12.710 Marine Geology and Geophysics
Forces that Drive Plate Motion

Turcotte and Schubert, Chapter 6
Kennett, Chapter 5
Fowler, Chapter 7

Introduction

Plate tectonics is the surface manifestation of thermal convection in the mantle. Motions of the Earth's tectonic plates are caused by the need to maintain a thermal balance between the hot interior and cool exterior. All parts of the circulation system must function in order to produce motion.

Note that **plate forces are sources of lithospheric stress**.
By definition stress equals force/(area over which the force is applied).

The Basic Question

Do plates passively ride along on top of the horizontally moving part of the convection cell -*OR*-- are the plates a fundamental part of the cell that drags along material beneath them?

We know that the mantle must convect to release heat so the mantle is not a passive participant in the plate motion process. However, the question remains as to how active a role the plates play. One significant issue is whether shear stress on the base of the plates aids or hinders plate motion.

<u>Tectonic Stresses in Plates</u>

The state of stress in the lithosphere provides a strong constraint on the forces that act on plates. We can compare stress orientations to plausible models of force distributions.

What kinds of forces act on plates?

- **gravitational** (buoyancy) – these are somewhat of a challenge to characterize because the depth distribution of internal density anomalies are not sufficiently well characterized.
- forces exerted on the **bottom surfaces** of plates - not too hard to characterize (area forces)
- forces exerted at plate **boundaries** - not too hard to characterize (line forces)

Units

- SI unit for pressure or stress is pascal=1 kg m^{-1} s^{-2}
- 1 MPa=10^6 Pa
- 1 MPa = 10 bars
- atmospheric pressure at earth's surface: 0.987 atmosphere = 1 bar
- strength of lithosphere near surface in tension: a few tens of MPa
- strength of lithosphere near surface in compression: ~100 MPa

Line forces I: Ridges

"Ridge push" - net compressive stress exerted on adjacent plates due to **"gravity sliding"** force from ridges. Ridge push occurs because at the ridge crest an oceanic plate is hot and so sits at a higher elevation than surroundings. This results in a "pushing" force away from the ridge manifest as near-ridge normal faulting (where the plate is in tension) and plate shortening away from the ridge (where the plate is in a state of net compression).

The ridge push force, F_r, can be expressed

$$F_r = \int_0^h (\rho_m - \rho_w) gz dz \tag{1}$$

where ρ_m and ρ_w are the densities of the mantle and seawater, z is depth, g is the acceleration due to gravity, and h is the lithosphere thickness The force due to the elevation of the ocean ridge depends on the age of the lithosphere. For a 100-km-thick lithosphere, F_r =3.9x10^{12} N. The stress level is about 30 MPa if integrated over a 100-km column depth.

Ridge push dominates over the resistance to spreading at the ridge axis. To convert stress to force, simply apply the stress over the length of the plate boundary (and the thickness, if one is constructing a 3-D model).

Line Forces II: Plate Convergence Forces

"Slab pull" - due to **"negative buoyancy"** associated with density contrast due to temperature difference between a subducted slab and surroundings. The subducting slab is colder than its surroundings and thus more dense. One may write the temperature dependence of density

$$\rho = \rho_o (1 - \alpha_v \Delta T) \tag{2}$$

where ρ is density, ρ_o is the density at a reference temperature (*i.e.* Earth surface temperature), α_v is the volumetric coefficient of thermal expansion (~3.5x10^{-5} K^{-1} for rocks), and ΔT is the temperature change. We may assume that the magnitude of this force is proportional to the length of the boundary segment. The force associated with temperature-related negative buoyancy is

$$F_{s1} = 2\rho_m g \alpha_v b \Delta T \left(\frac{\kappa\lambda}{2\pi u_o} \right)^{1/2} \tag{3}$$

where b is the depth scale, λ is the horizontal wavelength of mantle flow, κ is the thermal diffusivity, and u_o is the plate spreading velocity. For a spreading velocity of 50 mm yr^{-1}, b= 700 km, ΔT=1200 K and λ=4000 km, F_{s1}= 3.3x10^{13} N. There is an additional contribution to the total slab pull force due to the olivine-spinel phase change at about 660-km depth. This force is

$$F_{s2} = \frac{2\Delta T \gamma \Delta\rho_{os}}{\rho_o} \left(\frac{\kappa\lambda}{2\pi u_o} \right)^{1/2} \tag{4}$$

where $\Delta\rho_{os}$ is the density contrast associated with the phase change and γ is the slope of the Clapyeron curve for the phase change. For $\Delta\rho_{os}$=270 kg m^{-3} and γ=4 MPa K^{-1} F_{s2}=1.6 x10^{13} N. Thus the total slab pull force is F_s=4.9x10^{13} N. Note that that slab pull exerts a tensile stress on the plate of several hundred MPa. This is enough for tensile failure of the plate, which is, in fact, observed.

To be more correct, the effective force associated with subduction zones equals the net difference between (temperature- and phase-change-related) negative buoyancy and the various resistive forces that oppose subduction:

- drag at sides of slab
- shear resistance/friction along thrust planes
- excess pressure on leading edge of slab
- flexural resistance of slab (work done in plate bending)
- finite strength of plate

Evidence from subduction rates and earthquake focal mechanisms indicate that negative buoyancy and resistive forces approximately balance. The net slab pull is similar to or somewhat greater than the ridge push force.

Boundary forces at zones of active continental collisions are not the same as those at subduction zones.

- Elevated topography indicates that continental zones resist further convergence. (Recall that continental lithosphere is too buoyant to subduct. Collisional forces are much different in mountain belts and subduction zones).
- There is an ~100 MPa net compressive stress at mountain belts.

Drag Forces

Drag arises from viscous shear at base of a tectonic plate that is moving with respect to the asthenosphere. The magnitude of the force depends on the **plate velocity** and **asthenosphere viscosity**. The direction of motion depends on the velocity of local mantle flow relative to that of the plate, but generally occurs in direction opposite to plate motion. The drag force can be simply expressed

$$F_D = -Dv \tag{4}$$

where D is the drag coefficient and v is the plate velocity. The force is applied over the areal extent of the plates. The magnitude of F_D is ~1 MPa.

The **asthenosphere**, the zone of partial melt beneath the lithospheric plates, is probably not actively or passively dragging the plates but has complex motions.

Transform Fault Forces (small)

Shear stresses along transform boundaries resist plate motions. The magnitude is somewhat uncertain but comes from earthquake stress drops and models. One can also can use common-sense/back-of-the-envelope estimations to bound the magnitude: *i.e.*, if stresses of hundreds of MPa exist along major transforms then the ridge push and slab pull forces would not be adequate to drive the plates. Values of 10's of MPa are most likely.

Hot Spot Forces

These are somewhat similar to ridge pull but there are some notable differences. Hot spot forces:

- are oriented radially instead of linearly; and

• have a dynamic component (indicated by high geoid/topography ratios) whereas ridges are passive and the force is just due to elevation.

Global Stress or Plate Force Models

In a global stress or plate force model it is necessary to include the following:

- ridge push
- slab pull
- resistive drag from transforms
- resistive forces at orogenic belts and subduction zones
- push or pull on upper or downgoing plate at converging boundary
- hotspot contribution

Use observed plate motions and plate geometries to determine relative importance of forces.

What about the contributions of other kinds of stresses?

- Thermal stresses in cooling lithosphere ($\sigma \propto \alpha_v \Delta T$).
- Membrane stresses (due to change in radius of shell as plates move toward or away from the equator -- due to rotational flattening of Earth).
- Crustal thickness variations (stresses due to elevation differences)
- Loading
- Sea level changes
- Sedimentation and erosion
- etc.

All of these are small compared to the major plate forces.

Theoretical Issues

- The distribution of stresses in a plate is obtained by assuming that the lithosphere is a thin elastic shell.
- Bending moments are ignored as negligible since they are small in comparison to shear stresses in the plate (this is the basis for the thin plate assumption).
- Assume mechanical properties of plate are uniform throughout the shell.
- Perform a static balance of forces.

Model Assumptions

1) Each plate is in equilibrium, *i.e.,* vector sum of torques must = 0.

- forces at plate boundaries come from plate to plate interactions
- pull on slab comes from negative buoyancy
- drag on base of plates due to velocity relative to mantle below
- ridge elevation gives ridge push
- hotspots exert a push on the plates

2) The processes that determine forces on a segment of the boundary are characteristic of the boundary type. For example, the magnitude of a ridge push force is proportional to the length of the ridge segment. And the magnitude of the slab pull force is proportional to trench lengths.

Interlude: Calculating Torques

The motion of each plate has three degrees of freedom: (lat, lon, ω of pole of rotation), so the equations of static equilibrium provide three equations for each plate corresponding to three components of a torque vector taken about center of Earth.

To calculate torques:
- At each boundary segment calculate cross product of radius vector from center of Earth to segment with each coefficient vector for that segment. This provides a conversion from force to torque.
- Sum over all boundaries the contribution to the torque from each force parameter.
- Form a slab pull torque vector by summing individual slab pull torque vectors from all boundary segments that have a slab connected to them.

The vector equation has the form

$$a_{ij}x_j = b_i \qquad i=1,2,3 \tag{5}$$

where a_{ij} is the matrix of known coefficients, x_j is the vector of unknown forces, and b_i is the i^{th} component of the total slab pull torque. Equation (5) gives components for one plate. Multiply by the number of plates and this gives the number of equations that need to be solved as a linear system.

Solve system of equations by **least squares.** Define a residual

$$r_i = b_i - a_{ij}x_j \tag{6}$$

and **find the best fit by minimizing the square of the residual**

$$a_l a_{ik} x_k = b_i a_{il} \tag{7}$$

The magnitude of the residual torque vector represents the amount of torque that the model has failed to explain. The residual torque is due to:
- internal deformation of plates;
- lateral variations in plate strength;
- improper characterization of supposedly known quantities;
- etc.

Conclusions about plate forces based on observations and models

1) Downgoing slabs pull strongly. The slab pull force of subjecting lithosphere is the largest force, several times larger than any of the other forces that act on the plates.
- All plates with slabs move toward the slabs. This is true for plates with both large and small slabs.
- Plates with slabs have the fastest rates of motion.

2) Ridge forces are required to match the orientations of the observed stress field.

3) The resistive forces associated with the motion of the slab with respect to the mantle must nearly balance the gravitational potential of the slab.
- Slab pull is large in magnitude, hence resistive forces at slabs are significant.

- The upper limit on ratio of net slab to ridge forces may be increased by less than a factor of two if the net slab forces are reduced by the fastest moving plates by assuming that the resistance to subduction increases with plate convergence rate.

4) Forces that resist continued continental convergence along the Alpine and Himalaya belt are important for models of the intraplate stress field in Europe, Asia and the Indian plate.

5) The inclusion of resistive drag forces that act on the base of the lithosphere improve the stress orientation fits if the drag coefficient beneath old oceanic lithosphere is greater than that for young oceanic or continental lithosphere.
- Models in which viscous forces at the base of the plates drag rather than resist plate motions are in poor agreement with intraplate stress data.

6) In successful models slab forces act only on plates with subducted lithosphere and viscous drag forces balance the torques on each plate produced by ridge and trench forces.

7) Hot spot-related forces are not significant.

8) New observations of plate deformation that are being accumulated from regional studies are enabling more detailed and accurate characterization and interpretations about the magnitudes of plate forces.

The plates are not just passive players in the plate tectonics cycle. The negative buoyancy of slabs helps drive the circulation pattern in a significant way. Without negative buoyancy of slabs Earth could have mantle convection but might not have plate tectonics (cf. Venus).

Mantle Convection

How can we be certain that the mantle convects?

There are some straightforward calculations that prove the point. It is possible to engage in a physical exercise to demonstrate that mantle convection is an inevitable consequence of the Earth's need to rid itself of internal heat. Here we will derive a very simple expression for cooling of a medium of thickness d that is heated from below. The expression will include the fundamental parameters that are relevant to the process of internal cooling by mass movement:
- g - the acceleration of gravity;
- α_v - the volumetric coefficient of thermal expansion;
- ΔT - the temperature difference between the surface and lower boundary of the fluid;
- κ - thermal diffusivity= $\kappa/\rho C_p$, where k is thermal conductivity and C_p is specific heat; and
- ν - the kinematic viscosity of the medium, which describes how momentum diffuses.

We will show that these parameters can be expressed in dimensionless form as the **Rayleigh number**, *Ra*, which is a measure of the tendency of a medium to convect. The Rayleigh number **measures the relative importance of buoyancy forces (in the numerator) that drive convection and viscous resistance forces (in the denominator) that inhibit convection**. The Rayleigh number can be expressed in various forms that depend on the geometry of the medium and the nature of heating. However, a simple representation of the Rayleigh number can be determined from a balance of forces on a sphere in a uniform fluid medium. The sphere, of radius a ,will rise due to buoyancy forces at a velocity V. Buoyancy occurs because the sphere has a density $\rho-\Delta\rho$

that is less than the density of the fluid ρ. The buoyancy force scales with the volume of the sphere as

$$F_b = \frac{4}{3}\pi a^3 \Delta\rho g. \tag{7}$$

As the sphere rises there will be a viscous resistance to motion due drag over the spherical surface. This force can be written

$$F_r = \left(\mu \frac{V}{a}\right) 4\pi a^2 \tag{8}$$

where μ is the dynamic viscosity, which is a measure of the resistance of the medium. It is a ratio of the stress to the strain rate in a viscous medium. By equating the two forces we find the velocity at which the sphere will rise

$$V = \frac{\Delta\rho g a^2}{3\mu}. \tag{9}$$

If the sphere is less dense than its surroundings because it is warmer, we may express the velocity in terms of a temperature difference. We write

$$\Delta\rho = -\rho\alpha_v \Delta T. \tag{10}$$

Substituting into equation (9) we find

$$V = -\frac{\rho\alpha_v \Delta T g a^2}{3\mu}. \tag{11}$$

It is apparent that the sphere will keep rising until it cools sufficiently such that the buoyancy force is exceeded by viscous drag. The sphere will cool conductively at a time that is proportional to the thermal diffusivity of the medium and the size of the sphere as

$$\frac{\kappa t}{a^2} = c \tag{12}$$

where c is a constant. This is the **characteristic cooling time** of the sphere. Since velocity is just distance over time we may now solve for the distance the sphere will rise before it cools

$$v = \frac{d}{t} = \frac{\rho\alpha_v \Delta T g a^2}{3\mu} \tag{13}$$

For c=0.5, at which time the sphere will essentially be cool, the distance the sphere rises may be written

$$d = \frac{1}{6}\left(\frac{\rho\alpha_v g \Delta T a^3}{\mu\kappa}\right) a \tag{14}$$

The expression within the parentheses is dimensionless and is a ratio of buoyancy to drag. It is a simple form of the Rayleigh number, *Ra*. Often *Ra* is expressed in terms of the kinematic viscosity, which is related to the dynamic viscosity as $\upsilon = \mu / \rho$. Then

$$Ra = \frac{\alpha_v g \Delta T d^3}{\kappa \upsilon}. \tag{15}$$

There is a **critical value of *Ra*** that marks the onset of convection, Ra_c. The onset of convection also depends on whether the frictional surface boundary conditions are taken to be **no slip**, where the plates are assumed to resist convective motion along the top of the cell, or **free slip**, where the plates are assumed to be "along for the ride". For various geometries for a viscous slab of material that is heated from below, Ra_c falls in the range 100-1000**.**

The kinematic viscosity ν of Earth's mantle is estimated from the rate of **glacial rebound** of the crust. Remarkably, Earth's crust is still rising back into its pre-ice age shape following the most recent retreat of glaciers some 10,000 years ago. In 1935, **N.A. Haskell** used the ages of elevated beaches in Scandinavia to calculate that the viscosity of Earth's mantle is about 10^{23} times greater than the viscosity of liquid water.

Because of radioactive decay of heat producing nuclides (U, Th, K) in the mantle, it is appropriate to also consider the possibility that the Earth's mantle is heated from within, and not just from below. In its simplest manifestation, the Rayleigh number for a fluid slab heated from within takes the form

$$Ra = \frac{g \alpha_v d^3}{\kappa \upsilon} \frac{H d^2}{\kappa C_p} \tag{16}$$

where *H* is the internal heat generation per unit mass. With the value of ν determined from post-glacial rebound studies and various estimates of *H* and *d*, equation (16) as applied to Earth's mantle yields *Ra* in the range 10^6-10^9. This is orders of magnitude larger than the critical value for this problem Ra_c(~3000), which implies that **Earth's mantle is undeniably convecting**.

Observations of Earth's Interior Relevant to Convection: Seismic Tomography and Gravity

Detailed maps of mantle density structure derived from **seismic** tomography now resolve regions of hot upwelling and cold downwelling in the mantle. Certain important information on deep structure is also being gleaned from models of the Earth's gravity field. These investigations are now beginning to resolve the planform of mantle convection.

Based on inversions of many seismic waves that sample throughout the Earth, tomographic maps plot fast and slow wave speeds relative to some average Earth model, such as PREM. One of the fundamental observations of tomographic inversions is the recognition that many slabs penetrate the 660-km seismic discontinuity (Tonga, Marianas, Phillipine, Sunda) while others appear to be trapped in the upper mantle (Izu-Bonin, southern Kurile). The penetration of slabs beneath the 660-km discontinuity rules out the idea that mantle convection is fully stratified at this depth, but the severe deformation associated with deeply penetrating slabs dictates that the mantle does not have a simple whole-mantle convection pattern with single, well defined cells. The 660-km discontinuity clearly distorts mantle flow, but many slabs penetrate to at least 1700-km depth. There is a change in the heterogeneity spectrum of tomographic inversions between 1700 and 2300 km that is probably real, but its origin is unknown. This distinctive region, if it exists, would represent long-maintained differences in composition and/or phase chemistry.

A scenario favored by Louise Kellog, Rob van der Hilst and Brad Hager is that isolated "seismically distinctive" reservoirs exist in the deep mantle. They favor a mantle with the following components: First, there is undepleted and undegassed material in the lowest 1000 km of the mantle. For such regions to be preserved to the present they must have a slightly higher intrinsic density than the upper mantle. In contrast, the upper 2000 km of the mantle is involved in the recycling of slab material and represents the depleted and degassed source of mid-ocean ridge basalts. The upper mantle transition zone (400-1000 km depth) divides the depleted part of the mantle in a well-mixed, low viscosity domain and a deeper one with higher viscosity.

Seismic tomography has resulted in significant improvements in mantle structure. But there is much room for improvement. In the coming decade, better coverage may help delineate deep structure. Seismic anisotropy will be used to delineate mantle flow trajectories. And the combination of global seismic and gravity information and geodynamical modeling will clarify significantly the structure of the mantle, and further our understanding of plate dynamics.