12.002 Physics and Chemistry of the Earth and Terrestrial Planets
Fall 2008

Rheology and Seismology

Rheology of the Earth (and other Planets, but we know more about Earth)

The way that planets function internally and their topography and surface features depend largely on the "rhelogy" of the planet interiors. Rheology just means the way that particular rocks or materials deform under different pressure, temperature, hydration and other conditions. However, it is not a one-way street because the rheology of planetary interiors influences how well mixed a planetary interior is, and the mode of heat loss, so the rheology in turn influences the temperature. The rheology of a planet depends also on the material being deformed (e.g. crust, mantle or core) so that the size of the core and the size of the planet both play an important role in determining the rheology of the lower mantle.

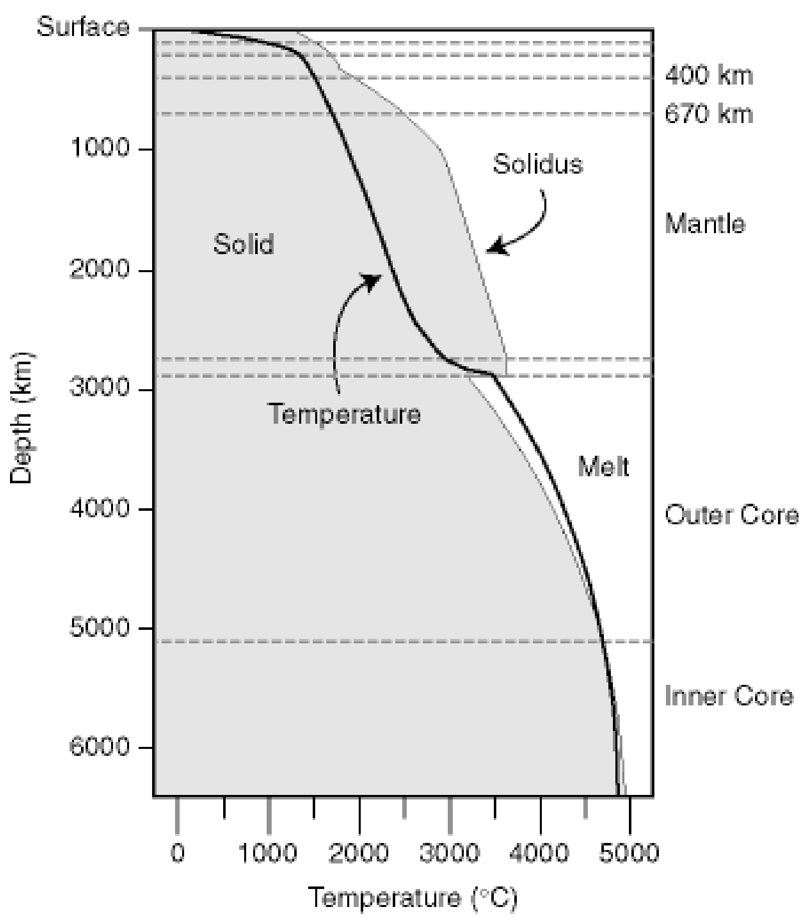
The three basic modes in which planetary materials deform can be broadly categorized as brittle, ductile and elastic. Within the brittle and ductile categories there are subsets of deformation, such as grain boundary deformation or diffusion creep, but we won't worry about that here. To make things more complicated, or more interesting depending on your perspective, deformation modes are quite sensitive to strain rate. If you deform something very rapidly it may behave in one mode, such as elastic, whereas if you deform the same material more slowly, it may respond by ductile failure, even at the same temperature and pressure conditions.

Elastic deformation is recoverable deformation. In elastic materials, stress is linearly related to strain. A material remains in a strained condition only as long as it is under stress. When the stress is removed, the material reverts to its initial shape.

This is different than ductile and brittle deformation where the deformation is more or less permanent. In **viscous** materials, stress is related to strain rate. This relationship is linear if it is a Newtonian viscous material, otherwise there is usually a power-law relationship, commonly with an exponent of 3 on the stress term. This means that a viscous material will stop deforming when we remove the applied stress, but the deformation that it has acquired up to that point remains in place.

Brittle deformation is familiar as the formation of cracks, fractures and slip on faults during earthquakes, as well as fault creep. It is also a permanent deformation but does not seem to be very strain-rate dependent. Instead, we can think of a critical yield stress that has to be reached before a fault, for example, can move. Many aspects of fault slip, including the physics of earthquake nucleation, remain very poorly understood. This is a big gap in our understanding of a planet whose surface deforms mainly by slip on faults.

In general, terrestrial planets are coolest at their surface and their temperature increases with depth. However, this does not happen in a very linear way. However, in the uppermost layers of the planet, where temperatures are lowest, the temperature typically increases downward in a fairly regular way. We will cover planetary heat flow later in the term, but for the moment it is worth looking at the basics of heat transfer by diffusion (conduction).



Conduction is the process by which heat diffuses through a material. For a material that is stationary and not producing heat internally, such as by radioactive decay, the relationship between heat loss, or heat flow, q, and the thermal gradient is:

$$q = K\frac{\partial T}{\partial z}$$

K is an empirically measured quantity called "thermal conductivity" and it has units of $Wm^{-1}K^{-1}$. In typical earth materials, at least crustal and mantle materials, K is commonly in the range of 2-10 W/mK.

The relationship between heat loss and thermal gradient means that if a planet is losing heat at a known rate, then its thermal gradient can be extrapolated downward through the lithosphere. What is the lithosphere? It is the strong outer shell of a planet that is not easily deformed. Its exact definition is a bit hazy and the lithosphere can be defined in different ways, but on earth you can think of it as the strong tectonic plates. The base of the lithosphere on earth is at ~1300-1400°C because above that temperature mantle materials become very weak at geologic stain rates (that is, at the rates of plate motions).

Because of plate tectonics, the earth is not losing heat everywhere at the same rate, for example heat loss is very high along mid ocean ridges and lower in the old ocean basins and continents. But if we take the background heat flow for the earth of 50 mW/m^2 (or .05 W/m^2) and an estimate for thermal conductivity of 5 W/mK, we calculate a thermal gradient for the lithosphere of 10°C/km. So, starting with a surface temperature of close to 0°C, we can estimate an lithospheric thickness of around 130-140 km, or more conservatively 100-170 km. In fact, this is not bad for old oceans. Continents can be thicker, up to 250 km or so, but we are still not far off.

So we can see that what controls the long-term thickness of planetary lithospheres is the heat flow through the outer shell of the planet and the surface temperature of the planet, at least if the thermal conductivities and rheology of the mantle materials is similar (remember about approximate chondritic compositions!) This also tells us that for surface temperatures less than 1000°C or so, all the terrestial planets should have a lithosphere. This will be underlain by softer, weaker and hotter mantle if the planet is not too cold or too small (but small planets cool more quickly, so those two quantities tend to go together.)

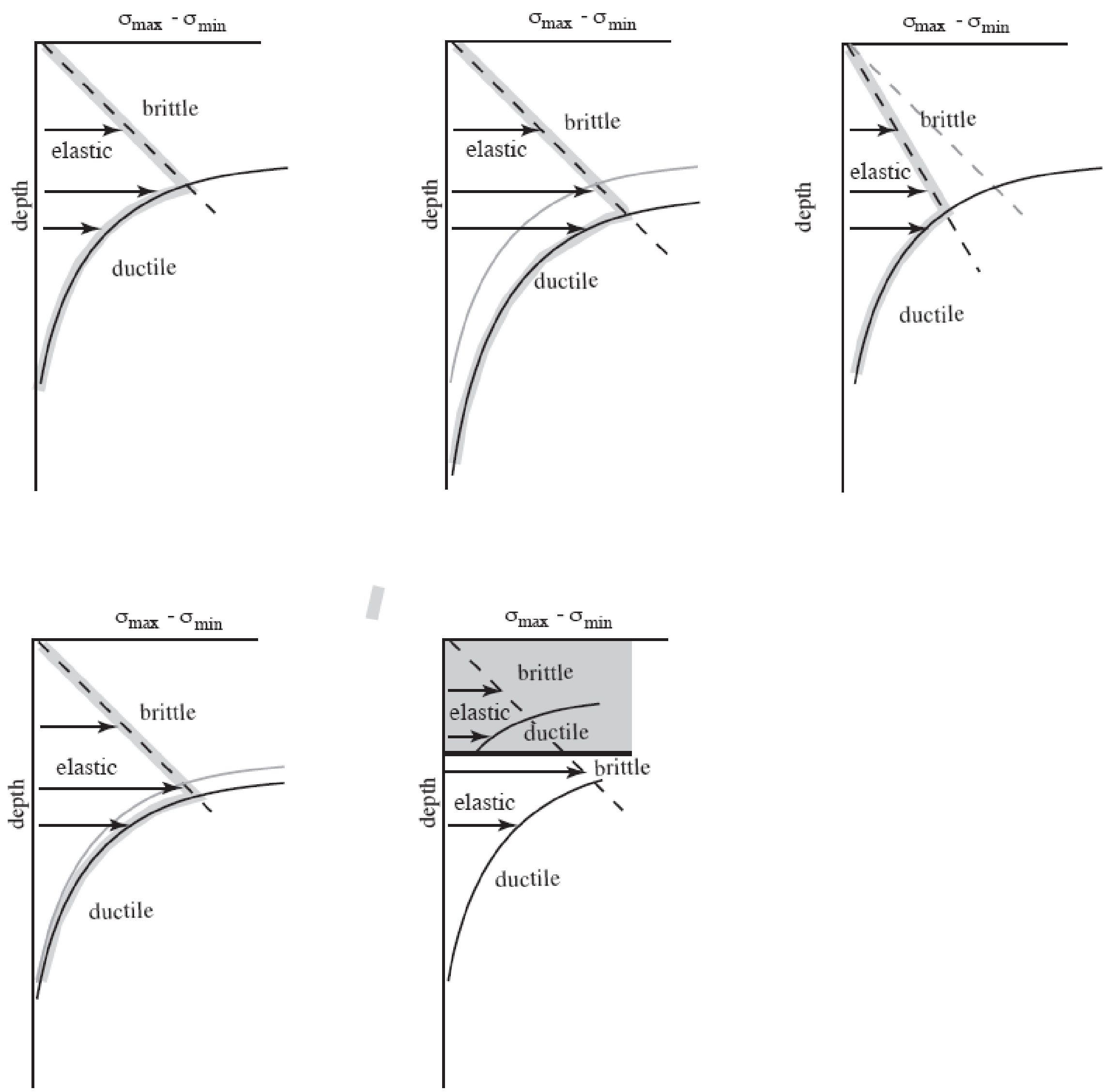
Yield Stress Diagrams for Planetary Lithospheres

Yield stress diagrams show where and how failure will occur throughout a column of a planet's lithosphere (containing crust and mantle), and at what stress difference. (Stress difference is the difference between the maximum and minimum compressional stress on a rock package, the orientation of those stresses will tell what kind of fault or strain will occur – ie extension, strike slip, etc.).

The yield criteria for brittle failure is close to linearly dependent on pressure. Since pressure goes up linearly with depth, so too does the stress difference required for brittle failure. Brittle yield criteria is not very sensitive to rock type, but it is lowered in the presence of high fluid pore pressure and is also lower on faults than on new fractures, on faults that have recently ruptured, and a whole host of other poorly understood things. The brittle failure criteria will also have a different depth dependence on different sized planets because g is linearly related to planetary radius and to average planetary density, so that pressure will be larger at a fixed depth below the surface on a large planet than on a small one.

Ductile failure can occur via several different deformation modes. In all modes, it is highly temperature dependent, with the stress difference for failure varying almost exponentially with temperature (rocks are weaker at higher temperatures). It is also highly strain rate dependent and rocks are much stronger at high strain rates than at low strain rates. This effect is not hugely important if we confine ourselves to geologic strain rate, much less important than temperature. However, if we consider the differences in strain rate between geologic strain rates and the strain rates associated with seismic waves traveling through the earth, there are many orders of magnitude difference so that material that is elastic on seismic time scales (seconds) can be ductile on geologic time scales (millions of years).

The diagrams below are yield stress diagrams that show how a planetary lithosphere will deform when stresses are imposed on it. The upper left hand diagram shows a reference case, and the subsequent cases, left to right, show what happens if you decrease the geothermal gradient, add fluids to weaken the material under brittle deformation, increase the strain rate, or put on a crust with a different (weaker) ductile yield criteria.



These yield diagrams show how a strong lithosphere can be constructed near a planetary surface, with an elastic core or sheet embedded in the lithosphere provided that the stress differences are not to large. On earth, and to some extent on the moon where there are a few seismometers, we can determine the thickness of the lithosphere seismically, but on other planetary bodies we have to use indirect means to estimate lithospheric thickness. One way to do this is to figure out how thick this effective elastic sheet is within the lithosphere, and use that as a measure of total lithospheric thickness, and thus of heat flow.