INTRAPLATE STRESS AND THE DRIVING MECHANISM FOR PLATE TECTONICS

by

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ABSTRACT

Long wavelength features of the intraplate stress field may be modeled in terms of the driving mechanism for plate tectonics. Estimates of the potential of various plate driving and resisting forces acting on the edge and along the base of the lithospheric plates indicate that topographic features such as mid-ocean ridges and mountain ranges at continental convergence zones may exert compressive deviatoric stresses of several hundred bars on the plates. Cool, dense subducted lithosphere at oceanic subduction zones may exert deviatoric tensile stresses of several kilobars on the plates. Viscous drag forces acting on the base of the lithosphere equivalent to shear stresses on the order of a few bars are suggested from estimates of mantle properties and absolute plate velocities of a few cm/yr. Whether these shear stresses drive or resist plate motions depends on the assumed absolute motions of the lithospheric plates with respect to the mantle.

Several long wavelength patterns for the orientation of horizontal principal deviatoric stresses have been established from a summary of the global intraplate stress field data based on earthquake mechanisms, in-situ strain and stress measurements, and stress sensitive geologic features. Maximum compressive stresses trend E-W to NE-SW for much of stable North America and E-W to NW-SE for continental South America. In western Europe, the maximum compressive stresses trend NW-SE, while in Asia the trend is more nearly N-S, especially near the Himalayan front. In the Indian plate, the trend varies from nearly N-S in continental India to more E-W in Australia. Horizontal stresses are variable in Africa, but tend to indicate a NW-SE trend for the maximum compressive stress in west Africa and an E-W trend for the minimum compressive stress in east Africa. Oceanic lithosphere away from plate boundaries is generally in a state of deviatoric compression, although few focal mechanisms can be constrained to define the orientation of the principal stresses.

Apparent stress and stress drop have been determined for a number of intraplate earthquakes, with typical values of a bar and a few tens of bars for apparent stress and stress drop, respectively. There is no significant difference in either apparent stress or stress drop between intraplate and plate boundary environments. The apparent stress and stress drop data are consistent with, but do not require, ambient tectonic stresses on the order of hundreds of bars rather than kilobars.
A linear elastic finite element method has been developed based on the wave-front solution technique. The method is first applied to a single plate analysis of stresses in the Nazca plate due to plate driving forces and large plate boundary earthquakes. Ridge pushing forces are required in all models that match the nearly east-west horizontal compression inferred from thrust earthquakes in the interior of the Nazca plate. The net pulling force of the subduction slab on the Nazca plate is at most comparable to ridge pushing forces. Based on the estimate of ridge pushing forces and the limit on the ratio of the net slab pulling force to ridge forces, regional intraplate deviatoric stresses are estimated to be on the order of a few hundred bars. Changes in the Nazca intraplate stress field due to the 1960 Chilean earthquake are, at most, a few tens of bars locally and about one bar at greater distances into the plate. Such small changes in stress levels are probably not significant, although the corresponding changes in the displacement field should be observable using precise geodetic measurement techniques.

The finite element technique is extended to global models of the intraplate stress field due to plate driving and resisting forces. Ridge pushing forces are required in all models that match the orientation of long wavelength features of observed global intraplate stresses. The net pulling force of subducted lithosphere is at most a few times larger than other forces acting on the plates. Resistive forces associated with trench thrust faults and motion of the slab with respect to the mantle must nearly balance the large gravitational potential of the slab. The upper limit on net slab forces may be increased by less than a factor of two if net slab forces are reduced for the fastest moving plates by assuming that the resistive component of the net slab force increases with subduction rate.

Forces acting at continental convergence zones along the Eurasian plate that resist further convergence are important for models of the intraplate stress field in Europe, Asia, and the Indian plate. The upper bound on net slab forces cannot be increased by including continental convergence zone forces.

Resistive viscous drag forces acting on the base of the lithosphere improve the fit between calculated and observed stresses in the Nazca and South American plates as long as the drag coefficient is non-zero beneath oceanic lithosphere. The calculated intraplate stress field is not very sensitive to an increased drag coefficient beneath old oceanic lithosphere compared to young oceanic or continental lithosphere. Increasing the drag coefficient beneath continents compared to oceanic lithosphere by a factor of five or ten has little effect on the overall fit of calculated stresses to observed stresses.

Models in which viscous drag forces drive, rather than resist, plate motions are in poor agreement with intraplate stress data, although this lack of fit may depend on the oversimplified model of mantle flow patterns that has been assumed.

A model of the driving mechanism where viscous drag forces are assumed to balance the torque on each plate due to symmetric forces at ridges and
continental convergence zones and slab forces which act only on the plate with subducted lithosphere produces stresses in good agreement with the data for most continental regions. The fit between calculated and observed stresses for this model is relatively poor for most oceanic regions, especially near subduction zones. This model suggests, however, that it is probably an oversimplification to assume that the net force exerted by the slab acts symmetrically about the plate boundary. While the role of drag forces in the driving mechanism remains poorly constrained, the finite element technique that has been developed may be applied in the future to any specific and realizable flow pattern that is proposed for the mantle.

Thesis Supervisor: Sean C. Solomon
Associate Professor of Geophysics
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CHAPTER 1: INTRODUCTION AND POSSIBLE PLATE DRIVING FORCES

SECTION 1.1 INTRODUCTION

The driving mechanism for plate tectonics is still largely unknown, more than three hundred years after observation of the similarity of Atlantic coastlines by Francis Bacon laid the foundation for the idea of continental drift. It is ironic that the now widely accepted theory of plate tectonics, so powerful in its ability to predict plate kinematics, is still without a well defined cause. The intent of this thesis is to investigate the relative importance of various potential components of the total driving mechanism. The approach adopted is to treat the lithosphere as an elastic body, capable of deformation in response to plate driving forces that act on the edge and along the base of the plate. The calculated intraplate stress field resulting from various assumed models of the driving mechanism is compared against available data on the state of stress in the plates to put constraints on possible driving mechanisms. Before considering the deformational approach chosen for this thesis, other approaches to studying the driving mechanism will be reviewed.

One approach to studying the driving mechanism is to consider the plates as rigid bodies in mechanical
equilibrium. The velocities of the plates relative to one another and the absolute velocities of the plates with respect to a fixed reference frame in the earth can then be related to the forces acting on the plates.

The relative velocities and geometries of most of the plates are reasonably well known (Minster and Jordan, 1978; Chase, 1978a). Making some assumptions about the dependence of various possible driving forces on the velocities and geometries of the plates allows the inverse problem of finding the relative strength of the various driving forces to be solved (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977). The actual quantity that is minimized in the inverse problem is the net torque acting on each plate. One limitation of this approach is that forces such as normal forces across transform faults and tangential forces across subduction zones are ignored. These forces are not simple functions of boundary type, but depend on the global state of stress. Even ignoring these forces, however, the net torque balance on each plate is small if the major forces acting on the system are due to gravitational sinking of the slab and resistance to the slab entering the mantle (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977). These two forces are an order of magnitude greater than any other force acting on the system. The net pull exerted on the surface plates by the sinking slab is the
sum of these two opposite, but nearly equal, forces. The net pull is on the same order as other forces acting on the plates.

Absolute plate velocities with respect to a presumably fixed lower mantle can be calculated from the relative velocities of the plates and the condition that the net torque exerted on the lithosphere is zero (Solomon and Sleep, 1974; Solomon et al., 1975). The torques applied to the plates are again assumed to depend on various possible driving and resisting forces such as the pull of subducted lithosphere, the push of ridges, drag between the lithosphere and the asthenosphere, and drag forces resisting horizontal translation of slabs. Balancing the net lithospheric torque instead of balancing the torque on each plate avoids the problems associated with normal forces across transform faults and tangential forces across subduction zones. Symmetric forces such as the push at ridges will exert no net torque on the lithosphere regardless of the magnitude of the forces. These forces cannot be resolved using this method.

The absolute velocities of the plates are similar for a wide range of possible driving forces (Solomon and Sleep, 1974; Solomon et al., 1975). Thus, the absolute velocities of the plates cannot be used to discriminate between the various possible driving forces. One
interesting aspect of the absolute plate velocities that may bear on the driving mechanism is the observation that ridges must migrate with a wide range of velocities with respect to the underlying source of new lithosphere (Solomon et al., 1975). This suggests that ridges are not surface expressions of the upwelling limbs of convection cells that have long term spatial stability in the mantle, as might be expected at trenches.

Another approach to the driving mechanism is to treat the mantle as a viscous fluid and study the relationship between plate motions and mantle flow. Convective flow may be important in driving the plates, especially for slow moving plates without subducting slabs (Richter, 1973; 1977; McKenzie et al., 1974). Shear stresses at the base of the plates due to convection may be important as a balancing term for other driving forces. Assuming that drag on the base of the plates balances other driving forces permits the calculation of a possible flow pattern in the mantle (Davies, 1978). The flow pattern in the mantle may be dominated by a counterflow associated with the transfer of mass in the mantle from trench to ridge (Chase, 1978b; Hager and O'Connell, 1978). Various assumptions about mantle rheology and possible interactions between plate motions and mantle flow lead to very different flow patterns in the mantle. Considerable insight into the driving mechanism will be
gained if it proves possible to determine the flow pattern in the mantle.

The approach to the driving mechanism adopted in this thesis is based on the premise that the plate will deform in response to forces acting on the plates. The deformation of the plates is reflected by the state of stress in the interior regions of the plates away from plate boundaries. In the first part of this thesis, the intraplate stress field inferred from seismic data, in-situ stress and strain measurements, and stress-sensitive geologic features will be presented.

Once the intraplate stress field has been inferred and a mechanical model of the lithosphere has been assumed, the stresses due to various possible driving forces can be calculated. In the second part of this thesis finite element numerical techniques are applied to the solution of equilibrium equations for membrane stresses in a thin, elastic shell approximation of the lithosphere. Comparing the inferred and calculated intraplate stress fields proves to be a powerful test of various possible driving and resisting forces.

The results of the modeling may be summarized as follows: ridge pushing forces are required of all models that match the inferred intraplate stress field. The net pulling force exerted on the plates by subducted lithosphere is at most a few times greater than other
forces acting on the plates, even if net slab forces decrease with increasing subduction rate to model resistive forces that depend on velocity. Forces acting at continental convergence zones that resist further convergence are important for models of the intraplate stress field in Europe, Asia, and the Indian plate. Viscous drag at the base of the plate is best modeled as a resistive force, with a drag coefficient that is non-zero beneath oceanic lithosphere, but which may be concentrated by a factor of ten beneath continental lithosphere. Models of the driving mechanism in which drag forces drive the plates or balance the torque on each plate due to boundary forces are in poor agreement with the observed data for most regions.

In the remainder of this chapter, I consider the potential of various forces to drive and stress the plates. These forces include a pull by subducted lithosphere on the surface plates, ridge pushing forces, and drag forces at the edge and along the base of the plate. Forces unrelated to plate motions that may stress the plates are also considered at the end of this chapter.

In Chapter 2, the techniques for inferring the intraplate stress field are reviewed and applied to a regional analysis of intraplate stresses. A global summary of the intraplate stress field is presented at the end of this chapter.

In Chapter 3, seismically determined apparent stress
and stress drops for intraplate stresses are presented. Inter- and intraplate seismic environments are compared and limits on the absolute level of ambient tectonic stresses are considered.

In Chapter 4, a finite element method applicable to analysis of intraplate stress for a single plate is developed. Intraplate stresses in the Nazca plate due to driving forces and large plate boundary earthquakes are presented.

The finite element method developed in Chapter 4 is extended to the calculation of the global intraplate stress field due to possible driving mechanisms in Chapter 5. The results of preliminary finite difference modeling of intraplate stresses are summarized and provide an introduction to the more refined finite element modeling that follows.

In the last chapter the results and conclusions of this thesis are summarized.

The potential of various forces to drive and stress the earth's lithospheric plates will now be considered.
SECTION 1.2 RIDGE FORCES

Forces due to horizontal variations in density beneath spreading centers may contribute to the driving mechanism. Typical ridge crest depths below sea level are on the order of 2.5 km, or about 3 km above the deep ocean basin floor. The ridge crest, according to the theory of plate tectonics, is supported by the upwelling of warm, low density mantle material. High heat flow and the lack of a major gravity anomaly across the ridge support this hypothesis. The source of potential energy at the ridge is excess heat in this upwelling zone. In this sense, any gravity forces due to density contrasts are thermally driven. In the old ocean basins, the lithosphere is cold and has a higher density than beneath the ridge. Thus, perpendicular to the ridge, there exists a horizontal density gradient. Horizontal forces arise that, with certain assumptions, are proportional to a vertical integral of stress differences within the plate due to the density variation. These forces, when averaged across that portion of the plate which can support stress, will produce a compressive state of stress in the plates and may be sufficient to push the plates away from the ridge toward the trench. An estimate of such forces is needed to test the potential of ridge forces to drive the plates.

Consider an idealized oceanic plate shown in Figure
1.1. With certain assumptions, the horizontal force per unit length of ridge is equal to an integral over depth of the difference between the lithostatic pressure beneath the ridge and beneath the old ocean basin. The depth over which the integral must be evaluated depends on the depth in the earth to which stress differences can be supported. This depth is known as the compensation depth. Near-zero free-air gravity anomalies across the ridge (Talwani et al., 1965) imply that the ridge is isostatically compensated at some depth. The compensation depth \( d \) is taken as the base of old oceanic lithosphere. Below this depth flow in the asthenosphere presumably will relax stress differences. The horizontal force due to the density variations normal to the ridge can then be expressed as

\[
F = \int_{0}^{d} P^R(z) - P^B(z) \, dz
\]

(1.1)

where \( P^R(z) \) and \( P^B(z) \) are the lithostatic pressures beneath the ridge and old ocean basin, respectively, and the integral is evaluated between sea level and the compensation depth \( d \).

The lithostatic pressure is assumed to be due to the weight of the overlying material, and can be expressed as

\[
P(z) = \int_{0}^{z} \rho(z')dz'
\]

(1.2)
where \( \rho(z') \) is the density and \( g \) is the gravitational acceleration, taken as a constant equal to 980 cm/sec\(^2\).

An idealized density profile for an old ocean basin is shown in Figure 1.1a, where it is assumed that the density of sea water is 1.03 g/cm\(^3\), the density of the crust is 2.8 g/cm\(^3\) and the density of cold lithosphere is 3.3 g/cm\(^3\). At the ridge, the density of sea water and the crust are taken as equal to the values for the old ocean basin. The density for the asthenosphere beneath the ridge is not a free parameter, but is determined by the condition of isostasy. Assuming Pratt isostasy, the density of the asthenosphere beneath the ridge is determined by the weight of a column of material in the old ocean basin. This is

\[
\int_0^d g \rho_R(z) \, dz = \int_0^d g \rho_B(z) \, dz \quad (1.3)
\]

where \( R, B \), refer to the ridge and old ocean basin, respectively. Assuming a compensation depth of 100 km and a crustal thickness of 7 km, the density of the asthenosphere beneath the ridge is 3.225 g/cm\(^3\). This idealized density profile for a ridge is also shown in Figure 1.1a.

Using the density profiles above leads to a value for the horizontal force per unit length of ridge in equation (1.1) of \( 3.3 \times 10^{15} \) dyne-cm. If this force can be supported across the entire 100 km thick oceanic
plate, it is equivalent to a compressive stress of 330 bars.

The horizontal stress derived above acts on a vertical plane at a distance away from the ridge beyond which the horizontal density variation is zero. The ridgepushing force decreases closer to the ridge until finally at the ridge crest the force is zero. Normal faulting earthquakes along the ridge crest result from upward relative movement of crustal blocks as the sea floor spreads from the axial valley (MacDonald and Atwater, 1978), perhaps in response to the lessening influences of viscous resistance to axial upwelling (Sleep, 1969; Sleep and Biehler, 1970). Normal faulting earthquakes along the ridge crest are entirely consistent with ridge-induced compressive stresses of several hundred bars away from the ridge. In-situ hydrofracture data along the landward extension of the Reykjanes ridge in Iceland indicate maximum non-lithostatic horizontal stresses on the order of a few tens of bars oriented approximately parallel to the ridge crest (Haimson and Voight, 1977). These authors believe that the stresses are related to thermoelastic processes, and conclude that stresses "associated with the driving mechanism of plate tectonics apparently do not dominate the observed stress pattern" (Haimson and Voight, 1977). Thermoelastic stresses on the order of tens of bars may well dominate the stress field very near the ridge.
Compressive stress of several hundred bars due to ridge elevation may dominate the stress field, however, in the interior regions of the plates.

In deriving the forces due to the ridge, it has been assumed that the problem is two-dimensional and that the shear stresses are small compared to the other two components of stress. The first assumption is an oversimplification for the earth, especially along ridge segments immediately adjacent to transform faults. The stress field in such areas is probably complicated. Away from these regions, however, the two-dimensional assumption is probably acceptable.

The second assumption implies that the lithosphere is free to slide over the mantle. Viscous drag between the plate and the asthenosphere and frictional stress on transform faults are assumed to be small. If the plates are driven from below by mantle convection, this condition obviously fails. If the plates are moving at a velocity of a few cm/yr with respect to a passive asthenosphere with a viscosity of $10^{20}$ poise, the viscous drag will be of the order of a few bars. This may be significant if averaged over the length of the major plates and would reduce the compressive stress that the ridge could exert on the plate. If shear stresses on transform faults are as high as several kilobars, the compressive stress due to the ridge could be reduced considerably. The nature
of the drag forces on the base of the lithosphere and shear stresses on transform faults are still the subject of considerable debate and will be considered in more detail later in this chapter. For a first order estimate of the potential stress due to the ridge, drag forces and shear stress along transforms are assumed to be small.

The forces associated with the ridge result from horizontal density contrasts. Most spreading centers have a depth below sea level of about 2.5 km, with the variation between fast and slow spreading centers less than a few hundred meters (Sclater et al., 1971). The forces due to the ridge are thus independent of spreading velocity to a first order approximation.

To summarize, an estimate of the force exerted on the plate per unit length of ridge is $3.3 \times 10^{15}$ dyne-cm. Averaged across a 100 km thick lithosphere, this is equivalent to a compressive stress of 330 bars. The ridge forces considered above are consistent with estimates of potential forces due to ridges by other authors (Hales, 1969; Frank, 1972; McKenzie, 1972; Artyushkov, 1973; Bird, 1976). In the next section potential forces due to sinking slabs are considered and found to be an order of magnitude greater than potential ridge forces.
SECTION 1.3 SLAB FORCES

One of the basic premises of plate tectonics is that oceanic lithosphere is consumed beneath the trench at subduction zones. Earthquake hypocenters often define a nearly linear trend extending to depths of at least 600 km (Isacks and Molnar, 1971), suggesting a high strength zone capable of supporting significant stresses. This zone is usually assumed to be the oceanic lithosphere that has rigidly entered the mantle as part of the process of plate motions. Oceanic lithosphere has a low thermal conductivity and will remain cooler than the surrounding mantle for an appreciable length of time for plate velocities on the order of a few cm/yr (Richter and McKenzie, 1978). There are many models of the thermal structure of the subducted slab (McKenzie, 1969a; Turcotte and Schubert, 1971; Toksoz et al., 1973). All of the models predict that the slab will remain several hundreds of degrees Celsius cooler than the surrounding lithosphere to depths of at least 400, and perhaps 700, kilometers.

Oceanic lithosphere is assumed to be roughly of the same composition as the mantle from which it is formed. The primary difference between oceanic lithosphere and upper mantle material is thus temperature. Subducted lithosphere, cooler than the surrounding mantle, will be
denser and will exert a negative buoyancy force on the system. If the slab is capable of transmitting this force to the overriding surface plate, it may be the dominant driving force for the plate tectonic driving mechanism. It will be shown below that the slab has the potential to exert a pulling force equivalent to a deviatoric tensile stress of several kilobars across a 100 km thick plate. Resistive forces may exist, however, that would reduce the potential of the slab by at least an order of magnitude.

To estimate the maximum value for potential slab forces, the following assumptions are made:

1) The slab can be considered as a two-dimensional system, so that there is no density variation along the strike of the trench axis.
2) Density variations within the slab are purely due to the thermal regime.
3) There is no shear resistance as the slab enters the mantle.

The first assumption ignores possible geometric effects at the edges of the trenches such as in the Fiji area (Isacks and Molnar, 1971). However, for many areas, this assumption is probably accurate. The second assumption implies that the density of the slab and mantle are equal at equal temperatures.

The final assumption permits the calculation of the
driving force as a vertical integral of stress, and hence, density differences within the slab. Viscous forces along the slab could increase the pull on the surface plate if mantle convection is faster than the slab is descending. Conversely, the effective pull of the slab may be reduced if there is viscous resistance to the slab entering the mantle. Resistive drag may significantly reduce the potential of the slab to pull the oceanic plate. The role of resistive drag forces is discussed later in this section.

Consider the idealized cross-section of a subducted slab in Figure 1.2. The z-axis is positive downward and the x-axis is parallel to the surface plate. The density is greater in the cooler slab, and hence the net excess weight per unit length across the thickness of the slab is given by

$$\Delta W = \int_A g \Delta \rho(x,z) dA \quad (1.4)$$

where the integral is evaluated over the cross-sectional area of the slab and $\Delta \rho(x,z)$ is the difference in density at any position between the slab and the mantle. Letting the slab thickness be $l$, the depth to which the slab penetrates be $h$, and the dip angle of the slab be $\phi$ leads to an approximate relation for $\Delta W$ given by

$$\Delta W \approx lh \csc(\phi) g \langle \Delta \rho \rangle \quad (1.5)$$
where $<\Delta \rho>$ is the average density difference between slab and mantle.

The force due to gravity acts in the $z$-direction. If this were the only force in the system, the slab should have a dip of $90^\circ$ and no horizontal force would be transmitted to the upper plate. Most slabs have dips around $45^\circ$ (Isacks and Molnar, 1971). If the slab is in static equilibrium, this implies a force normal to the dip of the slab that supports the slab. In this case, there is a component of force $D$ per unit length acting down-dip along the slab given by

$$D = \Delta W \sin \phi = \ell hg <\Delta \rho>$$ (1.6)

The force $D$ has a component in the horizontal direction given by

$$H = D \cos \phi = \ell hg <\Delta \rho> \cos \phi$$ (1.7)

where $H$ is the total horizontal force per unit length acting on the upper plate due to the excess mass of the slab.

It is now necessary to estimate the density difference between the slab and mantle. Assuming that the density differences are due totally to thermal contraction,
\[<\Delta \rho> = \alpha \rho_o <\Delta T> \quad (1.8)\]

where \(<\Delta T>\) is the average temperature difference between the slab and mantle, \(\alpha\) is the coefficient of thermal expansion for the slab, and \(\rho_o\) is the average density of the slab. Taking \(\rho_o\) as 3.3 g/cm\(^3\), \(\alpha = 3 \times 10^{-5}/^{\circ}\)K (Richter and McKenzie, 1978), and estimating the average temperature difference of several thermal models (McKenzie, 1969; Smith and Toksöz, 1971; Toksöz et al., 1973) to be approximately 500°C,

\[\Delta \rho \approx 0.05 \text{ g/cm}^3 \quad (1.9)\]

Substituting this value into equation (1.7) with \(h \approx 600\) km, \(\lambda \approx 100\) km, and \(\phi = 45^{\circ}\), gives

\[H = 2 \times 10^{16} \text{ dyne/cm} \quad (1.10)\]

Averaged across a 100 km thick plate, this is equivalent to a deviatoric tensile stress of 2 kb.

McKenzie (1969a) calculates the theoretical thermal structure of the slab and discusses the slab pulling force. Integrating the thermal equations throughout the slab instead of taking average temperature contrasts to obtain the densities, he obtains a deviatoric tensile stress of 1.8 kb for a similar dip angle. Turcotte and
Schubert (1971) obtained an estimate of approximately $10^{17}$ dyne/cm for $H$ using somewhat higher values for the temperature contrast and slab thickness. These estimates are on the same order as the value of $H$ in equation (1.10).

Phase transformations may affect the density distribution in the slab. The olivine-spinel transition, which would normally occur at a depth of approximately 400 km, may be elevated about 100 km in the cool slab (Turcotte and Schubert, 1971; Solomon and Paw U, 1975). In this case, the additional downward force per unit length is given by

$$T = g \Delta \rho A$$

(1.11)

where $\Delta \rho$ is the density difference between olivine and spinel and $A$ is the added area of the new, high density spinel phase. Assuming $\Delta \rho \approx 0.3$ g/cm$^3$ (Ringwood and Major, 1970) and $A \approx 7 \times 10^{13}$ cm$^2$ for the triangular wedge shown in Figure 1.2, gives

$$T \approx 2 \times 10^{16} \text{ dyne/cm}$$

(1.12)

Assuming, again, that the slab is in static equilibrium, this implies a horizontal pulling force given by

$$H_T \approx 1 \times 10^{16} \text{ dyne/cm}$$

(1.13)
for a slab dipping at 45°. This value is about a factor of two less than the slab pulling force due to thermal compression.

Phase transitions in the mantle at a depth of about 650 km between spinel and post-spinel phases may also be important (Anderson, 1967). There are several possible phase transitions in the spinel to post-spinel group. It is difficult to estimate the effect of these phase transitions on slab forces because it is not known whether the phase boundary is elevated or depressed in the slab (Turcotte and Schubert, 1971).

A first order approximation of the total potential force available to pull on the oceanic plate per unit length of slab is $3 \times 10^{16}$ dyne/cm. If this force can be supported across a 100 km thick plate, it is equivalent to a deviatoric tensile stress of 3 kb. This estimate of the force due to the slab is an order of magnitude greater than the force estimated for the ridge.

Resistive forces that were ignored in the derivation of the potential slab forces, however, may significantly reduce the pull exerted on the oceanic plate. Focal mechanisms for very deep earthquakes associated with the slab indicate compression parallel to the dip of the slab (Isacks et al., 1968; Isacks and Molnar, 1971). The slab may encounter resistance at this depth in the form
of a boundary or barrier to penetration (Richter, 1973; Forsyth and Uyeda, 1975). Viscous drag between the slab and the mantle may further reduce the slab forces. Smith and Toksöz (1972) compute the distribution of stresses in the downgoing slab as a function of gravitational forces, phase transformations, and shear resistance between the slab and the mantle. They found that the drag between the slab and mantle may be nearly as large as the gravitational force associated with the sinking slab.

A major resisting force may be a large non-hydrostatic pressure acting on the base of the subducted slab (Richter, 1977). This force increases with depth and may account for the down-dip compression mechanisms of very deep earthquakes.

The topography of trench-outer rise systems has been explained in terms of kilobars of horizontal compression (Hanks, 1971) and large bending moments (Caldwell et al., 1976) at subduction zones. High horizontal stresses are not required if oceanic lithosphere is modeled as visco-elastic (Lé Bremaecker, 1977). Neither high horizontal stresses nor large bending moments are required if oceanic lithosphere is modeled as a two layer plate with an elastic-plastic rheology (Forsyth and Chapple, 1978).

Another resistive force arises from shear stresses acting on the thrust faulting defining the plate boundary at the trench. Earthquake strain release for the earth
is concentrated along these thrust faults (Kanamori, 1977), indicating that there is significant resistance to subduction of the oceanic plate. The magnitude of this force depends upon the magnitude of the shear stress acting on the fault and the angle of underthrusting (Richter and McKenzie, 1978). If the shear stress acting on a thrust fault 200 km long with a dip of 20° is 200 bars, the resistive force per unit length of the thrust fault is approximately $1 \times 10^{16}$ dyne/cm. This force acts in the direction opposite to the slab pull and may reduce the effective pull on the oceanic plate by a factor of two. If the shear stresses on the thrust faults are higher, the effective pull may be further reduced.

It was assumed in the calculation of potential slab forces that the slab acts as an efficient stress guide. There is evidence that major earthquakes such as the Sanriku earthquake beneath Japan on 2 March 1933 and the Rat Island earthquake on 4 February 1965 may completely sever the slab (Kanamori, 1971; Abe, 1972). In this case, the force exerted on the surface plate is further reduced.

Using the observed plate motions and geometries, Forsyth and Uyeda (1975) concluded that the pulling and resisting forces of the slab are of comparable magnitudes. The difference between these two forces is the net force that the slab exerts on the surface plates. The
potential of slab gravitational forces is estimated to be $3 \times 10^{16}$ dyne/cm per unit length of slab. The net pull may be an order of magnitude less.

The magnitude of the net slab force probably depends on the rate of subduction. The component of the net pulling force due to the density of the slab is independent of velocity as long as the subduction rate is at least 3 cm/yr so that the slab stays colder than the surrounding mantle to several hundred kilometers depth (Richter and McKenzie, 1978). The resistive forces depend on the viscosity of the mantle and the subduction rate. There may also be a dependence on the age of the subducted lithosphere. Young subducted lithosphere may be warmer than old subducted lithosphere and hence the density contrast between slab and mantle may be reduced. This will reduce the net pulling force exerted by the slab. Thus, the net pull due to the slab consists of several parts, one of which is roughly independent of velocity and at least one which depends on the rate of subduction.

Both the ridge and slab forces are assumed to act in the direction of relative plate motion. If the plates are assumed to be in a state of static equilibrium, there must be other forces which resist plate motions. Possible forces include shear stresses on transform faults and viscous drag on the base of the lithosphere. The
role of transform faults will be considered later in this chapter. Drag on the base of the plate may drive or resist plate motions, depending upon the nature of flow in the mantle. Drag forces are considered in the next section.
SECTION 1.4 DRAG FORCES

Viscous forces will exist at the base of the plates if the lithosphere is in motion with respect to the asthenosphere. The magnitude of such forces will depend upon plate velocity and asthenosphere viscosity and the directions will depend upon whether the mantle flow is faster or slower than plate velocities. The mantle may be convecting and dragging the plates along as the top of convection cells, or may be passively resisting the motion of the plates. These two cases of drag forces are treated separately in the next two sections.

1.4.1 Driving drag

The speculation by Wegener (1912) that Africa and South America had once been a single continent led to investigation of possible sources of energy to explain the separation. Convection in the earth was suggested early (Holmes, 1928, 1929) and is an appealing source of energy because shear stress is applied over large areas along the bases of the plates. This overcomes the objection to edge forces at plate boundaries that must somehow be transmitted across thousands of kilometers without destroying the plate.

The flow structure of the mantle is the subject of considerable debate. There must be a flow of mass in the mantle to balance the displacement of material as a plate
moves from the ridge to the subduction zone. Whether this flow is restricted to the uppermost 600 km or so (McKenzie and Weiss, 1975) or whether it occurs across the entire mantle (Sammis et al., 1977) is not known. There is no consensus about a form of mantle convection that could drive the plates, although several forms have been hypothesized.

Rayleigh-Benard convection was one of the first forms of mantle flow suggested as a driving mechanism (Holmes, 1928). This form of convection takes place in a fluid layer that is heated from below. The fluid is assumed to have a uniform viscosity. The convection pattern consists of cells where the horizontal scale is comparable to the depth of the fluid layer. For the earth, the length scale of the plates varies by an order of magnitude between plates such as the Cocos and Pacific. If the depth of the convecting layer is constrained to be above 650 km, this implies that a large plate such as the Pacific plate must span many convection cells. Rayleigh-Benard convection cells are also periodic. For two neighboring cells, the flow direction will be clockwise in one and counterclockwise in the other. Thus, even if the flow of one cell can exert a drag on the base of a plate, the net drag on a large plate such as the Pacific plate will be small due to cancellation by neighboring cells. Further, an assemblage of cells cannot account for transfer of mass back from the trench
to the ridge. For these reasons, simple Rayleigh–Benard convection is an unacceptable driving mechanism (Richter, 1973).

More sophisticated convection models have included the effects of temperature-dependent viscosity, internal heating, and complicated geometries associated with the subducted slabs (Turcotte et al., 1973; McKenzie et al., 1974; Richter, 1978; Richter and McKenzie, 1978). These models are all two-dimensional in nature, and while they may give some information about the effects of various parameters, they cannot predict the flow pattern for a three-dimensional earth.

Plumes, or "hot spots", are another form of convection that has been proposed for the mantle (Morgan, 1971, 1972). Plumes in the middle of the plate cannot drive the plates by drag forces. Along ridges, the shear stresses associated with the plume would act on two plates and could drive the plates apart in this case. This may be important in places such as Iceland, but globally, plumes probably do not significantly contribute to the driving mechanism.

A major objection to convection as a driving mechanism is the amount of time required for the flow pattern to change. It is difficult to explain changes in spreading rates and directions that occur on the order of a few million years (e.g., Sclater and Fisher, 1974) with convection when estimates of the response time for
convection vary from at least tens of millions of years to the age of the earth (Richter, 1973).

The potential of convection to drive or stress the plates is difficult to estimate because the flow pattern within the mantle is unknown. If the flow pattern in the mantle can be established, the role of convection as a driving force may be estimated.

It should be noted at this point that the source of energy to drive the plates is almost surely excess heat in the earth. Whether the plates are driven by density contrasts at the ridge or in the slab, the energy source is still thermal. In this sense, ridge and slab forces are a form of thermal convection. However, for the purposes of clarity, the term "convection" as applied to the driving mechanism will refer only to tractions applied to the base of the plates by mantle flow capable of exerting a net force on the plate in the direction of plate motion.
1.4.2 Resistive drag

If the lithosphere is moving with respect to a passive asthenosphere there will be a viscous force resisting plate motion. The magnitude of the drag force will depend upon the velocity contrast between plate and asthenosphere and a law governing viscosity and strain rate \( \dot{\varepsilon} \). In the simplest case, viscosity is independent of strain rate and the drag force is linearly proportional to the velocity. If viscosity is dependent on temperature, shear heating due to the moving plate may reduce the viscosity and reduce the drag force. The exact form of the relationship between viscosity and strain rate for the mantle is unknown, although some form of a power law dependence seems likely (Stockert and Ashby, 1973).

A linear drag law may be considered for a first order estimate of the potential of drag forces to resist plate motions. A plate moving with respect to the asthenosphere with velocity \( V \) will have drag stress on the base of the plate given by

\[
\tau \sim \frac{\eta V}{h_a}
\]

(1.14)

where \( h_a \) and \( \eta \) are the thickness and viscosity of the asthenosphere, respectively. Taking \( h_a = 3 \times 10^7 \) cm, \( V = 10 \) cm/yr, and assuming an asthenospheric viscosity of approximately \( 10^{20} \) Poise (Cathles, 1975) gives a
shear stress of 1 bar. For a 100 km thick plate with a length of 5000 km, this is equivalent to a force of $5 \times 10^{14}$ dyne/cm per unit width of plate normal to the direction of absolute plate motion. This estimate is large enough that drag forces, when averaged over plate dimensions, can balance plate accelerations due to ridge pushing and slab pulling forces.

The estimate of drag forces depends on the viscosity of the asthenosphere, which may vary spatially. Specifically, drag forces beneath continents may be higher than beneath oceans because of a possible high viscosity layer beneath the continents (Jordan, 1975). Similarly, the viscosity beneath old ocean basins may be higher than the viscosity beneath young ocean floor (Chapman and Pollack, 1975).

Drag forces may play a significant role in the driving mechanism by providing a source against which ridge and slab forces, both assumed to act in the direction of plate motion, can do work. Both driving and resisting drag forces depend critically on the flow pattern in the mantle. Unfortunately, this flow pattern is poorly known. The flow pattern may be dominated by the flux of material from the trench to the ridge (Hager and O'Connell, 1978; Chase, 1978b). Some form of counterflow in the mantle is required to balance the transfer of mass from the ridge to the trench. Chase (1978b), assuming that subduction
zones and ridges act as sources and sinks, respectively, of material for the mantle, calculates the counterflow for simple mantle rheologies. This counterflow may dominate the flow pattern in the mantle. For most plates the calculated shear stresses applied to the base of the plate due to net mantle flow are nearly parallel and opposite to the direction of absolute plate motions, although one notable exception is North America (Chase, 1978b). The drag at the base of the plate is probably not a simple function of absolute plate velocities, but depends on the poorly known flow pattern in the mantle.

Another source of resistance to plate motions may be high shear stresses on transform faults. The relationship of transform faults to the driving mechanism is considered next.
SECTION 1.5 TRANSFORM FAULTS

Shear stress on transform faults separating two plates will resist the relative motion of the plates. The seismic activity observed on transform faults is a clear indication that transform faults resist plate motion. The magnitude of that resistance, however, is less clear.

Crucial to the role of transform faults in the driving mechanism is the level of the shear stress acting on the fault. The resistive force per unit length of transform with a shear stress $\tau_{TR}$ is given by

$$F_{TR} = h\tau_{TR}$$  \hspace{1cm} (1.15)

where $h$ is the depth over which the shear stress acts. If that depth is equal to 100 km, and the shear stress is 100 bars, then $F_{TR} \sim 10^{13}$ dyne/cm. When the focal depth of transform earthquakes can be constrained, depths are less than 10-20 km (Brace and Byerlee, 1970; Burr and Solomon, 1978). If all of the transform resistance is concentrated in this brittle zone, $F_{TR} \sim 10^{12}$ dyne/cm for a shear stress of 100 bars. The lack of earthquakes along transforms at depths greater than 10-20 km probably does not imply that the shear stress vanishes. More likely, some form of creep accounts for the non-seismic behavior of the plates as they move with respect to one another at depth. If the shear stress acting on the
fault is considerably greater, say 5 kb, then the resistive force at transforms may be as high as \(10^{15}\) dyne/cm. If similarly high shear stresses exist along major thrust faults at subduction zones, forces due to ridge pushing and slab pulling are insufficient to drive the plates and driving mantle convection must be involved (Hanks, 1977). Shear stresses associated with mantle convection that can balance the torque due to possible driving forces and large resistive forces have been calculated (Davies, 1978). These basal shear stresses, on the order of a few bars for the larger plates, may either drive or resist plate motion, depending upon the other forces acting on the plate.

The question of the level of shear stress on the transform faults, and hence the level of the ambient tectonic stress, is thus central to the role of transform faults. Laboratory measurement of the shear strength of rocks at pressures and temperatures corresponding to crustal depths indicates that several kilobars of shear stress are required to produce failure of samples, even when pre-existing faults exist (Brace and Byerlee, 1970; Stesky et al., 1974). The requirement of high shear stresses is relaxed if the effective stress is lowered by high fluid pore pressures. Laboratory measurements do not require that there be kilobar stresses acting on transform faults.

Low shear stresses on transform faults are indicated
by heat flow measurements across the San Andreas fault (Brune et al., 1969). These measurements, along with a model of shear heating, indicate a maximum shear stress of a few hundred bars acting on the fault.

Seismic stress drops for major earthquakes are typically on the order of 10 to 100 bars (Kanamori and Anderson, 1975). Unfortunately, the relationship between stress drop and ambient tectonic stress is not clear, as will be discussed in Chapter 3. These values for stress drop place a lower limit of 10 to 100 bars on the ambient tectonic stress, but cannot be used to determine whether 100 bars or kilobars act on transform faults.

The orthogonal relationship between transform faults and spreading centers has been used to conclude that resistance to motion along transforms is less than the resistance to plate separation at spreading centers (Lachenbruch and Thompson, 1972). To the extent that ridges are the site of upwelling, warm, soft material, this implies that the shear stresses on transforms are small. The actual resistance to spreading at the ridge axis is unknown, however, and low shear stresses on transforms are not required.

In-situ measurements of strain and stress in the earth's crust indicate non-lithostatic stresses of bars to hundreds of bars (Ranalli and Chandler, 1975; Haimson, 1977). To date, there are no in-situ measurements at depths of 5-10 km along major faults that could answer
the question of shear stress levels on the faults, although the importance of such measurements has been emphasized (Hanks, 1977).

One argument in favor of low shear stresses on transform faults is the inversion of observed plate motions and geometries by Forsyth and Uyeda (1975). They found that shear forces on transform faults were not an important part of the driving mechanism, principally because plate velocities do not depend on the length of transform for each plate.

The role of transform faults in the driving mechanism is not well known. A critical unknown is the shear stress acting on transforms and the ambient tectonic stress. If the forces due to ridges and slabs are each equivalent to stresses on the order of hundreds of bars, then resistive drag at the base of the plate equivalent to shear stresses of a few bars are sufficient to balance the forces acting on the plate. In this case, large resistance at transform faults are not required and the ambient tectonic stress should be on the order of hundreds of bars. Clearly, the level of the ambient tectonic stress is of major importance to the question of the driving mechanism.

Other forces besides plate driving forces may contribute to the overall stress field in the lithosphere. These forces are considered next.
SECTION 1.6 OTHER SOURCES OF STRESS IN THE LITHOSPHERE

Central to the idea that the intraplate stress field can be used to study the driving mechanism is the assumption that driving forces will stress the plates and that the effect of other, non-driving, forces can be removed or neglected. In the previous sections, forces at ridges and trenches and shear forces on the base of the plates and along transforms have been considered and all have the potential to stress the plates. Other sources of intraplate stress include thermal stresses associated with a cooling lithosphere (Turcotte and Oxburgh, 1973), stresses due to the motion of an elastic plate on an ellipsoidal earth (Turcotte and Oxburgh, 1973), crustal thickness inhomogeneities (Artyushkov, 1973), lithosphere loading due to glaciers, sea level changes, and volcanic constructs such as Hawaii (Walcott, 1970; Watts and Cochran, 1974), sedimentation and erosion (Voight, 1966; Haxby and Turcotte, 1976), and ancient tectonic events (Swolfs et al., 1974).

Tensile thermal stresses parallel to the ridge may arise as the plate moves away from the ridge and cools. This stress has been estimated to be potentially as large as 40 kb (Turcotte and Oxburgh, 1973). Thermal stresses must be relaxed by deformation on the grain size level because of anisotropic thermal expansion of mineral grains, a process that should also relax the larger scale
thermal stress without coherent lithospheric failure and large intraplate earthquakes.

The plates, as they move in a north-south direction, will experience a change in radius of curvature due to the ellipticity of the earth. The effect will be at a maximum at a latitude of ±45°, where estimates of the magnitude of tensile stresses at the edge of the plate due to ellipticity, based on the theory of thin shells, are as large as 6 kb (Turcotte and Oxburgh, 1973). Most of the earth's plates have velocities with large east-west components compared to north-south motions, which would minimize the effect of ellipticity. For northward plate velocities of a few centimeters a year, it would take about a hundred millions years for the plate to travel from the equator to 45°N. In this time, the plate may deform plastically and relieve these stresses. These membrane stresses may be important in Africa, where the East African rift system has been explained in terms of slow northward motion of the African plate over the last 100 million years (Oxburgh and Turcotte, 1974). Stresses due to ellipticity may be large, especially at mid-latitudes, but the uncertain nature of the relaxation process and the small north-south velocity of most plates may minimize the effects on the intraplate stress field for large areas of the earth's surface.

Horizontal variations in the thickness of the crust may result in stresses in the lithosphere. One form of
horizontal thickness variation has already been considered; namely, the ridge. Structures such as mountains, large plateaus, and continental shelves are other examples. Stresses arise even if the structures are isostatically compensated (Artyushkov, 1973). The horizontal variation in density between continental and oceanic lithosphere may produce deviatoric tensile stresses in the continents and deviatoric compressive stresses in oceanic lithosphere.

The magnitude of horizontal stresses due to crustal inhomogeneities is potentially as large as the load associated with the relief; that is, on the order of kilobars for mountains such as the Himalayas (Artyushkov, 1973; Bird, 1976). The net deviatoric compressive stress due to the Himalayas may be considerably less than kilobars if the effect of both the load associated with the mountain range and the horizontal density variation between continental and oceanic lithosphere are considered.

The addition and removal of mass from the surface of the lithosphere will cause stresses. Glaciation and corresponding changes in sea level periodically load and unload various portions of the lithosphere. A one km thick glacier places a load per unit area on the lithosphere of approximately $10^5$ g/cm$^2$. This is equivalent to a compressive stress of 100 bars at the base of the glacier. The mass of Hawaii is also a load on the lithosphere, and results in lithospheric flexure that indicates bending stresses of up to several kilobars (Walcott,
1970; Watts and Cochran, 1974).

Sedimentation and erosion may also produce stresses in the lithosphere (Voight, 1966; Haxby and Turcotte, 1976). For a Poisson's ratio of 0.25, the change in the horizontal stress at depth is one-third of the change in the vertical stress associated with the addition or removal of mass (Voight, 1966). For the case of erosion, this can lead to horizontal compressive stress on the order of hundreds of bars when several kilometers of crust have been removed. Competing against this effect, however, are thermal stresses associated with material cooling as it approaches the earth's surface. The effect may even dominate and the net result of erosion may be tensional stresses (Haxby and Turcotte, 1976). Considering the uncertainty of the effect of sedimentation and erosion, evidence for intraplate stresses due to the driving mechanisms should be selected from sites where these processes are not likely to be dominant.

There is evidence that strains may be locked into rocks after the forces producing the strains have ceased to operate (Swolfs et al., 1974). These strains, known as residual strains, may be the result of ancient tectonic processes and may mask strains due to present tectonic forces. The magnitudes of residual strains can be estimated using double overcoring techniques. In this manner, uncertainties in the origin of the stress
field may be minimized if residual strains are found to be small. Further, if the intraplate stress field is uniform over a large area including regions of very different geologic history, the effects of residual strains are probably minimal.

The techniques for inferring the stress in the lithosphere and the observed intraplate stress field are considered in the next chapter. Special emphasis is placed on those regions where stresses are most likely to represent the effects of plate driving forces, i.e., regions with a uniform stress pattern where the effects of the non-driving sources of stress discussed in this section may be minimal.
FIGURE CAPTIONS

Figure 1.1  a) Idealized density profiles beneath ridges and old ocean basins. The density of water and the uppermost crust are taken as 1.03 and 2.8 g/cm$^3$, respectively. The density of old oceanic lithosphere is taken as 3.3 g/cm$^3$ and the density of the asthenosphere is constrained by Pratt isostasy to be 3.225 g/cm$^3$.  b) Idealized cross-section of oceanic lithosphere between a ridge and an old ocean basin. $F_R$ is the horizontal force per unit length of ridge boundary.

Figure 1.2  Idealized cross-section of a subduction zone. $F_T$ is the potential force acting on the surface plate due to gravitational sinking of the slab. Shaded region represents additional portion of the slab in spinel phase due to the elevation of the phase boundary.
Figure 1.1
CHAPTER 2: OBSERVATION OF THE INTRAPLATE STRESS FIELD

In the first chapter the potential of various forces to drive and stress the plates were considered. In this chapter, common techniques used to infer the state of stress in the lithosphere are reviewed and a summary of intraplate stress data is presented.

Knowledge of the state of stress in the lithosphere is very important for the study of the driving mechanism for plate tectonics. Unfortunately, our knowledge of the state of stress is limited due to uncertainties in the methods used to infer stress and due to the uneven distribution of measurements. There are several areas on the earth's surface, however, where a consistent pattern for the orientation of principal stresses has been inferred using a variety of techniques. The methods used to infer the state of stress include fault plane solutions of large intraplate earthquakes, in-situ strain and stress measurements, and geologic features varying in size from $10^{-6}$ km to $10^2$ km.

The various techniques for inferring stress will be discussed in the next section. The intraplate stress field inferred from these techniques will then be summarized by region.
SECTION 2.1 TECHNIQUES FOR INFERRING STRESS IN THE PLATES

2.1.1 Fault plane solutions

Fault plane solutions of intraplate earthquakes provide information about the state of stress at the source. A fault plane solution is a procedure that maps onto a plane the first motion of P-waves from an earthquake as they are recorded at various seismograph stations. The distribution of first motions defines two orthogonal focal planes which are defined by the far field radiation pattern of a double couple source mechanism. One of the focal planes corresponds to the fault plane on which the earthquake occurred and the other focal plane is known as the auxiliary plane. Without other information such as aftershock locations or surface breakage, it is not possible to distinguish between the fault and auxiliary planes from a fault plane solution. Even with this uncertainty, a fault plane solution contains information about the orientation of the principal stresses at the source. If an earthquake occurs as a brittle fracture along a fresh, frictionless fault, the principal stress axes will bisect the angles between the two focal planes. The maximum and minimum compressive principal stresses will fall in the regions where the P-wave first motions represent dilatation and compression, respectively, at the station.

Laboratory experiments indicate that rocks will fail under stress along a plane that is inclined less than 45° to
the axis of maximum compression, depending upon friction in the failed plane (Byerlee, 1968). A wide range in the orientation of the principal stresses is possible if failure occurs along a pre-existing fault. In this case, the maximum compressive stress is only required to be located in the dilatational region. Coupled with the common uncertainty between fault and auxiliary planes, the error in assuming that the principal stress axes bisect the focal planes may be as large as $\pm 45^\circ$ (McKenzie, 1969b).

If a large intraplate region is characterized by consistent fault plane solutions, the uncertainties may be considerably smaller than the maximum of $\pm 45^\circ$. Earthquakes are assumed to occur on a plane which minimizes the resistance to failure. There are probably many pre-existing faults with a distribution of orientations in any large area. The inferred principal stress directions for earthquakes on pre-existing faults should reflect the ambient stress field for regions with a distribution of pre-existing faults. Thus over a large intraplate area with consistent fault plane solutions it is reasonable to assume that the effect of pre-existing faults may be minimized.

An isolated earthquake, even if the inferred principal stresses are correct, may not reflect the regional stress field. The local stress field that caused the earthquake may be due to heterogeneities in topography or loading history.
Most intraplate earthquakes have hypocentral depths of less than 30 km. While 30 km is deeper than most other techniques for inferring stress can sample, stresses inferred from intraplate earthquakes still may not represent the state of stress across the entire thickness of the plate. At trenches, for instance, the large bending moments associated with the downgoing slab may locally produce marked changes between the state of stress at the top and bottom of the plate (Engdahl and Scholz, 1977). For intraplate regions, away from topographic loads such as Hawaii, the bending moments are probably negligible (Turcotte, 1974). Thus, until techniques exist that can sample the state of stress across the entire plate thickness, it will be assumed that fault plane solutions represent the stress throughout the plate thickness.

Fault plane solutions provide information only about the orientation, and not the magnitude, of the stresses at the source. Other seismic parameters such as apparent stress and stress drop may contain some information about the magnitude of stresses; and will be discussed in detail in the next chapter.

Fault plane solutions for intraplate earthquakes provide most of the information about the intraplate stress field for large portions of the earth. For oceanic regions, fault plane solutions, along with some possible geological indicators, are the only source of data on the intraplate
stress field. In general, isolated intraplate earthquakes are assumed to represent at least the sense of the principal stresses. Unless a relationship to local conditions can be demonstrated, the orientation of the inferred stresses are assumed to be accurate to ±15° for regions where several fault plane solutions are consistent or where other techniques provide similar results.

2.1.2 In situ methods

There are two major types of in situ techniques used to estimate stress. The first, strain relief, uses the relaxation of a surface after stress has been removed to calculate the original strain. With additional information about the elastic properties of the sample, the stress can be determined. The second technique, hydrofracture, measures stress directly by applying pressure to a closed section of a borehole until the borehole fails.

Strain relief has been used extensively for many years, especially by engineers studying mines and tunnels. One common strain relief method is the CSIR 'door-stopper' technique (Leeman, 1971). This technique involves bonding a strain gauge rosette to the flattened end of a borehole. The borehole is overcored to relieve the ambient stress and strains are measured on the strain gauge. These strains are converted to stresses through Hooke's law after a section of core is returned to the lab for determination of the elastic parameters.
The assumptions involved in the 'door-stopper' technique are that the material being cored is isotropic, homogeneous and linearly elastic, and that one of the principal stresses is aligned with the borehole axis. The assumption about the principal stress direction can be removed if additional strain gauges are placed on the walls of the borehole. Then all six components of the strain ellipsoid can be calculated. The technique has several sources of uncertainty. If the technique is used in an anisotropic medium, such as bedded sediments, large errors are possible. The elastic parameters of the core are usually measured at zero pressure while the in situ stresses may be hundreds of bars. Errors will result if the elastic parameters are functions of pressure. If the overcoring process creates cracks, the technique will not be reversible and the linear elastic assumption will be violated. The magnitude of such errors are not known, but may be significant.

A major drawback of the 'door-stopper' technique, and of all other strain relief methods, is that the measurements must be made close to either the earth's surface or the surface of a mine wall. Sophisticated drilling techniques and instrumentation down borehole limit the length of the borehole to about 30 m. The stresses measured will include the effect of nearby mine cavities and local topography. While these effects are precisely what mine engineers wish to calcu-
late, the usefulness of strain relief data from mines for determining the regional stress field that may be tectonic in origin is questionable. However, if strain relief data are consistent over large regions, then the local effects are probably small compared to regional forces. Thus, caution should be used in the interpretation of isolated strain relief measurements.

Hydrofracture, unlike strain relief methods, measures stress directly. A sealed-off section of a borehole is hydraulically pressurized until the borehole fractures (Haimson and Fairhurst, 1967, 1970). Pressure at which fracture occurs, the pressure required to keep the fracture open, and the orientation of the fracture then give the minimum and maximum compressive stress in a plane perpendicular to the bore hole. In the theory relating these observations to the stresses, the effect of a hollow cylinder in an elastic medium and the effect of stresses on the orientation and internal pressure of opened cracks are calculated. The hydrofracture is assumed to open perpendicularly to the least compressive principal stress. The magnitude of this stress is a function only of the pressure required to keep the fracture open. The orientation of the fracture, and hence the orientation of the two principal stresses, is determined using either an impression packer or a down-hole camera. An impression packer is inserted down the hole and pressurized to make an impression of the fracture and is then returned to the
surface. The magnitude of the other principal stress, the maximum compressive principal stress, depends on the pressures required to initiate and maintain the fracture, as well as the pore pressure and the tensile strength of the rock. The effects of fluid permeating into the medium on the pore pressure may be included in the theory. The hydrofracture technique may then be extended to competent sandstones and shales.

The assumptions made in hydrofracture analysis are that the medium is linearly elastic, that rocks fail under brittle tension, and that one of the principal stresses is aligned with the borehole. The linear elastic assumption can hold only approximately in the earth; bedding, pre-existing faults, etc., will contribute to anisotropy. Laboratory studies of foliated and faulted samples indicate that under confining pressures common in the earth, fractures do not align with bedding or faulting, but align with the direction of the maximum applied stress, in agreement with the theory (Haimson and Avasthi, 1975). These experiments also show that the rocks failed predominately in tension, and not in shear. The assumption that one of the principal stresses aligns with the borehole, which is usually vertical, is difficult to test. When strain relief measurements enable the entire strain ellipsoid to be calculated, the most nearly vertical principal stress is often 20° or more from the vertical. This assumption is probably the weakest one in
terms of its effect on the measured stresses.

Hydrofracture has major advantages over strain relief methods in that elastic properties do not have to be measured and that the technique can be applied at great depth. In the Michigan basin, hydrofracture has been used to depths exceeding 5 km (Haimson, 1976). Thus hydrofracture is much less sensitive to the local topography that may mask tectonic stress near the surface. It should be noted, however, that 5 km is still relatively shallow compared to the thickness of the lithosphere and may not represent an average of the state of stress across the approximately 100 km thick lithosphere.

Interpretations of both strain relief and hydrofracture data require considerable care. As has already been mentioned local topography and loading histories may mask regional stresses, especially using strain relief data. Another source of uncertainty arises when both of the horizontal principal stresses are approximately equal. In this case, the azimuth of either stress is not well constrained. Often published data only include the average of the horizontal stresses, making it difficult to assess the reliability of such data.

Another major source of uncertainty is the effect of residual stresses. Residual stresses are stresses that are preserved in a medium after the sources causing the stress have been removed. Swolfs et al. (1974) have shown that
strain measured by overcoring a piece of quartz diorite over 2 m on a side did not agree with strains measured in the same piece of diorite after the length of the sides had been reduced. If the strains were a measure of a stress field that was constant over the length of the block, reducing the size of the block should have had no effect on the strains. They found that the orientation of the stresses varied by 55° after the block had been reduced in size. The effect of residual stresses on stress measurements may be to mask current tectonic stresses.

There are many uncertainties in the interpretation of stress measurements. Measurements are considered most reliable when they are consistent over large areas or where they are corroborated by other evidence. In such a case, the effect of local heterogeneities is probably small compared to the regional stress field. Fortunately, there are several large regions on the earth where a variety of techniques present a consistent picture of the state of stress in the lithosphere. Even when stress measurements are consistent on a regional scale, care and perhaps faith, must be used in inferring that the state of stress is due to plate driving forces.

2.1.3 Geologic methods

Geologic processes on a length scale which varies from $10^{-6}$ to $10^2$ km may be sensitive to the state of stress.
Geologic inferences about stress usually apply only to the time of formation of the feature, and may not contain any information about the current state of stress. However, if the geologic stresses are consistent with present day estimates of the stress, then current stresses may be a measure of a long term stable process.

Dislocation density, subgrain size and recrystallized grain size in deformed minerals may be sensitive to stress (Goetze and Kohlstedt, 1973; Mercier et al., 1977; Twiss, 1977).

A dislocation density on the order of $10^6$ cm$^{-2}$ for olivine crystals in an Hawaiian Island xenolith has been observed using a transmission electric microscope (Goetze and Kohlstedt, 1973). Using this density and assuming that the dislocation structure represents a state of equilibrium, Goetze and Kohlstedt (1973) obtained a differential stress of 40 bars at the depth of excavation.

The size of subgrains and recrystallized grains may also be sensitive to stress (e.g. Mercier et al., 1977; Mercier, 1978). Measurement of subgrain and recrystallized grain sizes in xenoliths from Africa and North America by Mercier et al. (1977) indicate that the maximum differential stress occurs at depths less than 50 km and has a value between 50 and 300 bars. At greater depths, the differential stress is less, with a value of about 30 bars (Mercier et al., 1977; Mercier, 1978).
The usefulness of the measurement of dislocation densities, subgrain, and recrystallized grain sizes for studying the regional intraplate stress field is limited. No estimates can be made about the direction of principal stresses. It is difficult to separate the effect of steady state ambient stresses from transient stresses due to emplacement. Thus these measurements are primarily useful for determining the depths over which differential stresses are concentrated and the depth to which differential stress can be supported. Even so, these measurements are one of the only sources of information about the state of deformation at depth in the stable interior regions of the continents.

On a somewhat larger scale, pillar-like structures called stylolites are found in carbonates and sandstones with lengths varying from millimeters to tens of centimeters. Stylolites result from pressure solution, and the long axis of each pillar should form parallel to the minimum compressive stress. Stress directions inferred from stylolites in Germany are consistent with other techniques for inferring stress (Greiner, 1975).

The orientation of fractures in fresh rock outcrops may also be related to stress. In theory, these orientations are related to a triaxial stress field whose principal directions are based on the Mohr-Coulomb theory of fracture (Kohlbeck and Scheidegger, 1977). This technique has been applied in Europe, N. America and the
Caribbean.

Evidence about the state of stress on a larger scale may be inferred from the orientation and structure of dikes, folds, and rift valleys. With these larger geologic features, it is difficult to determine whether the forces that were active during the formation of the structure are still active today. In any event, newly injected dikes should form perpendicular to the axis of the minimum principal stress to minimize the work necessary to open the dike to magma. There are linear dikes with lengths over 10 km in the southwest United States that are shaped like tear drops. The tear drop shape may result from a variable stress field, variable magma pressure, or variable elastic properties of the host rocks (Pollard and Muller, 1976). The major axis should still form perpendicular to the minimum compressive stress. The orientation and ratio of lengths of folds may also be related to the state of strain and stress (Johnson and Page, 1976).

Rift valleys and grabens may indicate regional extension. The lower Rhine graben in Germany is aligned with the axis of maximum compression inferred from recent earthquakes (Ahorner, 1975). This indicates that the stress field acting during the formation of this portion of the graben complex is similar to the present tectonic field. Similarly, the strike of the East African Rift system is normal to the inferred direction of minimum compressive stress from regional earthquakes (Sykes, 1967; Maasha
Volcanic features have been interpreted in terms of regional stress field. Linear island chains may be the expression of tensional cracks extending through the lithosphere (Turcotte and Oxburgh, 1973). Linear island chains with clear age progressions along the length of the chain, such as the Hawaiian-Emperor chain, have also been interpreted in terms of "hot spots" (Wilson, 1963; Morgan, 1972). Some island chains, such as the Line Islands in the Pacific, however, do not show an age progression along the length of the chain (Winterer, 1976; Jarrard and Clague, 1977). "Hot spots" seem an unlikely explanation in this case. The island chain may be similar to a dike, in which case the chain should be perpendicular to the least compressive stress. The Line Islands are at least 80 million years old, and probably do not reflect current stress patterns.

The orientation of flank volcanism may be an indicator of the stress field near volcanoes. If the flank volcanism forms as radial dikes from the main vent, then the orientation of the flank volcanism should be perpendicular to the minimum compressive stress. An analysis of the stress patterns has been made for volcanoes in the Aleutians and in Alaska (Nakamura et al., 1977).

Geologic evidence on the state of stress exists on a scale from millimeters to hundreds of kilometers. Most geologic evidence is related to the orientation and not
magnitude of the stresses in the earth. All geologic evidence must be carefully studied to determine whether or not there is any relationship to the current stress field.
SECTION 2.2 REGIONAL DESCRIPTION OF THE INTRAPLATE STRESS FIELD

In the previous section the methods used to infer the state of stress in the lithosphere were reviewed. In this section the observed intraplate stress field will be summarized by region.

2.2.1 African Plate

Continental East and South Africa are two regions where the state of stress in the lithosphere has been well studied. Geological evidence, earthquake mechanisms, in-situ measurements, and tiltmeter observations suggest different, but consistent, patterns for the stresses in these two areas.

The tectonics of East Africa are dominated by the East African rift system (Fig. 2.1), which extends from a few degrees north to at least 15° or 20°S (Fairhead and Girdler, 1971; Scholz et al., 1976). Normal faulting mechanisms for earthquakes in the area confirm that the rift valley is an extensional feature, with minimum compression trending approximately E-W (Sykes, 1967; Maasha and Molnar, 1972). One earthquake located near the eastern section of the rift southeast of Lake Victoria at 4°S, 35°E shows strike-slip faulting with maximum compression trending nearly E-W (Fairhead and Girdler, 1971). This earthquake is the only strike-slip event associated with the rift system. It is located near a normal fault (Fairhead and Girdler, 1971), and would be consistent with a
left lateral transform fault between normal faults.

The western branch of the northern rift system appears to be much older, with faulting and grabens that are at least Jurassic in age (McConnell, 1967). This branch clearly predates the current tectonics of the Red Sea and the Gulf of Aden, suggesting that at least the western branch is not a newly forming plate boundary (Girdler and Styles, 1974, 1978). The western branch is still active, with normal faulting earthquakes indicating minimum horizontal compression trending nearly E-W.

In the southern section of the rift system, the eastern and western branches merge and there is a general decrease in extensional tectonics with the rift valley becoming narrower and the walls lower. Normal faulting earthquakes are still common, consistent with extension trending E-W. By 20°S, the surface expression of the rift system dies out. A microearthquake study in Botswana between 18° and 21°S indicates that the rift system may be in the early stages of propagating further southward (Scholz et al., 1976).

While the rift system clearly indicates extensional tectonics, the question arises as to whether or not it is intraplate. If the rift system is the continental expression of a spreading center extending from the Afar triangle into Africa, then the state of stress in the region may not be related to forces acting on distant plate boundaries. Chase (1978a) has interpreted differences in spreading rates
between the Red Sea and the Gulf of Aden in terms of separation of African and Somalian plates along the East African rift system. Interpretation of the extensional nature of the rift system in terms of intraplate stresses is questionable.

The only earthquake in east Africa away from the rift system has a strike-slip mechanism with the P-axis trending NW-SE (event 1, Figure 2.1, Table 2.1). The orientation of the P-axis implies compression normal to the trend of the rift system and indicates that the state of stress may be very different in areas of east Africa away from the rift system.

South Africa has a different tectonic setting than East Africa. The rift system apparently ends by 20°S. Below this latitude observations of stress indicate maximum compression approximately E-W. These observations include earthquake mechanisms, tiltmeter observations, and in-situ measurements.

An earthquake on 29 September 1969 located at 33°S, 19°E, near Ceres, South Africa, had a strike-slip mechanism with maximum compression inferred to be nearly EW (Fairhead and Girdler, 1971; Maasha and Molnar, 1972). Aftershocks for this event trended WNW-ESE in agreement with the fault plane orientation.

In-situ measurements of stress in southern Africa have been primarily restricted to strain-relief methods in mines.
It is not always possible to remove the effect of the ore veins, nearby cavities, and tunnels when making strain relief measurements within a few tens of meters of the tunnel. Gay (1975) reviewed the existing data on in-situ measurements in South Africa and concluded that on a "regional scale, there appears to be no dominant trend in the orientation of the principal stresses over southern Africa." Examination of the data, however, suggests a very prominent trend if certain sources of error are considered. For example, if the difference between the two horizontal principal stresses is small, a small error in the measurement of the magnitudes of the stresses will result in a large error in the inferred azimuth. If this possibility is considered, 70% of his data have the azimuth of the greatest horizontal principal stress between 80° and 150° measured clockwise from north.

More measurements have been made recently and Gay now concludes that the "most striking aspect of the results is the consistent NS (001°, s.d. 25°) and E-W (271°, s.d. 25°) directions of the horizontal principal stresses" (Gay, 1977).

Tiltmeter observations in a deep gold mine in southern Africa near Johannesburg suggest the presence of significant non-hydrostatic stresses (McGarr and Green, 1975). While the orientation of the stresses is not known, the observations, along with measurement of the elastic
constants, imply deviatoric stresses of several hundred bars in this region. These values are consistent with the magnitudes of in-situ measurements in southern Africa.

In the western part of the African plate, two oceanic intraplate earthquakes and an in-situ strain-relief measurement indicate a NW-SE trend for the maximum compressive stress (Figure 2.1 and Table 2.1).

2.2.2 Indian Plate

The Indian plate is divided into three regions for discussion of the intraplate stress field. These regions include continental India, the Indian Ocean basin, and continental Australia.

The Indian plate with plate and continental boundaries is shown in Fig. 2.2. The stress direction data shown in Fig. 2.2 are listed in Table 2.2. The focal mechanisms for earthquakes in India are shown. On 10 December 1967 a reservoir-induced earthquake occurred near the Koyna Dam at 17.5°N, 73.8°E. The inferred P-axis trends NNW-SSE (Singh et al., 1975; Langston, 1976). This event occurred in a relatively aseismic area of the Deccan Traps and the P-axis is fairly consistent with other events in India. The effect of the reservoir may be to increase fluid pore pressure and reduce the effective stress. Thus, although the earthquake may have resulted from the filling of the reservoir, the stress directions inferred from the
fault plane solution probably reflect the prevailing stress field. Other earthquakes in India include two predominantly thrust events in western India and three predominantly thrust events in eastern India. Focal mechanisms for these events have been presented by Chandra (1977), and are shown in Fig. 2.2.

Much of the seismicity in the Indian Ocean away from plate boundaries occurs near the Ninety-east ridge, probably an ancient transform fault (Sclater and Fisher, 1974). Two events shown in Fig. 2.2 occurred near the ridge on 25 May 1964 and 10 October 1970 (see Table 2.2). Both events have strike-slip focal mechanisms with the P-axis trending NW-SE to NNW-SSE, respectively (Sykes, 1970; Fitch et al., 1973). A thrust fault earthquakes occurred to the west of the ridge on 25 June 1974. The P-axis for this event, shown in Fig. 2.2, also trends NW-SE (Stein and Okal, 1978). Another thrust fault earthquake occurred to the east of the ridge on 26 June 1971. The P-axis trends NW-SE (Sykes and Sbar, 1974). The style of faulting in the Indian Ocean seems to vary from strike-slip along the Ninety-east ridge to thrust on either side of the ridge. The orientation of the P-axis for all of the events, however, trends basically NW-SE. The stress field for the events may thus be similar, with the variation in style due to the availability of a zone of lithospheric weakness such as the Ninety-east ridge along which failure may occur.
A variety of stress sensitive measurements indicate a consistent, but different, stress field for Australia. Fault plane solutions exist for five earthquakes in Australia between 1968 and 1973 (Fitch et al., 1973; Stewart and Denham, 1974; Mills and Fitch, 1977). All events have either thrust or strike-slip mechanisms with the direction of maximum compressive horizontal stress nearly E-W in southern Australia and ENE-WSW to N-S in northern Australia. The directions of the P-axes inferred from the fault plane solutions are shown as solid circles with arrows in Fig. 2.2. The P-axes all have very small plunge angles, indicating that the maximum compressive principal stress is nearly horizontal.

Strain-relief in-situ measurements of stress have been made in Australia and Tasmania (Stephenson and Murray, 1970; Mathews and Edwards, 1969; Endersbee and Hofto, 1963). The directions of the maximum compressive stress are nearly horizontal and strike E-W in the north and more NW-SE in the south. The measurements were made at depths from 150 m to 650 m in mudstone and ore bodies. The maximum compressive stress varied from 150 to 330 bars, with an average of about 200 bars. These in-situ measurements are consistent with the general trend exhibited by the fault plane solutions of nearly horizontal maximum compression trending E-W in the south and more ENE-WSW in the north. The directions inferred from the in-situ measurement for maximum compress-
ion are shown by open circles with lines in Fig. 2.2 and listed in Table 2.2.

Most of the Indian plate away from plate boundaries is characterized by a state of compressive stress. On the Indian continent, the axis of maximum compression is nearly horizontal and trends basically N-S. The Indian Ocean basin appears to be consistent with a more NW-SE trending maximum compression axis and may represent a transition between the nearly N-S trend in the north and the nearly E-W to ENE-WSW trending axis of maximum compression inferred from fault plane solutions and in-situ measurements in Australia.

2.2.3 Eurasian plate

The state of stress in the Eurasian plate will be presented in two parts. In the first, Europe including parts of the Alps will be discussed. Asia will be considered separately.

The state of stress in western Europe has been well studied. There are numerous fault plane solutions, in-situ strain measurements, and a variety of geological indicators of past and recent states of stress. A summary of the data for western Europe is shown in Fig. 2.3 and listed in Table 2.3.

Fault plane solutions have been calculated for a large number of strike-slip and normal earthquakes in western Europe and have been summarized by Ahorner (1975).
The focal mechanism data in Fig. 2.3 are restricted to earthquakes with published fault plane solutions. These data are shown as solid circles in Fig. 2.3 with the directions of P-axes shown by inward pointing arrows. Poorly constrained solutions are shown by dashed lines. The directions for the maximum compressive stress, with few exceptions, trend NW-SE. These earthquakes occurred between 1933 and 1971 over a region nearly 500 kilometers a side, including parts of the Rhine graben and Alps. The consistency of the directions argues for a large scale source of the stress. Most of the inconsistent data are either pre-1964 and hence pre-World Wide Standard Seismograph Network, or are located in the geologically complicated Alps.

In-situ measurements of stress have been made in western Europe by many authors and have recently been summarized by Ranalli and Chandler (1975) and Greiner and Illies (1977). The magnitudes of the non-lithostatic stresses are generally on the order of tens of bars. These measurements are typically made using a strain-relief technique either near the earth's surface or in tunnels or mines. Although such measurements, taken singly, are suspect for reasons discussed in section 2.1.2, the consistency of the orientations of the maximum compressive stress as shown by the open circles in Fig. 2.3, and the agreement with the focal mechanism studies previously discussed, is remarkable.
Other indicators of stress, based on the orientation of geological features, are shown in Fig. 2.3 by open triangles. The orientation of the lower Rhine graben is indicative of the state of stress in the region at the time of formation. The strike is NW, indicating NE extension. This is consistent with the fault plane solutions of recent earthquakes, indicating that the current state of stress may have been stable over a long period of time.

The combination of fault plane solutions, in-situ measurements, and the orientation of stylolites and major geologic features such as the lower Rhine graben are all consistent with a regional stress pattern where the maximum compressive principal stress is horizontal and trends NW.

There are now focal mechanisms for many earthquakes in Asia. Each data point shown on Fig. 2.4 and listed in Table 2.4 represents the average of at least four closely spaced events with consistent mechanisms in various seismically-active areas of Eurasia (Molnar et al., 1973). Molnar et al. (1973) concluded that an "approximately N-S to NE-SW trending principal compressive stress appears to be transmitted across a large area north and east of the Himalayas." The Baikal rift zone, shown on Fig. 2.4, is similar to the East African rift system with evidence of extensional tectonics (Artemjev and Artyushkov, 1971). As with the East African rift system, it is difficult to determine whether such extensional tectonics represent the
formation of a new plate boundary or are truly intraplate in nature.

The data on the Eurasian plate east of Europe is presented for completeness. Most of the earthquakes occur along major faults defining broad zones of extensive deformation and folding between relatively stable blocks (Molnar et al., 1973). These blocks may be considered micro-plates separated by broad "plate boundaries." As has been pointed out, this approach is misleading because often the "plate boundaries" fail to completely enclose the micro-plates or die out in the stable interior of the micro-plates (Tapponnier and Molnar, 1977).

The origin of the stress field is the matter of considerable debate. The compression to the north of the Himalayas has been explained in terms of the collision of India with Eurasia (Tapponnier and Molnar, 1976, 1977). Both the Baikal extension and the lack of comparable deformation in India have been interpreted in terms of this theory. The collision of India with Eurasia almost surely effects the present day tectonics of Eurasia. The interpretation of Eurasian stresses in terms of intraplate tectonics is however, uncertain.
2.2.4 **American Plates**

Although relative motion between the North and South American plates is not observed along any well-defined plate boundary, the relative rotation of Africa with respect to the Americas is fit best using separate American plates (Minster and Jordan, 1978). The data on the American intraplate stress field will be presented in two parts. In the first, South American data will be presented. Then data for North America will be summarized.

To the best of my knowledge, there are no in-situ stress measurements for South America. There have been, however, about a dozen intraplate earthquakes for which focal mechanism studies have been made. The inferred P-axis for four shallow thrust earthquakes and the P-axis for five shallow strike-slip earthquakes located between 200 and 800 km away from the South America-Nazca plate boundary are shown in Fig. 2.5 and listed in Table 2.5 (Stauder, 1975). These mechanisms represent nearly E-W horizontal maximum compression on the western edge of the South American plate between 2°S and 14°S.

Further into continental South America there has been a series of earthquakes that have focal mechanisms consistent with NW-SE horizontal maximum compression (Mendiguren, 1978). These events, also shown in Fig. 2.5, are located in central and eastern Brazil. Most occurred before the establishment of the WWSSN. Although
not all of the focal planes are well constrained, taken together a consistent trend is visible. A poorly constrained P-axis is shown as a dashed arrow in Fig. 2.5, while two thrust earthquakes for which the nodal planes could not be determined are shown as solid circles.

Intraplate stress data are abundant for the North American plate. Perhaps this abundance contributes to the inferred complexity of the North American intraplate stress field. Much of continental North America east of the Rocky Mountains and north of the Gulf of Mexico is characterized by an E-W to ENE-WSW trend for the maximum horizontal compressive principal stress (Sbar and Sykes, 1973, 1977; Haimson, 1977). While there are exceptions, the bulk of the evidence, using a variety of techniques, strongly suggests a consistent regional pattern. The inferred stresses may be of tectonic origin, since local effects of topography and glacial history are unlikely to produce such a uniform pattern.

The techniques used in North America to infer the state of stress at depths from the surface to several kilometers include fault plane solutions, hydrofracture and strain-relief in situ measurements, and recent geologic features such as post-glacial buckles or pop-ups. The magnitudes of the maximum horizontal compressive principal stress are on the order of a hundred bars, with values varying from about 10 to 600 bars (Sbar and Sykes, 1973).
Fault plane solutions exist for a large number of earthquakes in North America away from plate boundaries. The stress patterns east and west of the Rocky Mountains represent different tectonic regimes. North America west of the Rocky Mountains and east of the San Andreas fault is exemplified by regional extension in the Basin and Range Province and normal faulting earthquakes along the Intermountain Seismic Belt (Sykes and Sbar, 1974). The origin of these extensional features is uncertain, but may well be related to ancient and recent changes in plate boundary types along western North America (Atwater, 1970; Burchfiel and Davis, 1975). The orientation of the T-axis for three normal faulting earthquakes along the Intermountain Seismic Belt that are typical of the several in the area are shown in Figure 2.6 and listed in Table 2.6 (Sykes and Sbar, 1974; Smith and Sbar, 1974).

Beginning in the Colorado Plateau and extending eastward across North America to the Appalachian Mountains, fault plane solutions tend to indicate either strike-slip or thrust faults with maximum compression trending E-W to NE-SW. P-axes for thrust and strike-slip earthquakes and T-axes for normal faulting earthquakes are shown in Fig. 2.6 as arrows pointing inward and outward, respectively, from filled circles. The data presented in Fig. 2.6 have been selected from summaries by Sykes and Sbar (1974), Raleigh (1974), Hashizume (1977), and McGarr and Gay (1978). The data in Fig. 2.6 do not represent all the focal mechanisms that have been done. Normal faulting
earthquakes on continental shelves have not been included (Hashizume, 1973). These events may be due to contraction of the continental margin (Sleep, 1971) or sediment loading, and hence may not be related to a regional stress field. Fault plane solutions for four small strike-slip earthquakes \( M_s 5.1-5.7 \) in a swarm in the Sverdrup Basin in Canada at \( 77^\circ N, 106^\circ W \) off Fig. 2.6 indicate NE-SW maximum compression (Hasegawa, 1977). Finally, whenever several earthquakes in the same locality have similar mechanisms, only a representative solution is shown. Dashed arrows indicate less well constrained data.

Although most of the earthquakes indicate E-W to NE-SW compression, there are exceptions. A well studied normal fault earthquake occurred on 21 October 1965 in southeastern Missouri at \( 37^\circ N, 91^\circ W \) (Mitchell, 1973; Street et al., 1974; Patton, 1976). The T-axis trends NW-SE, as shown in Fig. 2.6. It is located in an area of the continental United States that has a relatively high level of seismic activity. The mechanisms for small earthquakes in this area show a change in trend from normal faulting to thrust faulting over a distance of a few hundred km and may well be controlled by the local structure of the Mississippi Embayment and the Ozark Uplift (Street et al., 1974). A similar rapid change in type of faulting is seen in the Appalachians, which appear to mark the eastern edge of the E-W to NE-SW trend for most of eastern North America (Sbar and Sykes, 1977).
The eastern Missouri and Appalachian earthquakes indicate the complexity of the intraplate stress field in North America and the effects of local geological structures.

Numerous in-situ stress and strain measurements have been made in North America. In-situ measurements are shown as open circles with the line giving the direction of the maximum horizontal compressive stress in Fig. 2.6. Dashed lines indicate less well constrained measurements. The data presented in Fig. 2.6 have been selected from summaries by Raleigh (1974), Sykes and Sbar (1974), Ranalli and Chandler (1975), Haimson (1977), and McGarr and Gay (1978).

The E-W to NE-SW trend inferred from the in-situ measurements is consistent with the focal mechanism studies. This is encouraging because earthquakes are passive indicators of stress in the sense that we must wait for an earthquake to occur. In-situ measurements can be made in regions of interest, and thus hold the potential to greatly extend our knowledge of the intraplate stress field.

Post-glacial buckles or pop-ups are geologic evidence of the recent state of stress. Such buckles have been observed in New York, New England, Ohio, and eastern Canada (Sbar and Sykes, 1973). Often they are associated with quarrying where the lithostatic load has been recently removed. Many of the features trend NW, indicating an axis of maximum compressive stress approximately NE (Sbar and Sykes, 1973). Joint fractures in Ontario, Canada, indicate
one of the principal horizontal stress directions to be NE-SW, although it is difficult to determine which is the maximum and which is the minimum (Scheidegger, 1977c). Geologic indicators are shown in Fig. 2.6 as open circles with the line giving the direction of maximum compression.

2.2.5 Oceanic regions

Intraplate earthquakes which occurred in oceanic lithosphere for the African and Indian plates were summarized along with continental intraplate stress data in sec. 2.2.1 and 2.2.2, respectively. In this section, oceanic regions for which intraplate earthquakes are the primary source of information of the state of stress are considered.

For oceanic regions, information about the intraplate stress field is limited primarily to earthquake focal mechanisms. The number of data for oceanic regions is thus much smaller than for the continents. Further, it is impossible to corroborate the stress inferred from oceanic earthquake mechanisms using other techniques such as in-situ measurements. Fortunately, the geology, topography, and chemistry of oceanic plates are usually much less complicated than for the continents. Thus, I will assume that the oceanic data are indicative of the intraplate stress field.

There are a number of fault plane solutions for earthquakes in the Pacific plate. Several are so small that a
well constrained fault plane solution could not be obtained (Sykes and Sbar, 1974). The limited first motion data indicate thrust faulting, and these are shown as filled circles without arrows on Fig. 2.7. Two events on 24 September 1966 and 28 April 1968 at 12°N, 131°W and 45°N, 175°E, respectively, indicate thrust faulting with the maximum compressive stress nearly horizontal and trending NE-SW (Sykes and Sbar, 1974).

Focal mechanisms have been determined for two earthquakes on the island of Hawaii on 23 April 1973 and 29 November 1975 (Unger et al., 1973; Ando, 1976). The 1973 earthquake had a predominantly strike-slip mechanism with the P-axis trending NE-SW while the 1975 earthquake mechanism indicated a low-angle normal fault. Hawaii is not typical of a stable interior region of an oceanic plate. The stress field is probably complicated by large bending stresses associated with the load of Hawaii on the Pacific plate (Walcott, 1970; Watts and Cochran, 1974). Further, the events may be related to stresses associated with the magma chamber for the active Hawaii volcanoes. The relationship of Hawaiian earthquakes to the Pacific intraplate stress field is thus uncertain.

Most of the Pacific is characterized by thrust-type earthquakes where focal mechanisms can be well constrained. The maximum compressive stress trends NE-SW and is approximately horizontal.
The Nazca plate has had at least three intraplate earthquakes. These events are discussed in more detail in Chapter 4 where finite element models of the Nazca plate are developed. One of the events occurred on 25 November 1965 at 17°S, 100°W. The focal mechanism was constrained by P-wave first motions and the surface wave radiation pattern (Mendiguren, 1971). The P-axis is nearly horizontal and trends E-W. Two other nearby events which occurred in 1944 are compatible with the 1965 earthquake, but the focal planes could not be further constrained (Mendiguren, 1971).

Another well studied oceanic earthquake occurred on the Antarctic plate at 40°S, 105°W on 9 May 1971. The fault plane solution indicates a thrust mechanism with a horizontal P-axis trending ESE-WNW, as shown in Figure 2.7 (Forsyth, 1973a). Also shown in Figure 2.7 are oceanic thrust earthquakes in the Atlantic near the mid-Atlantic ridge at 44°N, 31°W and two near the Caribbean at approximately 20°N, 60°W (Sykes and Sbar, 1974).

Most of the seafloor older than 20 m.y. is characterized by a state of nearly horizontal compression (Sykes and Sbar, 1974). Some of the oceanic thrust earthquakes in Figure 2.7 occur in even younger seafloor. The state of compression may thus exist for most of the oceanic lithosphere away from active spreading centers.
In the next section, a global summary of the intraplate stress field based on the regional summaries in presented.
SECTION 2.3 SUMMARY OF THE GLOBAL INTRAPLATE STRESS FIELD

A summary of the global intraplate stress field is shown in Fig. 2.7 and listed in Table 2.7. Focal mechanisms are shown by filled circles with the P-axis for thrust earthquakes, the T-axis for normal earthquakes and P- and T- axes for strike-slip earthquakes. The P- and T- axes point inward and outward from filled circles, respectively. In-situ measurements and geologic indicators are shown by open circles with the line giving the orientation of the maximum horizontal compressive stress. The data shown in Figure 2.7 represent a summary of the more complete regional summaries presented in Sec. 2.2.1-2.2.6 and Figures 2.1-2.6. Some of the complexity of the intraplate stress field as summarized in Sec. 2.2.1-2.2.6 cannot be shown in a global summary. Dashed lines and arrows indicate less well constrained data. Only representative data are shown for some areas where the data concentration is high and the data are consistent.

Although the global intraplate stress field is complicated, several patterns can be seen. Much of stable North America is characterized by an E-W to NE-SW trend for the maximum compressive stress (Sec. 2.2.4). South America near the trench and perhaps further into the continent has horizontal compressive stresses trending E-W to NW-SE (Sec. 2.2.4). Western Europe north of the Alps is characterized by a NW-SE trending maximum horizontal compression (Sec.
2.2.3), while Asia has the maximum horizontal compressive stress trending more N-S, especially near the Himalayan front (Sec. 2.2.3). The Indian plate has a horizontal maximum compressive stress direction that varies from nearly N-S in continental India to more E-W to NE-SW in Australia (Sec. 2.2.2). Horizontal stresses are variable in Africa, but tend to be compressive with an E-W to NW-SE trend in the south and west and extensional with an E-W trend in east Africa (Sec. 2.2.1). Oceanic regions away from active spreading centers generally indicate a compressive state of stress (Sec. 2.2.5).

The consistency, or lack thereof, of the intraplate stress field depends on several factors. Each of the techniques used for inferring stress has uncertainties that are often large or unknown. This is further complicated by the fact that measurements are often taken very near the earth's surface or in mines and tunnels where the effects of nearby structures may be large. The geologic structure at the measurement site may control the state of stress. Thus it is not surprising that the inferred intraplate stress field is complicated. Perhaps more surprising is the fact that general trends for the orientations of principal stresses can often be found that are reasonably consistent over regional dimensions.

Calculated models of the intraplate stress field due to possible driving mechanisms will be compared in later
chapters with the observed intraplate stress data that has been summarized in this chapter.

In the next chapter apparent stresses and stress drops for intraplate earthquakes will be presented with the aim of constraining the level of ambient tectonic stress.
TABLE 2.1. INTRAPLATE STRESS DATA FOR THE AFRICAN PLATE

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<sup>1</sup> Numbers refer to Figure 2.1.

<sup>2</sup> Type: 1 = fault plane solution; 2 = strain-relief in-situ measurement.

<sup>3</sup> Azimuth and plunge of P-axis for fault plane solutions. Azimuth of maximum compressive stress for in-situ measurements. Azimuth in degrees measured clockwise from north.
References for Table 2.1.

a Fairhead and Girdler (1971)
b Gay (1977)
c Hast (1969)
d Maasha and Molnar (1972)
e Richardson and Solomon (1977)
f Sykes (1967)
g Sykes and Sbar (1974)
**TABLE 2.2.** INTRAPLATE STRESS DATA FOR THE INDIAN PLATE

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1 Numbers refer to Figure 2.2.

2 Type: 1 = fault plane solution; 2 = strain-relief in-situ measurement.

3 Azimuth and plunge of P-axis for fault plane solution. Azimuth of maximum compressive stress for in-situ measurements. Azimuth in degrees measured clockwise from north.
References for Table 2.2.

a Chandra (1977)
b Endersbee and Hofto (1963)
c Fitch (1972)
d Fitch et al. (1973)
e Langston (1976)
f Mathews and Edwards (1970)
g Mills and Fitch (1977)
h Stein and Okal (1978)
i Stephenson and Murray (1970)
j Stewart and Denham (1974)
k Sykes (1970)
l Sykes and Sbar (1974)
### TABLE 2.3. INTRAPLATE STRESS DATA FOR EUROPE

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1 Numbers refer to Figure 2.3.

2 Type: 1 = fault plane solution; 2 = strain-relief in-situ measurement; 3 = rock joint orientation; 4 = stylolite orientations.

3 Azimuth and plunge of P-axis for fault plane solutions. Azimuth of maximum compressive stress for in-situ measurements and geologic indicators. Azimuth in degrees measured clockwise from north.

4 Composite solution.

5 Local magnitude $M_L$. 
References for Table 2.3

a  Ahorner (1975)
b  Ahorner et al. (1972)
c  Ahorner and Schneider (1974)
d  Capozza et al. (1971)
e  Greiner (1975)
f  Greiner and Illies (1977)
g  Hast (1958)
h  Hast (1969)
i  Hast (1973)
j  Pavoni and Peterschmitt (1974)
k  Scheidegger (1977a)
l  Scheidegger (1977b)
m  Schneider et al. (1966)
TABLE 2.4. SELECTED EARTHQUAKES IN ASIA

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1 Numbers refer to Figure 2.4.

References

a Tapponnier and Molnar (1978)

b Tapponnier and Molnar (1977)
### TABLE 2.5. SOUTH AMERICAN INTRAPLATE EARTHQUAKES

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1 Numbers refer to Figure 2.5.

2 Composite fault plane solution.

* Fault plane solution indicates a thrust fault mechanism, but the focal planes could not be determined.
References for Table 2.5.

a. Mendiguren (1978)
b. Stauder (1975)
c. Wagner (1972)
TABLE 2.6. SELECTED INTRAPLATE STRESS DATA FOR NORTH AMERICA

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1 Numbers refer to Figure 2.6.

2 Type: 1 = fault plane solution; 2 = strain-relief in-situ measurement; 3 = hydrofracture in-situ measurement; 4 = geologic pop-up.

3 Azimuth and plunge of P-axis for fault plane solutions. Azimuth of maximum compressive stress for in-situ measurements and geologic indicators. Azimuth in degrees measured clockwise from north.

4 Composite fault plane solution.
References

a Fraser and Pettitt (1962)
b Haimson (1977)
c Hashizume (1977)
d Herget (1974)
e Hooker and Johnson (1969)
f Raleigh et al. (1972)
g Roegiers (1974)
h Sykes and Sbar (1974)
i Sbar and Sykes (1977)
j Smith and Sbar (1974)
k Stauder and Nuttli (1970)
l Street et al. (1974)
m Strubhar et al. (1975)
n von Schonfeldt (1970)
o Wetmiller (1975)
p Zoback and Healy (1977)
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1 Numbers refer to Figure 2.7.

2 Type: 1 = fault plane solution; 2 = strain-relief in-situ measurement; 3 = hydrofracture in-situ measurement; and 4 = stress-sensitive geologic features.

3 Azimuth and plunge of P-axis for fault plane solutions. Azimuth of maximum compressive stress for in-situ measurements and geologic features. Azimuth in degrees measured clockwise from north.

4 Composite fault plane solution.

5 Local magnitude M_L.

6 Fault planes could not be constrained, but indicate a thrust mechanism.
References for Table 2.7.

a Ahorner (1975)  
b Ahorner et al. (1972)  
c Ahorner and Schneider (1974)  
d Capozza et al. (1971)  
e Chandra (1977)  
f Endersbee and Hofto (1963)  
g Fairhead and Girdler (1971)  
h Fitch (1972)  
i Fitch et al. (1973)  
j Gay (1977)  
k Forsyth (1973a)  
l Greiner (1975)  
m Greiner and Illies (1977)  
n Haimson (1977)  
o Hashizume (1974)  
p Hashizume (1977)  
q Hast (1969)  
r Hast (1973)  
s Langston (1977)  
t Maasha and Molnar (1972)  
u Mathews and Edwards (1970)  
v Mendiguren (1971)  
w Mendiguren (1978)  
x Mills and Fitch (1977)  
y Tapponnier and Molnar (1978)  
z Pavoni and Peterschmidt (1974)  
aa Richardson and Solomon (1977)  
bb Roegiers (1974)  
cc Sbar and Sykes (1977)  
dd Scheidegger (1977b)  
ee Schneider et al. (1966)  
ff Smith and Sbar (1974)  
gg Stauder (1975)  
hh Steiner and Okal (1978)  
ii Stephenson and Murray (1970)  
jj Stewart and Denham (1974)  
kk Struthar et al. (1975)  
ll Sykes (1967)  
mm Sykes (1970)  
nn Sykes and Sbar (1974)  
oo Tapponnier and Molnar (1977)  
pp Unger et al. (1973)  
qq Wagner (1972)  
rr Zoback and Healy (1977)
FIGURE CAPTIONS

Figure 2.1  Intraplate stress field data for the African plate. Continental boundaries are shown as light lines and are taken as the thousand fathom bathymetric contour. Plate boundaries are shown as heavy lines. Filled and open circles represent fault plane solutions and in-situ measurements, respectively. The P- and T-axes for fault plane solutions point inward and outward, respectively. The P- and T-axes are shown for strike-slip earthquakes. The P-axis for thrust earthquakes, and the T-axis for normal faulting earthquakes, are shown. For in-situ data, the line gives the direction of the maximum compressive stress. Dashed lines indicate less reliable data. Numbers refer to Table 2.1.

Figure 2.2  Intraplate stress field data for the Indian plate. Continental boundaries taken as the thousand fathom bathymetric contour. The numbers refer to Table 2.2. For explanation of the symbols, see Figure 2.1.

Figure 2.3  Intraplate stress field data for Europe. Continental boundaries taken as the thousand fathom bathymetric contour. The open circles represent in-situ measurements and geologic indicators. The numbers
refer to Table 2.3. For explanation of the symbols, see Figure 2.1.

Figure 2.4 Intraplate earthquake data for Asia. Continental boundaries taken as the thousand fathom bathymetric contour. Each data point represents the average of at least four closely-spaced earthquakes with consistent focal mechanisms. Numbers refer to Table 2.4. For explanation of symbols, see Figure 2.1.

Figure 2.5 Intraplate earthquake data for South America. Because of the density of the data, only the P-axis is shown, although several of the mechanisms have a significant strike-slip component. Continental boundaries taken as the thousand fathom bathymetric contour. Numbers refer to Table 2.5. For other details, see Figure 2.1.

Figure 2.6 Selected intraplate stress field data for North America east of the Intermountain Seismic Belt. Continental boundaries taken as the thousand fathom bathymetric contour. Numbers refer to Table 2.6. For other details, see Figure 2.1.
Figure 2.7  Global summary of intraplate stress field data. Numbers refer to Table 2.7. For other details, see Figure 2.1.
Figure 2.2
Figure 2...
Figure 2.7
CHAPTER 3: APPARENT STRESS AND STRESS DROP FOR INTRAPLATE EARTHQUAKES

SECTION 3.1 INTRODUCTION

A variety of techniques exist for inferring the state of stress in the lithosphere. Earthquake focal mechanism studies are useful for determining the orientation, but not the magnitude, of principal stresses acting on fault surfaces. In situ measurements may provide both magnitude and orientation information, but are limited to the upper few kilometers of the continental crust. These techniques have been reviewed in the last chapter where a regional summary of the known properties of the intraplate stress field were presented.

In this chapter the apparent stress and stress drop of midplate earthquakes are discussed with the aims of placing bounds on the magnitude of intraplate deviatoric stresses and comparing intraplate and plate boundary stress environments.

Both the apparent stress and stress drop represent measures of some fraction of the shear stress acting during an earthquake and can be derived from seismic observations. The apparent stress is the product of the average shear stress \( \sigma \) on the fault before and after faulting and an unknown seismic efficiency factor \( \eta \), and is given by (Aki, 1966):
\begin{equation}
\eta \sigma = \mu \frac{E_S}{M_o} \tag{3.1}
\end{equation}

where \( \mu \) is the shear modulus, taken as \( 3 \times 10^{11} \text{ dyne/cm}^2 \), \( E_S \) is the radiated seismic energy in dyne-cm, and \( M_o \) is the seismic moment in dyne-cm.

The stress drop \( \Delta \sigma \) is the difference between the initial and final shear stress on the fault. For a vertical strike-slip earthquake the stress drop is given by (Knopoff, 1958):

\begin{equation}
\Delta \sigma = \frac{2M_o}{\pi Lw^2} \tag{3.2}
\end{equation}

and for a dip-slip earthquake it is (Aki, 1966):

\begin{equation}
\Delta \sigma = \frac{4}{\pi} \frac{(\lambda+\mu)}{\lambda+2\mu} \frac{M_o}{Lw^2} \tag{3.3}
\end{equation}

where \( L \) and \( w \) are fault length and width, respectively, for a rectangular fault, and \( \lambda \) is the Lamé constant. The quantities \( E_S \) and \( M_o \) can be determined or estimated from seismic wave amplitude measurements, and the parameters \( L \) and \( w \) can be determined either from field observations (surface rupture, aftershock locations) or from seismic measurements (corner frequency, directivity) and a model of the earthquake source.
SECTION 3.2 APPARENT STRESS FOR INTRAPLATE EARTHQUAKES

The midplate earthquakes for which determinations of apparent stress and stress drop have been made are listed in Table 3.1 and their epicenters are shown in Figure 3.1. Earthquakes from all the major plates are included. References for the fault plane solutions are listed in Table 3.1.

To determine the apparent stress from equation (3.1), both the seismic moment and seismic energy must be known. The seismic moment $M_0$ may be calculated either from long period surface waves or from long period body waves.

In this study, long period amplitude spectra of SH waves are used to calculate the moment when no published values are available. Long period horizontal components of shear waves recorded at WWSSN stations in the epicentral distance range of 32 to 67 degrees are digitized and rotated to isolate the SH component. The digitized signal with a time window of approximately 60 seconds is Fourier analyzed to obtain an amplitude spectrum. The resulting spectrum is corrected for the instrument response of WWSSN stations (Hagiwara, 1958), and the spectral amplitude $\Omega_0$ of the long period end of the spectrum is estimated (Hanks and Wyss, 1972). The seismic moment may be derived from $\Omega_0$ by the relation (Keilis-Borok, 1960)
where \( \rho \) and \( V \) are the density and shear wave velocity, respectively, at the source, \( R \) is a correction for radiation pattern, and \( G \) is a correction for geometrical spreading, attenuation, and the free surface. The radiation pattern correction, including superposition of the ss wave of appropriate source amplitude and free surface reflection coefficient, is taken from Hirasawa (1966) given the fault plane parameters of each earthquake. The propagation correction term is given by (Julian and Anderson, 1968)

\[
G = 2 \left[ \frac{cV^3 dt/d\Delta |d^2 t/d\Delta^2|}{\rho_o V r_o^2 r^2 \sin \Delta \cos i_0} \right]^{1/2} \exp \left[ -\pi f \int \frac{ds}{QV} \right] \tag{3.5}
\]

where \( r \) is the radius from the earth's center, \( \Delta \) is the epicentral distance, \( t \) is the travel time, \( i \) is the angle of incidence, the subscript \( o \) denotes the earth's surface, \( f \) is the frequency, \( 1/Q \) is the attenuation, and the integral is taken over the entire ray path. The \( Q \) model of Tsai and Aki (1969) and the shear wave velocity model SLUTD1 of Hales and Roberts (1970) are used in computing the attenuation correction and the travel time derivatives. The factor 2 is equation (3.5) accounts for the effect of the free surface at the station.
stations for seven intraplate earthquakes are listed in Table 3.2. The scatter in the data is fairly large, indicative of uncertainties in the procedure. The moments in Table 3.2 are somewhat higher than the values published in Richardson and Solomon (1977). The differences result from the estimation of the sS wave contribution to the digitized signal. The quality of the amplitude spectra are estimated visually. Figure 3.2 shows three spectra for event 9 with varying quality determinations for the measurement of the level amplitude $\Omega_0$. The radiation pattern correction probably introduces the greatest uncertainty. Stations are selected away from SH nodal planes where radiation pattern corrections are largest and most sensitive to errors in the estimation of the fault plane geometry. Inaccuracies in the attenuation model have the least effect at low frequencies where $\Omega_0$ is measured. Considering these uncertainties, the seismic moment calculated from equation (3.4) is probably accurate to about a factor of 3.

The mean moments, weighted by the quality of the determination, for these events are listed in Table 3.3. As a check on our procedure for calculating seismic moment from SH wave spectra, we may compare the computed $M_0$ for the 10 December 1967 Koyna Dam earthquake in India (event 12) with values obtained by Singh et al. (1975) from surface waves and by Langston (1976) from body waves.
Our method gives $M_0 = 12.7 \times 10^{25}$ dyne-cm while Singh et al. and Langston obtained $8.2 \times 10^{25}$ and $3.2 \times 10^{25}$, respectively. Maasha and Molnar (1972) estimated the moment for the 29 September 1969 Ceres, South Africa, earthquake (event 8) to be $2.0 \times 10^{25}$ dyne-cm from long period P waves. The moment determined here from SH waves is $6.4 \times 10^{25}$ dyne-cm. The two determinations agree to within the estimated errors in both calculations.

The radiated seismic energy $E_S$ is not, in general, a simple function of a single measurable parameter. Two estimates of $E_S$ in common use are the energy-magnitude relations of Gutenberg and Richter (1956);

$$\log_{10} E_S = 5.8 + 2.4 m_b \tag{3.6}$$

and, for large events:

$$\log_{10} E_S = 11.8 + 1.5 M_S \tag{3.7}$$

where $M_S$ and $m_b$ are the surface wave and body wave magnitudes, respectively. Equations (3.6)-(3.7) are based upon an implied similarity of seismic spectra for the frequencies at which $M_S$ and $m_b$ are measured. The magnitudes $M_S$ and $m_b$ are assumed to be related by (Gutenberg and Richter, 1956)

$$m_b = 2.5 + 0.63 M_S \tag{3.8}$$
over the entire range of magnitudes. Figure 3.3 shows a plot of $m_b$ determined by the International Seismological Centre (I.S.C.) versus various determinations for $M_S$ for intraplate earthquakes. There is a systematic deviation from the Gutenberg-Richter relationship in equation (3.8). Equation (3.8) was derived using instruments substantially different from those in use today. The P wave amplitude was measured at a period between 4 and 10 seconds (Geller, 1976), while today the period at which the amplitude is measured is very near 1 second. This systematically lowers the value of $m_b$ for a given $M_S$ and may account for at least part of the deviation observed in Figure 3.3.

$M_S$ is measured at a period of about 20 seconds. At a 20 second period, $M_S$ is a measure of the level amplitude $\Omega_0$ for $M_S$ less than about 6.0 and may not be a good measure of the radiated seismic energy (see also Figure 3.2). For very large earthquakes, $M_S$ is saturated and does not increase above about 8.3 (Kanamori, 1977). For intraplate regions where the earthquakes are generally small, this is not a problem. Many intraplate earthquakes have $M_S < 6$ and for these events $M_S$ is probably not a good measure of the seismic energy. For this study, $E_s$ will be estimated primarily with $m_b$ using equation (3.6). Estimates of $E_s$ using $M_S$ will be used for intraplate earthquakes with $M_S \geq 6.0$.

It is difficult to make an accurate quantitative
measure of energy because of the uncertainties in equations (3.6)-(3.8). However, to the extent that \( \bar{\eta \sigma} \) in equation (3.1) is regarded only as some measure of short period spectral level divided by a measure of long period spectral level, equations (3.6)-(3.8) may still be useful for the comparison of relative differences in apparent stress \( \eta \bar{\sigma} \) within a population of seismic events. Equations (3.6) and (3.7) are used below to measure an \( E_s \) and it is noted that this quantity may not be closely related to actual seismic energy. The apparent stresses thus calculated are comparable only to those determined in other studies using the same procedure to estimate \( E_s \).

Considering the uncertainties associated with the calculation of magnitudes and seismic moment, apparent stresses measured by a single procedure are accurate to about half an order of magnitude. Apparent stresses for 23 intraplate events from Table 3.1 are listed in Table 3.3. When \( m_b \) is used to estimate the seismic energy, apparent stresses \( \eta \bar{\sigma} \) vary from 0.004 to 3.0 bars with most values near 0.5 to 1.0 bar. When \( M_s \) is used to estimate the seismic energy for events with \( M_s \geq 6.0 \), apparent stresses \( \eta \bar{\sigma} \) vary from 2 to 50 bars, with 5 to 10 bars a representative value. There is no systematic difference in apparent stress between continental and oceanic events. Apparent stresses show no consistent
variation with increasing event magnitude. The smallest events show the widest range of apparent stresses, probably reflecting errors in determining magnitude and spectra accurately.
SECTION 3.3 STRESS DROP FOR INTRAPLATE EARTHQUAKES

The calculation of stress drop (equations 3.2-3.3) for midplate earthquakes is difficult because of the scarcity of information on fault dimensions and average fault displacement. Intraplate events are commonly small, usually occur far from seismological stations (which tend to be concentrated in more active areas), are not always accompanied by surface faulting, and rarely have aftershocks that can be located teleseismically. For these reasons, observation of surface faulting and accurate determination of aftershock locations, the most reliable measures of fault dimensions, are not always available.

The values of stress drop for 6 intraplate earthquakes for which estimates of fault length are available are given in Table 3.4. Fault widths were estimated from an assumed geometric similarity $L/w \sim 2.5$, weakly supported by limited seismic data on fault dimensions (Ellsworth, 1975; Geller, 1976). The range in $\Delta \sigma$ is 1 to 70 bars. The largest stress drop occurs for the single earthquake in oceanic lithosphere (event 14). The fault plane dimensions for the event were estimated from aftershock epicenters relocated using a master event technique. Because of the distance of this earthquake sequence from seismic stations, the stress drop for event 14 probably has the greatest uncertainty of the listed values.
Independent estimates of $\Delta \sigma$ are available for events 8, 10, and 12. Maasha and Molnar (1972) determined a stress drop of 25 bars for event 8 from the P wave corner frequency. Fitch et al. (1973) estimated $\Delta \sigma$ to be "about 100 bars" for event 10 using a somewhat lesser fault length and a width derived from the observed displacement across surface cracks in soil and weathered rock layers. Singh et al. (1975) estimated a stress drop between 6 and 20 bars for event 12 using a length determined from rupture velocity and a width determined from an empirical relationship between fault area and earthquake magnitude. Given the sensitivity of $\Delta \sigma$ to fault dimensions, especially the uncertain fault width, the fair agreement of previous determinations with those reported here is perhaps acceptable.
SECTION 3.4 DISCUSSION AND CONCLUSION

Kanamori and Anderson (1975) have suggested that, for events with $M_s$ 6 to 8, intraplate earthquakes have higher apparent stresses ($\sim$50 bars) than interplate earthquakes ($\sim$10 to 20 bars). Their population of 'intraplate' earthquakes, however, consists only of events very near plate boundaries and only in restricted areas of high activity (Japan, California-Nevada, Azores-Gibraltar). The values of apparent stress for midplate earthquakes obtained from $M_s$ in Table 3.3 are comparable to values obtained on plate boundaries (Kanamori and Anderson, 1975), and do not support a general distinction in stress level between intraplate and interplate events. The large $M_s$ events in Table 3.3 do not cover the entire range of $M_s$ from 6 to 8, but are limited to $M_s$ 6.0 to 6.8.

As a more comprehensive comparison of apparent stress for intraplate versus interplate earthquakes, we plot in Figure 3.4 $M_o$ versus $m_b$ for the events in Table 3.3 and for a large number of plate margin events. The data are summarized in Table 3.5. The plate boundary earthquakes include post-1963 events from the lists of intermediate depth and deep focus South American and Tonga island arc earthquakes from Wyss (1970a) and Molnar and Wyss (1972), and ridge crest and transform earthquakes from Wyss (1970b). All apparent stresses
for intraplate events were calculated from equations (3.1)
and (3.6), using $M_o$ from the original authors and $m_b$ from
the I.S.C.

The main conclusion to be drawn from Figure 3.4 is
that for all tectonic environments, including midplate
regions, the apparent stresses are very similar (see
also Table 3.5). Intraplate earthquakes have slightly
higher mean and median values than other environments,
but the differences are probably not significant. The
differences probably reflect larger uncertainties for
the usually small magnitude intraplate earthquakes.

Moment-magnitude relations do not support the contention
that intraplate earthquakes are characterized in general
by significantly higher stresses than plate boundary
earthquakes.

Several workers have suggested that intraplate earth-
quakes may be characterized by higher stress drops than
interplate events. Molnar and Wyss (1972) noted that
an earthquake in the Indian plate has a stress drop
(50 bars) higher than those of shallow Tonga arc earth-
quakes (1 to 25 bars) and that $\Delta \sigma$ was generally higher
in environments where new faulting might be expected
than where rupturing on older fault surfaces is likely.

Maasha and Molnar (1972) suggested that stress drops for
African earthquakes are higher (8 to 25 bars) for those
not associated with the East African rifting than for
those events (3 to 18 bars) that are. Kanamori and
and Anderson (1975), in their study of source parameters of earthquakes with $M_S$ 6 to 8, contended that large intraplate earthquakes have generally higher stress drops (c. 100 bars) than do large interplate earthquakes (c. 30 bars). The stress drops in Table 3.5, from a few bars to a few tens of bars, are generally in the range reported for interplate events. Thus while some intralithosphere regions may be characterized by relatively high stress drops and high inferred stresses, particularly regions near plate boundaries such as those studied in the works mentioned, there is no compelling evidence from the midplate earthquakes considered in this study that stress drops or stresses are generally greater in intraplate environments than along plate boundaries.

There is considerable uncertainty about exactly how apparent stress $\overline{\sigma}$ and stress drop $\Delta \sigma$ are related to the ambient stress field. The unknown seismic efficiency parameter $\eta$ is a measure of the amount of the total energy released in the earthquake that is radiated seismically. A significant portion of the energy may be used in inelastic deformation within the fault zone and in overcoming frictional and gravitational forces. Estimates of $\eta$ vary from 0.01 to as high as 0.15 (Wyss, 1970a; Savage and Wood, 1971; Wyss and Molnar, 1972). Such a range of $\eta$ makes interpretation of ambient stress from apparent stress a difficult procedure at
best. Coupled with large uncertainties in $E_s$, the value of apparent stress for determining ambient stress is questionable. Estimating $E_s$ from either $m_b$ or $M_s$ is an attempt to integrate the entire seismic energy spectrum from the value of the spectrum at one point (at a period of 1 sec or 20 sec for $m_b$ or $M_s$, respectively). In theory, the entire seismic spectrum should be integrated to obtain $E_s$. As a routine procedure this is impractical. When this has been done for selected events, the calculated seismic energy is often reasonably close to the estimate from equations (3.6) to (3.7) (Brune and Allen, 1967; Wyss, 1970c), although overestimation of $E_s$ by a factor of 10, using (3.6)-(3.7), has been observed (Hanks and Wyss, 1972). The uncertainty in $E_s$, either from integration of the seismic spectrum where uncertainties in high frequency attenuation in the crust may be large, or from equations (3.6)-(3.7), may be as large as a factor of 10.

Accepting the large uncertainties, the average value of about 1 bar for $\eta \sigma$ observed for intraplate earthquakes can be cautiously extrapolated to a lower bound estimate of the ambient average stress in the plates on the order of hundreds of bars. The arguments in favor of ambient stresses on this order are by no means compelling. The most important point is that ambient stresses of several kilobars are not required
to satisfy the apparent stress measurements. This may be important to the question of how efficient the subducted slab is at transmitting its potential of several kilobars of tensile stress to the overriding plate.

Interpretation of stress drops is not encumbered with the uncertainty associated with $\eta$, but is nevertheless a complicated issue. The average stress drop measured for intraplate earthquakes is on the order of a few tens of bars. There are two major questions regarding stress drop. The first has to do with the final shear stress $\sigma_0$ acting on the fault. If the final stress drops to zero, then $\Delta\sigma$ is an estimate of the ambient stress acting over the fault area. This may be the case for large events where frictional heating may produce melting in the fault and lower the frictional stress to zero (McKenzie and Brune, 1971). It is also possible that the final stress is not zero, in which case $\Delta\sigma$ is only a partial stress drop. In this case, the ambient stress field may be anywhere from tens of bars to kilobars.

The second, and more serious, question about stress drop is based on the fact that the seismic stress drop is a static, far field measurement. As such, it represents the stress drop averaged over time and over the entire area of the fault that fails. Dynamically, stress drops near the rupture front may be much higher
than the average over the entire fault. Also an earthquake may represent the fracture of isolated asperities within the fault. The stress drop may be large for each asperity that fails, but the average stress drop may be only a few tens of bars when the entire area of the fault is considered. For both of these reasons, the seismically measured stress drop is at best a lower bound estimate of the ambient stress field (Madariaga, 1977).

In conclusion, apparent stress and stress drop for intraplate earthquakes are on the order of bars and tens of bars, respectively. No evidence is found supporting the suggestion that intraplate environments are characterized by generally higher stresses than plate boundary regions. Interpreting $\eta_0$ and $\Delta \sigma$ in terms of ambient stresses indicates that tectonic stresses are bounded from below by tens of bars but may range from tens of bars to kilobars. At best, $\eta_0$ and $\Delta \sigma$ are consistent with a wide range of stress models.
TABLE 3.1. INTRAPLATE EARTHQUAKES USED IN THIS STUDY OF APPARENT STRESS AND STRESS DROP

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TABLE 3.2. SEISMIC MOMENTS CALCULATED FROM LONG-PERIOD SH WAVE SPECTRA

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$^a$ Subjective quality of determination of long period spectral level, $\Omega_o$. 
TABLE 3.3. APPARENT STRESSES FOR INTRAPLATE EARTHQUAKES

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<th>$(\overline{\eta \bar{\sigma}})^s$, bars</th>
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\( a \) ISC values for \( m_b \) unless otherwise referenced.

\( b \) Unless otherwise referenced, \( M_S \) is calculated for this study using \( M_S = \log(A/T) + 1.66 \log \Delta + 3.3 \), where \( A \) is the maximum zero to peak amplitude of the long period vertical component of the Rayleigh wave in microns in the period range of 18 to 22 seconds and \( \Delta \) is the distance in degrees.

\( c \) Moment calculated for this study unless otherwise referenced.

\( d \) USCGS

\( e \) Hashizume (1974)

\( f \) Hashizume (1973)

\( g \) Mitchell (pers. comm., 1976)

\( h \) Bhattacharya (1975)

\( i \) Mendiguren (1971)

\( j \) Fitch et al. (1973)

\( k \) Singh et al. (1975)

\( l \) Forsyth (1973b)

\( m \) Hashizume and Tange (1977)

\( n \) Stein and Okal (1978)

\( o \) Hasegawa (1977)

\( p \) Hashizume (1977)

\( q \) Mills and Fitch (1977)

\( r \) Patton (1978)

\( s \) Chen and Molnar (1977)
TABLE 3.4. STRESS DROPS FOR INTRAPLATE EARTHQUAKES

<table>
<thead>
<tr>
<th>Event</th>
<th>$M_o \times 10^{25}$ dyne-cm</th>
<th>Length, km</th>
<th>$\Delta \sigma$ (bars)$^a$</th>
</tr>
</thead>
<tbody>
<tr>
<td>5</td>
<td>0.013</td>
<td>7 (Rayleigh wave amplitude spectra)$^b$</td>
<td>2</td>
</tr>
<tr>
<td>8</td>
<td>6.4</td>
<td>27 (aftershocks)$^c$</td>
<td>13</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>25$^d$</td>
</tr>
<tr>
<td>10</td>
<td>6.1</td>
<td>37 (surface cracks)$^e$</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>45-120$^e$</td>
</tr>
<tr>
<td>12</td>
<td>12.</td>
<td>40 (finite rupture velocity)$^f$</td>
<td>7</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>6-20$^f$</td>
</tr>
<tr>
<td>14</td>
<td>9.0</td>
<td>8 (aftershocks)$^g$</td>
<td>70</td>
</tr>
<tr>
<td>21</td>
<td>0.057</td>
<td>10 (aftershocks)$^h$</td>
<td>1$^h$</td>
</tr>
</tbody>
</table>

$^a$ Stress drop calculated from equations (3.2) and (3.3), assuming $L/w = 2.5$.

$^b$ Hashizume (1974)

$^c$ Fairhead and Girdler (1971)

$^d$ Maasha and Molnar (1972)

$^e$ Fitch et al. (1973)

$^f$ Singh et al. (1975)

$^g$ Main shock used as master event to relocate aftershocks and estimate $L = 8$ km, $w = 12$ km.

$^h$ Main shock used as master event to relocate aftershocks and estimate $L = 10$ km, $w = 8$ km (Mills and Fitch, 1977).
TABLE 3.5. SUMMARY OF APPARENT STRESSES FOR VARIOUS TECTONIC ENVIRONMENTS

<table>
<thead>
<tr>
<th>Environment</th>
<th>No. of Events</th>
<th>Apparent Stress, bars</th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td>Range</td>
<td>Mean</td>
<td>Mean</td>
</tr>
<tr>
<td>Intraplate</td>
<td>23</td>
<td>.004-3.0</td>
<td>.37</td>
<td>.72</td>
</tr>
<tr>
<td>Shallow Island Arc</td>
<td>34</td>
<td>.01-7.6</td>
<td>.27</td>
<td>.70</td>
</tr>
<tr>
<td>Intermediate Depth</td>
<td>14</td>
<td>.05-1.6</td>
<td>.34</td>
<td>.54</td>
</tr>
<tr>
<td>Deep</td>
<td>8</td>
<td>.10-.52</td>
<td>.25</td>
<td>.29</td>
</tr>
<tr>
<td>Ridge Crest</td>
<td>8</td>
<td>.08-1.7</td>
<td>.37</td>
<td>.55</td>
</tr>
<tr>
<td>Transform Fault</td>
<td>10</td>
<td>.07-1.7</td>
<td>.23</td>
<td>.37</td>
</tr>
</tbody>
</table>
FIGURE CAPTIONS

Figure 3.1  Epicenters of intraplate earthquakes used in this study of apparent stress and stress drop. Event numbers are from Table 3.1. Solid lines are plate boundaries. Triangles denote thrust faulting, squares - normal faulting, and circles - strike-slip mechanisms. The horizontal projections of P and T axes are shown where well determined from the fault plane. Cylindrical equidistant projection.

Figure 3.2  Amplitude spectrum vs. frequency for WWSSN stations HLW, TAB, and TRN for event 9 (M_S = 5.8, see Tables 3.1 and 3.2), showing various quality determinations; log-log scales. Also shown is the frequency corresponding to period T = 20 seconds, showing that for M_S ≤ 6.0, T = 20 seconds is very near the level amplitude portion of the spectrum.

Figure 3.3  I.S.C. m_b vs. M_S for intraplate earthquakes. Numbered events correspond to earthquakes from Table 3.1. The solid line represents the Gutenberg-Richter m_b-M_S relationship from equation (3.7).
Figure 3.4  Relationship between seismic moment $M_0$ and body wave magnitude $m_b$ for intraplate and interplate earthquakes. Straight lines represent constant apparent stress $\eta\bar{C}$. Intraplate earthquake data are from Table 3.3. Interplate earthquake moments are from Wyss (1970a,b), Wyss and Molnar (1972), and Molnar and Wyss (1972). All $m_b$ are from the I.S.C.
Figure 3.1
Figure 3.2

- **Quality A**: Mo, T = 20 sec
- **Quality B**: TAB, T = 20 sec
- **Quality C**: TRN, T = 20 sec
Figure 3.4

APPARENT STRESS ($\eta_\theta$)

Log $M_0$

$m_b$

$\eta_\theta = 0.1$ bars

- = intraplate
- = shallow island arc
- = intermediate depth
- = deep
- = ridge crest
- = transform fault
CHAPTER 4: FINITE ELEMENT MODELING OF STRESS IN A SINGLE PLATE

SECTION 4.1 INTRODUCTION

The intraplate stress field serves as a test of models for plate driving forces acting on the lithosphere. Recent progress into the nature of the driving mechanism has been made with a finite difference analysis of the entire lithosphere using a fairly coarse grid (Solomon et al., 1975; Richardson et al., 1976; Solomon et al., 1977; see also Chapter 5). In this chapter a finite element analysis using the wave front solution technique is applied to models of the intraplate stress field in a single plate using a refined grid. Plate driving forces acting on the edges and base of the lithosphere may stress the plate. Finite element analysis of the intraplate stress field should help in placing limits on the relative importance and absolute magnitude of forces due to ridges, subducted slabs, and drag along the base of the lithosphere as the plate moves over the mantle. The Nazca plate is chosen for this initial study because ridge, subduction, and transform fault-type boundaries are all represented. The Nazca plate is relatively small, so that understanding the state of stress may require a technique with the ability to refine the grid to follow irregular plate geometries.

A portion of the Nazca plate is assumed to be in a
state of nearly east-west compression as inferred from the source mechanism of a thrust earthquake on 25 November 1965 and two other earthquakes consistent with the 1965 event (Mendiguren, 1971). Ridge pushing forces are required of all models consistent with the inferred intraplate stress field. The net pulling force exerted on the oceanic plate due to the subducted slab is at most comparable to ridge pushing forces. Based on the potential forces associated with the ridge, an estimate of a few hundred bars of horizontal deviatoric stress is suggested for intraplate regions.

Large plate boundary earthquakes may affect the state of stress at great distances into the plate. On 22 May 1960 a very large plate boundary earthquake ($M_s \sim 8.3$) occurred on the Peru-Chile segment of the Nazca-South America plate boundary. An estimated 1000 km of that plate boundary broke with average displacement between 20 and 30 m. Finite element analysis is used to study the effect of the 1960 earthquake on the Nazca intraplate stress field. Near the 1960 earthquake the intraplate stress field may change at most by a few tens of bars once the asthenosphere has relaxed. At greater distances into the plate, the change in stress is on the order of one bar. Such small changes in the intraplate stress field are probably not noticeable unless the lithosphere is very near failure.
SECTION 4.2 METHOD

A linear elastic medium may be modeled in a straightforward manner using the displacement method of finite elements. The displacement method is based on the principle of virtual displacements, which states that for any small, arbitrarily imposed virtual displacement, the internal virtual work is equal to the external virtual work. For an elastic medium, this is equivalent to minimizing the total potential energy of the system. This variational principle is a statement that the elastic body is in equilibrium; that is, all external loads are balanced by internal strains. The variational principle may be written as (Zienkiewicz, p. 26, 1971):

\[ \delta (W_I + W_E) = 0 \]  \hfill (4.1)

where \( W_I \) and \( W_E \) are the internal and external work, respectively, and \( \delta \) implies an arbitrarily small variation in displacement.

The variational principle may be rewritten for an elastic medium in terms of the virtual displacements \( \delta U \) and associated virtual strains \( \delta \varepsilon \) as (Bathe and Wilson, p. 87, 1976).

\[ \int_V (\delta \varepsilon) \sigma \, dV - \{ (\delta U_C) F_C + \int_V (\delta U_b) F_b \, dV + \int_S (\delta U_s) F_s \, dS \} = 0 \]  \hfill (4.2)
where \( V, S \) are the volume and surface of the medium, respectively, and the subscripts \( c, b, s \) refer to a point, the body and the surface of the medium, respectively. The \( \delta U \) terms are the virtual displacements and \( \delta \varepsilon \) is the associated internal virtual strain. The \( F \) terms are forces acting in and on the medium. The bracketed term is the total external virtual work and the first term in 4.2 is the total internal virtual work.

The approximation made in the finite element method is to divide a continuous medium into a discrete number of regions, each with a finite area, that are connected at a finite number of points. Each region is known as an element and the connection points between elements are known as joints or nodes. For each element, the displacement at every point in the interior of the element is approximated by some function of the displacements at the nodes. The internal displacements are usually cast as polynomial functions of the nodal displacements. The order of the polynomial depends on the number of nodes describing the element. This relationship for element \( i \) may be written as (Bathe and Wilson, p. 88, 1976):

\[
\{ U_i (x) \} = [P_i (x)] \{ U \} \tag{4.3}
\]

where \( x \) is position, \( \{ U_i (x) \} \) is a vector containing the internal displacements, \( \{ U \} \) is a vector containing the
nodal displacement and \([P_i(\vec{x})]\) is a matrix containing a system of polynomial relationships between the internal displacements \(\{U_i(\vec{x})\}\) and the various nodal displacements \(\{U\}\). The dimension of \(\{U\}\) is equal to the number of nodes times the number of types of displacement at each node. The possible types of displacement are called degrees of freedom. The dimension of the \(\{U_i(\vec{x})\}\) vector is equal to the number of degrees of freedom of a point in the interior of the element. Correspondingly, matrix \([P_i(\vec{x})]\) has dimensions of the number of degrees of freedom at a point by the total number of degrees of freedom for the element.

The strain at any point in an element is a function of the internal displacements in the element. Since the internal displacements may be written as a function of the nodal displacements using (4.3), it follows that the internal strains may be written as a function of nodal displacements. In the absence of initial strains, this may be written as (Bathe and Wilson, p. 88, 1976)

\[
\{\varepsilon_i(\vec{x})\} = [B_i(\vec{x})]\{U\} \tag{4.4}
\]

where \(\{\varepsilon_i(\vec{x})\}\) is a column matrix containing the internal strains and \([B_i(\vec{x})]\) is the strain operator matrix containing, in general, derivative operators on the nodal displacements.

Finally, through some stress-strain relationship,
the element stresses may be obtained. In the absence of initial stresses, the element stress vector may be written as (Bathe and Wilson, p. 88, 1976)

$$\{\sigma_i(\vec{x})\} = [C_i(\vec{x})]\{\epsilon_i(\vec{x})\}$$

$$= [C_i(\vec{x})][B_i(\vec{x})] \{U\}$$ (4.5)

where $$\{\sigma_i(\vec{x})\}$$ gives the stress at any point in the element $$i$$ and $$[C_i(\vec{x})]$$ is the elasticity matrix.

The variational principle in (4.2) may now be approximated as a sum over all the elements defining the medium. In this way, the arbitrary virtual displacement $$\delta\{U\}$$ may be isolated. The expression becomes (Bathe and Wilson, p. 89, 1976)

$$n \delta\{U\}^T \{[\sum_{i=1}^{n} \int_V [B_i]^T[C_i][B_i]dV] \{U\} -$$

$$\int_\Omega (F_c + \int_V [P_i]^T F_{dV_i} + \int_{S_i} [P_i]^T F_{dS_i})\} = 0$$ (4.6)

where the notation for the dependence on position $$\vec{x}$$ has been dropped and where $$n$$ is the total number of elements and the superscript $$T$$ refers to the transpose of the vector or matrix. The terms in this equation may be rewritten in shortened form as

$$\delta\{U\}^T \{[K]\{U\} - \{F\}\} = 0$$ (4.7)
where

\[ [K] = \sum_{i=1}^{n} [K_i] = \sum_{i=1}^{n} \int_{V_i} \{B_i\}^T[C_i]\{B_i\}dV_i \quad (4.7a) \]

and

\[ [F] = \sum_{i=1}^{n} \left( F_c + \int_{V_i} \{P_i\}^T F_b dV_i + \int_{S_i} \{P_i\}^T F_s dS_i \right) \quad (4.7b) \]

and \([U]\) is a column matrix containing all of the nodal displacements for the structure. The matrix \([K]\) is called the stiffness matrix and \([F]\) is the load matrix. The principle of virtual displacements can finally be applied to (4.7). Since \(\delta\{U\}^T\) is an arbitrary virtual displacement, it follows that the term in the brackets must be equal to zero. This gives rise to the equilibrium condition for a finite element approximation to an elastic medium (Bathe and Wilson, p. 89, 1976)

\[ [K][U] = [F] \quad (4.8) \]

where \([K]\) is the global stiffness matrix for the entire structure, \([U]\) is the matrix containing the nodal displacements for the entire structure in the global coordinate system and \([F]\) is the matrix containing all loading information for the structure. Element properties are usually defined in a local coordinate system natural for the element. For a triangular element, the
local coordinate system is conveniently defined as a plane with one axis parallel to one side of the triangle. Before equation (4.8) can be solved, a global coordinate system common to all of the elements must be defined. A transformation of coordinates between local and global coordinate systems is generally required for each element.

In general, the external loads are known, the stiffness matrix may be formed once the geometry and elastic properties of the system are known, and equation (4.8) must be solved for the unknown nodal displacements. There are numerous techniques available to solve this matrix equation. One particular solution technique will be discussed later.

The evaluation of equations (4.1)-(4.8) will now be considered for the constant-strain plane-stress triangular element. This element will be used to model the state of stress in the Nazca plate.

The element chosen for the study of intraplate stress is the constant-strain triangle. The lithosphere away from plate boundaries may be modeled as an elastic medium in a membrane state of stress (Turcotte, 1974). In this case, the constant-strain triangle, in a plane state of stress with only in-plane displacement degrees of freedom, may be used to model the state of stress. The constant-strain triangle was one of the first element types developed (Wilson, 1965). More
efficient types of element have since been developed, but for modeling an elastic lithosphere that is in static equilibrium, the constant-strain triangle has several advantages. First, the earth's surface is nearly spherical. A triangular element, unlike a quadrilateral element, will be trivially planar when the nodes are placed on the earth's surface. Second, to be able to cover the earth's surface with a resolution of about five degrees of latitude on a side requires about five thousand triangles with about six thousand degrees of freedom. This is already a very large problem. Higher order elements would increase the number of degrees of freedom, making an economical solution impossible.

Consider a typical triangular element. Each vertex is a node with two degrees of freedom, for a total of six degrees of freedom for the element. The two displacement degrees of freedom $u$ and $v$ for a point in the interior of the element are functions of the six nodal degrees of freedom.

For the case of plane stress, the non-zero strains for the $i$th element may be written as (Zienkiewicz, p. 51, 1971)

$$
\{\varepsilon\}_i = \begin{bmatrix}
\varepsilon_x \\
\varepsilon_y \\
\varepsilon_{xy}
\end{bmatrix} = \begin{bmatrix}
\frac{\delta u}{\delta x} \\
\frac{\delta v}{\delta y} \\
\frac{\delta u}{\delta y} + \frac{\delta v}{\delta x}
\end{bmatrix} = [B_i][U_i] \quad (4.9)
$$
where \([B_i]\) is the strain operator matrix for element \(i\). Because the internal displacements are linear in \(x\) and \(y\), the \([B_i]\) matrix will depend only on the area of the element and the relative difference in the locations of the nodes. That is, the strain will not depend on the location within the triangle. It is for this reason that the element is called a constant-strain triangle.

For the earth's lithosphere, the non-hydrostatic state of stress in the interior regions of the plate is assumed to be approximated by a state of plane stress. In the plane-stress state, only the three in-plane stresses may be non-zero. If the further assumption is made that the material is isotropic, Hooke's law relates stress and strain. This may be written in the matrix notation of (4.5) as (Zienkiewicz, p. 53, 1971)

\[
[C_i] = \frac{E}{1-\nu^2} \begin{bmatrix}
1 & \nu & 0 \\
\nu & 1 & 0 \\
0 & 0 & \frac{1-\nu}{2}
\end{bmatrix}
\]

(4.10)

where \(E\) is Young's modulus and \(\nu\) is Poisson's ratio.

The stiffness matrix \([K_i]\) for the \(i^\text{th}\) element can now be calculated using (4.7a) as

\[
[K_i] = \int_S [B_i]^T [C_i] [B_i] h \, dx \, dy
\]

(4.11)
where the integral over the coordinate direction normal to the surface of the element has been replaced by the thickness h. Once the local element stiffness matrix is known, it must be transformed into the global coordinate system and then added to the global stiffness matrix.

The solution of the equilibrium condition given in equation (4.8) for the Nazca plate is based on the wave front solution technique (Irons, 1970; Orringer, 1974). This method is based on Gauss elimination and utilizes both in-core and external computer storage space. In the wave front method the stiffness matrix is assembled element by element until all of the elements containing a particular degree of freedom j have been assembled. This degree of freedom will have a zero stiffness coefficient $k_{ij}$ for all subsequent degrees of freedom i that will be assembled when new elements are considered. At this stage, the degree of freedom j is eliminated and the back substitution coefficients computed from $k_{ij}$ are transferred to external computer storage. New elements are added and more degrees of freedom are removed, once all non-zero stiffness coefficients have been assembled for the particular degree of freedom. This process continues until the last degree of freedom is eliminated. Then the back substitution phase is begun, and each degree of freedom is determined in the reverse order from that by which it was eliminated. The required equations that were sent to external storage are recalled
one at a time until finally the first degree of freedom is recalled and calculated.

The maximum number of degrees of freedom that must be assembled at one time before a degree of freedom can be eliminated is the front width. The front width depends on the element numbering scheme, and not the nodal numbering scheme. For Nazca plate models, a front width of 30 and approximately 5 seconds CPU time on an IBM 370/168 were needed to solve a problem with approximately 250 degrees of freedom.

The implementation of the finite element analysis with the wave front solution technique involves the use of EGL (Element Generator Library) and FRAP (Frontal Analysis Program), made available through the Aeronautical and Structures Laboratory at M.I.T. (Orringer and French, 1977; Orringer, 1974). EGL calculates the local stiffness matrix and FRAP consists of two subroutines that eliminate each degree of freedom and perform the Gauss back substitution phase.
4.2.1 Test case with an analytic solution

In this section an analytic solution is developed from the general membrane equilibrium stress equation for a special case of a thin spherical shell with zero radial displacement. The planar element used on the sphere has, by definition, no displacement perpendicular to the plane of the element. The analytic solution consists of a thin shell with latitudinal forces chosen in such a manner that the radial displacement is identically zero.

The equilibrium membrane equations for a spherical membrane with no azimuthal loading can be expressed as (after Kraus, p. 100, 1967):

\[
\frac{\partial N_\phi}{\partial \phi} + 2 \cot \phi N_\phi + R(q_\phi + q \cot \phi) = 0 \quad (4.12a)
\]

\[N_\phi + N_\theta + qR = 0 \quad (4.12b)\]
where $\phi, \theta$ are colatitude and longitude, respectively, $N_\phi, N_\theta$ are stress resultants, $q_\phi, q_\theta$ are the colatitudinal and radial tractions, respectively and $R$ is the radius to the midplane of the spherical shell. Equation 4.12a is of the general form

$$\frac{3 N_\phi(\phi)}{\sin^2 \phi} + f(\phi) N_\phi(\phi) + g(\phi) = 0. \tag{4.12c}$$

Equation 4.12c has a solution for $N_\phi$ of the form

$$N_\phi = [C - \int g \exp(\int f(\phi)d\phi)d\phi] \exp(-\int fd\phi) \tag{4.12d}$$

which leads to the following expression for $N_\phi$

$$N_\phi = \frac{C}{\sin^2 \phi} - \frac{R}{\sin^2 \phi} \int_0^\phi (q \sin \phi \cos \phi + q_\phi \sin^2 \phi) d\phi \tag{4.12e}$$

where $q$ and $q_\phi$ are as yet unspecified and $C$ is a constant of integration to be determined later.

The specific analytic solution we seek has a radial displacement equal to zero. To determine $q, q_\phi$ in 4.12e, we must rewrite 4.12e in terms of displacements and apply the condition that the radial displacement be zero. The stress resultants can be expressed in terms of strains.
as (Kraus, p. 88, 1967)

\[ N_\phi = K(\epsilon_\phi + \nu \epsilon_\theta) \]  
\[ N_\theta = K(\epsilon_\theta + \nu \epsilon_\phi) \]  

(4.13a)

where \( K = \frac{E \cdot h}{(1-\nu^2)} \), \( \epsilon_\phi, \epsilon_\theta \) are strains, \( h \) is the plate thickness, \( E \) is Young's modulus and \( \nu \) is Poisson's ratio. The strains can be expressed in terms of displacements as (Kraus, p. 101, 1967):

\[ \epsilon_\phi = \frac{1}{R} \left( \frac{\partial u}{\partial \phi} + w \right) \]  
\[ \epsilon_\theta = \frac{1}{R \sin \phi} \left( u \cos \phi + w \sin \phi \right) \]  

(4.13b)

where \( u, w \) are colatitudinal and radial displacements, respectively. Isolating \( u \) in equation 4.13b and using 4.13a and 4.12b gives

\[ \frac{\partial u}{\partial \phi} - u \cot \phi - \frac{R(1+\nu)}{E h} (2N_\phi + qR) = 0 \]  

(4.13c)
which is again of the form of equation 4.12c. Thus,

\[ u = \sin \phi [C_1 + \frac{R(1+\nu)}{E \cdot h} \int_{0}^{\phi} \frac{q R^2 + 2 N}{\sin \phi} d\phi] \quad (4.13d) \]

where \( N_\phi \) depends on \( q, q_\phi \) from equation 4.12e and \( C_1 \) is a constant of integration. Isolating \( w \) from 4.13b and using 4.13a and 4.12a gives:

\[ w = \frac{R}{E \cdot h} (-q R - (1+\nu)N_\phi) - u \cot \phi. \quad (4.13f) \]

Setting \( w=0 \) in equation 4.13f and expressing \( u \) and \( N_\phi \) in terms of the integrals given by equations 4.13d and 4.12e and specifying that \( u=0 \) at the top of the dome and relating \( q, q_\phi \) through 4.12b permits the non-unique solution of \( q, q_\phi \) as

\[ q = (1+\nu) \cos \phi \left[ \frac{\sin^2 \phi}{3+\nu} + \frac{a \sin^3 \phi}{4+\nu} \right] \quad (4.14) \]

\[ q_\phi = \frac{2 \sin \phi \cos^2 \phi}{3+\nu} - \frac{\sin^3 \phi}{4} + \frac{3 \sin^2 \phi \cos^2 \phi}{4+\nu} - \frac{a \sin^4 \phi}{5} \]

where

\[ a = - \frac{5}{4} \frac{4+\nu}{3+\nu} \]
The stress resultants and displacements are then given by:

\[ N_\phi = R\{-\cos\phi \sin^2\phi \left( \frac{1}{4} + a \frac{\sin\phi}{5} \right) \} \]

(4.15)

\[ N_\theta = R\{\cos\phi \sin^2\phi \left( \frac{1}{4} + a \frac{\sin\phi}{5} \right) -(1+v)\cos\phi \left( \frac{\sin^2\phi}{3+v} \right) \]

\[ + a \frac{\sin^3\phi}{4+v} \} \]

and

\[ u = \sin^3\phi (1-v^2) \left( \frac{1}{4 (3+v)} + a \frac{\sin\phi}{5 (4+v)} \right) \frac{R^2}{E h} \]

\[ w \equiv 0. \]

Because of the azimuthal symmetry of the problem, only a portion of a hemisphere need be considered. Fig. 4.1 shows a very coarse finite element grid with three divisions between the equator and the north pole used as a starting model for the analytic test case. The loads defined by eqn. 4.14 are specified as equivalent nodal forces and the displacement and stresses are calculated using the solution technique described in sec. 4.2.

The results are shown in Fig. 4.2-4.4 along with the analytic solution for the latitudinal displacement \( u_\phi = -u \) and the stresses \( \sigma_\phi, \sigma_\theta \) which are given by \( N_\phi/h \) and \( N_\theta/h \), respectively. For convenience, values for \( E \), \( h \), and \( R \) have been taken as \( 10^6 \) dyne/cm\(^2\), 1 cm, and 100 cm, respectively. Model D3 with 30 degree grid spacing clearly lacks the resolution to accurately model the displacement.
or stresses. However, refining the grid spacing to 10 degrees in model D10 and 5 degrees in model D19 results in vastly improved solutions. The 5 degree grid spacing corresponds to the grid spacing used in the next section for modeling the Nazca plate. The errors using this grid spacing for the analytic test case where the stresses vary as fourth powers of sines and cosines are less than 1% for displacements and stresses.
SECTION 4.3 NAZCA PLATE MODELS

The Nazca plate is reasonably well suited for stress analysis. The three major plate boundary types are represented in the Nazca plate, with the East Pacific Rise to the west and the Peru-Chile trench to the east. The Nazca-Cocos plate boundary to the north is a ridge-type boundary, while the Chile Rise to the south has a significant portion of transform fault-type boundary. The Nazca plate boundaries are shown in Figure 4.5 where ridges are denoted by a double line, subduction zones by a solid line, and transform faults by a dashed line. The boundaries have been taken from seismicity maps.

The tectonics of the oceanic Nazca plate may be less complicated than the tectonics of plates with significant amounts of continental crust. For continents, stresses associated with previous continent-continent collisions, mountain-building episodes, or back-arc accretion processes may mask stresses associated with present-day tectonics. The tectonics of oceanic plates in general may be more clearly related to present-day plate driving forces. A fossil ridge located in the Nazca plate may complicate interpretation of the intraplate stress field. The possible effects of the fossil ridge will be considered later.

The relatively small area of the Nazca plate facilitates the generation of a fairly refined grid and allows a wide range of models to be studied economically.
Previous modeling of stress in the earth's lithosphere as a whole using finite difference techniques with a coarse grid was not very successful in matching the inferred state of stress in the Nazca plate (Richardson et al., 1976). The finite element technique, as applied to a single plate analysis of the Nazca plate, may improve our understanding of the driving mechanism.

The state of stress for a large portion of the Nazca plate is unknown. For oceanic plates in general, intraplate earthquakes are the major source of information concerning the state of stress in the plate. There has been one relatively large intraplate earthquake in the Nazca plate for which a focal mechanism study has been made. Careful study of an event with an $m_b$ magnitude of 5.75 located 17.1° S, 100.2° W on 25 November 1965 using both body and surface wave data indicates an essentially east-west maximum horizontal compression axis (Mendiguren, 1971). The compression axis is assumed to bisect the angle between the two focal planes. Laboratory experiments indicate that rocks will fail under stress along a plane that is inclined less than 45° to the axis of maximum compression, depending upon friction in the failed plane. If the fault plane is frictionless, the maximum compression axis will be inclined 45° to the failed surface. If failure occurs along a pre-existing fault, the maximum compression axis may differ significantly
from the one inferred by the bisector of the focal planes. For the Nazca plate event, it is not possible to determine which of the focal planes was the fault plane. Nor is it possible to determine if failure occurred along a pre-existing fault. It is, therefore, assumed that failure occurred on a fresh, frictionless fault. The location of the event and the orientation of the inferred pressure axis are shown in Figure 4.5, along with the location of two events which occurred in 1944. The limited data available for the 1944 events are consistent with the 1965 focal mechanism solution (Mendiguren, 1971). While it is difficult to extrapolate these data to the entire plate, there is evidence that most oceanic lithosphere older than about 20 million years is under deviatoric compression (Sykes and Sbar, 1973). Most of the Nazca plate east of the 1965 and 1944 events is older than 20 million years (Herron, 1972). For driving force models discussed in this chapter, primary emphasis is placed on matching the 1965 east-west thrust mechanism and secondary emphasis is placed on matching a general deviatoric compressive state of stress near the 1944 events.

An assumption inherent in the modeling is that the inferred state of stress in the Nazca plate is the result of plate driving forces acting on the edges and base of the lithosphere. The topography and magnetic anomalies of the Nazca plate suggest that the ridge axis jumped from the
Galapagos Rise to the East Pacific Rise about six million years ago (Anderson and Sclater, 1972). The Galapagos Rise is located about 1500 km east of the East Pacific Rise and is elevated as much as a kilometer above the surrounding seafloor. This excess topography may affect the state of stress and complicate the interpretation of plate driving forces acting on the boundaries. The topographic high associated with the Galapagos Rise extends from about 5°S to 15°S. A profile along 18°S shows a much less developed depth anomaly of about a hundred meters (Anderson and Sclater, 1972). The 1965 event is located between the two rises, about 500 km west of the southern extreme of the topographic expression of the Galapagos Rise. The 1944 events are located on the eastern flank of the Galapagos Rise and their mechanism suggests that a general compressive state of stress extends across the fossil ridge.

The effect of the Galapagos Rise on the state of stress is uncertain, but may be small since the 1944 thrust events occur near the fossil ridge crest where the Galapagos Rise forces should be at a minimum. Thus because the 1965 and 1944 events occur in regions where other sources of stress may be present, caution is suggested in interpreting the inferred stress in terms of driving force models for the Nazca plate until further corroborative data are gathered.

There are other possible sources of stress in the plates, including thermoelastic effects in the cooling lithosphere and latitudinal motion of the lithosphere on
an ellipsoidal earth (Turcotte and Oxburgh, 1973; Turcotte, 1974). For the Nazca plate, where plate motion is primarily east-west, the effects of latitudinal motion are minimal. Thermoelastic stresses may be large, but there is reason to believe that the stresses may be relaxed on the grain boundary level over time (Solomon et al., 1975). A final assumption is that the stress inferred from the 1965 event is representative of the state of deviatoric stress over the thickness of the entire lithosphere. This assumption may be tested if the depths of future earthquakes can be well determined. Until that time, the assumption of uniform stress acting across the plate thickness will have to suffice.

The grid for the Nazca plate is shown in Figure 4.6. The constant-strain triangular element with two in-plane displacement degrees of freedom per node has been used throughout. Plate interior regions are divided into triangular elements that have dimensions of 5x5x7 degrees. The plate boundary nodes are defined approximately every three degrees along a digitized record of the boundaries. There are 210 elements using 132 nodes for a total of 264 degrees of freedom before boundary conditions are applied. Each element is assumed to have a thickness of 100 km. Although lithosphere thickens with age as it cools, the variation in plate thickness in the interior regions of the plate where stresses are calculated is probably small
enough to justify the uniform thickness assumption. The magnitude of the stresses in a 50 km thick plate will be twice as large as the stresses in a 100 km thick plate for the same forces acting on the boundaries. The elastic properties are assumed to be everywhere isotropic and uniform with values for Young's modulus $E$ and Poisson's ratio $\nu$ of $7 \times 10^{11}$ dyne-cm$^{-2}$ and 0.25, respectively. The calculated stresses are independent of the choice of $E$ for all models where boundary forces are specified.

Several possible boundary driving forces are included in the models and are shown schematically in Figure 4.7. A ridge pushing force $F_R$ per unit length of boundary is modeled as a nodal force in the direction of relative plate spreading taken from the RM1 model of Minster et al. (1974). Initial values for $F_R$ of $10^{15}$ dyne-cm$^{-1}$ are equivalent to a compressive stress of 100 bars across a 100 km thick plate and are consistent with other estimates of the potential ridge driving force due to excess topography at the ridge (Hales, 1963; Frank, 1972; Jacoby, 1970; McKenzie, 1972; Artyushkov, 1973; Bird, 1976).

The net pulling force per unit length at trenches $F_T$ may be much larger than $F_R$. Thermal models of the slab indicate density contrasts between the mantle and the slab that have the potential to exert a force per unit length as large as $5 \times 10^{16}$ dyne-cm$^{-1}$ on the oceanic plate. This is equivalent to a deviatoric tensile stress of five kilo-
bars across a 100 km thick plate, and is consistent with maximum estimates of potential slab forces (McKenzie, 1969a; Jacoby, 1970; Turcotte and Schubert, 1971; Solomon and Pau U, 1975). For the Nazca plate models, $F_T$ is input in the direction of relative plate motion. The value of $F_T$ varies from model to model.

There will be a large net torque acting on the Nazca plate if only ridge and subducted slab forces, both of which act primarily in an eastward direction, are included. The net torque acting on a plate must vanish, however, if the plate is to be in a state of mechanical equilibrium. While plate velocities have changed during geologic history, on a time scale of a few million years plate velocities generally appear to be nearly constant (McKenzie and Sclater, 1971; Pitman and Talwani, 1972; Molnar et al., 1975), suggesting that the earth's lithosphere, as well as each individual plate, is in a state of equilibrium.

There must be other forces acting on the Nazca plate that balance any torque due to pushing forces at ridges and pulling forces at trenches if the plate is in mechanical equilibrium. These forces may be transmitted across plate boundaries from more distant boundaries. In this case, it is not possible to solve for the state of stress in the Nazca plate uniquely as a function of Nazca plate forces. For a first order model of the Nazca plate, it may be reasonable to ignore distant plate boundary forces. Forces acting on the western edge of the Pacific plate are one to
two Nazca plate widths away from the Nazca plate and are separated by the East Pacific Rise. Ridge boundaries, as zones of weakness, probably do not transmit forces across the ridge crest very efficiently. This would tend further to reduce the effect of western Pacific plate boundaries. The nearest ridge boundary to the east is the mid-Atlantic ridge, nearly two plate widths distant. To the north, the Cocos and Caribbean plates may apply forces to the Nazca plate, but the effects are difficult to estimate since both plates and plate boundaries in question are small with respect to the Nazca plate. It will be assumed that the contribution of the Cocos and Caribbean plates to the torque balance of the Nazca plate can be ignored. The problem of distant plate boundary forces does not arise if the earth's entire lithosphere is considered. Then equilibrium is satisfied if the net torque on the whole lithosphere vanishes. A thorough study of the effects of various plate boundary forces on the state of stress in the interior of the plates should include the entire lithosphere. It is useful, however, as a first order model of an oceanic plate to study the Nazca plate.

Forces at transform faults may contribute to the net torque balance for the Nazca plate. Transform fault earthquakes generally have hypocenters with depths less than 10 km when depths can be constrained (Burr and Solomon , 1978). If the forces acting on transform faults are constrained to the uppermost 50 km to include the possible
contributions of both a shallow brittle zone and a shear zone at depth, then large stresses are required to balance reasonable values of plate boundary forces. For example, if the only boundary forces are at ridges with a force equivalent to 100 bars across a 100 km thick plate, the net torque is on the order of $7 \times 10^{32}$ dyne-cm. There is approximately 5000 km of transform fault boundary for the Nazca plate with most occurring along the Nazca-Antarctica plate boundary (see Figure 4.5). To balance the ridge force, shear stress on the order of several kilobars is required. While this value is possible, estimates of the maximum shear stress acting on the San Andreas fault are on the order of 100-200 bars (Brune et al., 1969). Transform faults alone are probably not sufficient to balance the torque due to both ridge and trench forces. Although transform faults may contribute to the total torque balance, models of the Nazca plate in this paper will not include imposed forces at transform faults.

Drag forces at the base of the plate due to the motion of the plate over a viscous mantle may contribute to the total torque balance. The Nazca plate has an area on the order of $10^{17}$ cm$^2$. To balance the torque considered above due to ridges alone, a shear stress on the order of a few bars acting on the base is required. This level of shear stress is consistent with estimates of drag stress given by

\[
\tau = \frac{nV}{H_a}
\]

(4.16)
where \( \tau \) is the shear stress on the base of the plate, \( h_a \) and \( \eta \) are the thickness and viscosity of the asthenosphere, respectively, and \( V \) is the plate velocity. Taking \( V = 10 \) cm/yr, \( h_a = 3 \times 10^7 \) cm and \( \eta = 10^{20} \) poise, \( \tau \) is 1 bar. Converting this amount of shear stress to an in-plane resistive force per unit area of \( L/h \) bars, where \( L \) and \( h \) are the plate length and thickness, respectively, produces 30 bars for the Nazca plate. Thus, drag forces have the potential to balance plate boundary forces.

The force due to drag is not treated as a free parameter, but will be chosen so that the torque due to drag balances boundary torques. The drag forces will be calculated assuming that drag is proportional to the absolute velocity of the plate. The direction of drag forces is assumed to be opposite to the absolute velocity so that drag will resist plate motion. A power law relationship may exist between velocity and drag (Stocker and Ashby, 1973). Results of previous modeling indicate that the intraplate stress field cannot be used to distinguish between linear and power law drag (Solomon et al., 1977). For Nazca models, a linear drag law is therefore assumed. The drag force is given by

\[
\vec{F}_D = -C \vec{V}_{\text{abs}} = -C \vec{\omega}_{\text{abs}} x_r \tag{4.17}
\]

where \( \vec{F}_D \) is the force per unit area, \( \vec{V}_{\text{abs}} \) and \( \vec{\omega}_{\text{abs}} \) are the absolute velocity and rotation pole for the plate, respec-
tively, $\mathbf{r}$ is the radius vector and $C$ is a constant. The value of $C$ is chosen such that a velocity of 1 cm/yr produces a shear stress of 1 bar, in qualitative agreement with the estimate of drag stress from equation (4.16).

Then, $\mathbf{\dot{w}}_{\text{abs}}$ is calculated to produce drag forces that will balance the torques due to any boundary forces. Only the product $CW_{\text{abs}}$ is constrained to balance the torques. The variation in the magnitude of $\mathbf{\dot{w}}_{\text{abs}}$ for various models may only reflect a variation in the choice of $C$. The value of $\mathbf{\dot{w}}_{\text{abs}}$ for each model is listed in Table 4.1.

The stiffness matrix $[K]$ will be singular if rigid body motion of the plate is not constrained. The minimum boundary condition required to constrain rigid body motion involves constraining three suitably chosen degrees of freedom. For example, it is sufficient to constrain one grid point and one degree of freedom at another grid point that corresponds to rotation about the fixed point. Once all forces have been specified for a model such that the net torque vanishes, the solution for stress within the Nazca plate is not sensitive to the choice of points used to constrain rigid body motion.

For all the models in this study, the grid point at the Nazca-Antarctic-Pacific triple junction, and the longitudinal displacement of the grid point at the Nazca-Cocos-South America triple junction are constrained to be zero.

Loads associated with ridge and slab forces are applied to plate boundary elements at the nodes to the
immediate interior of plate boundary nodes. When whole earth models are considered in Chapter 5, this will enable forces due to a single plate boundary to act on the two adjoining plates. For models considered in this study of the Nazca plate, the effect is to have the boundary forces acting on the interior edge of the plate boundary element rather than at the plate edge. The actual plate boundary nodes are free to deform and the plate edge is approximately a free surface. If boundary nodes are constrained, a significant portion of the ridge and slab loads artificially deform the boundary elements and less force is transmitted to the interior of the plate. The orientation of the calculated stresses are very similar whether or not the boundary nodes are constrained. With the boundary nodes free, the load exerted on the interior of the plate is the net load due to the driving mechanism. Since the interior nodes of the boundary elements parallel the plate boundary nodes, the effect of having the edge free and applying forces at the interior nodes on the intraplate stress in the interior regions of the model will be small.

Once the nodal loads corresponding to forces at ridge and trench boundaries have been chosen for a particular model of the driving mechanism, the balancing drag forces may be determined and the state of stress in the Nazca plate can be calculated using the wave front
technique. The calculated model stresses, although referred to as deviatoric stresses, are measured with respect to the vertical stress. In classical continuum mechanics, deviatoric stresses are measured with respect to the hydrostatic stress, rather than the vertical stress. It is implicitly assumed that the vertical stress at any depth does not vary spatially and that the difference between the horizontal and vertical stress is constant across the thickness of the plate. Although the term deviatoric stress, as used in this thesis, may create some confusion, it is retained to emphasize that a calculated tensile stress does not necessarily imply net tension in the earth. Particular model cases will be presented next.

4.3.1 Driving force models

Models of the state of stress in the Nazca plate involve varying the possible driving forces $F_R$, $F_T$, and $F_D$. Even with the simplifying assumptions about the nature of the forces $F_R$, $F_T$, and $F_D$, there are still many possible models. Various models will now be considered.

In the first set of models, ridge, trench, and drag forces ($F_R$, $F_T$, and $F_D$, respectively) are considered. When $F_T=0$, and $F_R$ equivalent to a force per unit length of $10^{15}$ dyne-cm$^{-1}$ is applied to the ridge boundaries shown in Fig. 4.5, the predicted stress field is as shown for model R100 in Figure 4.3 were stresses in adjacent
elements forming a quadrilateral have been averaged and assigned to the midpoint of the common boundary. The magnitude of the stresses in the interior of the plate is on the order of 40 to 100 bars, comparable to the value of $F_R$ averaged across the plate thickness. The magnitude of the intraplate stresses scales linearly with the choice of values for $F_R$. The plate is almost entirely in a state of deviatoric compression, with the maximum horizontal compressive stress axis trending nearly east-west. The error in azimuth between the calculated stress and the inferred state of stress from the 1965 earthquake is less than $10^\circ$. This is less than the uncertainty in the orientation of the principal stresses from a fault plane solution.

For comparison, model T100 with $F_R = 0$, and $F_T = 10^{15}$ dyne-cm$^{-1}$ for trench boundaries shown in Figure 4.5, is shown in Figure 4.9. The entire plate is in a state of deviatoric tension. The magnitude of the stresses vary from 100 bars in the eastern part of the plate to 30 bars in the west. Near the epicenter of the 1965 thrust earthquake, the calculated maximum horizontal deviatoric stress is tensile and oriented nearly east-west. This model does not satisfy the data.

Models that have $F_R$ equal to $10^{15}$ dyne-cm$^{-1}$ and small amounts of $F_T$ are qualitatively similar to models with $F_T = 0$ until $F_R/F_T \approx 2$, when the orientation of the principal stresses begins to vary for small changes in the ratio $F_R/F_T$. 
and the eastern part of the plate begins to show deviatoric tension. Model R100T100 with both \( F_R \) and \( F_T \) equal to \( 10^{15} \) dyne-cm\(^{-1}\) is shown in Figure 4.10. The deviatoric stress near the 1965 event has dropped from about 70 bars for the pure ridge model to about 50 bars for this model. The azimuthal error has increased, but is still acceptably small. For \( \frac{F_R}{F_T} = 1.0 \), the azimuths of the principal stresses vary most rapidly with changing \( \frac{F_R}{F_T} \) for interior regions of the plates. This is not surprising since the length of ridge and trench boundaries is approximately equal and most of each boundary type trends north-south.

The best whole earth models of Richardson et al. (1976) centered around a ratio of ridge-to-trench driving forces near unity. The finite element analysis clearly shows that this is precisely the region in \( \frac{F_R}{F_T} \) space when small uncertainties in \( \frac{F_R}{F_T} \) or the effect of distant forces may produce rapid change in the orientation of the principal stresses. This may account for some of the inability of the finite difference models to match the state of stress in the Nazca plate.

Model R50T100 with \( \frac{F_R}{F_T} \) equal to 0.5 has most of the plate under deviatoric tension and is shown in Figure 4.11. The sense of the stress near the 1965 earthquake can no longer be accommodated. The magnitude of deviatoric stresses near the 1965 event is small, on the order of 15 bars. Decreasing \( \frac{F_R}{F_T} \) below 0.5 further degrades the comparison between calculated and inferred stresses. A lower
limit of \( F_R/F_T \approx 0.5 \) is established for models that are consistent with the 1965 earthquake.

Some models were calculated for cases where ridges pull on the plate and slabs push on the plate, exactly opposite to the cases considered so far. Models where the slab push exceeds the ridge pull also match the data fairly well. This basic non-uniqueness is not surprising, since both boundary types run essentially north-south. These models have been discarded, however, because no physical mechanism has been suggested that will allow the subducting lithosphere to push on the oceanic plate.

4.3.2 Large plate boundary earthquakes

The effect of large plate boundary earthquakes on the intraplate stress field can be studied using the finite element techniques that were utilized for driving mechanism models. A very large (\( M_s \approx 8.3 \)) earthquake occurred along the Chile trench segment of the Nazca–South American plate boundary at approximately 42°S, 75°W on 22 May 1960. This earthquake is probably the largest earthquake ever recorded (Kanamori and Cipar, 1974), and provides an excellent opportunity to study the possible effects of large plate boundary earthquakes.

The location of the epicenter is shown in Figure 4.5, along with the portion of the plate boundary that ruptured (Plafker, 1972). Approximately 1000 km of plate boundary ruptured along a strike of N10°E. Focal mechanism studies, surface wave analysis and surface deformations indicate a
low angle thrust faulting event (Kanamori and Cipar, 1974; Plafker and Savage, 1970; Plafker, 1972). The dip angle is between 10° and 20° and the slip angle is near 90° indicating primarily thrust with a small amount of strike-slip motion possible. The depth of the event was less than 60 km, indicating that the earthquake occurred on the boundary between the Nazca oceanic plate and the South American continental plate. The average displacement along the fault has been estimated to be between 20 and 30 m using dislocation models and triangulation data (Kanamori and Cipar, 1974; Plafker, 1972). The seismic moment for the event is $2.7 \times 10^{30}$ dyne-cm (Kanamori and Cipar, 1974). The stress drop can be estimated from the seismic moment $M_0$ and the fault dimensions as (Aki, 1966)

$$\Delta \sigma = \frac{4(\lambda+\mu)M_0}{\pi(\lambda+2\mu)Lw^2} \quad (4.18)$$

where $L$ and $w$ are fault length and width, respectively, $\mu$ is the shear modulus and $\lambda$ is the Lamé constant. Taking values of $3 \times 10^{11}$ dyne-cm$^{-2}$ for $\lambda$ and $\mu$ and 1000 and 200 km for $L$ and $w$, respectively, gives a value of 58 bars for the stress drop.

An earthquake with magnitude, displacements, seismic moment, and stress drop as large as those observed for the 1960 Chilean earthquake may alter the intraplate stress
field for large areas of the Nazca plate.

Two types of models have been considered for the possible effect of the Chilean earthquake on the intraplate stress field for the Nazca plate. In the first, model CH30, a displacement of 30 m in the direction of fault motion has been specified along the grid corresponding to the 1000 km of plate boundary that failed. The maximum value of 30 m is used, assuming that all deformation occurred within the Nazca plate, to place an upper bound on the possible effects of the earthquake. All other boundary grid points are fixed and the stresses are calculated and shown in Figure 4.12. The plate is in a state of deviatoric tension, with the largest stresses on the order of 10 bars in the southeast corner of the plate near the applied displacements. A qualitative estimate of the stress due to a displacement of 30 m gives a strain over a 3000 km plate of $10^{-5}$ which implies a stress on the order of 10 bars for reasonable values of Young's modulus.

The response of the Nazca plate to the imposed displacement in model CH30 is that of an elastic plate over a completely relaxed viscous asthenosphere. In the earth, the effect of imposing a displacement in an elastic plate overlying a viscous asthenosphere is a time dependent visco-elastic problem. The time required for the effect of boundary displacement due to the Chilean earthquake to propagate approximately 3000 km across the Nazca plate can be estimated from a simple model of an elastic plate on a
viscous fluid overlying a rigid half-space. The displacement of a point in the interior of the plate due to a boundary displacement is given by the diffusion equation (Elsasser, 1969). The mean distance $x$ that a disturbance propagates in time $t$ is then given by

$$x \approx (4\kappa t)^{1/2}$$

(4.19)

where $\kappa$ is a constant given by $h h_a E/\eta$, $h$ and $h_a$ are the thickness of the elastic plate and viscous asthenosphere, respectively, $E$ is the Young's modulus of the plate and $\eta$ is the viscosity of the asthenosphere. Equation 4.19 can be solved for time $t$ given $x = 3000$ km, $h = 100$ km, $h_a = 300$ km, $E = 7 \times 10^{11}$ dyne-cm$^{-2}$, and $\eta = 10^{20}$ poise to obtain a characteristic time of several hundred years for a boundary displacement to travel across the Nazca plate.

The stresses shown in Figure 4.12 for model CH30 are the maximum stresses due to the Chilean earthquake displacements. On a time scale of several hundred years required for all of the plate to respond to the boundary displacement, other plate boundary earthquakes, minor variations in spreading rates, and the effects of distant boundary forces may mask deviatoric stresses on the order of 10 bars. Unless the lithosphere is very near the yield stress, it seems unlikely that the few bars due to the displacements of an isolated plate boundary earthquake will have much effect on the seismicity and state of stress in the interior of the plate.
A second approach to modeling the effect of the Chilean earthquake is to estimate the change in driving forces for the slab due to the earthquake. An upper bound is placed on the additional slab forces by the elastic strain energy released during the Chilean earthquake. Additional forces due to the slab cannot produce more strain energy in the Nazca plate than was made available during the earthquake.

The elastic strain energy \( W \) released during an earthquake can be estimated from the source parameters as (Knopoff, 1958)

\[
W = A \overline{D} \overline{\sigma} \tag{4.20}
\]

where \( A \) is fault area, \( \overline{D} \) is the average slip along the fault, and \( \overline{\sigma} \) is the average of the initial and final shear stress acting on the fault. If the stress acting on the fault drops to zero during an earthquake, the average stress is given by one-half the stress drop \( \Delta\sigma \). If the stress drop is complete, \( W \) is given by (Kanamori, 1977)

\[
W = \frac{1}{2} A \overline{D} \Delta\sigma = \frac{1}{2} \frac{M_o}{\mu} \Delta\sigma \tag{4.21}
\]

where \( M_o \) is the seismic moment and \( \mu \) is the shear modulus. Taking \( \Delta\sigma = 58 \) bars, \( \mu = 3 \times 10^{11} \) dyne-cm\(^{-2} \), and \( M_o = 2.7 \times 10^{30} \) dyne-cm, the total elastic strain energy \( W \)
for the Chilean earthquake is $2.4 \times 10^{26}$ dyne-cm.

The elastic strain energy $E_p$ for an idealized plate can be estimated using a simple analogy as

$$E_p = \frac{1}{2} \int_V \varepsilon \sigma dV$$  \hspace{1cm} (4.22)

where $V$ is the volume of the plate, $\varepsilon$ is the strain, and $\sigma$ is the stress acting on the edge of the plate. Assuming a Young's modulus of $7 \times 10^{11}$ dyne-cm$^{-2}$ and considering a plate 3000 km across by 1000 km long by 100 km thick, gives $E_p = 3 \times 10^{11} \sigma^2$.

Equating the elastic strain energy $W$ released during the earthquake with the elastic strain energy $E_p$ in deforming the oceanic plate provides an estimate of 35 additional bars acting across a 100 km thick plate. In equating $W$ and $E_p$, it has been assumed that the stress drop is complete. The elastic strain energy $W$ from equation (4.20) increases linearly with ambient stress if the stress drop is only a small fraction of the ambient stress. In such a case, equation (4.21) significantly underestimates the value of $W$. The value of $E_p$ also increases linearly with ambient stress if the stress drop is small compared to the ambient stress. Equating $W$ and $E_p$ in the limit as stress drop becomes very small with respect to the ambient stress results in an estimate of 20 bars for the additional stress acting on the plate. The estimate of 35 bars for $\sigma$
based on equating $E_p$ and $W$ is only valid if the stress drop is complete. There is some justification for assuming that the stress drop is complete for the Chilean earthquake. For large earthquakes, frictional heating on the fault surface may produce melting and reduce the frictional stress to zero (McKenzie and Brune, 1972). It is assumed that the stress drop is complete for the very large, $M_s \sim 8.3$ Chilean earthquake. The estimate of 35 bars is about one-half of the calculated stress drop for the Chilean earthquake. If the stress drop of 58 bars is used to determine additional slab forces, the strain energy in the Nazca plate exceeds the strain energy released during the earthquake by about a factor of four, assuming a complete stress drop.

A model with forces equivalent to 35 bars across a 100 km thick plate along the segment of the Peru-Chile trench that failed during the 1960 earthquake is shown in Figure 4.13. The intraplate stresses for model C35 are similar in orientation to the stresses for model CH30 where displacements are imposed along the same boundary segment. The magnitude of the stresses is several times greater for model C35 than for model CH30, with values of about 25 bars in the southeastern part of the plate and one bar near the epicenter of the 1965
earthquake. As for model CH30, this model represents the response of an elastic plate over a completely relaxed viscous asthenosphere. The asthenosphere relaxation time is on the order of several hundred years, and the stresses due to the 1960 earthquake may be masked by other processes acting during that period.
SECTION 4.4 DISCUSSION AND CONCLUSIONS

Static finite element analysis using the wave front solution technique has been used to model the state of stress in the Nazca plate. Although only static elastic models have been explicitly calculated, certain limiting cases of an elastic plate over a viscous asthenosphere have also been modeled. It is important to consider the geophysical assumptions that have been made in the modeling before discussing the conclusions.

The lithosphere has been treated as a linear elastic isotropic plate. This is an obvious oversimplification, but uncertainties in data, anisotropy, and failure criteria may not warrant more sophisticated modeling techniques. The thickness of oceanic lithosphere increases as the plate cools, varying most rapidly near the ridge crest (Forsyth, 1975). Model stresses are calculated for the interior regions of the Nazca plate, where variations in plate thickness are small compared to regions near the ridge crest. If the plate is thinner, the stresses will be somewhat higher, but the orientation of the stresses will remain similar. It might be useful to include a failure criterion. Plate boundary forces are input as nodal loads at grid points and adjoining elements always have the highest stresses. These elements are most likely to satisfy a failure criterion.

Static elastic analysis of the intraplate stress field
is satisfactory if boundary forces have not changed for a sufficiently long period of time for the viscous asthenosphere to have completely relaxed. The characteristic time has been shown to be on the order of several hundred years for the Nazca plate. Evidence of fairly uniform spreading rates for the Nazca plate for the last four to six million years suggest that the boundary forces may be modeled as static forces.

The model stresses represent the difference between the vertical and horizontal stresses in an elastic plate that is initially unstressed. The vertical stress is assumed to be constant everywhere at the same depth. In the lithosphere, the vertical stress may vary laterally due to topography or initial stresses. Since the complete stress tensor is required to predict the type of faulting, the model stresses cannot be used to determine the nature of failure without some assumption about the vertical stress. There is no ambiguity, however, in deciding whether the model horizontal deviatoric stresses are consistent with those inferred from earthquakes. For model stresses to be consistent with the 1965 earthquake, the maximum horizontal compressive deviatoric stress must be oriented essentially east-west.

Several simplifying assumptions about the nature of the driving forces have been made. Ridge and slab forces have been taken as constant per unit length of boundary. For the ridge, where gravitational forces are proportional
the elevation of the ridge, this assumption is reasonable. The depth to the ridge crest varies less than a kilometer for most of the ridge boundary. Slab forces, however, may vary regionally depending on the amount of slab under tension, the subduction rate, and the age of the subducted lithosphere. Fault plane solutions for earthquakes in the subducted slab show at least three regimes along the extent of the Nazca-South America plate boundary (Santo, 1969; Isacks and Molnar, 1971; Stauder, 1973, 1975). The northern section, from about 0° to 16° S, is characterized by a shallow zone, dipping 10-15° eastward to 200 km depth, of earthquakes with the T-axis down dip, a gap in seismicity between 200 and 500 km depth, and a cluster of events at about 600 km with P-axis down dip. Between 16° S and about 25°S, intermediate depth earthquakes are common in the depth range 75-150 km with T-axis down dip and no activity is known at great depths. South of 25°S, there is another trend similar to the northern section, except that the upper most zone dips at about 30° and extends to 300 to 350 km depth and the cluster of earthquakes at about 600 km is displaced about 500 km to the east.

It has been suggested that the depth to the transition in a subducted slab between events with T-axis down dip and events with P-axis down dip may be an indicator of the amount of force the slab exerts on the overriding plate (Chapple and Tullis, 1977). This hypothesis may be impor-
tant for other subduction zones, but cannot readily be applied to the Nazca plate. The crossover depth is not well determined for the three slab regimes described. In the north, it may be anywhere between 200 and 500 km. In the south, the transition may be anywhere between 350 and 500 km. In the middle, no transition can be defined. The possible variation in crossover depth along the subduction boundary does not justify more sophisticated modeling of slab forces.

The relative rotation pole for the Nazca-South America subduction plate boundary is located at 59.1°N, 91.4°W with a rate of 12.6 x 10^{-7} deg/yr (Minster et al., 1974). With this pole, the subduction rate only varies between 11 cm/yr in the north and 14 cm/yr in the south. The age of the lithosphere being subducted shows a decrease in age from Eocene in the north to Oligocene in the south (Herron, 1972). Older subducted lithosphere may be cooler, and hence denser, than young lithosphere and may exert a greater force per unit length of boundary. Also, a high rate of subduction may imply greater forces since more slab that is cold will enter the mantle. For the Nazca plate, the effects of lithosphere age and subduction rate compete against each other. If either effect dominates, it should be noticeable in a change in the depth to the crossover from T-axis down dip to P-axis down dip events in the slab. As previously discussed, this depth has an uncer-
tainty of about a factor of two along the trench axis and does not justify a more sophisticated model of slab forces than a constant force per unit length of boundary.

The most critical assumption made in the stress analysis is the nature of the forces balancing boundary force torques. For all models of possible driving mechanisms considered in this paper, drag forces along the base of the plate have been chosen to balance any boundary torques. Other forces may be important in the total torque balance. Specifically, plate boundary forces acting on other plates and resistance along transform faults may contribute to the net torque balance. Plate boundary forces at great distances such as the western Pacific subduction zones and mid-Atlantic ridge may have a minimal effect on the state of stress locally. By St. Venant's principle, variations in boundary forces should propagate into the plates a distance comparable to the wavelength of the variation. Thus, the effect on the Nazca plate of forces acting on the Cocos and Caribbean plates may be minimized because those boundary segments are relatively short. Such forces are separated from the Nazca plate primarily by ridge boundaries which may act as zones of weakness and further reduce the effect of forces acting on neighboring plates.

It is more difficult to determine the role of transform faults in the total torque balance. Estimates of stresses on transform faults are as high as several
kilobars (Hanks, 1977). Such high shear stresses on Nazca transform faults are marginally sufficient to balance boundary torques due to ridge and slab forces equivalent to 100 bars. Further, the amount of resistance at transform boundaries is likely to be a function of the normal stresses acting on the fault, and hence not strictly a function of boundary type. Drag forces on the base of the plate equivalent to a few bars are sufficient to balance the boundary torques. For this study, drag forces are assumed to balance the net torque acting on the plate.

The intraplate stress field is not likely to be sensitive to the type of drag law chosen for drag forces. For most of the models, the rotation pole is located near 30°N and 90°W. For an absolute rotation rate of $8 \times 10^{-7}$ deg/yr, the velocity of various points on the plate range from 5 to 10 cm/yr between the northernmost and southernmost portions of the plate, respectively. The force per unit velocity may vary if shear heating reduces friction (Schubert and Turcotte, 1972; Froidevaux and Schubert, 1975), or if dislocation creep models describe the flow mechanism for the mantle (Stocker and Ashby, 1973). As long as the drag force acts in the direction opposite to absolute plate velocity, the qualitative results of the models will remain similar.

It is interesting to note that the rotation poles
calculated to balance boundary torques for possible driving force models are in rough agreement with various absolute velocity models. The absolute rotation of the Nazca plate, defined by a zero net torque, is located at 47°N, 83°W with a rate of 9.8 x 10^{-7} deg/yr (Solomon et al., 1975). An absolute reference frame based on an inversion of hot spot traces gives a Nazca rotation pole at 45°N, 100°W with a rate of 8.8 x 10^{-7} deg/yr (Minster et al., 1974). For all of the driving force models considered in this paper, the absolute rotation pole varied between 36°N, 63°W and 23°N, 96°W. While this may be coincidental, it suggests that using drag forces to balance boundary torques is consistent with various definitions of an absolute reference frame for the Nazca plate.

Several conclusions may be drawn from the driving mechanism study of the Nazca intraplate stress field within the assumptions that have been made. Ridge forces play a significant role in driving the Nazca plate. All of the acceptable models included ridge forces. Net slab forces are, at most, comparable to net ridge forces. For ridge forces equivalent to 100 bars across a 100 km thick Nazca plate, an upper limit of about 200 bars due to subducting slabs is established for models satisfying the thrust mechanism of the 1965 earthquake. The model stresses near the 1965 event are most sensitive to ridge and slab forces in the northern half of the Nazca plate.
Forces acting on the Nazca-Antarctic and southern Chile segments of the plate boundary may be changed without considerably affecting the stresses near the 1965 event. It would be difficult to extrapolate to the earth's lithosphere the conclusion drawn for just the Nazca plate that net ridge and slab forces are comparable. Previous modeling of driving forces acting on the entire lithosphere, however, suggested that net ridge and slab forces are comparable (Solomon et al., 1975; Richardson et al., 1976). The finite element modeling of the Nazca plate in this chapter supports this conclusion. The single plate finite element analysis of the Nazca plate presented in this chapter removes the inability to match the stress inferred from the 1965 event for models with comparable net ridge and slab forces.

A net force equivalent to a few hundred bars due to the subducting slab is considerably less than estimates of potential buoyancy forces acting on the slab. The large negative buoyancy forces are nearly balanced by resistive forces opposing subduction (Forsyth and Uyeda, 1975; Chapple and Tullis, 1977). The net pull due to the slab transmitted to the surface plates is comparable to the net push due to the ridge.

The effects of large plate boundary earthquakes have been studied using the 1960 Chilean earthquake as a model. Models have included imposing boundary displacements comparable to the displacement observed and
inferred for the 1960 event. Also, stresses have been calculated for models with the slab forces increased along the 1000 km failed portion of the plate boundary. These models represent the maximum effect in an elastic plate over a completely relaxed viscous asthenosphere. An estimate of several hundred years for the boundary displacements to diffuse across the Nazca plate is about two orders of magnitude too large to account for proposed correlations on a time scale of a few years between ridge and trench strain release events (Berg and Sutton, 1974). A recent study of stress diffusion in the lithosphere which treated the asthenosphere as visco-elastic indicated that diffusion rates for distances on the order of a thousand kilometers might be as short as tens of years (Smith et al., 1976). The maximum effect of the imposed boundary displacements is a few bars, which is probably too small to account for the possible correlation between strain release events. Such small stresses may be important if portions of the lithosphere are near failure. Otherwise, stress changes of a few bars will have little effect on the intraplate stress field compared to stresses due to plate boundary forces.

Equating the elastic strain energy of the earthquake with the elastic strain energy due to deforming a simple plate indicates a possible increase of slab pulling forces equivalent to 35 bars across a 100 km thick plate. This additional pulling force results in
stress on the order of 25 bars in the southeastern part of the Nazca plate and only about a bar near the epicenter of the 1965 earthquake. The addition of slab forces equivalent to 35 bars produces boundary displacements in excess of the 30 m displacement inferred for the earthquake. Only about one-fourth of the elastic strain energy for the 1960 earthquake was used to deform the Nazca plate if the 30 m of displacement observed represents the maximum allowable boundary displacement. The rest of the energy must do work elsewhere, perhaps in deforming the South American plate or the subducted slab.

The effect of large plate boundary earthquakes on the intraplate stress field is probably less than a few tens of bars locally and decreases to a few bars at greater distances. Locally, the effect may be significant for seismicity, but at distances comparable to the distance between the 1960 event and the epicenter of the 1965 event, the effect on the intraplate stress field is probably not detectable unless the lithosphere is very near failure. Displacements within the plate vary between a few meters in the southwestern part of the Nazca plate and a few cm in the northwestern part of the plate for model CH30. The displacements are about five times larger for model C35. Precise geodetic techniques such as lunar laser ranging (Bender and Silverberg, 1975) and very long baseline interferometry (Coates et al.,
1975) should be able to measure displacements of a few cm over a few years between distant points. The displacements within the Nazca plate due to large plate boundary earthquakes are measurable and may provide important information about deformation within the plate.
### Table 4.1. Description of Various Nazca Plate Models

<table>
<thead>
<tr>
<th>Model</th>
<th>$F_R \times 10^{13}$ dyne-cm$^{-1}$</th>
<th>$F_T \times 10^{13}$ dyne-cm$^{-1}$</th>
<th>Absolute Velocity for Drag $\omega \times 10^{-7}$ deg/yr</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>R100</td>
<td>100</td>
<td>0</td>
<td>36°N 63°W 5.3</td>
<td>Figure 4.8</td>
</tr>
<tr>
<td>T100</td>
<td>0</td>
<td>100</td>
<td>23°N 96°W 6.0</td>
<td>Figure 4.9</td>
</tr>
<tr>
<td>R100T33</td>
<td>100</td>
<td>33</td>
<td>33°N 73°W 7.1</td>
<td>Figure 4.10</td>
</tr>
<tr>
<td>R100T100</td>
<td>100</td>
<td>100</td>
<td>30°N 82°W 10.9</td>
<td>Figure 4.11</td>
</tr>
<tr>
<td>R66T100</td>
<td>66</td>
<td>100</td>
<td>28°N 85°W 9.2</td>
<td>Figure 4.11</td>
</tr>
<tr>
<td>R50T100</td>
<td>50</td>
<td>100</td>
<td>28°N 87°W 8.4</td>
<td>Figure 4.11</td>
</tr>
<tr>
<td>R33T100</td>
<td>33</td>
<td>100</td>
<td>26°N 89°W 7.6</td>
<td>Figure 4.11</td>
</tr>
<tr>
<td>CH30</td>
<td>0</td>
<td>0</td>
<td>-- -- --</td>
<td>Displacement of 30 m on Peru-Chile boundary. See Figure 4.12.</td>
</tr>
<tr>
<td>CH35</td>
<td>0</td>
<td>35$^2$</td>
<td>13°S 93°W 2.0</td>
<td>Figure 4.13</td>
</tr>
</tbody>
</table>

1 The magnitude of $\omega$ corresponds to the constant $C$ arbitrarily chosen so that a velocity of 1 cm/yr results in drag stress of 1 bar.

2 A force of $35 \times 10^{13}$ dyne-cm$^{-1}$ was applied along that portion of the Peru-Chile subduction zone that failed during the 22 May 1960 earthquake, and no force was applied to the section that did not fail.
Figure Captions

Figure 4.1  Coarse finite element grid of one quarter of a hemispherical dome for comparison between computed and analytic solutions for test case presented in section 4.2.1.

Figure 4.2  Analytic solution for latitudinal displacement $u_\phi$ and computed values of $u_\phi$ for test cases as a function of latitude. Models D3, D10, and D19 correspond to successively refined grid spacings of 30, 10, and 5 degrees, respectively.

Figure 4.3  Analytic solution for latitudinal stress $\sigma_\phi$ and computed values of $\sigma_\phi$ for test cases with various grid spacings. For other details, see Figure 4.2.

Figure 4.4  Analytic solution for longitudinal stress $\sigma_\theta$ and computed values of $\sigma_\theta$ for test cases with various grid spacings. For other details, see Figure 4.2.

Figure 4.5  The Nazca plate with plate boundaries defined by earthquake epicenters. Ridge boundary is denoted by a double line, subduction boundary by a single solid line, and transform boundary by a dashed line. The orientation of the horizontal compression axis inferred from the focal mechanism of the 25 November 1965 earthquake is shown by an
arrow. The location of the two 5 August 1944 earthquakes is shown by a filled circle. The segment of the Peru-Chile trench boundary that failed during the 22 May 1960 earthquake is hatched. The 1960 epicenter is shown by an open circle. Mercator projection.

Figure 4.6 Finite element grid for the Nazca plate. There are 210 constant-strain triangular elements with six degrees of freedom per element and a total of 132 nodes. Grid spacing of 3 degrees along boundaries and 5 degrees in the interior. Mercator projection.

Figure 4.7 Schematic illustration of possible driving forces acting on the Nazca plate. $F_R$ and $F_T$ are forces per unit length of ridge and trench, respectively, and act in the direction of relative plate motions. $F_D$ is drag force per unit area and is opposite the velocity $V$ of the plate with respect to the mantle. The choice of $V$ is not free, but is fixed so that torques due to drag forces balance those due to boundary loads.

Figure 4.8 Principal horizontal deviatoric stresses in the Nazca plate predicted for Model R100 (Table 4.1). Principal stresses with arrows pointing inward denote deviatoric compression; those pointing out-
ward denote deviatoric tension. The location of the 1944 and 1965 earthquakes and the orientation of compression axis for the 1965 event are shown. In cases where the absolute value of the principal stress is small, the arrowhead may be larger than the value of the principal stress.

Figure 4.9 Principal horizontal deviatoric stresses in the Nazca plate predicted for model T100 (see Table 4.1). For other details, see Figure 4.8.

Figure 4.10 Principal horizontal deviatoric stresses in the Nazca plate predicted for model R100T100 (see Table 4.1). For other details, see Figure 4.8.

Figure 4.11 Principal horizontal deviatoric stresses in the Nazca plate predicted for model R50T100 (see Table 4.1). For other details, see Figure 4.8.

Figure 4.12 Principal horizontal deviatoric stresses in the Nazca plate predicted for model CH30 (see Table 4.1). A displacement of 30 m has been specified along the hatched section of the Peru-Chile trench that failed during the 22 May 1960 earthquake. The scale for the stresses is ten times smaller than for the models in Figures 4.8-4.11.
Figure 4.13 Principal horizontal deviatoric stresses in the Nazca plate predicted for model C35 (see Table 4.1). For other details see Figure 4.8. The scale for stresses is one-half the scale used for models in Figures 4.8-4.11.
FINITE ELEMENT GRID

Figure 4.1
Figure 4.2

The graph shows the relationship between $U \Phi$ and $\Phi$. The data points for D3, D10, and D19 are marked with different symbols. The Analytic curve is also plotted.

The axes are labeled as follows:
- Y-axis: $U \Phi$
- X-axis: $\Phi$

The figure is labeled as Figure 4.2.
Figure 4.3
Figure 4.4
NAZCA PLATE GRID
Chapter 5: Numerical Modeling of Global Intraplate Stress

Section 5.1 Introduction

The global intraplate stress fields predicted by various models of the driving mechanism are compared in this chapter with the observed intraplate stresses summarized in Chapter 2. A first order approach to modeling the global intraplate stress field based on finite difference techniques using a coarse grid has already been attempted (Solomon et al., 1975, 1977; Richardson et al., 1976). By way of an introduction to the finite element modeling which supercedes the previous finite difference models, a brief summary of the techniques, results, and limitations of the early attempts to model the global intraplate stress field with finite differences will be presented first. Because of the coarse grid used in the finite difference analysis, and because of some problems with the numerical solution of the equilibrium equations, the finite difference models must be considered as a preliminary study to the global finite element models which follow.

The global finite element models presented in this chapter represent an extension to the entire lithosphere of the technique developed in Chapter 4 for the analysis of stresses in a single plate. These global finite element
models use 5246 elements for approximately a four-fold increase in grid resolution compared to the previous finite difference models. The relative importance of forces due to ridges, subducted slabs, continental collision zones and drag along the base of the lithosphere may be studied by comparing the calculated intraplate stress field with the observed intraplate stresses considered in Chapter 2 and summarized again for convenience in this chapter as Figure 5.1.

Comparison between calculated and observed intraplate stresses for the finite element models proves to be a sensitive test of the importance of possible driving forces. A net ridge push is required in all models that match the observed stresses. The net slab pulling force is at most a few times greater than other forces acting on the plates. Forces which resist further convergence at continental convergence zones along the Eurasian plate are an important component of models that agree with the data for Europe, Asia, and the Indian plate. Viscous drag at the base of the lithosphere is best modeled as a resistive force, with a drag coefficient that is non-zero beneath oceanic lithosphere, but which may be concentrated by a factor of ten or more beneath continental lithosphere. Models of the driving mechanism in which drag forces drive plate motions or balance the net torque on each plate due to boundary forces are in
poorer agreement with the data than are models in which drag resists plate motions.
Section 5.2 Finite Difference Analysis

The global intraplate stress field due to possible plate driving and resisting forces may be calculated by treating the lithosphere as a thin elastic shell in the membrane state of stress. Equilibrium equations relating the stress resultants to applied tractions may be derived and solved using approximate numerical methods. In the first section, the derivation of the equilibrium equations is presented and a finite difference solution technique is described.

5.2.1 Methods

The equilibrium condition for membrane stresses in a thin elastic shell follows from Hamilton's Principle, which states that

\[ \delta \{ \int (P.E. - K.E.) dt \} = 0 \]  

(5.1)

where P.E. and K.E. are potential and kinetic energy, respectively, and are given as

\[ P.E. = U - \int_S \mathbf{T} \cdot \mathbf{\dot{u}} dS - \int_V \mathbf{F} \cdot \mathbf{\dot{u}} dV \]  

(5.2a)

\[ K.E. = \frac{1}{2} \int_V \rho \mathbf{u} \cdot \mathbf{\dot{u}} dV \]  

(5.2b)
where \( U \) is the strain energy given by \( U = \int_V PdV \) and 
\( P = \frac{1}{2} \sigma_{ij} \varepsilon_{ij} \) is the strain energy density function. \( S \) and \( V \) refer to the surface and volume of the body, respectively. \( \mathbf{F} \) and \( \mathbf{F} \) are body and surface forces, respectively, \( \mathbf{u} \) and \( \mathbf{u} \) are displacement and velocity, respectively, and \( \rho \) is the density, \( \sigma \) and \( \varepsilon \) are stress and strain tensors, respectively, and \( \delta \) implies a small variation in P.E. and K.E.

The forces \( \mathbf{F} \) and \( \mathbf{T} \) may be replaced by statically equivalent tractions \( \mathbf{q} \) which act on the reference surface of the thin shell. Replacing \( \mathbf{F} \cdot \mathbf{u} \) and \( \mathbf{T} \cdot \mathbf{u} \) by \( W \), the work done by the statically equivalent tractions, implies that equation (5.1) may be rewritten as

\[
\delta \{ \int (U - W - K.E.) dt \} = 0
\]  
(5.3)

The variations implied in equation (5.3) are performed one by one and can be briefly summarized

\[
\delta U = \int PdV = \int \delta P dS = R^2 \int_S \{ N_\phi (\delta (\varepsilon_\phi \sin \phi)) + N_\theta (\delta (\varepsilon_\theta \sin \phi)) \} dS
\]  
(5.4)

\[
\delta W = R^2 \int_S (\mathbf{q} \cdot \delta \mathbf{u}) \sin \phi dS
\]

\[
\delta K.E. = \rho h \int_S \mathbf{u} \cdot \delta \mathbf{u} dS
\]
where \( h \) is the shell thickness, \( R \) is the radius to the reference surface, \( \phi, \theta \) are the spherical coordinates colatitude and longitude, respectively, and \( N_\phi, N_\theta \) are stress resultants given by integrals of the stresses \( \sigma_\phi, \sigma_\theta \) across the thickness of the shell.

The integral equation given by equations (5.3) and (5.4) can be expressed as a partial differential equation for the stress resultants in terms of the applied tractions (after Kraus, p. 88, 1967)

\[
\frac{\partial}{\partial \phi} (N_\phi \sin \phi) + \frac{\partial N_\phi}{\partial \theta} - N_\theta \cos \phi + q_\phi Rs \sin \phi = 0 \quad (5.5a)
\]

\[
\frac{\partial}{\partial \phi} (N_\phi \sin \phi) + \frac{\partial N_\theta}{\partial \theta} + N_\theta \cos \phi + q_\theta Rs \sin \phi = 0 \quad (5.5b)
\]

\[
N_\phi + N_\theta + qR = 0. \quad (5.5c)
\]

The stress resultants are related to the strains by (Kraus, p. 88, 1967)

\[
N_\phi = K[\varepsilon_\phi + \nu \varepsilon_\theta] \quad (5.6a)
\]

\[
N_\theta = K[\nu \varepsilon_\phi + \varepsilon_\theta] \quad (5.6b)
\]

\[
N_{\phi \theta}^\varepsilon = \mu h \varepsilon_{\phi \theta} \quad (5.6c)
\]
where \( K = \frac{hE}{(1-\nu^2)} \) is the extensional rigidity, \( E \) and \( \nu \) are Young's modulus and Poisson's ratio, respectively, and \( \mu \) is the shear modulus. Values of \( 10^{12} \) dyne/cm\(^2\) and 0.25 are assumed for Young's modulus and Poisson's ratio, respectively. The strains may be expressed in terms of the displacements as (Kraus, p. 88, 1967)

\[
\varepsilon_\phi = \frac{1}{R} \left( \frac{\partial u_\phi}{\partial \phi} + w \right) \quad (5.7a)
\]

\[
\varepsilon_\theta = \frac{1}{R} \left( \csc \phi \frac{\partial u_\theta}{\partial \theta} + u_\phi \cot \phi + w \right) \quad (5.7b)
\]

\[
\varepsilon_{\phi\theta} = \frac{1}{R} \left( \frac{\partial u_\theta}{\partial \phi} - u_\theta \cot \phi + \csc \phi \frac{\partial u_\phi}{\partial \phi} \right) \quad (5.7c)
\]

where \( u_\phi, u_\theta, \) and \( w \) are colatitudinal, longitudinal, and radial displacements, respectively.

Using equations (5.6)-(5.7), equation (5.5) may be rewritten in terms of the displacements as

\[
\sin \phi \left\{ \frac{\partial^2 u_\phi}{\partial \phi^2} + (1+\nu) \frac{\partial w}{\partial \phi} + q \frac{R^2}{\phi K} \right\} + \cos \phi \frac{\partial u_\phi}{\partial \phi} + \csc \phi \left\{ -\nu u_\phi + \frac{(1-\nu)}{2} \frac{\partial^2 u_\phi}{\partial \theta^2} + \cot \phi \right\} = 0 \quad (5.8a)
\]

\[
\left(-\frac{3-\nu}{2} \frac{\partial u_\theta}{\partial \theta} - (1-\nu) u_\phi \cos \phi \right) + \frac{1+\nu}{2} \frac{\partial^2 u_\theta}{\partial \phi \partial \theta} = 0
\]

\[
\left(\frac{1-\nu}{2}\right) \sin \phi \left\{ \frac{\partial^2 u_\theta}{\partial \phi^2} + \frac{2}{1-\nu} q \frac{R^2}{\phi K} \right\} + \cos \phi \left( \frac{1-\nu}{2} \frac{\partial u_\theta}{\partial \phi} \right) \quad (5.8b)
\]

\[
+ \csc \phi \left\{ \frac{1-\nu}{2} u_\theta - (1-\nu) \cos^2 \phi u_\phi + \frac{\partial^2 u_\theta}{\partial \phi^2} \right\} + \cot \phi \left\{ \frac{3-\nu}{2} \frac{\partial u_\phi}{\partial \theta} \right\} + \frac{1+\nu}{2} \frac{\partial^2 u_\phi}{\partial \phi \partial \theta} + (1+\nu) \frac{\partial w}{\partial \theta} = 0
\]

\[
(1+\nu) \left\{ 2w + \frac{\partial u_\phi}{\partial \phi} + \csc \phi \frac{\partial u_\theta}{\partial \theta} + u_\phi \cot \phi \right\} + q \frac{R^2}{K} = 0 \quad (5.8c)
\]
The displacement $w$ is a dependent variable if the radial traction $q$ is assumed to be a radial restoring force due to gravity. Long wavelength elevation variations are isostatically compensated in the earth, and thus, the radial traction is given by the hydrostatic relation

$$q = -\rho gw$$  \hspace{1cm} (5.9)$$

where $\rho = 2.3 \, \text{g/cm}^3$, because the process of isostasy occurs mainly under water, and $g = 980 \, \text{cm/sec}^2$. With this assumption, equations (5.8a-c) reduce to two independent equations in the unknown displacements $u_\phi, u_\theta$ in terms of applied tractions $q_\phi$ and $q_\theta$.

The partial differential equations for $u_\phi, u_\theta$, and $w$ in equation (5.8) are solved using standard finite difference techniques. Derivatives of a function $f$ at some point '0', with respect to the coordinate directions $\phi$ and $\theta$, are approximated in terms of the values of the function at neighboring points. Introducing the following grid to identify the neighboring points and the coordinates $\phi, \theta$,
where the points a, b, c, and d are used to evaluate constants between grid points, implies

$$\frac{\partial f_0}{\partial \theta} = \frac{f_3-f_1}{2\Delta \theta}$$  \hspace{1cm} (5.10a)

where $f_3 = f(3) = f$ evaluated at point '3'.

Similarly,

$$\frac{\partial^2 f_0}{\partial \theta^2} = \frac{f_3-2f_0+f_1}{\Delta \theta^2}$$  \hspace{1cm} (5.10b)

and

$$\frac{\partial^2 f_0}{\partial \phi \partial \theta} = \frac{(f_6-f_7)-(f_6-f_5)}{4\Delta \phi \Delta \theta}$$  \hspace{1cm} (5.10c)

The finite difference equations for the displacements $u_\phi = u$, $u_\theta = v$, and the dependent displacement w in equation (5.8) are then given by:

$$u_0 = \left[ \frac{S_dK_dU_dS_bK_bU_b}{\Delta^2 \phi} + \frac{(V_8-V_7)\nu_4 K_4 - (V_6-V_5)\nu_2 K_2}{4 \Delta \phi \Delta \theta} + \frac{\nu_4 K_4 C_4 U_4 - \nu_2 K_2 C_2 U_2}{2 \Delta \theta} \right] + \frac{1}{2} \left[ \frac{(V_8/S_8 - V_6/S_6)K_3 (1-\nu_3)S_3 - (V_7/S_7 - V_5/S_5)K_1 (1-\nu_1)S_1}{4 \Delta \phi \Delta \theta} \right]$$  \hspace{1cm} (5.11a)

$$+ \frac{1}{2} \left[ \frac{(V_8/S_8 - V_6/S_6)K_3 (1-\nu_3)S_3 - (V_7/S_7 - V_5/S_5)K_1 (1-\nu_1)S_1}{4 \Delta \phi \Delta \theta} \right] + \frac{1}{2} \frac{1}{S_0} \left[ K_c (1-\nu_c)U_c + K_a (1-\nu_a)U_a \right] / \Delta^2 \theta - \frac{C_o K_o}{2 \Delta \phi} \frac{(V_3-V_1)}{2 \Delta \theta} + \nu_0 \frac{U_4-U_2}{2 \Delta \phi}$$

$$+ q \phi (0) R^2 S_o \left[ \frac{S_dK_dS_bK_b}{\Delta^2 \phi} + \frac{1}{2} \frac{(K_c (1-\nu_c) + K_a (1-\nu_a))}{\Delta^2 \theta} + C_o \cot(0) K_o \right]$$
\[ v_0 = \left( \frac{1}{2} \right) \left( \frac{s^2 K_d (1-v_d) v_4 / S_4 + s^2 K_b (1-v_b) v_2 / S_2}{\Delta^2 \phi} \right) \]

\[ + \frac{1}{2} \left( \frac{(U_6-U_7)K_4 (1-v_4)-(U_6-U_5)K_2 (1-v_2)}{4\Delta \phi \Delta \theta} \right) \]

\[ + \frac{1}{2} \left( \frac{C_0 S_0 K_0 (1-v_0) (V_4 / S_4 - V_2 / S_2)}{2\Delta \phi} \right) \]

\[ + \frac{1}{2} \left( \frac{C_0 (K_3 U_3 - K_1 U_1)}{2\Delta \phi} \right) \]

\[ + \frac{1}{2} \left( \frac{K_0 (1-v_0)' (V_3 - U_1) / S_0}{2\Delta \theta} \right) \]

\[ + \frac{1}{2} \left( \frac{q_\theta(0) R^2 S_0}{1 + \frac{S^2 d (1-v_d) + s^2 b (1-v_b)}{\Delta^2 \phi} + \frac{1}{S_0} \frac{(K_c + K_a)}{\Delta^2 \theta}} \right) \]

\[ w_0 = -\left( \frac{1}{2} \right) \left( \frac{U_4 - U_2}{2\Delta \phi} + \frac{V_3 - V_1}{2S_0 \Delta \theta} + \frac{C_0}{S_0} u_0 \right) \]

\[ \left( \frac{R^2 (1-v_0)}{2Eh \rho g} \right) \]

where \( s_d = \sin(\phi_d), c_0 = \cos(\phi_0), K_d = K(\phi_d, \theta_d), \) etc.

Equations (5.11a, b, c) may be solved iteratively for \( u, v \) and \( w \) as functions of \( q_\phi, q_\theta \) once a grid has been chosen and boundary conditions specified. The earth's lithosphere is divided into a \( 10^\circ \times 10^\circ \) grid in \( \phi \) and \( \theta \) as shown in Figure 5.2. Plate boundaries pass midway between grid points. With such a coarse grid size, only the seven largest plates are included. The Philippine plate is incorporated into the Pacific, the Arabian into the Indian, the Caribbean into the American, and the Cocos into the Nazca.

The only boundary condition on a shell that is closed with respect to \( \theta \) is periodicity of the solution

\[ f(\theta + 2\pi) = f(\theta) \]
Equations (5.11a,b,c) are solved iteratively using pointwise successive over-relaxation of the form

\[ f_i = (1-WT)f_{i-1} + g(\phi, \theta) \]  

(5.13)

where \( f \) is the function to be evaluated, \( i \) is the \( i^{th} \) iteration, \( g(\phi, \theta) \) is the function defining \( f \) (i.e., equations (5.11a,b,c) for \( u,v,w \)) and \( WT \) is a weighting factor between 1.0 and 2.0. For the finite difference models summarized later in this chapter, convergence was fastest for \( WT \approx 1.7 \) and approximately 1000 iterations were required to insure convergence.

Equation (5.11a,b,c) contain a \( 1/\sin(\phi) \) singularity at \( \phi = 0 \) and \( \pi \), making the direct evaluation of displacements at the coordinate poles impossible. The grid spacing is refined near the coordinate poles so that displacements and derivatives of displacements will vary more smoothly in these regions. Displacements along the innermost ring of grid points are extrapolated linearly from the displacements along surrounding rings of grid points expressed in Cartesian coordinates.

The displacements calculated from equations (5.11a,b,c) are converted to stresses using equations (5.7) and (5.6). Principal stresses are calculated from the stresses and are plotted for comparison with the observed intraplate stress field.
The displacements in equations (5.11a,b,c) are functions of the applied tractions \( q_\phi, q_g \), which depend on a model of the driving mechanism. In the next section, the parameterization of various possible driving and resisting forces is considered.

5.2.2 Specification of Forces

At grid points immediately adjacent to subduction zones and spreading centers, tractions equivalent to imposed tangential forces \( F_T \) and \( F_R \) per unit area, respectively, are imposed. These boundary forces are assumed to act in the direction of relative plate velocity. This assumption leads to a small net torque on the lithosphere from both ridge and trench forces because adjacent grid points across a boundary do not generally lie on a small circle about the spreading pole. This net torque, on the order of the torque due to the traction applied at a single grid point, is balanced for most models by making small arbitrary adjustments of a few percent to the imposed tractions at several ridge and trench boundaries. A few models in which this torque imbalance is offset by altering the absolute plate velocities slightly are also considered, though this procedure is reasonable only for large values of the viscous drag coefficient.

The driving force \( F_R \) is always symmetric about the
ridge boundary; $F_T$ may be larger on the subducting plate side of the trench boundary. Tractions equivalent to the drag force $F_D$ acting on a unit area of the lower surface of the lithosphere are applied at all grid points. The force $F_D$ is proportional to, and generally in the direction opposite to, the absolute velocity $\dot{\mathbf{v}}$ of the lithosphere at that point,

$$F_D = -D\dot{\mathbf{v}}$$

(5.14)

The drag coefficient $D$ in (5.14) has units of stress/velocity. A schematic illustration of available forces is shown in Figure 5.3.

The linear relationship (5.14) between plate velocity and drag stress at the base of the lithosphere may not be appropriate if deformation in the mantle is controlled by dislocation creep or if shear heating lowers the shear stress at high plate velocities (e.g., Schubert et al., 1976). The effect of nonlinear rheologies on absolute plate velocities and on intraplate stress has been considered by Solomon et al. (1977). Neither plate motions nor intraplate stress can distinguish a $v^{1/n}$ dependence of viscous drag from (5.14) if $n < 10$, as would be appropriate for dislocation creep ($n \sim 3$). Viscous drag laws in which shear stress increases with $v$ more slowly than $v^{1/10}$ or decreases with $v$ produce
absolute plate motions in conflict with paleomagnetic limits on net polar wander.

5.2.3 Summary of Model Results

The intraplate stress field has been determined for a number of diverse driving force models by Richardson et al. (1976). In this section, I will present a summary of the models and conclusions that are described in Richardson et al. (1976). The finite difference analysis presented in this chapter and in Richardson et al. (1976) represents a preliminary solution of the global intraplate stress field. The finite element techniques introduced in Chapter 4 will be extended to global intraplate stress models later in this chapter. These finite element models, for a variety of reasons including greatly improved grid resolution, the ease of specifying driving forces, and the lack of an artificial singularity at the coordinate poles, supplant the finite difference models. The finite difference models do, however, provide preliminary estimates of the relative importance of various driving forces.

The various models from Richardson et al. (1976) and Solomon et al. (1977) are summarized in Table 5.1. Each model may be parameterized by the relative magnitudes of six quantities: \( F_R \) is the symmetric force exerted at ridges; \( F_T \) is the symmetric force exerted at trenches;
and \( D_o \) and \( D_c \) are the viscous drag coefficients beneath oceanic and continental lithosphere and \( D_n \) is the non-linear drag coefficient. The forces \( F_R \), \( F_T \), and \( F_S \) are positive if they drive plate motion, and \( D_o \), \( D_c \), and \( D_n \) are positive if drag resists plate motion.

The results of one of the preferred models of the driving mechanism is shown in Figure 5.4. For model S1, \( F_R \) and \( F_T \) have equal magnitudes and resistive drag at the base of the plate is scaled by \( D_v/F_T = 0.003v \), where \( v \) is the velocity in cm/yr.

Although this model is rather simple, a comparison of Figures 5.1 and 5.4 reveals a satisfactory fit between predicted and observed stress in many regions. In North America the calculated stresses are compressive with trends of NE-SW to E-W and agree quite well with the observed stresses. In Europe, both the greatest and the least compressive stresses are in good correspondence with those shown in Figure 5.1. The observed stress fields in these two regions are among the best studied, and agreement between calculated and observed stress in these areas is critical in evaluating any model. Other acceptable features of the model include compression throughout most of the Pacific, a good fit to strike-slip and thrust faulting earthquakes in the northern Indian plate and in western South America, and the NW-SE tension in south and east Africa. Regions where the fit
is less than adequate include the Nazca-Cocos plate and the western African plate. In both regions, at least one of the principal stresses is calculated to be tensile, but the few observed earthquake mechanisms indicate thrust faulting.

Other models included varying the ratio of ridge to slab forces, varying the amount of drag forces, and the ratio of continental to oceanic drag forces, and concentrating the slab forces on the side with the downgoing slab.

Comparison of the observed and calculated intraplate stress fields for the finite difference models allows preliminary limits to be placed on the relative importance of various driving and resisting forces (Richardson et al., 1976). The net driving push at spreading centers is found to be at least comparable in magnitude to other forces acting on the lithosphere and in particular is 0.7 to 1.5 times the net driving pull at convergence zones. Subducting lithosphere, which from seismic and thermal evidence has more potential energy available to drive plates than does a spreading center, thus converts relatively little of this energy to a net force acting on the surface plates. The drag coefficient at the base of the lithosphere may be greater by a factor of 3 to almost 10 beneath continents than beneath oceans without substantially affecting the fit between calculated and
observed stress fields. Intraplate stresses calculated for models in which viscous drag at the base of the lithosphere acts in the direction of absolute plate velocity to drive plate motion are in much poorer agreement with observed stresses than are those calculated for models in which drag resists plate motions.

These finite difference models provide a reasonable starting point for the finite element models developed and discussed in detail in the remainder of this chapter.
Section 5.3 Finite Element Analysis

In the remainder of this chapter the finite element analysis technique presented in the last chapter for calculating the intraplate stress field in the Nazca plate is extended to whole lithosphere models. The entire lithosphere is treated as an assemblage of subregions or elements in a membrane state of stress. The solution of the equilibrium finite element matrix equations is based on the wave front technique introduced in Chapter 4.

The finite element technique used in this section has several advantages over the finite difference technique summarized in the last section. Irregular boundaries are easily modeled by varying the size and orientation of elements using the finite element technique. With finite difference techniques it is difficult to vary the grid spacing in regions of interest. The finite element procedure also avoids the $1/\sin\phi$ singularity that was an artifact of the finite difference coordinate system (see section 5.2.1). In the finite element technique, nodal forces equivalent to surface tractions and line forces acting on each element are particularly easy to specify.

The finite element technique presented in section 4.2 will now be briefly summarized and extended to a
treatment of the global intraplate stress field.

5.3.1 Methods

If the lithosphere is approximated by a linear elastic solid, the equilibrium condition between internal deformation and external loads follows from the principle of virtual displacement. This variational principle, based on minimizing the total potential energy, leads to the standard matrix equation for finite elements (Bathe and Wilson, p. 74, 1976; and Section 4.2)

\[ [K][U] = [F] \] (5.15)

where \([U]\) is the matrix containing the unknown nodal displacements, \([K]\) is the stiffness matrix which depends on the geometry and elastic properties of the structure, and \([F]\) is the matrix containing all loads acting on the structure.

There are many solution techniques for equation (5.15) that take advantage of the special properties of the stiffness matrix. One technique that is particularly useful when the dimensions of the stiffness matrix are very large is Irons' wave front solution technique (Irons, 1970; Orringer, 1974). This technique is based on Gauss elimination and back-substitution, but avoids prohibitively large in-core computer storage requirements.
by utilizing external storage space. Rather than assembling all of the linear equations represented by equation (5.15) at one time, only enough equations are considered to eliminate one degree of freedom or variable at a time. This information is then sent to external storage to await retrieval during the back-substitution phase. This technique was applied to the Nazca plate models in Section 4.3, although the total number of degrees of freedom for these models was only about 250. For whole earth models, the number of degrees of freedom is approximately 5000. If an in-core solution were attempted, over a megabyte of core storage would be required. It is difficult to find and prohibitively expensive to use a computer with this amount of available core.

The earth's lithosphere is divided into 5246 triangular plane stress elements as shown in Figures 5.5-5.3, where Figures 5.5 and 5.6 show the mid-latitude regions and Figures 5.7 and 5.8 show the north and south polar caps, respectively. Away from plate boundaries, each triangular element has dimensions of 5 x 5 x 7 degrees. Nodal spacing is approximately three degrees along plate boundaries. Twelve plates have been included and comparison with the finite difference grid in Figure 5.2 shows a definite improvement in the spatial resolution both of stress within the plate and of plate boundary definition. Each element has three nodes, each of which has two in-plane
degrees of freedom. These degrees of freedom correspond to the latitudinal and longitudinal components of displacement in the global coordinate system. There are 2625 nodes associated with the 5246 elements, giving a total of 5250 degrees of freedom.

Each element is assumed to have a thickness of 100 km. The elastic parameters of the elements away from plate boundaries are assumed to be constant. Values for Young's modulus $E$ and Poisson's ratio $\nu$ are taken as $7 \times 10^{11}$ dyne/cm$^2$ and 0.25, respectively. The calculated stresses are independent of the specific choice for $E$ as long as the Young's modulus is everywhere uniform. If the Young's modulus varies spatially, only the contrast, and not the actual values, will affect the calculated stresses.

Ridges are the site of upwelling, relatively warm mantle material and are thus probably of lower elastic moduli than older lithosphere. Stresses are probably not transmitted across a ridge as efficiently as across stable lithosphere. In an attempt to model the presence of the relatively warm mantle material at ridges, the Young's modulus for ridge elements is assumed to be $3.5 \times 10^{10}$ dyne/cm$^2$, or a factor of twenty less than the Young's modulus for stable lithosphere. For test cases, the calculated state of stress away from plate boundaries was not very sensitive to the choice of $E$ for the ridge, as long as the contrast between various regions was less
than about three orders of magnitude. For greater contrasts in $E$, numerical problems arise from essentially isolating regions from one another.

A subduction zone may be another region where stresses are inefficiently transmitted from one plate to another. Some stress transmitted across the oceanic plate may deform the subducted slab rather than act on the neighboring plate. Also, inter-arc spreading and island arc volcanism indicate the presence of warm material near the surface. As a first order model of the elastic parameters at subduction zones, the Young's modulus for subduction boundary elements is rather arbitrarily chosen equal to the value at ridges.

Continental convergence zones may be regions where there is a significant amount of shear heating taking place (Bird, 1976). Again, rather arbitrarily, the Young's modulus for continental convergence boundary elements is chosen to be $3.5 \times 10^{10}$ dyne/cm$^2$.

Transform faults may also act as soft regions. This may be particularly true for transform faults along the major mid-ocean ridges. For these transforms, the lithosphere is usually young and therefore warmer than old lithosphere. In all of the models considered in this chapter, transform faults along the major mid-ocean ridges are assumed to have a Young's modulus equal to the value chosen for ridge elements. Transform faults in
continental lithosphere may be nearly as rigid as stable plate regions. Deformation along the San Andreas fault appears to be strongly concentrated within a few tens of km of the fault (Savage and Burford, 1970). The fault probably does not represent a broad zone of weakness. Normal stresses may be transmitted rather efficiently across these transform faults while shear stresses may be less efficiently transmitted. The elastic constants may thus be anisotropic along continental transform faults such that the values for Young's modulus and the shear modulus parallel to the strike of the fault are less than the values perpendicular to the strike. Future modeling should include some form of anisotropic elements. For models considered in this chapter, continental transform fault boundary elements are assumed to have the same uniform elastic constants as interior elements.

The parameterization of various possible plate driving and resisting forces are considered in the next section.

5.3.2 Specification of Forces

Forces are applied at nodal points immediately adjacent to subduction zones and spreading centers. The nodal points adjacent to the plate boundaries were chosen to lie on small circles about the pole of rotation for the plates. The relative plate velocities are taken from
the RM2 model of Minster and Jordan (1978), except for the Philippine plate, where the Eurasian-Philippine angular velocity of Fitch (1972) has been used. Although more up-to-date information on the relative motion of the Philippine plate is now available (Chase, 1978a), the results of the global modeling are unlikely to change significantly for a change in the orientation of forces acting on the small Philippine plate. The relative plate velocities adopted for this thesis are listed in Table 5.2.

Forces at spreading centers and subduction zones are assumed to act in the direction of relative plate motion and are proportional to the length of plate boundary. Forces due to the ridge should act perpendicular to the strike of the ridge (see Section 1.2). For most ridges, the difference between the direction defined by perpendicular to the strike of the ridge and the direction of relative plate motion is small (see Figure 5.9). For computational ease, it is much more convenient to specify the direction of ridge forces along a plate boundary with a single rotation pole than it is to know the strike of the ridge everywhere along the boundary. At subduction zones the situation is less clear. It is still convenient to specify the direction of forces with a single rotation pole for each plate boundary. For many areas such as South America and the Kurils, the direction defined by perpendicular to the strike of the
trench and relative plate motions are very similar (see Figure 5.12). Specifying the direction of slab forces by convergence directions may be preferable to perpendicular to the strike for regions such as the western Aleutians and Sumatra as a way of accounting, at least in terms of directions, for the shorter length of the slab in these areas compared to the eastern Aleutian and Java regions. Future modeling might take into account variations in the age of subducted lithosphere, curvature of subduction zones, and length of subducted slab. For the present modeling, it seems sufficient to assume that forces at subduction zones act in the direction of plate convergence. The models will include, however, a possible dependence on the rate of subduction.

Forces associated with the various plate boundary types are applied at nodes located symmetrically about the boundary. A net torque may arise because forces are not applied exactly on the boundary. For computational reasons which may be justified physically, the forces actually applied at nodes on either sides of the plate boundary act along great circles tangent to the small circles which define the direction of relative plate motions at the plate boundary, rather than along the small circles themselves. For great circle path forces, any symmetric forces acting on the boundary will create no net torque on the lithosphere. The difference between the directions inferred from great and small circle paths
is less than a few degrees for all distances greater than 20 degrees from the pole of rotation. For distances less than 20 degrees, the difference increases, but the directions inferred from great circle paths are perhaps more appropriate for these regions anyway. Specifically, forces defined by great circle directions will remain orthogonal to a plate boundary that lies on a meridian about the pole of rotation as the boundary approaches the pole. This should more closely approximate the directions of gravitational forces at ridges near spreading poles than would forces defined by small circles about the rotation pole.

The driving force $F_R$ at ridges is always symmetric about the ridge boundary. Initially, the net force $F_T$ exerted on the surface plates by the subducted slab is assumed to act symmetrically about the trench boundary. This condition may be relaxed by specifying drag forces to balance the torque exerted by forces which are concentrated on the subducting plate, and will be considered later. The magnitude of $F_T$ may vary spatially to include a dependence on the rate of subduction as described in Section 1.3. Also, forces acting at continental convergence zones such as along the Himalayas may be different than forces acting at oceanic convergence zones with well defined Benioff zones.

Drag forces due to motion of the lithosphere with
respect to the mantle may be specified at each nodal point. The drag force $F_D$ is assumed to be proportional to plate area and absolute plate velocity. The direction of the force is opposite to the absolute plate velocity $\vec{v}$ if the force resists plate motions, and given by

$$F_D = -D\vec{v} \tag{5.17}$$

where $D$ is a drag coefficient which may vary spatially between continental, young oceanic, and old oceanic lithosphere. Drag forces are not specified for plate boundary elements. At plate boundaries the motions between the plate and the mantle material below are probably very complicated. At least for the ridge, the mantle material may have a relatively low viscosity and hence the drag forces there may be negligible.

No forces are specifically applied along transform faults. If drag forces dominate the driving mechanism, shear stresses along transform faults will be high naturally as a consequence of the rapid variation in the direction of drag forces across the transform. If the drag forces drive plate motions, shear stress on the transforms may provide significant resistance to plate motions (Hanks, 1977).

A schematic illustration of possible driving and resisting forces is shown in Figure 5.3.
The effect of various possible driving and resisting forces on the calculated intraplate stress field will now be considered.
5.3.3 Models

In this section various models of the driving mechanism are considered. The first class of models includes only symmetric forces acting on plate boundaries. In later models, drag forces on the base of the plates are included. A representative list of all models computed is given in Table 5.3. The various force parameters include $F_R$, the symmetric force exerted at ridges; $F_T$, the symmetric force exerted at trenches; $F_S$, the additional force exerted on the subducted plate; $F_V$, the velocity dependent symmetric force at trenches; $F_C$, the symmetric force at continental convergence zones; and $D_C$, $D_O$, $D_Y$, the drag coefficients beneath continental, old oceanic, and young oceanic lithosphere, respectively. The forces $F_R$, $F_T$, $F_S$, $F_V$ are positive if they drive plate motion, and $F_C$, $D_C$, $D_O$, $D_Y$ are positive if they resist plate motion.

5.3.3.1 Forces at Plate Boundaries

The simplest models each include only one of the force parameters. While all of these models are too simple to match the inferred intraplate stress field, they each isolate the effect of one of the possible forces.

The distribution of forces associated with ridges is shown in Figure 5.9, where each arrow corresponds to
the direction of relative plate motions given in Table 5.2. The number shown along the plate boundary refers to the plate pair given in Table 5.2. The distribution of elements associated with plate boundaries that have lowered Young's moduli is shown in Figure 5.10. The predicted deviatoric intraplate stress field for model El, with $F_R$ equivalent to 100 bars compression, is shown in Figure 5.11. The calculated stresses are referred to as deviatoric stresses, although they are better thought of as differences between the horizontal and vertical stresses. In classical continuum mechanism, deviatoric stresses are measured with respect to the hydrostatic stress, rather than the vertical. Although the term deviatoric stress, as used in Section 4.3 and in this chapter, may create some confusion, it is retained to emphasize that a calculated tensile stress does not necessarily imply a net tensional state of stress in the earth. Principal stress arrows in Figures 5.11 - 5.39 that either point inward or lack an arrowhead denote deviatoric compression, while arrows that point outward denote deviatoric tension. Deviatoric compression is predicted for most of the lithosphere for model El in Figure 5.11. The magnitude of the calculated stresses is on the order of 50 bars. Away from plate boundaries the stress field varies smoothly. Across ridge boundaries the maximum compressive stress is generally aligned with the direction of plate motions. Both of these features are as one would expect, and lend
confidence to the numerical techniques that have been used. Comparison with the observed intraplate stress field shown in Figure 5.1 shows moderate agreement in several places. In eastern North America the predicted principal stresses trend N-S to NE-SW. In the western portion of the Nazca plate a state of deviatoric compressive stress is predicted, although the direction is poorly constrained because the two principal stresses are of nearly the same magnitude. The southern portion of the Indian plate shows a NW-SE to nearly E-W trend for the maximum compressive stress. The Pacific plate is generally in a state of compression.

Another of the simplest models (E2) has only symmetric forces $F_T$ at trenches equivalent to a deviatoric tensile stress of 100 bars. The distribution of the trench forces is shown in Figure 5.12, where the lines give the direction of relative plate convergence. The plate boundary numbers refer to Table 5.2. The predicted stress field for model E2 is shown in Figure 5.13. Most of the earth's surface is predicted to be in a state of deviatoric tension. The stress field varies slowly away from plate boundaries. The maximum stresses occur near trench boundaries and are aligned with the direction of plate motions. Typical values for the deviatoric tension in the middle of the plates are on the order of 50 bars.

The predicted stresses for this model are inconsistent
with the observed stress field. Although this is a very simple model, it indicates that slab pulling forces cannot completely dominate the intraplate stress field.

Slab forces may depend on the subduction rate, as described in Section 1.3. Intraplate stresses for a simple model (E3) with symmetric forces that depend on subduction rate is shown in Figure 5.14. For this model, the slab force is assumed to be given by

\[
F_v = 1 \times 10^{15} \text{ dynes cm}^{-1} \times \begin{cases} 
\frac{v}{4} & 0 \leq v \leq 4 \text{ cm/yr} \\
\frac{12-v}{8} & 4 < v \leq 12 \text{ cm/yr} \\
0 & v > 12 \text{ cm/yr}
\end{cases}
\]  

(5.18)

where \(v\) is the subduction rate as determined from Table 5.2. The motivation for such a velocity dependence is as follows. The gravitational force on the slab may remain constant as long as the rate of subduction is greater than about 4 cm/yr (Richter and McKenzie, 1978). For lesser subduction rates, there is likely to be a lesser gravitational force due to the reduced thermal contrast between slab and mantle. The net pulling force due to the slab is the sum of the gravitational force and the resisting forces acting on the slab as it moves with respect to the mantle. The viscous drag on the slab may increase with velocity until, at some terminal velocity, the resisting and pulling forces balance (Forsyth
and Uyeda, 1975). To model this effect, $F_v$ is taken to decrease linearly to zero for $v$ between 4 and 12 cm/yr. $F_v$ has a maximum value equivalent to a deviatoric tensile stress of 100 bars at $v = 4$ cm/yr. As shown in Figure 5.14, the predicted stress field is dominated by deviatoric tension. The largest stresses, on the order of 50 bars, are in the southern Pacific and the southeast Indian plates. The main difference between models E2 and E3 is in the relative magnitude of the stresses for various plates. In model E3, the difference in the magnitude of the stresses between the western Pacific and Nazca plates is greater than for model E2. This may be important in later modeling because focal mechanisms in the Nazca plate suggest a state of nearly east-west compression (e.g., Section 2.2.5, Section 4.3, and Figure 5.1).

Convergence zone forces are probably different for zones of continental convergence than for regions of oceanic lithosphere subduction. The Indian-Eurasian and African-Eurasian plate convergence zones do not have a well-defined slab that can pull on the plates. The uplifted Himalayan region along the Indian-Eurasian plate boundary may exert a compressive stress on the plates and a net resistance to convergence rather than a driving pull on the plates. This compressive stress arises in much the same fashion as stresses associated with the ridge; a horizontal variation in density causes horizontal forces. For the
Himalayas, these compressive stresses may be on the order of a kilobar (Bird, 1976). The plate boundary in the Mediterranean between the African and Eurasian plates is not well defined. Part of the difficulty in defining the plate boundary is that the pole of rotation for the two plates is very near the Mediterranean. The Alps are located just to the north of the plate boundary and, like the Himalayas, may exert a resistance to further convergence. The magnitude of the associated compressive stresses is probably less than for those associated with the Himalayas because of the smaller topographic expression for the Alps. However, as a first order approximation, the forces along the African-Eurasian and Indian-Eurasian plate boundaries are assumed to be equal. These forces are shown in Figure 5.15, where the directions of the forces are defined by the relative plate velocities. The directions of the forces are nearly perpendicular to the strike of the topography associated with the continental convergence zones. These forces are assumed to compress, rather than extend, the plates as do the forces in models E2 and E3. The calculated stresses for this model (E4) are shown in Figure 5.16, where $F_C$ has been assumed to be equivalent to a deviatoric compressive stress of 100 bars. The largest stresses, on the order of 50 bars, occur in the Eurasian and Indian plates near the Himalayas. As expected, the magnitude of the stresses
decreases away from the plate boundaries. In North America, for example, the stresses are on the order of a few bars.

Comparison between Figures 5.16 and 5.1 shows several interesting features. First, the directions of the calculated maximum compressive stresses in northern India and Asia are similar to the focal mechanisms for these areas. Also, the predicted maximum compressive stress varies from N-S to NW-SE between northern and central portions of the Indian plate. In the northern portions of the Asian plate the component of deviatoric tension is consistent with Baikal rift tectonics (e.g., Section 2.2.3 and Figure 5.1). The calculated stresses in Asia for model E4 are qualitatively similar to the results obtained by Tapponnier and Molnar (1976, 1977) from a slip line analysis which treated India as a rigid indentor impinging upon Asia.

The calculated maximum compressive deviatoric stress in Europe for model E4 trends NW-SE, also in agreement with the data for this area. Model E4 is locally consistent with the intraplate stress field. The calculated stresses at great distances from the African-Eurasian and Indian-Eurasian plate boundaries are very small. For models considered later in this chapter, forces associated with continental convergence appear to be important for regional intraplate stresses, but
have little effect at great distances.

For this simple class of models with only plate boundary forces, it is not possible to consider forces which are concentrated on one side of subduction zones. The force $F_S$, because it is not applied symmetrically about the plate boundary, exerts a net torque on the lithosphere. The equilibrium condition which forms the basis for the numerical techniques requires that the net torque on the lithosphere be zero. In later models, $F_S$ may be included by requiring drag forces to balance the torque.

The next class of models includes linear combinations of models E1-E4. For the finite element model of the lithosphere that has been developed in this thesis, the deformation of the lithosphere is linear in the applied forces. If the forces are doubled, the magnitude of the stresses double while the orientations remain unchanged. Thus, to consider the effects of both ridge and convergence zone forces, the results from models E1-E4 may be linearly superposed.

The role of increasing the relative strength of $F_T$ with respect to $F_R$ can be seen with models E5-E7 in Table 5.3. These models cover the range of $F_T/F_R$ from 1:2 to 3:1, scaled such that $F_R$ is equivalent to a compressive stress of 100 bars. Calculated stresses for model E5 are qualitatively similar to those for model E1.
and indicate that adding a small trench force does not affect the intraplate stresses away from the trenches significantly. Calculated stresses for model E6, with $F_T/F_R = 1:1$, are shown in Figure 5.17. The differences between calculated stresses for E1 and E6 are noticeable, especially near trenches. The calculated stresses in the Nazca plate show a component of deviatoric tension in the E-W direction that is inconsistent with the data shown in Figure 5.1. The calculated stresses in Europe are small, but do show a NW-SE trend for the maximum compressive stress. The maximum compressive stresses in eastern North America are smaller for model E6 than for model E1. Model E6 is less acceptable than model E1, and indicates that either ridge forces dominate trench forces or, more likely, other forces must be included to realistically model the intraplate stress field.

Increasing $F_T/F_R$ to 3:1 produces large calculated deviatoric tensile stresses, especially in oceanic regions, as shown for model E7 in Figure 5.18. The fit to the observed stresses is poor for the Pacific and Nazca plates. The northern Indian plate does show some agreement with the inferred N-S principal stresses. There is similar agreement in east Africa, where the least compressive stress trends E-W. The calculated stresses in Europe and eastern North America are inconsistent with the observed stress field. This model
indicates a limit for the ratio of $F_T/F_R$ of less than 3:1, as long as no other forces are active.

Slab forces may depend on the subduction rate. In the next set of models, the role of increasing the ratio of $F_V$ to $F_R$ is considered. The results for models E8-E10 are summarized in Table 5.3 for $F_V/F_R$ increasing from 1:1 to 5:1. Calculated stresses for model E8, with $F_V/F_R = 1:1$ are similar to those for model E6 ($F_T/F_R = 1:1$), except near trenches where the subduction rate is very different from 4 cm/yr. Calculated stresses for model E9, with $F_V/F_R = 3:1$, are shown in Figure 5.19. The fit between calculated and observed stresses for this model is better than for the corresponding model with slab forces that are independent of subduction rate (E7). In eastern North America the calculated principal stresses are compressive. The component of NW-SE compression in Europe is greater than for model E7, which predicted predominantly deviatoric tension trending NE-SW. In the Nazca plate the dominant stress is deviatoric tension, although there is a small component of compression not present in model E7. In the Antarctic plate south of the Nazca plate, the calculated stresses are compressive, in agreement with the sense, although not the orientation, of a thrust fault earthquake in this region. Increasing the ratio of $F_V/F_R$ to 5:1 produces stresses as shown in Figure 5.20 for model E10.
Most regions show deviatoric tension. This model indicates an upper limit of 5:1 for $F_V/F_R$. This limit is somewhat higher than the corresponding limit for $F_T/F_R$ of about 3:1 and reflects reduced forces in several regions as a result of the assumed subduction rate dependence of slab forces.

Ridge, slab, and continental convergence forces are combined in the next class of boundary force models. To test whether the upper limit on the slab forces may be increased, models Ell-13 include increasing forces along the continental convergence zone in Eurasia. The forces associated with this convergence zone are assumed to create deviatoric compression in the plates, as shown for model E4 in Figure 5.16. The additional compressive stresses associated with the continental convergence zone may allow a larger slab force contribution to the driving mechanism without degrading the fit to the predominantly compressive observed intraplate stress field. Model Ell is similar to model E7 ($F_T/F_R = 3:1$), except that forces equivalent to 200 bars of compressive stress along the Himalayan and Mediterranean plate boundaries have been included. Comparing Figures 5.18 and 5.21 indicates an improved fit between calculated and observed stresses in Europe and Asia for model Ell. Further away from the continental convergence zone, however, the effect of the additional forces is
minimal. Thus, in the Pacific, Nazca, and South American plates, the calculated stresses are still in poor agreement with the data. Increasing the forces along the continental convergence zone to $F_C/F_R = 5:1$ in model E13 produces calculated stresses as shown in Figure 5.22. The differences in the predicted stresses between models E11 and E13 are confined almost completely to regions near the continental convergence zone. The fit to the observed data in other parts of the lithosphere is not appreciably improved by increasing the forces acting along continental convergence zones. Thus, additional forces acting along continental convergence zones are not sufficient to increase the upper bound limit on $F_T/F_R$ established for boundary force models. The situation is similar for model E14, with predicted stresses shown in Figure 5.23, where slab forces are dependent upon subduction rate. The ratio of $F_V/F_R$ is limited to less than 5:1 for boundary force models.

Models E15-E16 represent an attempt to obtain a best fit to the observed intraplate stress field using only boundary forces. For model E15, forces at ridges and continental convergence zones are equivalent to deviatoric compressive stresses of 100 bars and forces at subduction zones are equivalent to deviatoric tensile stresses of 100 bars. The agreement between calculated (Figure 5.24) and observed stresses is quite good in
Europe, Asia, and the Indian plate. The calculated stresses for other parts of the world are in agreement with the sense, although not always the orientation, of the observed stress data.

The results of the boundary force models may be briefly summarized before considering models with drag forces on the base of the plates. Models that produce a reasonable fit to the observed data all include ridge forces. For models with boundary forces only, an upper limit of about 3:1 for $F_T/F_R$ and about 5:1 for $F_V/F_R$ is established. Forces at continental convergence zones in Eurasia are important for the regional intraplate stress field, but have insufficient effect in other parts of the lithosphere to increase the upper bound limits established for the ratios $F_T/F_R$ and $F_V/F_R$.

The boundary forces considered above, with the exception of continental convergence forces, are assumed to drive, rather than resist, plate motions. Although the net torque exerted on the lithosphere is zero for all of the boundary force models considered thus far, forces acting to resist plate motions have not been explicitly included. The next set of models will consider the role of drag forces acting on the base of the lithosphere.

Drag forces on the base of the lithosphere depend on the velocity of the plate with respect to the
mantle and the viscosity of the mantle. As discussed in Section 1.4, the motion of the plates with respect to the mantle is a poorly known quantity. One approach to modeling drag forces is to determine an absolute reference frame for motions between the plates and the mantle. Unfortunately, there are a variety of possible definitions of an absolute reference frame. One possible reference frame is defined by 'hot spots' which are presumably fixed in the lower mantle (Morgan, 1972). Other reference frames may be defined by setting the net torque due to various forces acting on the plates to zero (Solomon and Sleep, 1974; Solomon et al., 1975). For the first set of drag models considered in this chapter, a linear drag law of the form given in equation (5.17) will be assumed. The absolute velocities are determined from the condition that the drag forces exert no net torque on the lithosphere. The drag coefficient $D$ is positive if $F_D$ resists plate motions. The possible parameterizations of drag forces have also been considered in Section 4.3.

Model E17 is characterized by drag forces of the form given in equation (5.17), where the drag coefficient is taken to be everywhere uniform. The calculated absolute plate velocity field is summarized in Table 5.4. Only the Pacific absolute plate velocity is given, since all other absolute plate velocities follow from the relative plate velocities given in Table 5.2.
The calculated stresses for model E17 are shown in Figure 5.25. The largest deviatoric stresses are tensile in the western Pacific and Asia and compressional in the eastern Pacific and the Nazca plate. The stresses are small in eastern North America and Europe. The calculated stresses are in poor agreement with the observed stress field in the western Pacific, Asia, and the Indian plate.

The drag coefficient may be concentrated beneath continents (Jordan, 1975). Calculated stresses for model E18, with drag forces beneath continents only, are shown in Figure 5.26. The continental boundaries are taken as the one thousand fathom bathymetric contour. The magnitudes of the calculated stresses for model E18 are smaller than for E17, primarily because a smaller area is subject to drag forces. The major differences between models E17 and E18 occur in the eastern Pacific and the Nazca plate, where deviatoric tension in model E18 replaces large deviatoric compressive stresses in model E17. The compressive stresses in eastern North America and Europe are greater relative to other regions for model E18 than for model E17, although the calculated SW-SE direction in eastern North America is not supported by the data. For oceanic regions a uniform drag model is better than a continental drag model.
Drag forces may also be concentrated beneath old oceanic lithosphere (Chapman and Pollack, 1975). Model E19 has drag forces only beneath oceanic lithosphere which is older than 80 million years. The 80 million year isochron was taken from Sclater et al. (1978). The stresses for E19 are shown in Figure 5.27, and are qualitatively similar in orientation to those for model E17. The main difference is that the predicted deviatoric tension is larger in Europe relative to other regions for model E19 compared to model E17.

Models E17-E19 have been considered in an attempt to isolate the effects of various parameterizations of the drag coefficient. If only drag forces act on the plates, the drag coefficient beneath oceanic lithosphere must be non-zero if compressive stresses are required for oceanic regions. The calculated continental stresses are qualitatively similar for uniform drag and old oceanic drag force models. The drag coefficient beneath continents may be higher than beneath the oceans without significantly affecting the continental stresses.

A model in which drag forces drive rather than resist plate motions can be constructed from model E17 by assuming that the drag forces act in the direction of plate motions. For such a case, the sense of stress at each point in Figure 5.25 is reversed and the magnitude of the stress is unchanged. Such a model (E20)
predicts deviatoric compression in Eurasia, the western Pacific and Indian plates and deviatoric tension in the eastern Pacific, Nazca and South American plates.

Models with only drag forces cannot be expected to match the observed intraplate stresses very well, since the contribution of plate boundary forces due to ridges, subduction zones, and continental convergence zones have been neglected. In the next class of models, both boundary and drag forces are included with the aim of establishing limits on the various contributions to the driving mechanism and of finding a best model.

Models E21-E23 are designed to test whether the limit of about 3:1 for $F_T/F_R$ established for the boundary force models may be increased by adding resistive drag forces. Stresses for model E21, in which a small amount of drag has been added to model E7 (see Table 5.3) are shown in Figure 5.28. The calculated stresses are very similar to those for model E7, with small differences confined primarily to the Pacific plate. The effect of increasing the drag coefficient can be seen in Figure 5.1) for model E22. The magnitude of the calculated deviatoric tensile stresses in Asia and the western Pacific has increased, while the magnitude of the compressive deviatoric stresses has increased for the eastern Pacific and Nazca plates. The stresses in North America and Europe are not changed appreciably.
Calculated stresses for model E23, with a large drag coefficient are shown in Figure 5.30. Deviatoric tension dominates in the western Pacific, Asia, and southern Indian plates. The predicted stresses are compressive in the Nazca plate, and agree with the earthquake data for this plate (e.g, Figure 5.1). The maximum compressive stress is small and oriented NW-SE in eastern North America.

Models E21-E23 clearly indicate that adding uniform drag forces does not increase the upper bound of about 3:1 for $\frac{F_T}{F_R}$. The only areas where adding drag forces improves the fit to the data are the western Pacific and Nazca plates.

Concentrating the drag coefficient by a factor of six beneath continents produces stresses as shown in Figure 5.31 for model E24. The stresses are very similar to model E21, where the drag coefficient was assumed to be uniform, with the main difference being that the compressive component of the stresses is slightly increased for continents and slightly decreased for oceanic lithosphere for model E24 compared to model E21. The fit to the observed data is only marginally changed. The drag coefficient beneath continents is enhanced by a factor of 11 compared to oceanic lithosphere for model E25, with the calculated stresses shown in Figure 5.32. The stresses are very similar to those for model E24, with only minor changes
in South America and the Pacific.

The fit to the observed stresses in Europe, Asia, and India can be improved over models E21-E25 by including continental convergence zone forces as is shown for model E26 in Figure 5.33. The fit to the observed data is essentially unchanged away from the continental convergence forces.

Models E21-E26 (see Table 5.3) indicate that resistive drag forces acting on the base of the plate cannot significantly improve the fit to the observed intraplate stress data for the limiting case of slab to ridge forces. The drag coefficient beneath the continents may be larger by a factor of ten with little effect as long as drag forces do not dominate the intraplate stress field.

In the previous models drag forces were assumed to resist plate motions. A simple model (E27) in which drag forces drive plate motions with resistance to plate motion provided by compressive forces at continental convergence zones and at trenches predicts stresses as shown in Figure 5.34. The fit to the data is poor for the eastern Pacific, Nazca, and South American plates. The calculated stresses in eastern North America are acceptable. This probably results from the resistive forces at subduction zones, however, since Figure 5.25 indicates that stresses in eastern North America due to drag forces based on a calculated absolute velocity, whether resistive or plate driving, are small compared to other regions.
The orientations of the maximum compressive stresses in Europe and Asia are inconsistent with the data. The calculated stresses for model E27, in general, are only in moderate agreement with the data.

The effect of increasing the drag coefficient for driving drag models is shown in Figure 5.35 for model E28. Compared to E27, model E28 predicts larger deviatoric tension in the eastern Pacific, Nazca, and South American plates. Increasing the drag coefficient will further degrade the fit between calculated and observed stresses in these regions. It is unfair to conclude that viscous drag does not drive the plates from simple models such as E27-28. However, for the assumptions made about the forces acting on the plate boundaries and the relationship between drag forces and plate velocities with respect to the mantle, models with resistive drag forces are in better agreement with the observed intraplate stresses than are models with plate driving drag forces.

It is probably an oversimplification to assume that drag forces either uniformly resist or drive plate motions. An alternative approach, taken in Chapter 4 for Nazca plate models and suggested by Davies (1978), is to suppose that drag forces balance the torques on each plate due to boundary forces. The equilibrium condition imposed in models E1-E28 is zero net torque on the entire lithosphere. In model E29 the torque
on each plate is set to zero by specifying drag forces to balance boundary forces. The boundary forces include symmetric forces at ridges and continental convergence zones and forces at subduction zones that act only on the plate with subducted lithosphere (see Table 5.3). Symmetric forces at subduction zones have been assumed for all previous models. The absolute rotation poles for the plates are no longer required to satisfy the relative plate velocities given in Table 5.2. The absolute velocity pole for each plate in model E29 is given in Table 5.5, along with the magnitude of the shear stress on the base of the plate required to balance torques. Davies (1978) assumed that kilobars of shear stress act on transform faults and used the results of Chapple and Tullis (1977) to estimate slab forces. Because of large assumed resistive forces at trenches, Davies concluded that plates without significant subducted slabs such as South America and Eurasia were driven from below by drag forces. For model E29, no forces have been specified on transform faults or on the non-subducting plate at subduction zones. For example, no drag forces are specified on the Caribbean plate for model E29. Thus, no direct comparison can be made between Davies' (1978) model and model E29. Model E29 does represent, however, an attempt to model subduction zone forces that act only on the subducted plate. The stresses for model E29 are shown in Figure 5.36. For several areas the predicted stresses agree very well with the data.
In the North American and Nazca plates, the orientation of the maximum compressive stress is well matched by the model. The fit is almost as good in Europe and in Asia north of the Himalayas. In the Indian plate, compressive stresses trend NW-SE in continental India, in agreement with the data, but the fit is poorer in Australia. In South America, the maximum compressive stress trends E-W, in only moderate agreement with the data. In oceanic regions near subduction zones and in the eastern part of the African plate the agreement with the data is poor. On the whole, the model provides a better fit to continental than oceanic data, and suggests that any force pulling the overthrust plate toward the trench is probably lower in magnitude than the net pull on the subducted plate. The assumed drag forces are significantly different for model E29 than for earlier drag models (E17-20). An acceptable fit between calculated and observed stresses can be achieved, especially for continental regions, with various parameterizations of drag forces. Thus, the role of drag forces in the driving mechanism remains poorly constrained.

The last models for this chapter represent an attempt to find a best model, rather than limiting models, of the driving mechanism. Model E30, shown in Figure 5.37, has symmetric forces at ridges and continental convergence
zones equivalent to a deviatoric compressive stress of 100 bars across a 100 km thick lithosphere, symmetric forces at subduction zones equivalent to a deviatoric tensile stress of 100 bars, and drag forces on the base of the lithosphere with a drag coefficient that is concentrated by a factor of four beneath continents (see Table 5.3). The model stresses are in good agreement with the data for eastern North America, Europe, Asia near the Himalayas, and the Indian plate. The fit to the data is good in South America, especially away from the trench, and in western Africa and is acceptable in most of the Pacific plate. The orientation of the calculated maximum compressive stress in the Nazca plate for model E30 is only in moderate agreement with the orientation inferred from the single fault plane solution available (Mendiguren, 1971). The fit to the data in the northern Pacific, eastern Asia, and east Africa is rather poor. The fit to the data in the northern Pacific and eastern Asia could probably be improved if subduction zone forces were decreased along the western Pacific plate margin. No attempt, however, has been made to vary plate boundary forces locally to match inferred stresses. If such an approach were attempted, most observed stresses could probably be matched but the solution for the driving mechanism would be unjustifiably arbitrary and non-unique. The approach used in this thesis has been to treat each
potential driving and resisting force acting on the plates by class, and make estimates of the importance of each type of force. Thus, while model E30 is a good model in terms of the fit between calculated and observed stresses, the actual driving mechanism for plate tectonics is not precisely constrained to the particular forces that have been assumed for this model. For example, model E31, with calculated stresses shown in Figure 5.38, differs from E30 in that the drag coefficient beneath continents has been increased to a factor of six compared to oceanic lithosphere. The calculated stresses are very similar to those for model E30. Also, model E32, with the force at continental convergence zones doubled compared to model E31 (Figure 5.39), predicts stresses that are very similar to those for models E30-31. The predicted compressive stresses in Asia, Europe, and the Indian plate have increased, but the only change in the fit of the model to the data is a slight improvement in the Eurasian plate.

The role of modeling intraplate stresses is best suited to testing the relative importance of various forces acting on the plates, rather than determining the exact balance of forces that make up the driving mechanism. In the next section, the results of the modeling will be briefly discussed and summarized.
Section 5.4 Discussion and Conclusions

The calculated intraplate stress field is very sensitive to forces applied at plate boundaries and on the base of the plates. Varying the forces assumed to be components of the driving mechanism results in significantly different calculated stress fields. Ridge pushing forces are required of all models that match the orientations of intraplate stresses. The net driving force due to subducted lithosphere is less than a few times larger than other forces acting on the plates for all acceptable driving mechanism models. The upper limit on net slab forces may be extended by less than a factor of two if slab forces are assumed to depend on the subduction rate such that the force acting on the fastest moving plates (e.g., The Pacific and Nazca plates) is reduced.

Forces acting at continental convergence zones along the Eurasian plate are important for intraplate stresses, and improve the fit to the data, in Europe, Asia, and the Indian plate. Models with compressive forces acting at continental convergence zones are in better agreement with the data than all those without forces at these boundaries. The upper bound limit on net slab forces is not affected by forces acting at continental convergence zones.

Resistive viscous drag forces acting on the base of the plate improve the fit to the intraplate stress field
for the Nazca and South American plates as long as some drag acts beneath oceanic lithosphere. The intraplate stress field is not very sensitive to an increased drag coefficient beneath old oceanic lithosphere compared to young oceanic or continental lithosphere. Increasing the drag coefficient beneath continents by a factor of five or ten changes the calculated stresses only slightly and has little effect on the overall fit of calculated stresses to observed stresses as long as some resistive drag acts beneath oceanic lithosphere.

Models in which drag forces drive rather than resist plate motions are in poor agreement with the data. This poor agreement may depend on the oversimplified model of the interaction between the plate and the asthenosphere that has been assumed. The actual flow pattern in the mantle and resulting drag forces acting on the base of the plates may be considerably more complicated than has been assumed in these models.

An alternative approach to modeling drag is to require drag forces to balance the torques due to boundary forces. This approach was used in Chapter 4 to model intraplate stresses in the Nazca plate, and Davies (1978) has suggested that it may be important for the entire plate system. The drag forces depend on the assumed boundary forces, and the calculated stresses for
one choice of boundary forces, with symmetric forces at ridges and continental convergence zones and forces exerted only on the subducted plate at subduction zones, are in good agreement with the data for the Nazca and several continental plates but in poor agreement with the data for oceanic plates near subduction zones. Most models in this chapter can be formed as linear combinations of simple one-parameter models. Each different specification of boundary forces requires that a complete new model be calculated when drag forces are assumed to balance boundary torques. It is thus impractical to consider a suite of such models, since each new complete model requires approximately three hours of computer CPU time. It is thus difficult to assess the validity of the approach of balancing torques with drag on the base of the plates.

The best models of the driving mechanism have comparable net forces at ridges, subduction zones, and continental convergence zones, and resistive drag forces modeled by a drag coefficient which must be non-zero beneath oceanic lithosphere but which may be concentrated beneath continental lithosphere.

The calculated stresses for several regions suggest limitations of the modeling and indicate improvements that might be included in future work. Most models failed to predict extensional deviatoric stresses in east Africa. East Africa is surrounded by ridge and continental convergence plate boundaries which are assumed to compress the plates. The absolute velocity of the African plate is small compared to most plates, and drag forces are assumed to be proportional to velo-
city. Increasing the drag coefficient sufficiently to affect African intraplate stresses degrades the fit to the data in most other regions. If the assumed form of boundary and drag forces is correct, then some other forces must dominate the tectonics of east Africa. Forces associated with northward motion of Africa on an ellipsoidal earth during the past hundred million years may produce the tensional tectonics of east Africa (Oxburgh and Turcotte, 1974). East Africa may be a plate boundary between the African and Somalian plates, rather than an intraplate feature (Chase, 1978a). Future modeling of the intraplate stress field might treat the east African rift system as a spreading plate boundary.

Some models predict deviatoric tension in the Nazca plate while predicting reasonable stresses elsewhere. It may be that the inactive Galapagos Rise, ignored in this modeling, may contribute to the state of compression inferred from the thrust earthquakes in the Nazca plate (see also Section 4.3). Alternatively, the slab forces acting along the per-Chile trench may be less than have been assumed. There is some evidence that this might be the case. Earthquakes beneath the Peru-Chile trench do not form a linear Benioff zone that might define a continuous slab (Isacks and Molnar, 1971).
Rather, earthquakes with T-axes down-dip extending to a depth of about 200 km indicate a shallow dipping slab while the deepest events below 500 km have P-axes down-dip and indicate a steeply dipping slab. There is very little seismic activity between the depths of about 200 and 500 km. The portion of the slab able to pull on the Nazca plate may thus be confined to depths above 200 km, and the net pull may be less than has been assumed in the simple models of slab forces used in this thesis (see also Section 4.4).

For most models, the calculated stresses for South America and eastern Asia have a large component of deviatoric tension that is not supported by the data. This may result from the oversimplification that slab forces act symmetrically about the trench axis. Model E29 with slab forces only on the plate with subducted lithosphere-produced stresses for South America and eastern Asia in much better agreement with the data, although the agreement with the data was poor in oceanic regions near subduction zones. Future modeling should include asymmetric subduction zone forces with other methods for balancing the torque due to slab forces besides drag forces acting on each plate.

The Andes may represent another possible source for compressive stresses in South America. In much the same manner that the Himalayas and Alps may stress the plates, the
horizontal density variation associated with the Andes may produce compressive stresses in South America. Future modeling might include the effects of major mountain ranges and perhaps the effects of horizontal density variations associated with continental margins.

The number of parameters associated with driving mechanism models is already fairly large (e.g., Table 5.3). It is likely that if even more parameters were considered, such as subduction zone forces that varied from region to region, the fit to the data could be improved. The models are already non-unique, and the philosophy behind the modeling in this thesis has been to describe the dominant features of the driving mechanism that are functions of boundary type with as few free parameters as possible.

To briefly summarize the results of the driving mechanism models, net forces at ridges, subduction zones, and continental convergence zones are comparable for the best models. Slab forces are at most a few times greater than other forces acting on the plates. Viscous drag is best modeled as a resistive force, with a drag coefficient that is non-zero beneath oceanic lithosphere, but which may be concentrated beneath continental
lithosphere by a factor of five or ten.

In the final chapter, the results of the thesis are summarized.
TABLE 5.1. Description and Ratings of Finite Difference Driving Force Models

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*Force Parameters
†Fit to Observed Stresses

Fig. 4, Richardson et al., 1976
Fig. 5, Richardson et al., 1976
Fig. 6, Richardson et al., 1976
No F<sub>S</sub> along Alpine-Himalayan belt: Fig. 7, Richardson et al., 1976
### TABLE 5.1. continued

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* Units for force parameters are arbitrary.

† AME is the American plate; EUA, Eurasian; AFR, African; IND, Indian; PAC, Pacific, NAZ-COC, Nazca-Cocos; and ANT, Antarctic. The numeral 2 indicates a fairly good visual fit to all observations in Figure 5.1; 1, a fairly good fit to at least one observation in Figure 5.1; and 0, no fit.
### TABLE 5.2. Adopted Values for Relative Plate Motions

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TABLE 5.2. Notes

1 Unless otherwise noted, relative rotation vectors are taken from Minster and Jordan (1978).

2 NAM is the North American plate; PAC, Pacific; COC, Cocos; NAZ, Nazca; EUA, Eurasian; INDI, Indian; ANT, Antarctic; AFR, African; CAR, Caribbean; SAM, South American; ARB, Arabian; and PHL, Philippine.

3 The convention for the relative rotation vectors is such that, seen from above the rotation pole, the first named plate moves counterclockwise with respect to the second named plate.

4 Rotation poles taken from Fitch (1972).
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**TABLE 5.3. Finite Element Driving Mechanism Models**

\(^1\) Force Parameters

\(^2\) Fit to Observed Stresses
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<td>.1</td>
<td>2 1 1 1 2 1 1 1</td>
<td>5.37</td>
</tr>
<tr>
<td>E31</td>
<td>1</td>
<td>1</td>
<td>0</td>
<td>1</td>
<td>0</td>
<td>.6</td>
<td>.1</td>
<td>.1</td>
<td>2 1 1 1 2 1 1 1</td>
<td>5.38</td>
</tr>
<tr>
<td>E32</td>
<td>1</td>
<td>1</td>
<td>0</td>
<td>2</td>
<td>0</td>
<td>.6</td>
<td>.1</td>
<td>.1</td>
<td>2 1 1 2 2 1 1 1</td>
<td>5.39</td>
</tr>
</tbody>
</table>

$^1$ Units for \( F_R, F_T, F_Y, F_C, \) and \( F_S \) are \( 1 \times 10^{15} \) dynes cm\(^{-1}\), which is equivalent to a stress of 100 bars across a 100 km thick plate. Units for \( D_C, D_O, \) and \( D_Y \) are \( 1 \times 10^6 \) dynes cm\(^{-2}\), which is equivalent to a shear stress of 1 bar for an absolute plate velocity of 1 cm/yr.

$^2$ NAM is the North American plate; SAM, South American; EUA, Eurasian; IND, Indian; AFR, African; PAC, Pacific; NAZ, Nazca; and ANT, Antarctic. The numeral 2 indicates a fairly good visual fit to general pattern of the observed intraplate stress directions in Figure 5.1; 1, a reasonable
fit to the sense, if not the exact orientation, of the data in Figure 5.1; and 0, a poor

Calculated stresses for NAM, SAM, PAC, NAZ, and ANT are so small with respect to EUA, IND, and

Drag forces specified for each plate as listed in Table 5.5.
### TABLE 5.4. Absolute Plate Velocities for Drag Models

<table>
<thead>
<tr>
<th>Model</th>
<th>Description</th>
<th>Lat °N</th>
<th>Long °E</th>
<th>$\omega \times 10^{-7} \text{ deg/yr}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>E17</td>
<td>Uniform drag coefficient beneath all plates</td>
<td>-62.37</td>
<td>111.08</td>
<td>6.99</td>
</tr>
<tr>
<td>E18</td>
<td>Drag beneath continents only</td>
<td>-59.43</td>
<td>109.84</td>
<td>9.48</td>
</tr>
<tr>
<td>E19</td>
<td>Drag beneath old oceanic lithosphere only</td>
<td>-57.72</td>
<td>147.00</td>
<td>3.28</td>
</tr>
</tbody>
</table>

1 Listed is the absolute velocity vector for the Pacific plate with respect to a presumably fixed mantle. Convention as in Table 5.2. Absolute velocities of other plates follow from this table and the relative plate velocities given in Table 5.2.
TABLE 5.5. Shear Stress on the Base of the Plates for Model E29

<table>
<thead>
<tr>
<th>Plate</th>
<th>Pole</th>
<th>Shear Stress, bars</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Lat 0°N</td>
<td>Long 0°E</td>
</tr>
<tr>
<td>PAC</td>
<td>-57.3</td>
<td>71.4</td>
</tr>
<tr>
<td>NAM</td>
<td>-44.3</td>
<td>-35.5</td>
</tr>
<tr>
<td>SAM</td>
<td>-58.2</td>
<td>142.7</td>
</tr>
<tr>
<td>EUA</td>
<td>43.2</td>
<td>-56.1</td>
</tr>
<tr>
<td>AFR</td>
<td>66.1</td>
<td>77.8</td>
</tr>
<tr>
<td>IND</td>
<td>46.8</td>
<td>2.0</td>
</tr>
<tr>
<td>ANT</td>
<td>11.7</td>
<td>-148.1</td>
</tr>
<tr>
<td>NAZ</td>
<td>43.0</td>
<td>-78.5</td>
</tr>
<tr>
<td>COC</td>
<td>38.4</td>
<td>170.8</td>
</tr>
<tr>
<td>CAR</td>
<td>--</td>
<td>--</td>
</tr>
<tr>
<td>ARB</td>
<td>36.1</td>
<td>46.5</td>
</tr>
<tr>
<td>PHL</td>
<td>-36.9</td>
<td>-24.7</td>
</tr>
</tbody>
</table>

1 For list of plate abbreviations, see Table 5.2.

2 Absolute rotation pole of plate with respect to presumably fixed mantle. Convention as in Table 5.2.

3 Basal shear stress $-\mathbf{\omega} \times \mathbf{r}$ at 90° distance from the rotation pole, where $\mathbf{\omega}$ is the absolute angular velocity and $\mathbf{r}$ is the radius from the center of the Earth.
FIGURE CAPTIONS

Figure 5.1 Principal horizontal deviatoric stresses inferred from intraplate earthquake mechanisms, in-situ strain and stress measurements, and stress sensitive geologic features. Filled circles are earthquake mechanisms and open circles are in-situ measurements and geologic indicators. Arrows for filled circles denote the horizontal projection of P (compressional) and T (tensional) axes. Solid line for open circles denotes the orientation of the maximum horizontal compressive stress axis. Numbers refer to Table 2.7. For regional description of intraplate stress data, see Figures 2.1-2.6.

Figure 5.2 Plate boundaries adopted for finite difference intraplate stress calculation (from Richardson et al., 1976). All segments of plate boundaries fall along either meridians or parallels and pass midway between adjacent grid points on a 10°x10° spherical grid. Ridges are denoted by double lines, subduction zones by single solid lines, and (implicit) transform faults by dashed lines. Forces at plate boundaries are applied in the direction of relative plate velocities, shown by arrows at selected points.
Figure 5.3 Schematic illustration of forces potentially important for the driving mechanism.

Figure 5.4 Principal horizontal deviatoric stresses in the lithosphere predicted for the finite difference model S1 (see Table 5.1). Principal stresses with arrows pointing inward and outward denote deviatoric compression and tension, respectively. The length of each stress axis is proportional to the magnitude of the principal stress; all lengths are arbitrary to within a multiplicative constant.

Figure 5.5 Finite element grid for mid-latitudes, Part 1. Constant strain triangular elements have dimensions of 5x5x7 degrees for plate interiors and 3x3x4 degrees for plate boundaries. There are a total of 5246 elements for the entire grid.

Figure 5.6 Finite element grid for mid-latitudes, Part 2. For other details, see Figure 5.5.

Figure 5.7 Finite element grid for north polar region.

Figure 5.8 Finite element grid for south polar region.
Figure 5.9 Forces at ridge boundaries. Forces are applied in the direction of the arrows. Numbers refer to Table 5.2.

Figure 5.10 Location of plate boundary elements with lowered Young's modulus.

Figure 5.11 Principal horizontal deviatoric stresses in the lithosphere for model E1 (see Table 5.3). Principal stress axes without arrowheads denote deviatoric compression while axes with arrowheads pointing outward denote deviatoric tension.

Figure 5.12 Forces at subduction zone boundaries. Lines denote orientation of forces acting toward the trench axis from either side of the plate boundary. Numbers refer to Table 5.2.

Figure 5.13 Principal horizontal deviatoric stresses in the lithosphere predicted for model E2 (see Table 5.3). For other details, see Figure 5.11.

Figure 5.14 Principal horizontal deviatoric stresses in the lithosphere for model E3 (see Table 5.3). Principal stress axes with arrows pointing inward and outward denote deviatoric compression and tension, respectively.
Figure 5.15 Forces at continental convergence zones. Arrows denote orientation of forces. Numbers refer to Table 5.2.

Figure 5.16 Principal horizontal deviatoric stresses in the lithosphere for model E4 (see Table 5.3). For other details, see Figure 5.11.

Figure 5.17 Principal horizontal deviatoric stresses in the lithosphere for model E6 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.18 Principal horizontal deviatoric stresses in the lithosphere for model E7 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.19 Principal horizontal deviatoric stresses in the lithosphere for model E9 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.20 Principal horizontal deviatoric stresses in the lithosphere for model E10 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.21 Principal horizontal deviatoric stresses in the lithosphere for model E11 (see Table 5.3). For
other details, see Figure 5.14.

Figure 5.22 Principal horizontal deviatoric stresses in the lithosphere for model E13 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.23 Principal horizontal deviatoric stresses in the lithosphere for model E14 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.24 Principal horizontal deviatoric stresses in the lithosphere for model E15 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.25 Principal horizontal deviatoric stresses in the lithosphere for model E17 (see Table 5.3). For other details, see Figure 5.11.

Figure 5.26 Principal horizontal deviatoric stresses in the lithosphere for model E18 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.27 Principal horizontal deviatoric stresses in the lithosphere for model E19 (see Table 5.3). For other details, see Figure 5.14.
Figure 5.28 Principal horizontal deviatoric stresses in the lithosphere for model E21 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.29 Principal horizontal deviatoric stresses in the lithosphere for model E22 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.30 Principal horizontal deviatoric stresses in the lithosphere for model E23 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.31 Principal horizontal deviatoric stresses in the lithosphere for model E24 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.32 Principal horizontal deviatoric stresses in the lithosphere for model E25 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.33 Principal horizontal deviatoric stresses in the lithosphere for model E26 (see Table 5.3). For other details, see Figure 5.11.

Figure 5.34 Principal horizontal deviatoric stresses in the lithosphere for model E27 (see Table 5.3). For other details, see Figure 5.14.
Figure 5.35 Principal horizontal deviatoric stresses in the lithosphere for model E28 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.36 Principal horizontal deviatoric stresses in the lithosphere for model E29 (see Table 5.3). For other details, see Figure 5.11.

Figure 5.37 Principal horizontal deviatoric stresses in the lithosphere for model E30 (see Table 5.3). For other details, see Figure 5.14.

Figure 5.38 Principal horizontal deviatoric stresses in the lithosphere for model E31 (see Table 5.3). For other details, see Figure 5.11.

Figure 5.39 Principal horizontal deviatoric stresses in the lithosphere for model E32 (see Table 5.3). For other details, see Figure 5.14.
Figure 5.1
MODEL S1
Figure 5.5
Figure 5.6
FINITE ELEMENT GRID
SOUTH POLE

Figure 5.8
RIDGE FORCES

Figure 5.9
Figure 5.10
SUBDUCTION ZONE FORCES

Figure 5.12
Figure 5.14
CONTINENTAL CONVERGENCE ZONE FORCES

Figure 5.15
MODEL E6

Figure 5.17
MODEL E9

Figure 5.19
Figure 5.21
Figure 5.22
Figure 5.24
Figure 5.25

MODEL E17

250 bars

324
Figure 5.28
Figure 5.31
Figure 5.32
Figure 5.33
Figure 5.34
Figure 5.37
Figure 5.39
CHAPTER 6: SUMMARY AND CONCLUSIONS

Long wavelength features of the intraplate stress field may be modeled in terms of various plate driving and resisting forces which act on the edge and along the base of the lithospheric plates. Estimates of the potential of various forces acting on the plates to drive and stress the plates indicate that topographic features such as mid-ocean ridges and mountain ridges at continental convergence zones may exert compressive deviatoric stresses of several hundred bars on the plates while cool, dense subducted lithosphere at oceanic subduction zones may exert deviatoric tensile stresses of several kilobars on the plates. Viscous drag forces acting on the base of the lithosphere depend on a constitutive relation between strain rate and deviatoric stress, the viscosity of the asthenosphere and the absolute velocity of the lithospheric plate with respect to the mantle. Shear stresses on the order of a few bars are suggested from estimates of mantle properties and absolute plate velocities of a few cm/yr. Whether these shear stresses drive or resist plate motions depends on the assumed absolute motions of the lithospheric plates with respect to the mantle.

The global intraplate stress field based on earthquake mechanisms, in-situ strain and stress measurements...
and stress sensitive geologic features, has been summarized and several long wavelength patterns for the orientation of horizontal principal deviatoric stresses have been discussed. Much of stable North America is characterized by an E-W to NE-SW trend for the maximum compressive stress. Maximum compressive stresses for continental South America trend E-W to NW-SE. Western Europe north of the Alps is characterized by a NW-SE trending maximum horizontal compression, while Asia has the maximum horizontal compressive stress trending more N-S, especially near the Himalayan front. The Indian plate has a horizontal maximum compressive stress direction that varies from nearly N-S in continental India to more E-W to NE-SW in Australia. Horizontal stresses are variable in Africa, but tend to indicate a NW-SE trend for the maximum compressive stress in west Africa and an E-W trend for the minimum compressive stress in east Africa. Oceanic lithosphere away from plate boundaries is generally described by a state of deviatoric compression, although few focal mechanisms can be constrained to define the orientation of the principal stresses.

Intraplate and plate boundary environments have been compared using seismically determined apparent stress and stress drop. No significant differences in the absolute level of stresses for the two environments can be inferred from the data. The apparent stress and stress
drop data are consistent with, but do not require, ambient tectonic stresses on the order of hundreds of bars rather than kilobars.

A finite element method applicable to analysis of intraplate stress for a single lithospheric plate due to plate driving forces and large plate boundary earthquakes has been developed and applied to the Nazca plate. Ridge pushing forces are required in all models that match the nearly east-west horizontal compression inferred from thrust earthquakes in the interior of the Nazca plate. The net pulling force of the subducting slab on the oceanic plate is at most comparable to ridge pushing forces, indicating that subducted lithosphere transmits only a small portion of its gravitational potential to the surface plate. Based on the estimate of ridge pushing forces and the limit on the ratio of the net slab pulling force to ridge forces, regional intraplate deviatoric stresses are estimated to be on the order of a few hundred bars. Changes in the Nazca intraplate stress field due to the 1960 Chilean earthquake are, at most, a few tens of bars locally and about one bar at greater distances into the plate. Such small changes in stress levels are probably not significant, although the corresponding changes in the displacement field should be observable using precise geodetic measurement techniques.

Calculation of intraplate stresses due to plate
driving and resisting forces has been extended from one plate models of the Nazca plate to global models in Chapter 5. The wave front and finite element solution technique has been employed. Ridge pushing forces are required in all models that match the inferred intraplate stresses summarized in Chapter 2. The net pulling force of subducted lithosphere is at most a few times larger than other forces acting on the plates. Resistive forces associated with trench thrust faults and motion of the slab with respect to the mantle must nearly balance the large gravitational potential of the cool, dense slab. The upper limit on net slab forces may be increased by less than a factor of two if net slab forces are reduced for the fastest moving plates (e.g., the Pacific and Nazca plates) by assuming that resistive forces acting on the slab increase with subduction rate.

Forces acting at continental convergence zones along the Eurasian plate that resist further convergence are important for models of the intraplate stress field in Europe, Asia, and the Indian plate. The upper bound on net slab forces cannot be increased by including continental convergence zone forces.

Resistive viscous drag forces acting on the base of the lithosphere improve the fit between calculated and observed stresses in the Nazca and South American plates
as long as the drag coefficient is non-zero beneath oceanic lithosphere. The calculated intraplate stress field is not very sensitive to an increased drag coefficient beneath old oceanic lithosphere compared to young oceanic or continental lithosphere. Increasing the drag coefficient beneath continents compared to oceanic lithosphere by a factor of five or ten changes the calculated continental stresses only slightly, and has little effect on the overall fit of calculated stresses to observed stresses.

Models in which viscous drag forces drive, rather than resist, plate motions are in poor agreement with intraplate stress data, although this lack of fit may depend on the oversimplified model of mantle flow patterns that has been assumed.

A model of the driving mechanism in which slab forces act only on the plate with subducted lithosphere and where viscous drag forces are assumed to balance the torque on each plate due to boundary forces produces stresses in good agreement with the data for the Nazca plate and several continental regions. The fit between calculated and observed stresses for this model is poor for oceanic regions near subduction zones. This model suggests, however, that it is probably an oversimplification to assume that the net force exerted by the slab acts symmetrically about the
The primary contribution of this thesis toward answering the question of the driving mechanism for plate tectonics is the demonstration that observed intraplate stresses can be explained in terms of the forces acting on the edge and along the base of the lithospheric plates. Comparison of observed intraplate stresses with calculated stresses for various models of the driving mechanism is an effective test of the relative importance of plate driving and resisting forces. The modeling can be extended in the future to include distributed forces due to topographic features such as continents, thermal forces associated with regions of anomalously thin or thick lithosphere, and shear resistance along transform faults. The role of drag forces can be further tested for specific and realizable flow patterns in the mantle. Additional data on the intraplate stress field, particularly for oceanic regions, should help to define regional patterns that are now poorly known. Of special importance for further constraining models of the plate tectonic driving mechanism is the reliable determination of the magnitudes of principal lithospheric stresses on a regional scale.
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The author's contributions to science include:


