magnitude of mean longitudinal velocity within the equatorial band is larger than in Fig. 5 for a number of alternative poles and is sensitive to the width of the equatorial band. Thus the non-zero values of westward drift in Fig. 5 could easily be the result of chance.

The probability that two points such as the pole giving maximum equatorial $v_{\text{rms}}$ and the geographic pole fall within 15 degrees, however, is less than 3 per cent and cannot be readily ascribed to chance. It is interesting to repeat that the magnetic pole, and presumably the rotation axis, has undergone no resolvable net rotation...
over the past 55 My with respect to a co-ordinate system in which there is no net lithosphere rotation. Explanations for Fig. 6 involving Coriolis forces and tidal torques can readily be eliminated by simple physical arguments (e.g. Richter 1973b; Jordan 1974).

One possible explanation is that the geographic pole may seek the pole of the axis of maximum non-hydrostatic inertia (Goldreich & Toomre 1969). A net pole wander of about 90° would be expected in the time for the density inhomogeneities to re-order themselves (< 200 My if connected with plate tectonics). It is not clear how rapid plate motion could create non-hydrostatic inertia, although more rapid subduction and lithosphere generation imply lateral density differences and hydrodynamic forces occur where flow is induced. The axis of minimum non-hydrostatic inertia (15° W) does not coincide with the pole of minimum equatorial rms velocity (45° E) as might be expected from the Goldreich and Toomre concept.

As tidal dissipation in the solid earth may account for up to several per cent of the Earth's heat budget, an asymmetry in plate motions may result from a concentration of this heat at low latitudes. A locking of the net motion of the lithosphere to the geographic pole in this situation could occur from the Goldreich and Toomre mechanism because of the tidally heated region or because of high viscosity in the lower mantle retaining a fossil equatorial bulge (McKenzie 1966). The 20 per cent excess rms velocity near the equator is probably too high, however, to result from an equivalent excess of tidal heating.

Another possibility is that long transform faults tend to stabilize the pole of relative motion between two plates. If by chance two or more such relative motion poles lie near the Earth's geographic pole, they would tend to remain there, giving higher absolute velocities in equatorial regions than in polar regions for the plates involved.

A simple physical explanation for the coincidence of the geographic pole and the pole giving the largest equatorial rms velocity is thus not evident. There is a definite tendency, however, for plates to go fast near the equator and slow near the poles. This is not an artifact of our method for determining absolute velocity since the calculation is invariant to co-ordinate transformations. It will be interesting to see if low plate velocities were present near the poles in the geologic past.

Rms velocities of plate boundaries

The rms velocities of boundaries between plates, in particular of spreading centres and of the overthrust side of convergence zones, are of interest for a number of reasons. Kaula (1974) has argued that the condition that these velocities, or at least their components perpendicular to the plate boundaries, be minimized determines absolute plate motions. In addition, the absolute velocity of boundaries may provide explanations of such features as the variable dip angles of Benioff zones and the propensity of some spreading centres to jump large distances in a short time.

The rms velocities of subduction zone boundaries are given in Table 6 for absolute velocity model B4, a model among these closest to the hot spot models and to Kaula's (1975) preferred solutions (4) and (5). Three velocities are given for each of the two plates abutting each subduction zone:

\[
 v_{\text{rms}} = \left[ \frac{1}{L_t} \int_{\text{line}} v(r) \cdot v(r) \, dl \right]^{\frac{1}{2}} \tag{18}
\]

\[
 (v_{\perp})_{\text{rms}} = \left[ \frac{1}{L_t} \int_{\text{line}} (\omega_{pm}(r) \cdot \frac{dl}{dl})^2 \, dl \right]^{\frac{1}{2}} \tag{19}
\]

\[
 (v_{||})_{\text{rms}} = [v_{\text{rms}}^2 - (v_{\perp})_{\text{rms}}^2]^{\frac{1}{2}} \tag{20}
\]
where the integrations are carried out along the subduction zone $t$, $L_t$ is the total length of the zone, and $\omega_{pm}$ is the absolute angular velocity of the appropriate plate.

An alternative view of the transverse rms velocities of subducting and overthrust plates along the trenches is given in Fig. 7. Morgan (1972a) has suggested that both plates move toward the trench and that there might be a linear relationship between the trenchward components of the subducting and overthrust plates. The former statement is true except for several exceptions that arise from geometrical constraints. PHL cannot move towards both PHLT and MART (it moves away from the latter) and CAR cannot move towards both PRT and CENT (it moves away from the former). Neither EUA nor AME can simultaneously move toward all trenches on their respective perimeters; in model B4, EUA moves slowly away from MEDT, ARBT and HIMT, and AME moves slowly away from ALUT and KURT. A completely general rule, we should note, is that subducted plates always move toward the subduction zone.

Morgan's second statement is not substantiated by Fig. 7. The rms perpendicular velocity $v_\perp$ is near 1 cm/yr for the overthrust plate irrespective of $v_\perp$ for the subducting plate. Two contradicting points are at MART, where the high $v_\perp$ for PHL is misleading since the Philippine plate is moving away from the Marianas subduction zone, and at SOLT, where $v_\perp$ is smaller than $v_\parallel$ for PAC and would be much smaller still if we did not include the Fiji plateau as part of the Pacific plate (see Chase 1971). A value of 1 cm/yr for $v_\perp$ for overthrusting plate boundaries appears to be generally consistent with the results of Kaula (1975).

![Fig. 7. Rms absolute velocities of plates at subduction zones. Shown is the component $v_\perp$ of rms velocity perpendicular to the subduction zone for the overthrust plate vs that for the subducted plate. Open symbols are for subduction zones at which the overthrust plate is moving away from the plate boundary. Filled symbols are for subduction zones at which the overthrust plate is moving toward the plate boundary. Several lines of constant $v_\perp/v_\parallel$ (where $v_\parallel$ and $v_\perp$ are $v_\parallel$ for the subducted and overthrust plates, respectively) are shown for reference.](https://academic.oup.com/gji/article-abstract/42/2/769/652631)
Since the trenchward component of absolute velocity of the overthrust plate at subduction zones is nearly a constant for all absolute place motion models considered, it is not surprising that the dip angle of subducted slabs cannot be explained simply as a direct consequence of the absolute velocity of the overthrust plate and a constant or nearly constant vertical sinking rate for slab material (Luyendyk 1970). It might be noted, however, that Table 6 and Fig. 7 do not include the effects of interarc spreading.

As an aside, the relative magnitudes of $v_{\parallel}$ and $v_{\perp}$ at the subduction zones listed in Table 6 can be viewed as a measure of the ratio of the forces transmitted through the plates over large distances to the local forces associated with the subducted lithosphere.

The rms velocities of spreading centre boundaries are given in Table 7, again for absolute velocity model B4. In that table,

$$v_{\text{rms}} = \left( \frac{1}{L_s} \int_{s_{\text{line}}} \left( \omega_{\text{sm}} \times r \right) \cdot \left( \omega_{\text{sm}} \times r \right) \, dl \right)^{\frac{1}{2}}$$

$$\langle v_{\perp} \rangle_{\text{rms}} = \left( \frac{1}{L_s} \int_{s_{\text{line}}} \left( \frac{\left( \omega_{\text{sm}} \times r \right) \cdot \left( \omega_{12} \times r \right)}{|\omega_{12} \times r|} \right)^2 \, dl \right)^{\frac{1}{2}}$$

$$\langle v_{\parallel} \rangle_{\text{rms}} = \left[ v_{\parallel}^{\text{rms}} - \langle v_{\perp} \rangle_{\text{rms}}^2 \right]^{\frac{1}{2}}$$

where the integrations are carried out along the spreading centre $s$, $L_s$ is the total length of the spreading centre (including ridge crest and transform fault segments), and $\omega_{12}$ and $\omega_{\text{sm}}$ are the relative angular velocity and the mean of the absolute angular velocities of the adjacent plates, respectively. With these definitions, $v_{\parallel}$ is the component of $v$ in a direction generally perpendicular to ridge crest segments and always parallel to transform faults.

Table 6

<table>
<thead>
<tr>
<th>Subduction zone</th>
<th>Length*</th>
<th>Name</th>
<th>$v$</th>
<th>$v_\perp$</th>
<th>$v_\parallel$</th>
<th>Overthrust plate</th>
<th>Name</th>
<th>$v$</th>
<th>$v_\perp$</th>
<th>$v_\parallel$</th>
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<tr>
<td>ALUT</td>
<td>0.601</td>
<td>PAC</td>
<td>5.64</td>
<td>4.26</td>
<td>5.09</td>
<td>AME</td>
<td>1.38</td>
<td>0.72</td>
<td>1.18</td>
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<td>MEXT</td>
<td>0.346</td>
<td>COC</td>
<td>5.44</td>
<td>5.36</td>
<td>0.94</td>
<td>AME</td>
<td>1.61</td>
<td>0.80</td>
<td>1.40</td>
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<tr>
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<td>0.300</td>
<td>COC</td>
<td>9.01</td>
<td>8.99</td>
<td>0.56</td>
<td>CAR</td>
<td>1.06</td>
<td>0.50</td>
<td>0.93</td>
<td></td>
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<td>CHIT</td>
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<td>NAZ</td>
<td>9.27</td>
<td>8.41</td>
<td>3.90</td>
<td>AME</td>
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<td>0.91</td>
<td>0.67</td>
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<td>AME</td>
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<td>1.13</td>
<td>1.20</td>
<td>CAR</td>
<td>0.90</td>
<td>0.61</td>
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<td>0.93</td>
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<td>1.14</td>
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<td>3.76</td>
<td>1.62</td>
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<td>4.76</td>
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<td>EUA</td>
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<td>0.70</td>
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<tr>
<td>JAYT</td>
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<td>5.91</td>
<td>3.89</td>
<td>EUA</td>
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<td>0.83</td>
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<tr>
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<td>4.91</td>
<td>2.42</td>
<td>PAC</td>
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<td>6.91</td>
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<td>IND</td>
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<td>1.28</td>
<td>2.66</td>
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<tr>
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<td>7.03</td>
<td>3.58</td>
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<td>0.32</td>
<td>1.18</td>
<td></td>
</tr>
<tr>
<td>JAPT</td>
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<td>PAC</td>
<td>8.54</td>
<td>7.23</td>
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<td>EUA</td>
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<td>1.37</td>
<td>0.31</td>
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<tr>
<td>PHLT</td>
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<td>PHL</td>
<td>7.74</td>
<td>5.58</td>
<td>5.37</td>
<td>EUA</td>
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<td>1.14</td>
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<tr>
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<td>PHL</td>
<td>7.20</td>
<td>5.73</td>
<td>4.36</td>
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</tbody>
</table>

For explanation of $v$, $v_\perp$ and $v_\parallel$, see equations (18)–(20). Velocities are in cm/yr.

* Earth's radius equals 1.

† COC abuts CENT only along northermost segment of length 0.163. See Fig. 1.
There is an order of magnitude variation in the absolute velocity of spreading centres. Perhaps more importantly, there is a similar variation in \( v_L \), a measure of how fast the ridge is moving with respect to its immediate source of new lithospheric material in the underlying mantle. It is difficult to generalize from Table 7. The rms perpendicular velocity for all spreading centres taken as a group is 2.55 cm/yr, a figure similar to one obtained by Kaula (1974). More than half of the entries in Table 7, but less than half of the total length of spreading centres, have \( u_L \) greater than this. It is not evident from the Table why minimization of \( u_L \) for spreading centres should be an important constraint for determination of absolute plate motions.

It is tempting to speculate that the conditions for ridge crest 'jumps' (Sclater, Anderson & Bell 1971) and perhaps also asymmetric spreading (Weissel & Hayes 1971) are more favourable when the ridge crest is moving rapidly with respect to its immediate source of new ocean floor than when it is relatively stationary. Such a ridge might be expected to jump in that direction which reduces \( v_L \), i.e. towards the site of its former source of new lithosphere. The ridge segment with the largest value of \( v_L \) in Table 7 is the Chile rise (also Chile ridge). There is no evidence for any jump within the last 5 My (Klitgord et al. 1973), though constraints on the earlier history of the ridge crest are rather weak. The Chile rise was part of the east Pacific spreading system that underwent a complex pattern of jumps between 5 and 10 My ago (Sclater et al. 1971; Anderson & Sclater 1972; Herron 1972; Molnar et al. 1975). Some of the presently active ridge segments in the eastern Pacific also show relatively large values (3 to 4 cm/yr) for \( v_L \). The Pacific–Antarctic ridge, with \( v_L \) equal to 3 cm/yr, is one of the few documented cases of asymmetric spreading (Weissel & Hayes 1971, 1974) also in the sense to be expected if the velocity of the ridge with respect to its source of new lithosphere were a controlling parameter. On the other hand, the Carlsberg ridge, with \( v_L \) equal to 4 cm/yr, shows no evidence for either jumps or asymmetries (McKenzie & Sclater 1971). It does not appear, therefore, that a large value of \( v_L \) is a sufficient condition for ridge jumps or for asymmetric spreading, though jumps may be more likely when \( v_L \) is large than when it is small.

To summarize this section, the rms absolute velocities of subduction zones and of spreading centres are generally less (roughly 2 cm/yr) than are the rms absolute velocities of the plates (5 cm/yr). This observation may have an important physical
significance (Kaula 1975). On the other hand it may be due solely to a sampling problem: the slow moving plate margins tend to persist longer in time than the fast moving ones, which collide with other features and are annihilated. Thus at any given time, we are more likely to observe more slow moving than fast moving margins even if both types are produced at roughly equal rates.

**Intraplate stress**

A powerful constraint on models for the forces driving plates is the intraplate stress field. In this section we first review the available observational data pertinent to the present state of stress within the lithosphere. We then show how the intraplate stress field may be calculated from a driving force model. Finally, we present the calculated stresses for several force models to illustrate how comparison of the predicted with the observed stress fields can serve as a test of the acceptability of each model.

**Observations of intraplate stress**

Within every large lithospheric plate, earthquakes occasionally occur. In a few mid-plate regions, usually within continents, earthquakes are common. If it is assumed that these earthquakes represent brittle failure in the lithosphere along fresh, frictionless faults, then the directions of the three principal stress axes prior to failure can be recovered unambiguously from the fault-plane solution for each earthquake (Mendiguren 1971; Forsyth 1973; Fitch, Worthington & Everingham 1973). An excellent summary of about 80 mid-plate earthquake mechanisms and the inferred lithospheric stress pattern was given by Sykes & Sbar (1973). Some of their results are shown in Fig. 8.

![Fig. 8. Principal horizontal deviatoric stresses inferred from midplate earthquake mechanisms, after Sykes & Sbar (1973) and L. R. Sykes, private communication (1974). Triangles are thrust-fault mechanisms, squares are normal-fault mechanisms, and circles are strike-slip mechanisms. Arrows denote the horizontal projection of relevant \( P \) (compressional) and \( T \) (tensional) axes inferred by Sykes & Sbar (1974) from fault plane solutions. Symbols without arrows are events for which the type of faulting is evident but the location of the two fault planes is not. The earthquakes shown are those given by Sykes and Sbar but excluding normal-faulting events near ridges, all mechanisms very near subduction zones, normal-faulting events possibly associated with subsidence of continental margins, and events very close to ones shown. The mechanism for the April 1973 earthquake near Hawaii (Unger, Koyanagi & Ward 1973) has been added. Cylindrical equidistant projection, with plate boundaries from Fig. 1.](https://academic.oup.com/gji/article-abstract/42/2/769/652631)
There are a number of reasons to be sceptical of the stress field inferred from a single isolated earthquake. The rupture may occur along a pre-existing zone of weakness and thus bear no simple relationship to the stress field (McKenzie 1969b). The inferred stress axes may vary markedly over a small region because of topographic or structural effects (e.g. Street, Herrman & Nuttli 1974).

Nonetheless there are several regions where, by demonstration that the inferred stress field is uniform over a large area, or by corroboration with measurements from independent techniques, the lithospheric stress field is well established. Throughout much of the European continent north of and including parts of the Alps, earthquake mechanisms and other data imply that the direction of greatest compressive deviatoric stress is roughly horizontal and strikes NW-SE, approximately parallel to the lower Rhine graben (Ahorner 1970; Ahorner, Murawski & Schneider 1972; Schäfer 1974). Similarly, over a large area in eastern North America, a number of independent techniques show that the greatest compressive deviatoric stress is generally horizontal and strikes ENE-WSW (Voight 1969; Sbar & Sykes 1973). Sykes & Sbar (1973) state that large horizontal compressive stresses are typical of oceanic lithosphere older than 20 My and of many continental regions as well.

In addition to intraplate earthquake mechanisms, there are several other techniques for determining the state of stress in the lithosphere. In situ measurement of stress in deep holes, by several different methods, holds great promise for defining both the directions and magnitudes of the principal deviatoric stresses in at least the continental plates (e.g. Hast 1969; Voight 1969; Kropotkin 1972). A number of large-scale geologic features can be indicative of recent stress patterns. Horizontal tensional stress has been invoked to account for linear volcanic chains in old oceanic lithosphere (Jackson & Wright 1970; Jackson, Silver & Dalrymple 1972; Turcotte & Oxburgh 1973; Solomon & Sleep 1974) and for continental rift structures such as the Rhine graben and the east African rift system (Turcotte & Oxburgh 1973; Oxburgh & Turcotte 1974). Zonal faulted and folded sediments can be indicators of compression of oceanic lithosphere (Eittreim & Ewing 1972). Schäfer (1974) has argued that stylolites can be used to deduce compressive stresses in the crust at various geologic times. Finally, the magnitude of deviatoric stresses in the Earth may be bounded, certainly by the fracture strength of rock, and also by the heat-flow anomaly associated with a fault (Brune, Henyey & Roy 1969) and by the earthquake source parameters stress drop, effective stress, and apparent average stress (Brune 1970; Trifunac 1972), all three of which may be estimated from the amplitudes of appropriate seismic waves.

Other causes of deviatoric stress in the lithosphere

There are several reasons why one must be cautious about attributing any particular measured stress to the forces driving plates. The stress may not be representative of the entire thickness of the plate. The depths of very few mid-plate earthquakes have been determined with much accuracy, for instance, and in situ measurements are limited to the upper few kilometres of the crust. More important, there may be a number of mechanisms unrelated to the forces driving plates that can produce significant deviatoric stress in the lithosphere. Among these are crustal thickness inhomogeneities (Artysukhov 1973); thermoelastic stresses in the cooling oceanic lithosphere (Turcotte & Oxburgh 1973; Turcotte 1974b; Collette et al. 1974), translation of plates over an ellipsoidal earth (McKenzie 1972; Turcotte & Oxburgh 1973; Oxburgh & Turcotte 1974), and time-dependent loading associated with glaciation, sedimentation, and erosion.

That stresses due to plate motion on a non-spherical earth produce all or most rift structures and volcanic island chains can be easily challenged on purely geometrical grounds (Solomon & Sleep 1974), though such a mechanism may be operative for a few such structures (Oxburgh & Turcotte 1974). Regions in which large stresses are produced by glacial unloading and by rapid sedimentation or erosion are probably
restricted in areal extent, and effects of such processes can probably be minimized by using stress determinations which are uniform over large regions. Stresses due to crustal thickness inhomogeneities within continents and between continent and ocean are not considered below, but these stresses can be calculated from known crustal structure and add linearly to the stresses associated with driving forces.

It may also be argued that thermoelastic stresses are not important in the lithosphere far from active plate margins. Newly-formed oceanic lithosphere is hot except very near the surface (Parker & Oldenburg 1973; Sleep 1974). As the lithosphere spreads and ages, the middle portion of the lithosphere cools substantially but the base of the lithosphere cools only slightly. Thermoelastic stress results from the contraction of the middle portion of the lithosphere with respect to the top and bottom. As with a columnar dike, fractures much more closely spaced than the thickness of the lithosphere are needed to relax the stress. If most of this deformation occurs at depth the surficial area of the plate will be unchanged.

Thermoelastic stresses on the level of grain size are also the same order as the total thermoelastic stress, because common mantle minerals such as pyroxenes and olivine have thermal expansion coefficients which are anisotropic to as much as a factor of two (Skinner 1966). These stresses must be relaxed by intragranular and intergranular cracks. The crumbly texture of ultramafic xenoliths is in part evidence of this process. Recrystallization often observed in these xenoliths would also relax stress.

Horizontal thermoelastic stresses produced by lateral contraction of the plate can cause a large, isolated fracture only if stress relaxation cannot occur throughout the material. As mentioned above, however, the relaxation of thermoelastic stress must be homogeneously distributed simultaneously on the scales both of grain size and of plate thickness. The ubiquitous cracks and fractures formed to relax vertical and granular stresses would also serve to relax the larger scale horizontal stresses. It also should be noted that no stress remains once thermoelastic stresses are relaxed by a small percentage of creep, in contrast to the case of density heterogeneities, where small non-elastic deformations cannot alter the mass distribution significantly.

The self-consistent alignment of seamount chains with present spreading directions rather than with fossil magnetic lineations and the creation of most of the relief of fracture zones at the ridge axis cannot simply be attributed to a thermal-elastic mechanism. Central rift and fracture zone topography is probably related to viscous forces in the conduit feeding the ridge (Sleep & Biehler 1970). We discuss below the alternative possibility that seamount chains are related to the driving forces acting at plate boundaries.

Our philosophy, then, is that the long-wavelength features of the global stress field, particularly those features which have been verified by a number of independent types of measurement, are most likely a result of the forces currently acting on plate boundaries. While this premise deserves continuous scrutiny, we feel it has value as a working hypothesis because of the clear potential for elucidating the driving mechanisms for plate tectonics.

Calculation of intraplate stress

Intraplate stress is calculated from the theory of thin, elastic shells, adopting the assumptions described earlier for determination of absolute velocities and making several additional simplifications.

Because the thickness of each lithospheric plate is much less than its radius of curvature (the radius of the Earth), the state of stress in the lithosphere may be approximated by the membrane state of stress (Turcotte 1974a). In the membrane theory, bending moments are small compared with stress resultants, and the equili-
On the forces driving plate tectonics

The equilibrium equations for a spherical shell are (e.g. Kraus 1967, p. 88)

\[
\begin{align*}
\frac{\partial}{\partial \phi} (N_\phi \sin \phi) + \frac{\partial N_\phi}{\partial \theta} - N_\theta \cos \phi + q_\phi R \sin \phi &= 0 \\
\frac{\partial}{\partial \phi} (N_\phi \sin \phi) + \frac{\partial N_\theta}{\partial \theta} + N_\phi \cos \phi + q_\theta R \sin \phi &= 0 \\
N_\phi + N_\theta + q_r R &= 0
\end{align*}
\]

(24)

where \(\phi, \theta, r\) are the spherical co-ordinates co-latitude, longitude, and radius, and \(R\) is the radius of the reference surface of the shell. The quantities \(q_\phi, q_\theta,\) and \(q_r\) are applied tractions, statically equivalent to the body and surface forces exerted on the plates at their edges and on their bases. The stress resultants \(N_\phi, N_\theta,\) and \(N_\phi\theta\) are integrals of the stress components \(\sigma_\phi, \sigma_\theta,\) and \(\sigma_\phi\theta\) across the thickness of the shell.

The equilibrium equations (24) may also be written in terms of the components of displacement \(u_\phi, u_\theta, w\) in the co-ordinates \(\phi, \theta, r\) respectively. The displacements are related to strains by

\[
\begin{align*}
\varepsilon_\phi &= \frac{1}{R} \left( \frac{\partial u_\phi}{\partial \phi} + w \right) \\
\varepsilon_\theta &= \frac{1}{R} \left( \csc \phi \frac{\partial u_\phi}{\partial \theta} + u_\phi \cot \phi + w \right) \\
\varepsilon_\phi\theta &= \frac{1}{R} \left( \frac{\partial u_\theta}{\partial \phi} - u_\theta \cot \phi + \csc \phi \frac{\partial u_\phi}{\partial \theta} \right)
\end{align*}
\]

(25)

and the strains are in turn related to the stress resultants by

\[
\begin{align*}
N_\phi &= K(\varepsilon_\phi + \nu \varepsilon_\theta) \\
N_\theta &= K(\nu \varepsilon_\phi + \varepsilon_\theta) \\
N_\phi\theta &= \mu \varepsilon_\phi\theta
\end{align*}
\]

(26)

where

\[K = \frac{hE}{1 - \nu^2}\]

and where \(h\) is the plate thickness, \(E\) is Young’s modulus, \(\nu\) is Poisson’s ratio, and \(\mu\) is the shear modulus.

The governing equations (24–26) amount to a two-dimensional description of intraplate stress, albeit on a spherical surface. Once the applied tractions \(q_\phi, q_\theta,\) and \(q_r\) are specified, then the equilibrium equations may be solved for displacements and for stresses, regarded as stress resultants divided by plate thickness \(h\). The tangential tractions \(q_\phi\) and \(q_\theta\) are arbitrary and depend on the assumed model for driving forces. The radial traction \(q_r\) is not arbitrary, because long-wavelength elevation variations are isostatically compensated in the Earth. We take

\[q_r = -\rho gw,\]

(27)

where \(\rho = 2.3 \text{ g cm}^3\) and \(g = 980 \text{ cm s}^{-2}\). As a result of (27), \(w\) is generally much smaller than \(u_\phi\) or \(u_\theta\).

To solve equations (24)–(27) given \(q_\phi\) and \(q_\theta\), we rewrite the equations in terms of finite differences and solve the difference equations numerically using a successive over-relaxation technique. A full description of the difference equations, including treatment of singularities in the equations at the co-ordinate poles and tests of the accuracy of the equations, is given in a separate paper (Richardson, Solomon & Sleep 1975).
Fig. 9. Plate boundaries adopted for intraplate stress calculation. All segments of plate boundaries fall along either meridians or parallels and pass midway between adjacent grid points on a $10^\circ \times 10^\circ$ spherical grid. Ridges are denoted by double lines, subduction zones by single solid lines, and transform faults by dashed lines. Cylindrical equidistant projection.

The Earth is divided into a $10^\circ \times 10^\circ$ grid in $\phi$ and $\theta$ (Fig. 9). Plate boundaries pass between grid points. With such a coarse grid size, only the seven largest plates are included. PHL is incorporated into PAC, ARB into IND, CAR into AME, and COC into NAZ.

The elastic constants $E$ and $v$ and the shell thickness $h$ are taken to be everywhere uniform ($10^{12}$ dyne cm$^{-2}$, 0.25, and $10^7$ cm, respectively). This amounts to treating the earth's lithosphere as a continuous, homogeneous shell. A quite different approach would be to consider each plate in isolation, i.e. to regard every plate edge as a free boundary. The problem would then be to specify adequately the forces acting on each plate boundary. When such forces are parameterized to depend only on the type of boundary and on the relative plate velocity at the boundary, the torques on individual plates do not balance without making a number of ad hoc adjustments to the forces on most of the plates (Tullis & Chapple 1973; Forsyth & Uyeda 1975; Harper 1975). Clearly this is due to omission of forces transmitted across boundaries from one plate to another. The approach taken in this paper is admittedly at the other extreme. The forces transferred across plate boundaries are perhaps accentuated unduly. In future models, an intermediate stance can be taken by lowering the elastic constants within narrow regions along each plate boundary.

Traction forces are applied at grid points immediately adjacent to subduction zones and spreading centres equivalent to imposed body forces $F_T$ and $F_R$ per unit length, respectively. The body forces are assumed to act in the direction of relative plate velocity, away from the boundary at ridges and toward the boundary at trenches. $F_R$ is always symmetric about the ridge boundary; $F_T$ may be larger on the subducting plate side of the trench boundary. The force imposed at ridges may be envisioned as due to cutting the lithosphere along all ridge crests, adding a fixed amount of new material per unit length of cut, and then welding the cuts closed. The force imposed at trenches may be visualized as due to cutting, removing material, and then welding. No explicit account is taken of transform faults.

Traction forces equivalent to the drag force $F_D$ acting on the lower surface of the lithosphere are applied at all grid points. The force $F_D$ is proportional to, and in the
direction opposite to, the absolute velocity $v$ of the lithosphere at that point:

$$F_D = -Dv.$$  \hfill (28)

Even with all of the simplifications made, there are still many degrees of freedom in the driving force models we might examine. We consider the ratios $F_T/F_R$ and $D/F_R$ (see equation (28)) arbitrary, as is the ratio of pull on subducting plate to pull on overthrust plate at trenches. $D$ may differ for continents and oceans. $F_T$ may not be a constant for all subduction zones but may depend on convergence rate or on the age of the subducted lithosphere, though we do not consider such possibilities in the models below. The various parameterizations are not completely arbitrary. Conservation of energy may serve to limit some of them and simple mechanical and/or thermal models for plate boundaries may also provide guidance.

**Intraplate stress for several force models**

The intraplate stress field was calculated for a number of simple driving force models following the guidelines given above. Several such models are presented here in detail, and a number of variations on these models are discussed as well. Additional models are given in Richardson, Solomon & Sleep (1975).

The simplest class of models which contain many of the potentially important forces consists of buoyancy forces $F_R$ and $F_T$ exerted symmetrically about ridges and about trenches, respectively, and viscous drag at the base of the lithosphere governed by a uniform drag coefficient $D$. The midplate stress field for one such model is shown in Fig. 10. For that model, the magnitudes of $F_R$ and $F_T$ are equal and the tractions $q$ in (24) associated with the drag force $F_D$ in (28) are scaled to the tractions exerted at plate boundaries by $D = 0.003 F_T$. The absolute velocity $v$ in (28) is essentially the velocity from model A3, except that the summations in equation (2) are over grid points; the absolute angular velocity vector for PAC is $7.25 \times 10^{-7}$ deg/yr about $64.8^\circ$ S, $115.5^\circ$ E instead of that given in Table 4.

![Fig. 10. Principal horizontal deviatoric stresses in the lithosphere predicted by a model with symmetric driving forces $F_R$ and $F_T$ at ridges and trenches, respectively, and drag beneath all plates given by $F_D = -Dv$, where $v$ is the absolute plate velocity. Scaling of various imposed forces is as shown. Principal stress axes with arrows denote tensional deviatoric stress; those without denote compression. The length of each stress axis is proportional to the magnitude of the principal stress; all lengths are arbitrary to within a multiplicative constant. Plate boundaries and map projection from Fig. 9.](https://example.com/image.png)
In spite of the simplicity of the model, the directions of principal stresses (Fig. 10) in many regions compare quite favourably with those in Fig. 8 inferred from midplate earthquake mechanisms (Sykes & Sbar 1973). The agreement is very good in eastern North America and in western Europe, the two regions in which intraplate deviatoric stress has been most reliably determined. Agreement is also good in the eastern Pacific, eastern Asia and parts of the Indian plate. The match is less than adequate, however, in west Africa and in the Nazca plate. The largest stresses (not at grid points immediately adjacent to ridge or trench segments) are in the eastern and western Pacific and near the Alpine–Himalayan belt. The large tensional stresses in oceanic plates near several subduction zones (e.g. the south-west Pacific) are not corroborated by normal faulting earthquakes, but may be partially relieved by tensional fractures in the lithosphere that give rise to volcanic island chains.

How reasonable are the ratios of driving forces at ridges to those at trenches and of drag forces to ridge or trench driving forces? Certainly there is a great deal more potential energy available at subduction zones to drive plate motions than at ridges (McKenzie 1969a; Smith & Toksoz 1972; Press 1973), so one might expect \( F_R/F_T \) to be much less than unity. Both \( F_T \) and \( F_R \), however, are the net force exerted at plate edges, i.e. the difference between the available buoyancy force and the viscous forces opposing them. Forsyth & Uyeda (1975) and Tullis & Chapple (1973) both maintain from plate velocity considerations that driving and resistive forces at subduction zones, though both larger than other forces in the plate tectonic system, nearly cancel each other. \( F_R/F_T \) near unity may therefore not be unreasonable, though a constant ratio is doubtless an oversimplification. We consider other ratios below.

The scaling of drag to driving forces for the model of Fig. 10 is permitted by conservation of energy (e.g. equation (15)) as long as other resistive forces are available against which the plates must perform work. Such additional resistive forces are implicit rather than explicit in the model and include forces at transform faults and forces transmitted from one plate to another at other plate boundaries. We considered models identical to that in Fig. 10 except with different ratios of \( D/F_T \). Models with that ratio less than about 0.001 did not match the stresses inferred from the Eurasian and Indian plates as well as the model in Fig. 10. Models with \( D/F_T \) higher than about 0.01 also provide a less satisfactory fit to the data of Fig. 8.

All stresses shown in Fig. 10 and for subsequent models are arbitrary to within a multiplicative constant, since we have not specified the magnitude of \( F_R, F_T \) or \( D \), but only their ratios. It is not intended in this paper to present a completely quantitative model of stress in the Earth, but we can at least demonstrate the plausibility of the model in Fig. 10 by fixing the deviatoric stress at one point and thus determining the stress everywhere else. It has often been remarked (Andrews 1972; Morgan 1972; Artyushkov 1973) that horizontal compressive stresses are on the order of few hundreds of bars in the oceanic lithosphere near ridges. If this figure is adopted for the model in Fig. 10, then the applied tensional tractions at subduction zones are also a few hundred bars and the tractions due to drag range from perhaps one to ten bars. The drag force may be converted to a rough estimate of the viscosity \( \eta \) of the asthenosphere. In the simplest model of drag

\[
F_D = - \frac{\eta}{h} v
\]

where \( h \) is the thickness of the asthenosphere. Taking \( F_D = 10^6 \), \( h = 10^7 \), and \( v = 10^{-7} \) (all in cgs units) gives \( \eta = 10^{20} \) poise, not an unreasonable value. All of the stresses shown in Fig. 10 for this scaling are on the order of hundreds of bars, within the lower range of, or perhaps slightly below, the deviatoric stress magnitudes thought necessary to produce occasional midplate earthquakes.

One of the important conclusions to be drawn from the class of models represented by Fig. 10 is that midplate stress is quite sensitive to the viscous drag at the base of
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the lithosphere. This follows directly from membrane stress theory. Tractions applied more or less uniformly over the reference surface of a shell produce membrane stresses which are larger than the tractions by a factor of $R/h$. A commonly cited example of this is the expression for membrane stresses in a spherical shell loaded by its own weight (Kraus 1967; p. 102). Thus for the Earth, the effect of drag forces of membrane stress in the lithosphere is magnified by a factor of perhaps 50 to 100 relative to the effect of edge forces at ridges and trenches.

The intraplate stress for a model similar to that in Fig. 10 but with negligible driving force exerted at ridges ($F_R/F_T = 0$) is shown in Fig. 11. The principal stresses are tensional throughout much of the Earth, and are in contradiction to Fig. 8 for all regions except parts of Africa and eastern Asia. Agreement is not satisfactory in eastern North America and in western Europe, the two regions in which stress is best determined. This model is unacceptable. No model of the general type considered in this paper, in fact, could match the principal stresses in North America and Europe unless driving forces at ridges are at least comparable in importance to the driving forces exerted at subduction zones.

A model in Fig. 12 tests the sensitivity of intraplate stress to increased drag at the base of the continental lithosphere. The model has $F_R/F_T = 1$, $D/F_T = 0.003$ beneath oceans and 0.009 beneath continents. The absolute velocity field is similar to model B4, except the summations in equation (2) are over continental grid points so that the absolute angular velocity for PAC is $7.91 \times 10^{-7}$ deg/yr about 64°2'S, 113°7'E. Except for some small but noticeable rotations in the directions of principal stresses and some changes in the magnitudes of the principal stresses, the intraplate stress field in Fig. 12 is quite similar to that in Fig. 10. It is clear that at our present level of knowledge about stress in the lithosphere, the mid-plate stress field cannot be used to decide whether drag resistance is greater beneath continental than oceanic plates. Viscous drag at the base of oceanic plates is probably not negligible, however. In models with no drag beneath oceans, the fit of the principal stresses in oceanic plates to the data of Fig. 8 is somewhat worse than the fit for the models in Figs 10 and 12.

We do not wish to imply that we have a unique or even a best model for the plate tectonic driving forces, but rather to show that the intraplate stress field can provide a sensitive test of possible force models. Within the limited framework of models

![Diagram](https://academic.oup.com/gji/article-abstract/42/2/769/652631)
considered, viscous drag at the base of the lithosphere is clearly an important contributor to midplate stress and the driving forces exerted on plates at spreading centres appear to be of comparable magnitude to the forces at subduction zones. That this last conclusion is assumption dependent cannot be ruled out since we have not considered several possible additional forces such as drag associated with convective motions in the asthenosphere (e.g. the secondary convection scale of Richter (1973b)), forces at omitted plate boundaries (e.g. in East Africa), or more complicated dependence of buoyancy forces on the velocity, age or constitution of the respective lithosphere.

Conclusions

The concept that the plates of the Earth can be treated as distinct mechanical entities, as well as providing the foundation for the kinematics of plate tectonics, is also valuable for deciding among the possible driving mechanisms for plate motions. In this paper the absolute velocities of the plates and the intraplate stress field were determined for a number of possible sets of driving forces.

The absolute plate motions which follow from a wide variety of independent physical or geological constraints are all very similar. Limited evidence suggests that the class of absolute velocity fields encompassed by such models probably contains the true absolute velocity field, but there is presently no means to choose among the models and thus quantitatively characterize the physical process controlling the motions. Plates move faster, in a root-mean-squared sense, then do most plate boundaries and mantle 'hot spots', probably because tectonic boundary features and hot spots are related to lateral temperature inhomogeneities in the asthenosphere which cannot move as fast as can the plates.

The rms absolute velocities of individual plates are most consistent with the driving mechanism if pull by subducted lithosphere is an important driving force and if increased drag at the base of continental lithosphere is an important resisting force. The rms plate velocities also appear to be somehow related to the Earth's rotation, in that plate velocities are significantly higher near the equator than near the poles, a conclusion that is not true for co-ordinate poles much different from the Earth's
geographic pole. Detailed examination of the rms absolute velocities of plate boundaries produced two new findings: (1) Almost all subducted slabs have nearly the same horizontal translation velocity perpendicular to strike, about 1 cm/yr. The dip of the various slabs is not, therefore, a simple consequence of a nearly constant vertical sinking rate and the absolute velocity of the overthrust plate. (2) Ridges apparently can move fast or slow relative to their underlying source of new lithospheric material. The fast moving ridges are not necessarily those most prone to jumping.

The intraplate stress field, calculated using membrane theory for thin shells, depends explicitly on all driving and resistive forces acting on the plates. Comparison of the calculated stresses with carefully screened estimates of midplate stress determined from earthquake mechanisms, in situ measurements and stress-induced geological structures is a powerful test of plate tectonic driving mechanisms.

The simple force models considered in this paper can provide a surprisingly good fit to measured or inferred stresses. At this stage we cannot find an acceptable force model that does not include a driving force component at ridges that is at least comparable in importance to the driving force component at trenches. Intraplate stresses are extremely sensitive to the nature of viscous drag at the base of the lithosphere, so that stress measurements may ultimately provide one of the better indicators of absolute plate velocities.

Clearly more detailed models of stress in the lithosphere are required. A finer spatial resolution and the ability to include smaller plates would be desirable. Additional forces that merit testing are explicit resistive forces at transform faults and driving forces at subduction zones that vary with the convergence rate and/or the age of the subducted lithosphere. The effect of less coupling between plates at boundaries needs to be considered, possibly by modelling boundaries as narrow zones of low elastic moduli. A contrast in elastic properties or thickness between continental and oceanic lithosphere may also be important.

We expect that as our knowledge of the state of stress in the lithosphere improves and as our ability to model plates and the forces acting on them is refined, the possible driving mechanisms for plate tectonics can be greatly restricted.

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