

3 Gravity and the lithosphere

The Earth is self-gravitating in as much as it creates its own gravitational field. The gravitational body force exerted on material of unit volume in this field is given by the product of its density and the acceleration due to gravity (which is function of the mass distribution in the Earth). The way mass is distributed in the lithosphere is central to the notion of *isostatic compensation* and the consequences of gravity acting on the mass distribution in isostatically compensated lithosphere is central to all of geodynamics.

3.1 Isostasy

It has long been recognized that the surface elevation of the continents is in some way related to the density distribution in the subsurface. The gravity field at the surface of the earth reflects the distribution of mass at depth and gravity measurements across mountain belts show that regions of high elevation generally have a deficiency of mass at depth. That is, there is some specific depth beneath the mountain range where the rocks have a lower density compared with rocks at the same depth beneath the low lying regions flanking the mountain range. The gravity field shows that the deficiency of mass at depth is, to a first approximation, equal to the excess mass in the mountains, implying that at some great depth within the earth, termed the depth of *isostatic compensation*, the mass of the overlaying rock is equal and independent of the surface elevation. From the gravity data alone, it is not possible to determine exactly how the density is distributed in order to compensate the topography, and two rival isostatic models have been proposed, referred to as Pratt Isostasy and Airy Isostasy (Figure ??). These models were proposed long before the concept of the lithosphere was formalized and in the original formulation both models considered that isostatic compensation was achieved entirely within the crust.

While modern seismic methods have shown that the structure of mountain ranges closely approximates the Airy model, it is important to dispel the notion that compensation is generally achieved at the bottom of the crust. The logical place for compensation to take place is beneath the lithosphere, and in this course we will initially assume, and then attempt to demonstrate, that this is the general case. Indeed the very existence of large lateral temperature gradients in the oceanic mantle lithosphere leads to significant horizontal density gradients. These thermally induced density changes lead to corresponding changes in the surface elevation of the oceanic lithosphere, through a kind of thermal isostasy. The horizontal force resulting from the ocean floor topography, termed *ridge push*, provides one of the

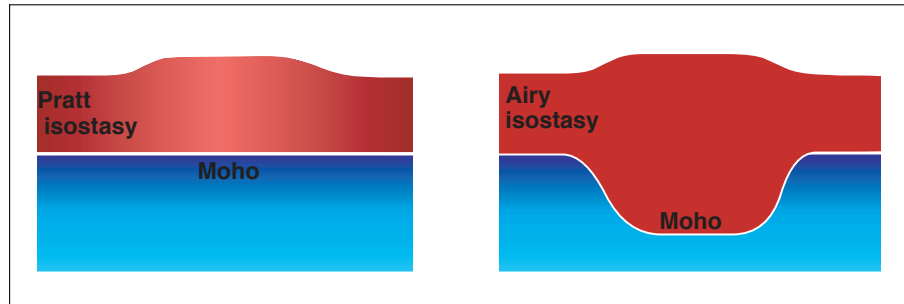


Figure 3.1 Pratt (a) and Airy (b) isostasy assume rather different subsurface density distributions to compensate the excess mass associated with the additional topography in mountains. Density distribution is proportional to stippling. Seismic studies beneath mountains such as the Himalayas show that the density structure approaches that of Airy Isostasy, although it is now assumed that isostatic compensation occurs beneath the lithosphere and not at the Moho.

fundamental driving forces for the motion and deformation of the lithosphere. Isostatic compensation can be achieved because the lithosphere essentially *floats* on a relatively *inviscid* substrate: the weak peridotite of the asthenosphere. Changes in the buoyancy or elevation of the lithosphere are accommodated by displacement of asthenospheric mantle. However, asthenospheric mantle is not completely inviscid (that is, its viscosity is not negligible), and its displacement in response to lithospheric loading or unloading must take a finite length of time, related to its effective viscosity. An insight into the timescales for the isostatic response of the asthenosphere to loads is provided by the rebound of continental lithosphere following the removal of glacial icecaps. The rebound following the removal of the Pleistocene Laurentide icecap shows that the time scales appropriate to this isostatic response are of the order of $10^4 - 10^5$ years. Since mountain belts are built and decay on the time scale of $10^7 - 10^8$ years, the isostatic response is effectively instantaneous.

3.2 The flexural strength of the lithosphere

An important question concerning isostasy is the horizontal length scale on which isostatic compensation is achieved. Gravity measurements across mountain belts suggest that, on the scale of several hundred kilometers, the lithosphere often approaches isostatic equilibrium. However the same measurements show that the mass excess associated with small-scale topography, for example, individual mountain peaks within mountain ranges, is generally not compensated. The length-scale of isostatic equilibrium (viz., regional versus local isostatic compensation) relates to the strength of the lithosphere: on small length-scales departures from isostatic equilibrium are supported by the flexural (or elastic) strength of the lithosphere.

We illustrate the interaction between the flexural strength of the lithosphere and the horizontal length scale of isostatic compensation by considering a hypothetical infinitely strong lithosphere floating on a completely inviscid substrate. Subject to a substantial, localized load such as a mountain range, such a hypothetical lithosphere will be depressed such that the mass of the displaced asthenosphere is equivalent to the mass of the load. In this case the mass of the mountains is compensated over the regional horizontal dimension of the lithosphere. Of course gravity shows us that this is not the case for the earth, because the density distribution beneath mountains somehow compensates the excess mass of the mountains on a length-scale of comparable dimension to the mountains (Figure 3.2b). However, on the smaller length scale of individual mountain peaks compensation is not achieved; that is the lithosphere is sufficiently strong to distribute the load of an individual mountain over a length scale which is large compared to the lateral dimension of the individual mountain.

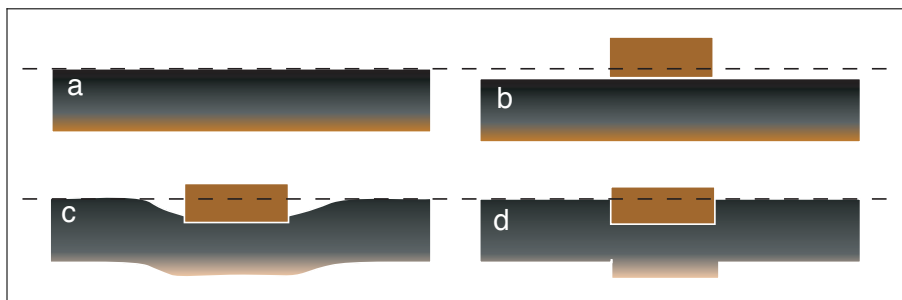


Figure 3.2 The response of the lithosphere (a) to applied loads depends on its flexural strength (or rigidity). An infinitely strong lithosphere compensates loads regionally over the lateral dimension of the whole lithosphere (b). At the other end of the spectrum compensation may occur completely beneath the load, so called local isostatic compensation (d). The behaviour of the lithosphere, which has some finite flexural strength, lies somewhere between these two extremes (c), implying that this flexural response of the lithosphere is an important component of deformation associated with loading. For a visco-elastic lithosphere in which the response to loading is time dependent these figures illustrate schematically the expected temporal evolution following loading as the elastic response gives way to a viscous response through time.

As we shall see, the lithosphere has a finite flexural strength, in as much as it can support a limited amount of loading without permanent deformation. The effective flexural strength of the lithosphere is characterized by the thickness of the elastic lithosphere, which in turn is dependent on the thermal and compositional structure of the lithosphere. Importantly the flexural response of the lithosphere to an applied load may be time dependent (as would be expected for a visco-elastic material);

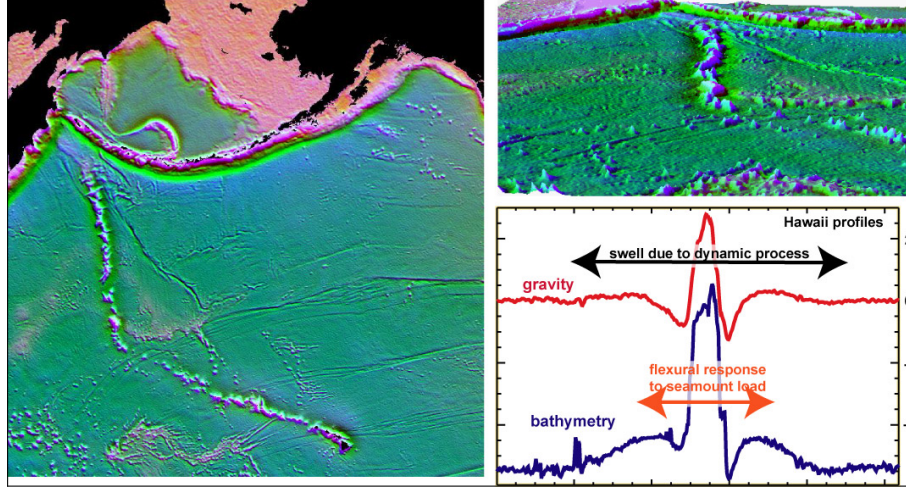


Figure 3.3 Example of lithospheric flexure associated with loading from seamounts on the Emperor-Hawaiian seamount chain in the NW Pacific ocean

after the emplacement of a load, the effective elastic thickness of the lithosphere may decay with time as the elastic stresses are relieved by permanent deformation.

3.3 Gravitational potential energy

Isostasy is a vertical stress balance!

Local isostasy requires that at the level of isostatic compensation the mass of overlying rock is equalized, independently of the density distribution above this level. Equivalently the normal stress in the vertical direction σ_{zz} is equalized at the depth of compensation. Necessarily, the vertical density structure at, and below, the level of compensation must everywhere be equal. The vertical stress resulting from gravitational *body forces* acting on the mass, or body, of the rock. The vertical stress, σ_{zz} , at depth h is given by:

$$\sigma_{zz} = g \int_0^h \rho \, dz \quad (3.1)$$

where ρ is the density.

Under hydrostatic conditions ($\sigma_{zz} = \sigma_{xx} = \sigma_{yy}$) the lithostatic pressure, P , is equal to σ_{zz} . Thus for constant density (depth independent) hydrostatic conditions:

$$P = \sigma_{zz} = g \rho h \quad (3.2)$$

Isostasy is not a complete stress balance!

A complete stress balance requires that horizontal stresses as well as vertical stresses must be balanced. In order to achieve this in the presence of a gravitational field all density interfaces must be horizontal. This is clearly not the case for the lithosphere which is characterized by significant lateral variation in the density structure. For example, significant topography characterizes the density interfaces at the Earth's surface (rock-air), the Moho (mantle-crust) and within the mantle lithosphere. Variations in lateral density structure within the lithosphere contribute to variations in the lithospheric gravitational potential energy, U_l . The gravitational potential energy of a lithospheric column of unit area is given by the integral of the vertical stress from the surface of the earth to the base of the lithosphere (i.e., depth of isostatic compensation):

$$\begin{aligned}
 U_l &= \int_0^h \sigma_{zz} dz \\
 &= g \int_0^h \rho dz \\
 &= g \int_0^h (\rho z) dz
 \end{aligned} \tag{3.3}$$

Because the lithospheric potential energy is defined in terms of a column of unit area it has dimensions Jm^{-2} or equivalently, Nm^{-1} . Lateral variations in gravitational potential energy, ΔU_l , induce substantial horizontal forces, termed *buoyancy forces*. The difference in potential energy between two columns, 1 and 2, integrated over the depth h , is (Figure 3.4):

$$\begin{aligned}
 \Delta U_l &= \int_0^h \sigma_{zz1} dz - \int_0^h \sigma_{zz2} dz \\
 &= g \int_0^h (\Delta \rho z) dz
 \end{aligned} \tag{3.4}$$

where $\Delta \rho$ is $(\rho_1 - \rho_2)_z$.⁹

Equations 3.3 and 3.4 simply state that in the presence of a gravitational field there is a tendency to reduce fluctuations in gravitational potential energy (*note that these equations apply equally to the forces operating on an ageing brie, as it spreads to lower its centre of gravity*). While variations in potential energy generally correlate with surface elevation it is important to realise that this is not necessarily so, as

⁹ The term $\int_0^h (\Delta \rho z) dz$ is referred to as the dipole moment of the density distribution.

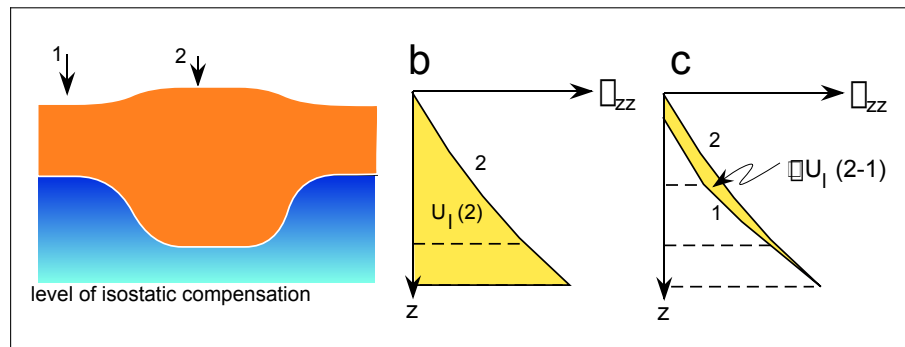


Figure 3.4 The magnitude of the buoyancy forces arising from variations in potential energy generally correlate with topographic gradients on density interfaces in the lithosphere is equal to the area between the σ_{zz} vs. depth curves for both regions. Note that σ_{zz} is equalized at the level of isostatic compensation, independent of the overlaying density structure.

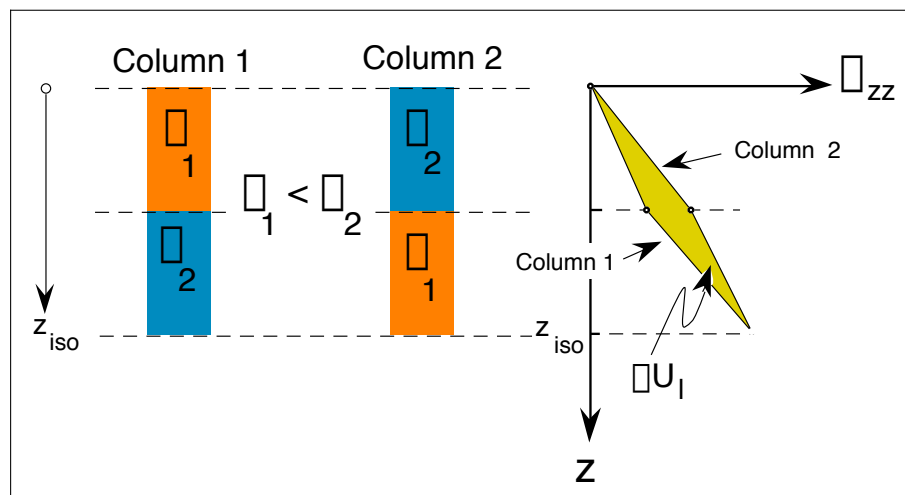


Figure 3.5 While variations in potential energy generally correlate with variations in elevation, they need not do so as illustrated in this scenario. Clearly column 2 is gravitationally unstable. In a geological context this scenario may have some relevance to old ocean lithosphere (see Chapter 7)

illustrated in Figure ??.¹⁰ In the absence of an externally applied stress field, regions of high potential energy will experience tension, and regions of low potential energy will experience compression.

¹⁰ In Chapter 9 we show that, ignoring contributions due to variations in the thermal structure of the mantle lithosphere, for Airy isostasy the surface elevation, h , varies linearly with crustal thickness, z_c , while potential energy, U_l varies with z_c^2 (see also Chapter 14, p 290, in Brown, Hawkesworth and Wilson).

The buoyancy forces arising from variations in gravitational potential energy are large (in fact they provide the fundamental driving forces for the horizontal motion and deformation in the lithosphere). These forces can be sustained because rocks have finite strength *just as an immature brie does*. An understanding of the rheology of rocks, that is the way in which they respond to applied forces, is fundamental to developing quantitative tectonic models.

Gravity acts on mass and as mass is distributed throughout the volume of a body the forces resulting from the action of gravity are termed *body forces* (other types of body forces such as magnetic forces are largely irrelevant in tectonics). As we have seen in the previous section the effect of gravitational body forces acting on two columns with different density distributions produces forces acting on the surface between the two columns. Such *surface forces* are dependent on the area the orientation of the surface. Since the sum of the forces acting on a body are equal to the mass times acceleration (Newton's second law) it is relatively simple to derive the equations of motion, relating displacements and forces (see Appendix A.1). The equations of motion are necessarily couched in terms of the components of the stresses (or, more correctly, the stress tensor) and therefore they must be rendered useful through combination with equations relating stresses to displacements. Such equations, termed *constitutive equations*, are material dependent and their study is the stuff of *rheology*.

3.4 Lithospheric potential energy and geoid anomalies

The contributions of lateral variations in the density structure of the lithospheric to observed geoid anomalies can be directly related to the dipole moment of the density distribution:

$$\Delta N = -\frac{2\pi G}{g} \int_0^h (\Delta\rho z) dz \quad (3.5)$$

where ΔN is the geoid anomaly in metres. Consequently, variations in lithospheric potential energy are reflected in variations in the geoid, with an anomaly of 5 m corresponding to a variation in potential energy of about $1 \times 10^{12} \text{ N m}^{-1}$.

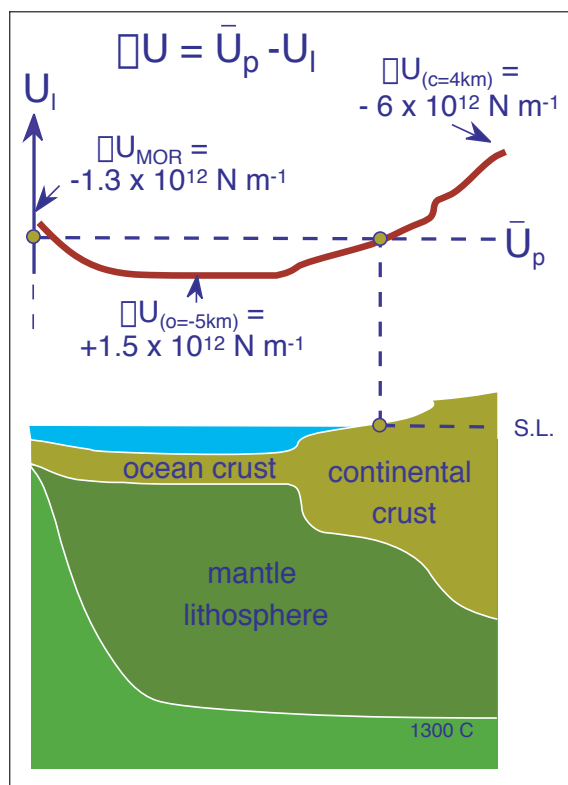


Figure 3.6 Illustration of the way in which lithospheric gravitational potential energy varies across ocean and continental lithosphere)