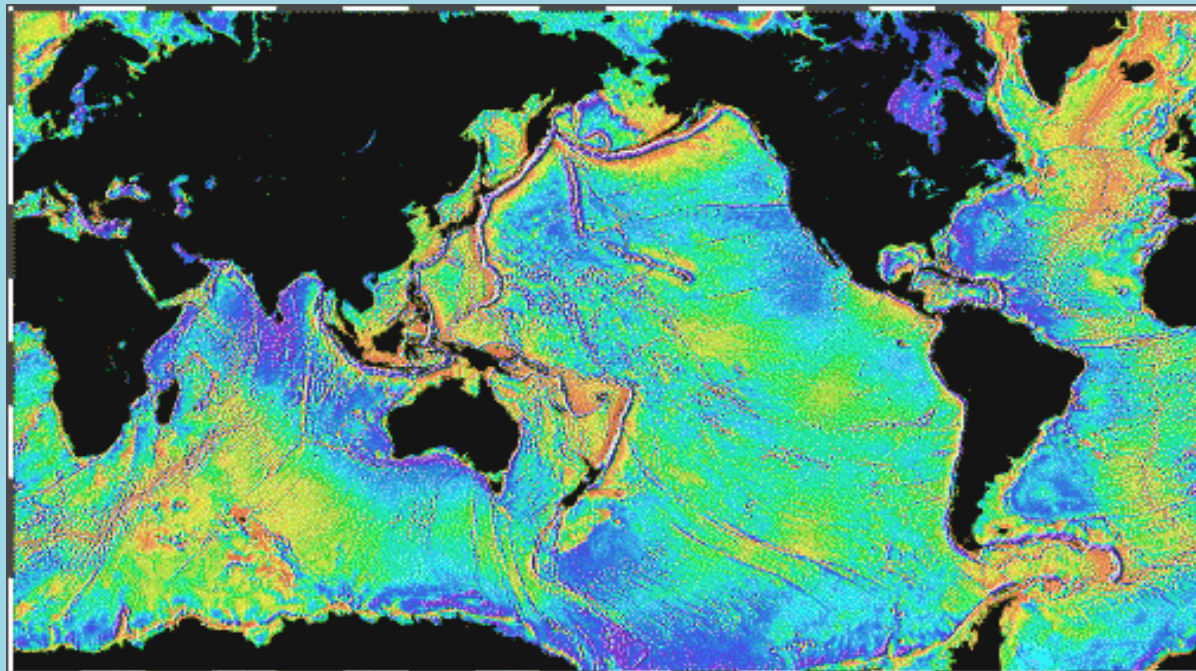


Lecture 2: Gravity, Isostasy and Flexure

Read pages 42-51 in KK&V

You can access all of KK&V through eReserves
(password: yf160)

Homework 1 is due next Thursday



Gravity: theory and measurements

gravity - "weighing" the earth

$$g = \frac{GM_e}{R_e^2}$$

g = acceleration caused by attraction of one mass to Earth's mass

$g = 980 \text{ cm/sec}^2 = 980 \text{ Gals} = 980,000 \text{ mGals (milligal)}$

*This is for a stationary earth. For a rotating earth there are terms for centrifugal acceleration and nonspherical shape of earth amounting to <1%.

G = gravitational constant ($6.67 \times 10^{-11} \text{ Nm}^2/\text{kg}^2$)

M_e = mass of Earth ($5.973 \times 10^{24} \text{ kg}$)

R_e = radius of Earth

equatorial radius = $6.378139 \times 10^6 \text{ m}$

polar radius = $6.35675 \times 10^6 \text{ m}$

Cavendish experiment (1798)

average density = 5.448 Mg/m^3

modern value = 5.517 Mg/m^3

At high pressures density of silicate minerals increases significantly

e.g. olivine at 1.7 Mbar, density = 5.5 Mg/m^3

pressure at center of earth: $P = 3.6 \times 10^{11} \text{ Pa} = 3.6 \times 10^6 \text{ bar} = 3.6 \text{ Mbar}$

So denser material (primarily Fe) needed to account for average density of earth

composition of crust known from direct sampling

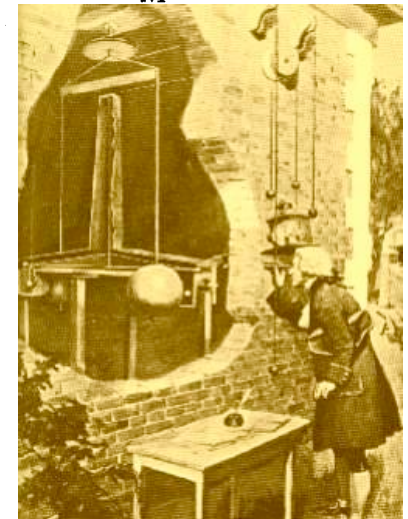
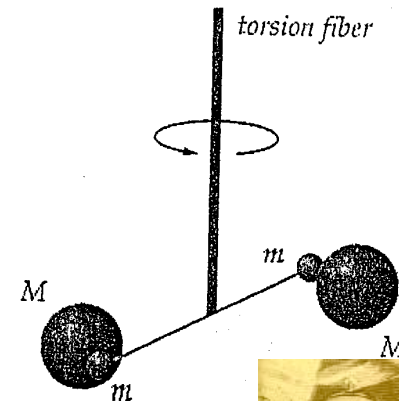
meteorites, composition of solar system

core: 95+% Fe, some Ni and possibly light elements (O,S,K)

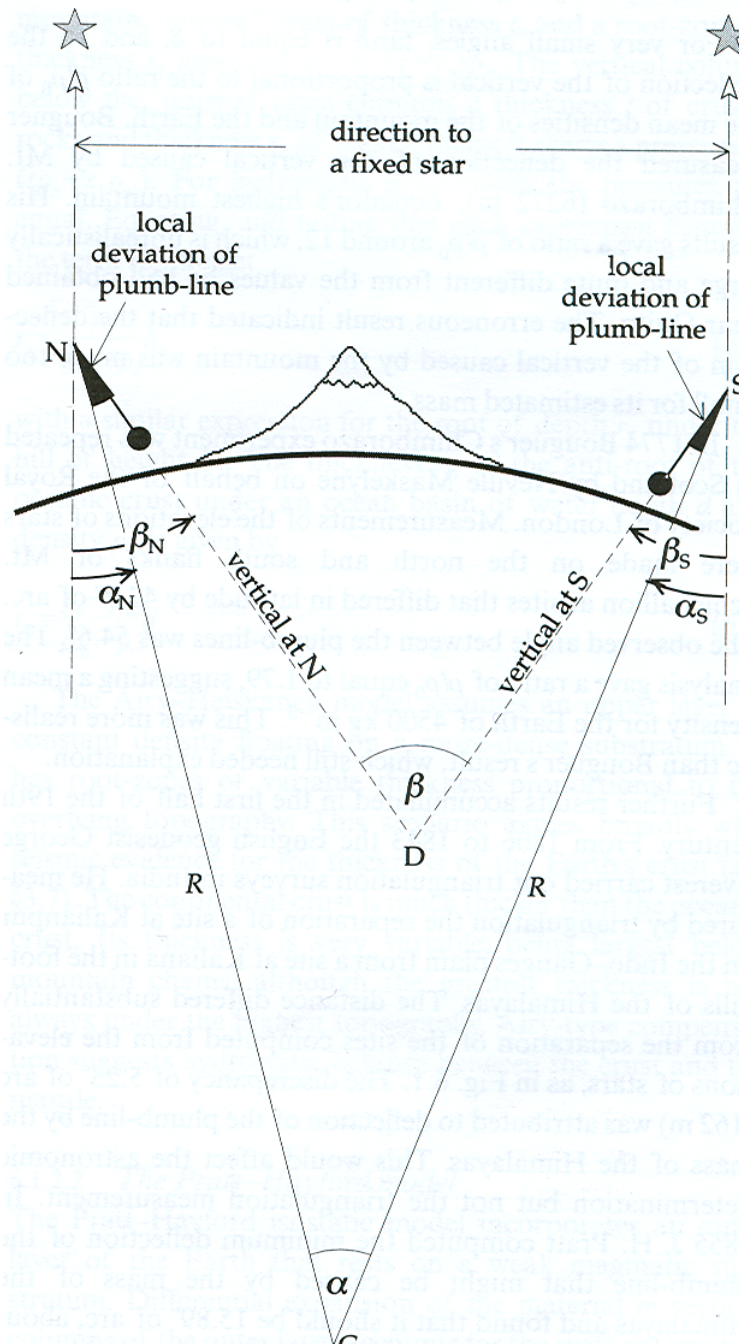
supported by presence of magnetic field

Newton: $F = G (m_1 m_2)/r^2$

and $F = ma$

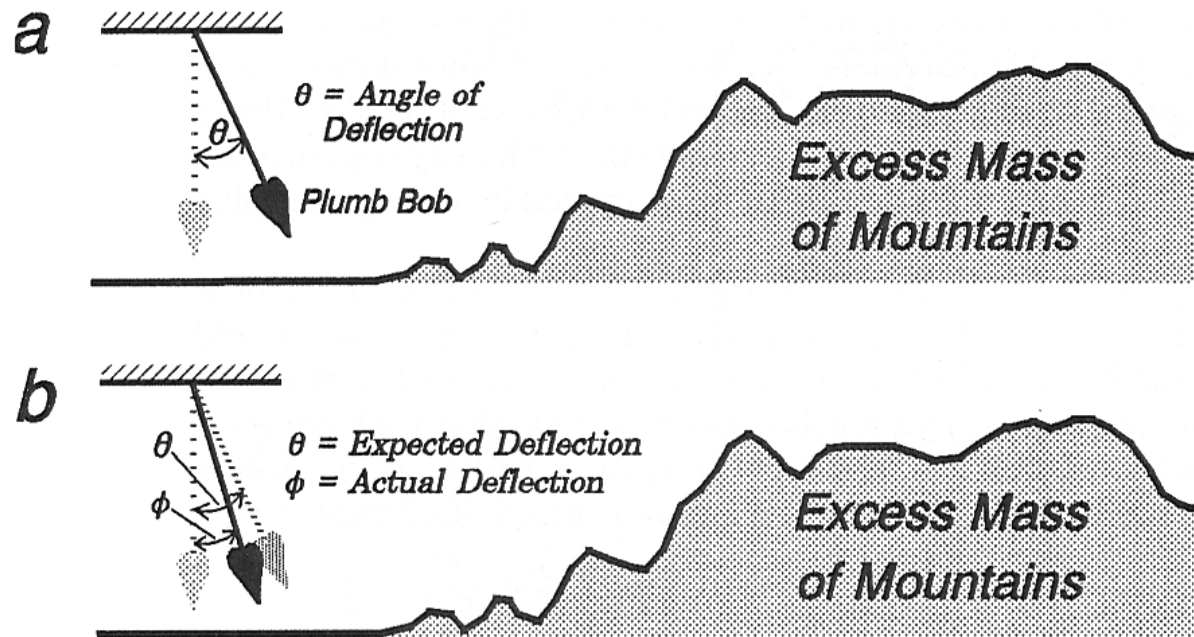


In XVIII and XIX centuries, surveys set out to measure the shape of the Earth

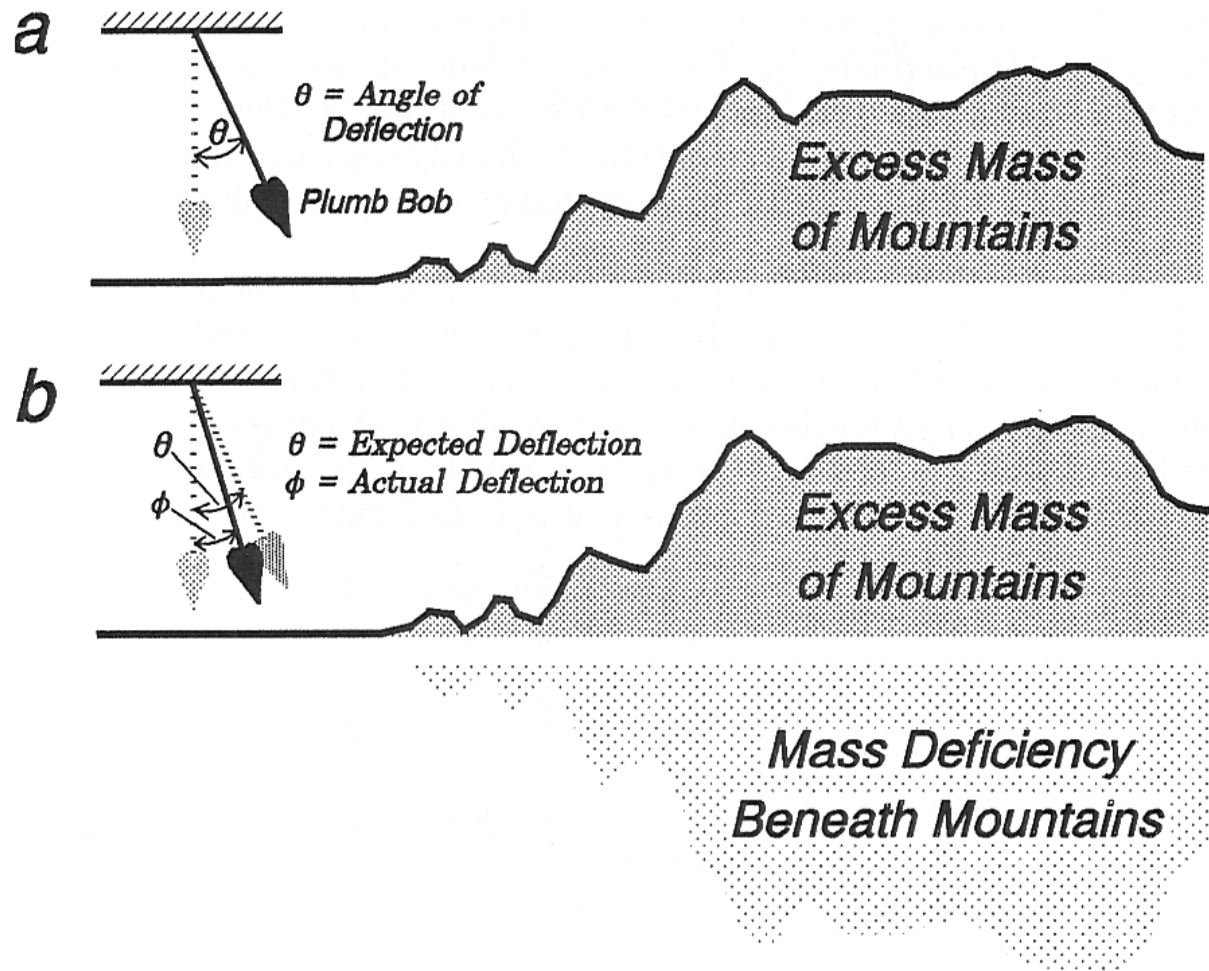


They used plumb bobs and expected them to be attracted toward adjacent mountains such as Andes and Himalayas

... but the plumb bobs did not deflect as much as expected

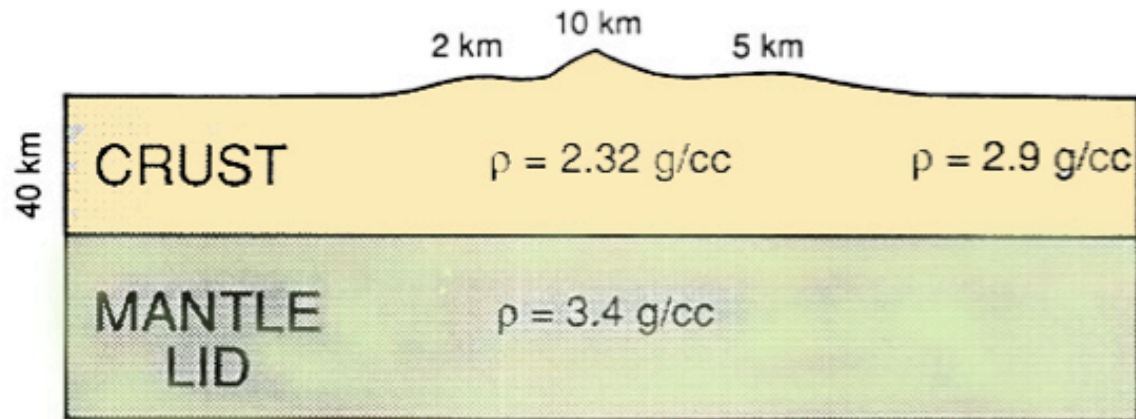


... smaller than predicted deflection could result from either overestimated mass (density) of the mountains, or some deficit of mass below the mountains



Isostasy – the state of gravitational equilibrium between the Earth's crust and mantle, as if the (lighter) crust were floating on a (heavier) mantle

Pratt Isostasy



Airy Isostasy

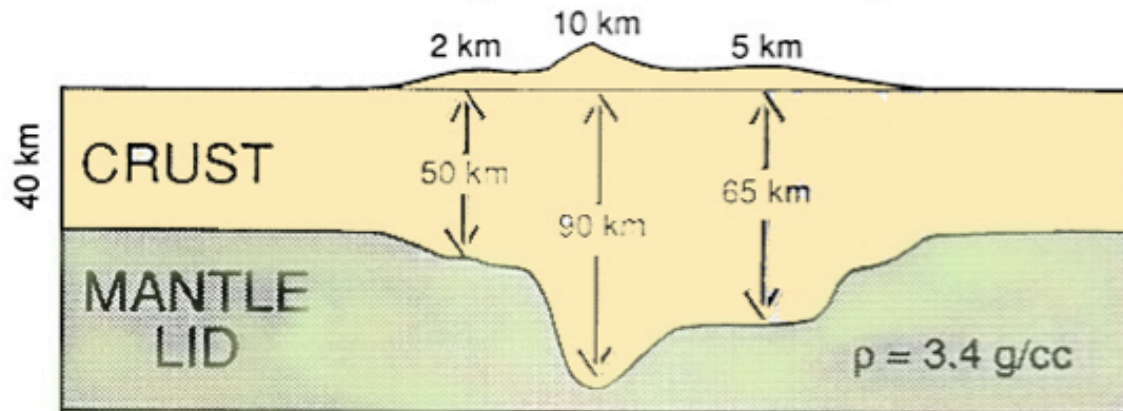


Fig. 1.10. John Henry Pratt 1 year after his appointment as Chaplain to the Bishop of Calcutta. Portrait courtesy of D. P. McKenzie of Cambridge University.



where G is the universal gravitational constant and d is the distance between the two masses. In the case that M and P are points on the surface of a spherical Earth, then it follows (Fig. 1.11) that the component of the attraction at a station P due to the elementary mass at M *in the direction of gravity* is given by

$$dg = \frac{G \, dm \, \sin(\xi)}{d^2}$$

where

$$\xi = \frac{\theta_s}{2} \quad \text{and} \quad d = 2r_e \sin\left(\frac{\theta_s}{2}\right)$$

$$\therefore dg = G \, dm \frac{\sin\left(\frac{\theta_s}{2}\right)}{4r_e^2 \sin^2\left(\frac{\theta_s}{2}\right)} \quad (1.2)$$

Pratt used Eq. (1.2) to calculate the gravitational effect of the Himalayas at Kaliana and Kalianpur, and published the results in a 75-page long paper (Pratt, 1855). He found that the gravitational effect of the mountains is large enough to deflect the plumb line by $15.885''$, more than three times the observed value (Fig. 1.12). Pratt was satisfied, despite problems with not knowing the detailed topography of the Himalayas, that he had correctly computed the effect of the mountains at Kaliana and Kalianpur. He concluded his paper by saying that he did not understand the cause of the discrepancy and that the problem should be investigated further.

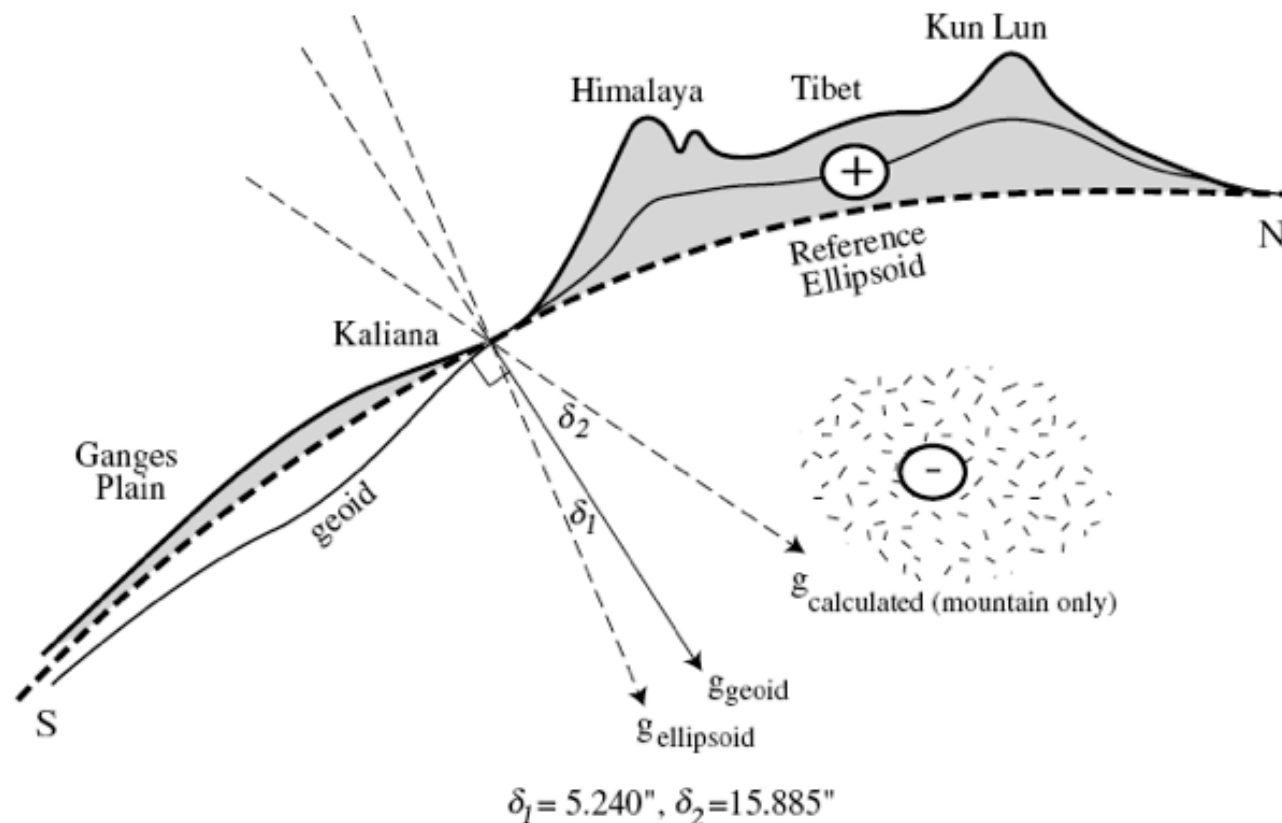
Fig. 1.13. George Biddell Airy, Professor of Astronomy at Cambridge University and Astronomer Royal. Portrait courtesy of D. P. McKenzie of Cambridge University.



1.4 Isostasy According to Airy

Shortly after Pratt's paper, G. B. Airy (1801–92; Fig. 1.13), the Astronomer Royal, presented a paper to the Royal Society in which he offered an explanation for the discrepancy. Unlike Pratt, Airy was not surprised by the discrepancy. Indeed, he thought that it should have been anticipated.

Airy's argument was based on his belief that the outer layers of the Earth consisted of a thin crust that overlay a fluid layer of greater density than the crust. He referred to the fluid layer as "*lava*". Airy compared the state of the crust lying on the lava to timber blocks floating on water. He wrote (Airy, 1855, p. 103):



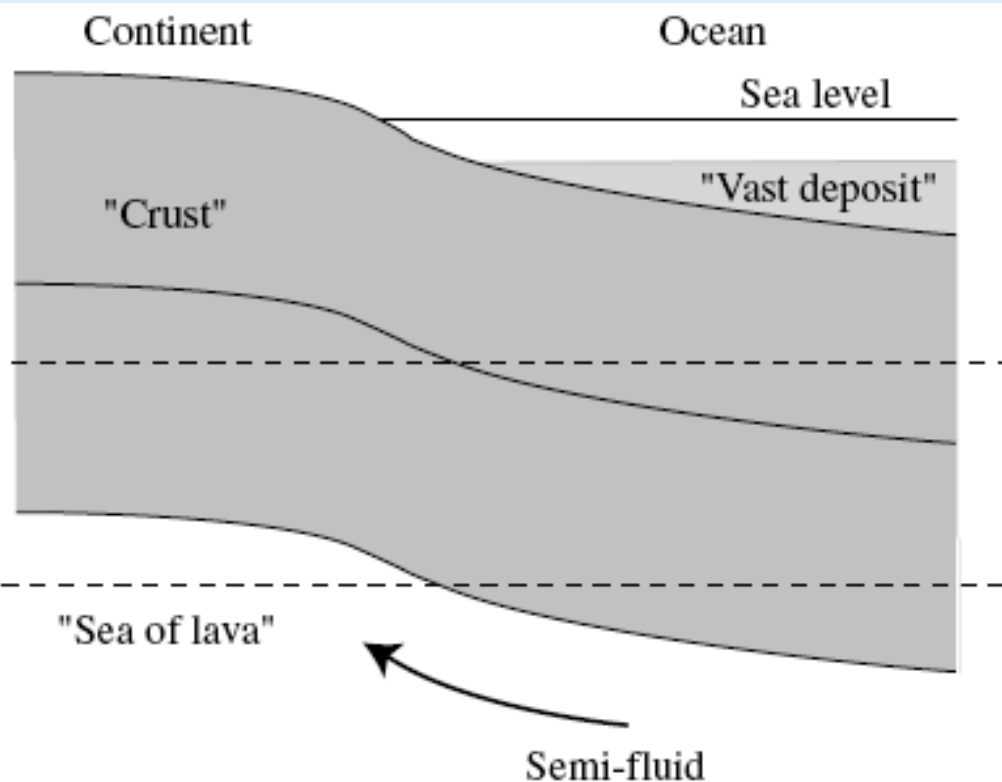


Fig. 1.6. The adjustment of the crust to a "vast deposit" by flow in the underlying "sea of lava". Reproduced from a figure in Herschel (1836) with permission of the Royal Society.

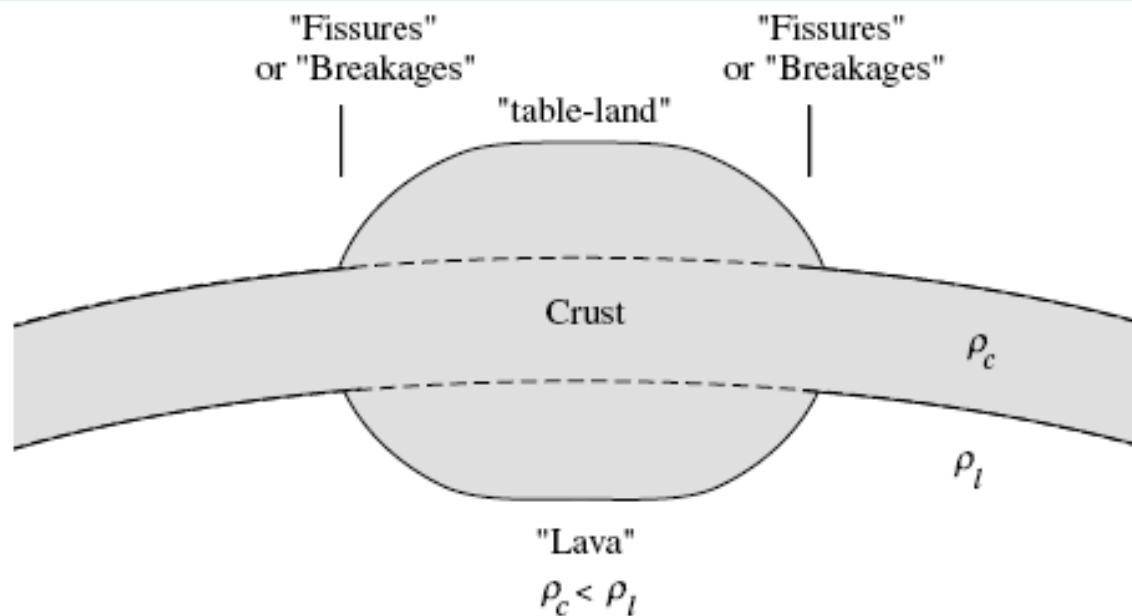
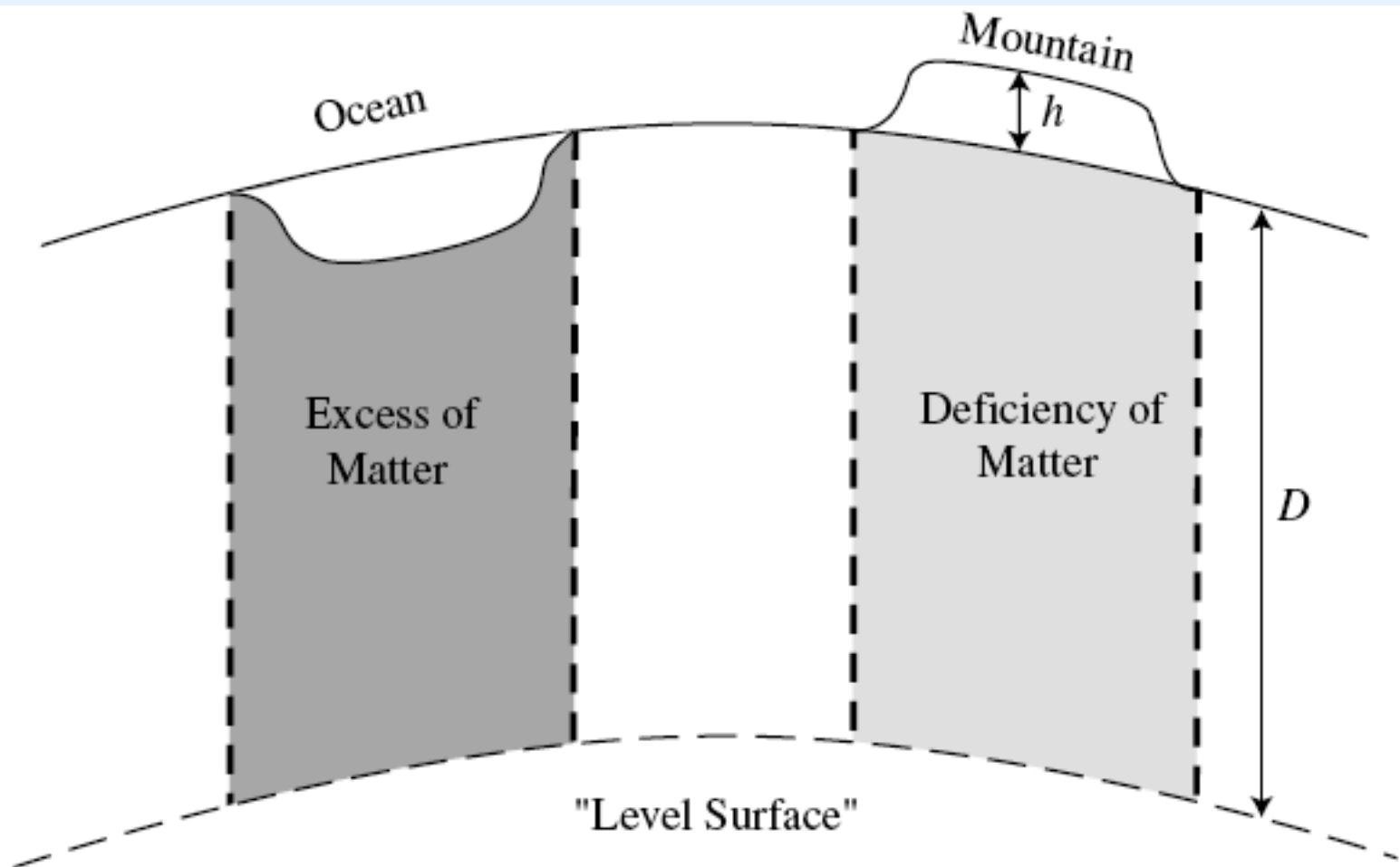
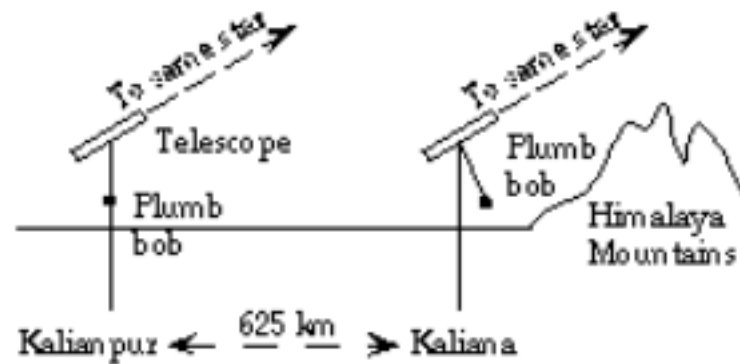
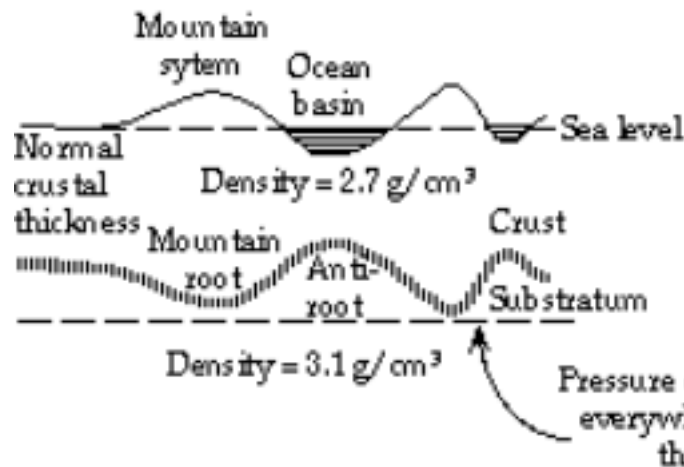


Fig. 1.14. Airy's hypothesis of a crust that "floats" upon "lava". Reproduced from Fig. 2 of Airy (1855) with permission of the Royal Society.

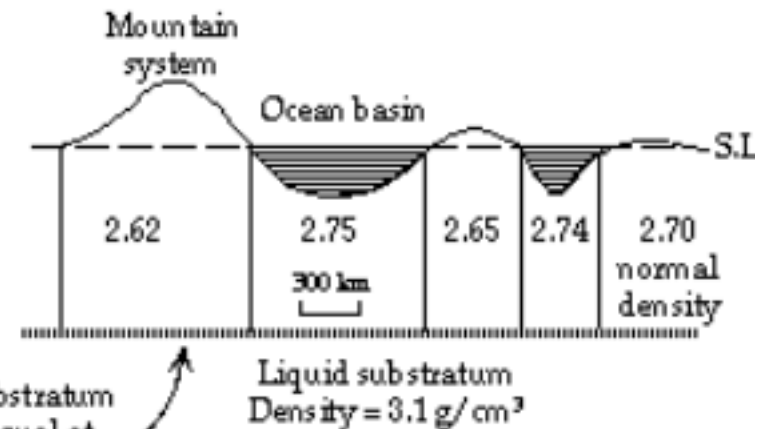




AIRY'S MODEL



PRATT'S MODEL

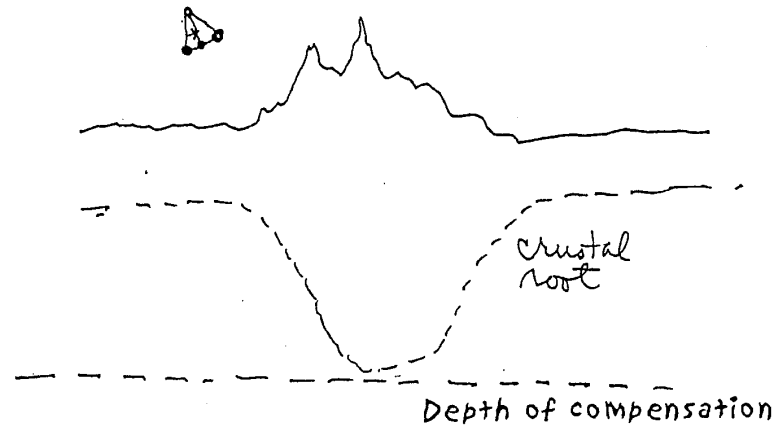


Isostasy

iso = equal
stasis = standing

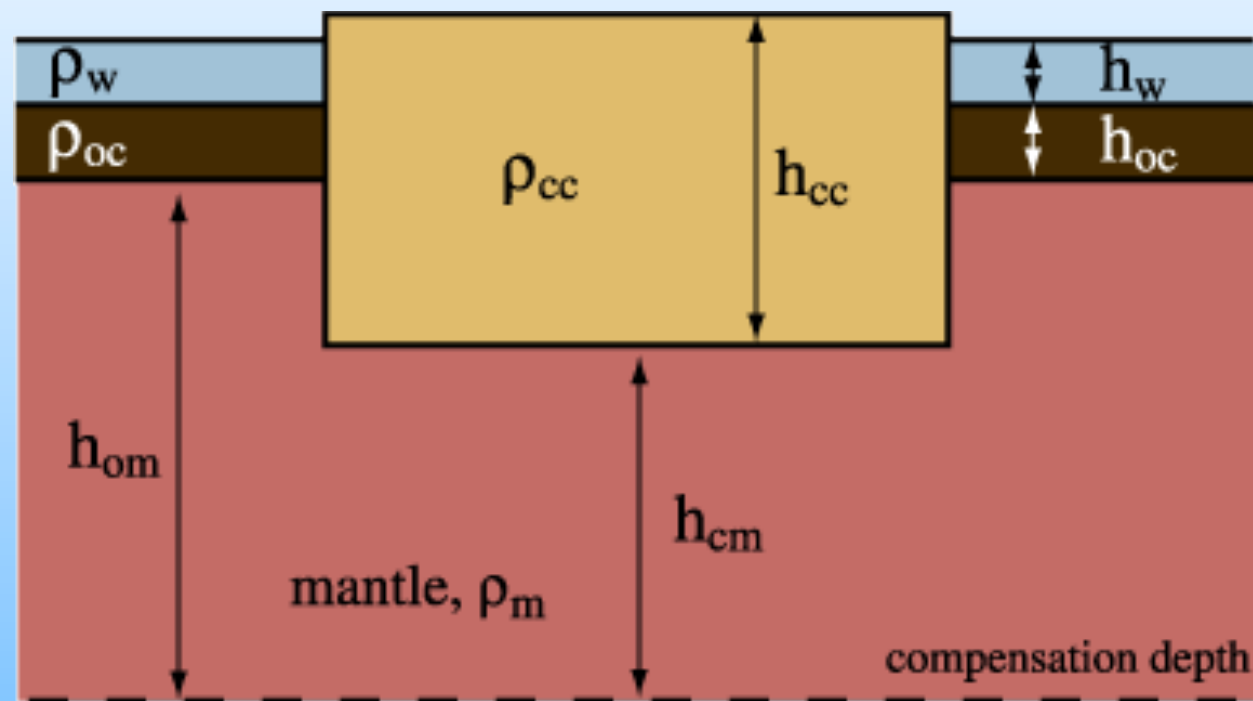
French surveyors in Peru in the 1740's noticed that the Andes deflected a plumb bob much less than anticipated.

Also surveys in India in the 1800s showed compensation of the Himalayas.



Principle: Beneath a certain depth, known as the depth of compensation, the pressures generated by all overlying materials are everywhere equal.

Think of an iceberg.



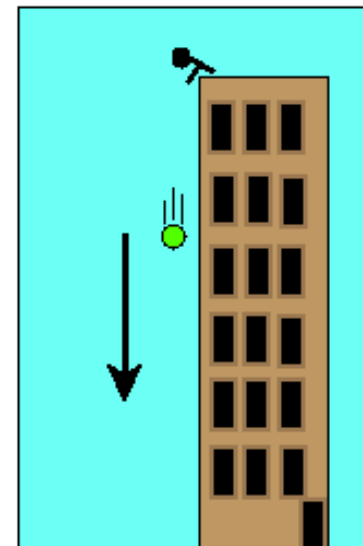
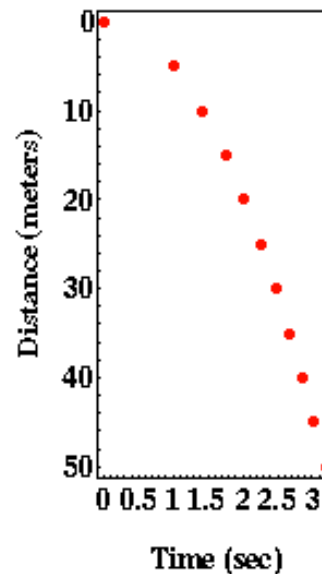
$$\rho_w h_w + \rho_{oc} h_{oc} + \rho_m h_{om} = \rho_{cc} h_{cc} + \rho_m h_{cm}$$

Falling Body Measurements



The gravitational acceleration can be measured directly by dropping an object and measuring its time rate of change of speed (acceleration) as it falls. By tradition, this is the method we have commonly ascribed to Galileo Galilei. In this experiment, Galileo is supposed to have dropped objects of varying mass from the leaning tower of Pisa and found that the gravitational acceleration an object undergoes is independent of its mass. He is also said to have estimated the value of the gravitational acceleration in this experiment. While it is true that Galileo did make these observations, he didn't use a falling body experiment to do them. Rather, he used measurements based on pendulums.

It is easy to show that the distance a body falls is proportional to the time it has fallen squared. The proportionality constant is the gravitational acceleration, g . Therefore, by measuring distances and times as a body falls, it is possible to estimate the gravitational acceleration. To measure changes in the gravitational acceleration down to 1 part in 40 million using an instrument of reasonable size (say one that allows the object to drop 1 meter), we need to be able to measure changes in distance down to 1 part in 10 million and changes in time down to 1 part in 100 million!! As you can imagine, it is difficult to make measurements with this level of accuracy.



$$\text{Distance} = \frac{1}{2} g \text{ time}^2$$

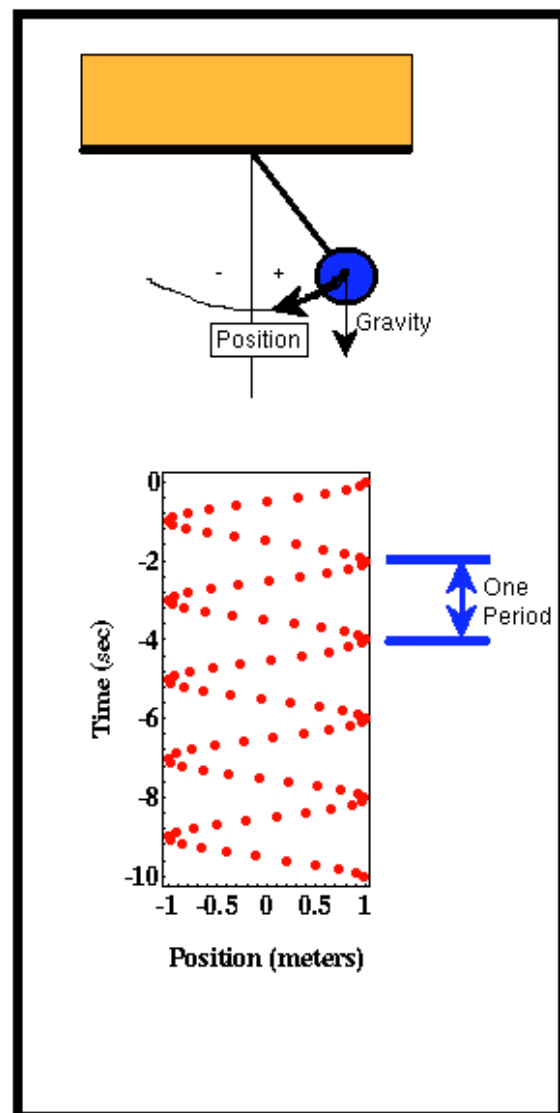
$$g = \frac{2 \text{ Distance}}{\text{time}^2}$$

It is, however, possible to design an instrument capable of measuring accurate distances and times and computing the absolute gravity down to 1 microgal (0.001 mgals; this is a measurement accuracy of almost 1 part in 1 billion!!). [Micro-g Solutions](#) is one manufacturer of this type of instrument, known as an *Absolute Gravimeter*. Unlike the instruments described next, this class of instruments is the only field instrument designed to measure *absolute gravity*. That is, this instrument measures the size of the vertical component of gravitational acceleration at a given point. As described [previously](#), the instruments more commonly used in exploration surveys are capable of measuring only the change in gravitational acceleration from point to point, not the absolute value of gravity at any one point.

Although absolute gravimeters are more expensive than the traditional, relative gravimeters and require a longer station occupation time (1/2 day to 1 day per station), the increased precision offered by them and the fact that the [looping strategies](#) described later are not required to remove instrument drift or tidal variations may outweigh the extra expense in operating them. This is particularly true when survey designs require large station spacings or for experiments needing the continuous monitoring of the gravitational acceleration at a single location. As an example of this latter application, it is possible to observe as little as 3 mm of crustal uplift over time by monitoring the change in gravitational acceleration at a single location with one of these instruments.

Pendulum Measurements

Another method by which we can measure the acceleration due to gravity is to observe the oscillation of a pendulum, such as that found on a grandfather clock. Contrary to popular belief, [Galileo Galilei](#) made his famous gravity observations using a pendulum, not by dropping objects from the Leaning Tower of Pisa.



If we were to construct a simple pendulum by hanging a mass from a rod and then displace the mass from vertical, the pendulum would begin to oscillate about the vertical in a regular fashion. The relevant parameter that describes this oscillation is known as the period* of oscillation.

*The period of oscillation is the time required for the pendulum to complete one cycle in its motion. This can be determined by measuring the time required for the pendulum to reoccupy a given position. In the example shown to the left, the period of oscillation of the pendulum is approximately two seconds.

The reason that the pendulum oscillates about the vertical is that if the pendulum is displaced, the [force of gravity](#) pulls down on the pendulum. The pendulum begins to move downward. When the pendulum reaches vertical it can't stop instantaneously. The pendulum continues past the vertical and upward in the opposite direction. The force of gravity slows it down until it eventually stops and begins to fall again. If there is no friction where the pendulum is attached to the ceiling and there is no wind resistance to the motion of the pendulum, this would continue forever.

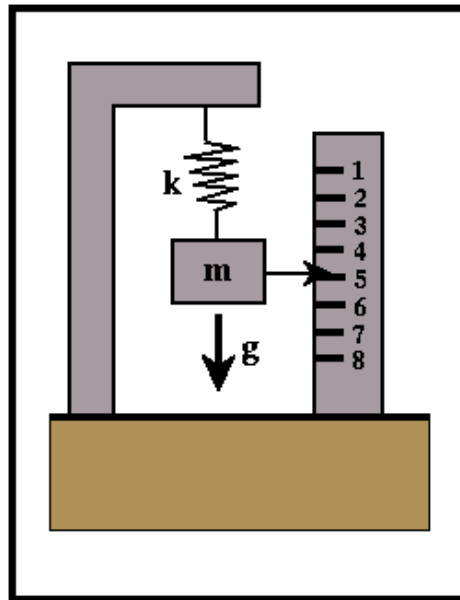
Because it is the force of gravity that produces the oscillation, one might expect the period of oscillation to differ for differing values of gravity. In particular, if the force of gravity is small, there is less force pulling the pendulum downward, the pendulum moves more slowly toward vertical, and the observed period of oscillation becomes longer. Thus, by measuring the period of oscillation of a pendulum, we can estimate the gravitational force or acceleration.

It can be shown that the period of oscillation of the pendulum, T , is proportional to one over the square root of the gravitational

$$T = 2\pi\sqrt{\frac{k}{g}}$$

acceleration, g . The constant of proportionality, k , depends on the physical characteristics of the pendulum such as its length and the distribution of mass about the pendulum's pivot point.

Mass and Spring Measurements

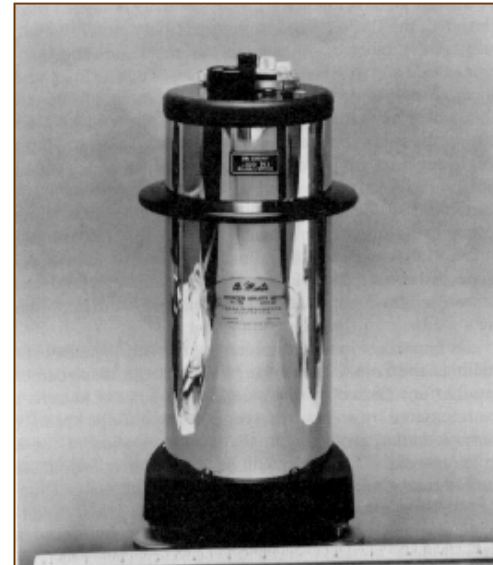


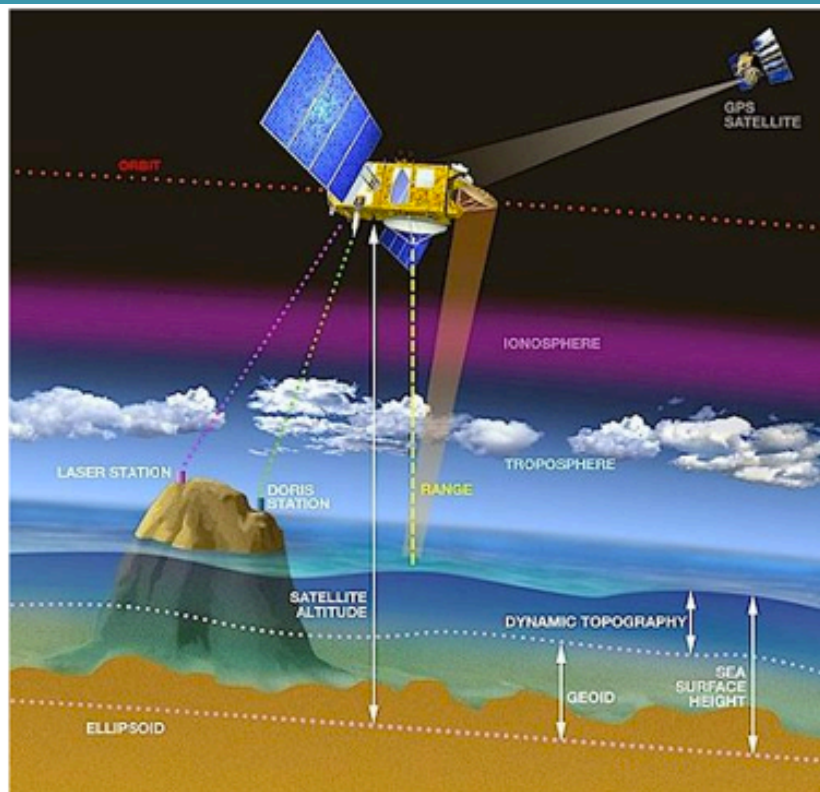
The most common type of gravimeter* used in exploration surveys is based on a simple mass-spring system. If we hang a mass on a spring, the force of gravity will stretch the spring by an amount that is proportional to the gravitational force. It can be shown that the proportionality between the stretch of the spring and the gravitational acceleration is the magnitude of the mass hung on the spring divided by a constant, k , which describes the stiffness of the spring. The larger k is, the stiffer the spring is, and the less the spring will stretch for a given value of gravitational acceleration.

Like pendulum measurements, we can not determine k accurately enough to estimate the absolute value of the gravitational acceleration to 1 part in 40 million. We can, however, estimate variations in the gravitational acceleration from place to place to within this precision. To be able to do this, however, a sophisticated mass-spring system is used that places the mass on a beam and employs a special type of spring known as a zero-length spring.

$$x = \frac{m g}{k}$$

Instruments of this type are produced by several manufacturers; *LaCoste and Romberg*, *Texas Instruments (Worden Gravity Meter)*, and *Scintrex*. Modern gravimeters are capable of measuring changes in the Earth's gravitational acceleration down to 1 part in 100 million. This translates to a precision of about 0.01 mgal. Such a precision can be obtained only under optimal conditions when the recommended field procedures are carefully followed.





The principle of altimetry
(Credits [CNES](#)/D. Ducros)

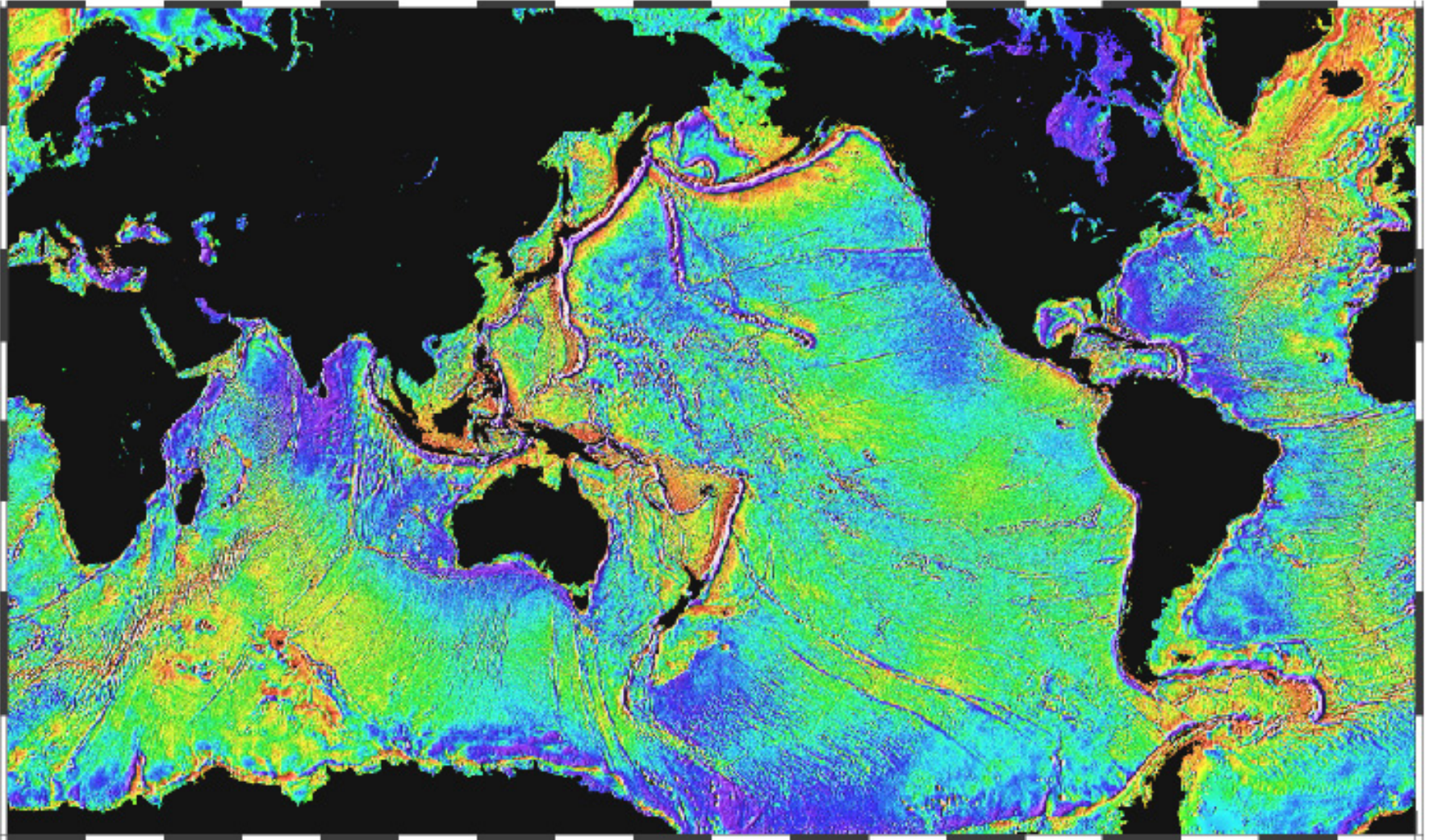
Satellite-to-surface distance: Range

Radar altimeters on board the satellite transmit signals at high frequencies (over 1,700 pulses per second) to Earth, and receive the [echo from the surface](#) (the "waveform"). (the 'waveform'). This is analysed to derive a precise measurement of the time taken to make the round trip between the satellite and the surface. This time measurement, scaled to the speed of light (the speed at which electromagnetic waves travel), yields a **range R** measurement (see [From radar pulse to altimetry measurements](#) for Further details).

However, as electromagnetic waves travel through the atmosphere, they can be decelerated by water vapour or ionisation. Once these phenomena have been corrected for, the final range can be estimated with great accuracy (see data [processing](#)).

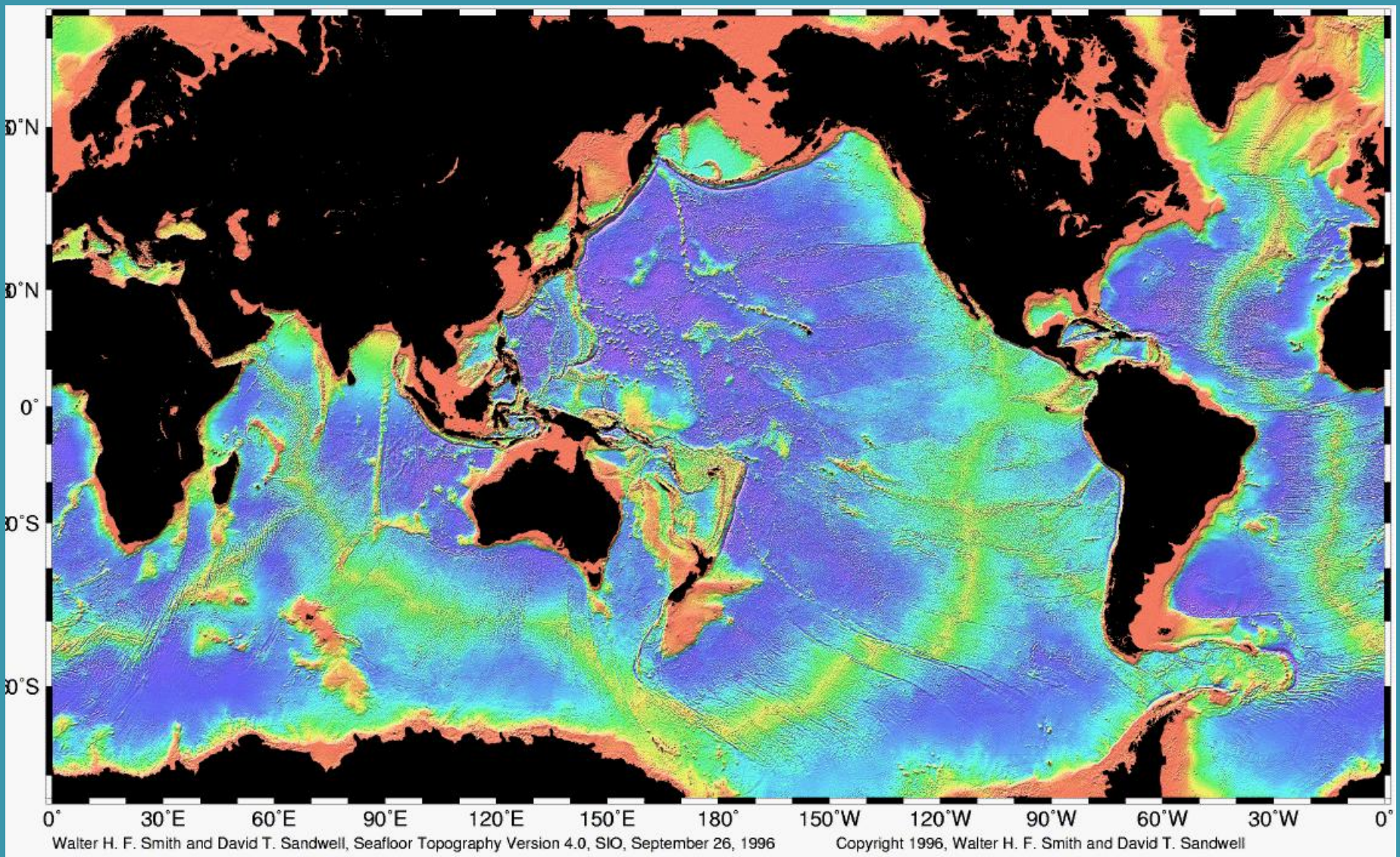
The ultimate aim is to measure surface height relative to a terrestrial reference frame. This requires independent measurements of the satellite's orbital trajectory, i.e. exact latitude, longitude and altitude coordinates.

A straight forward calculation converts sea surface height to gravity



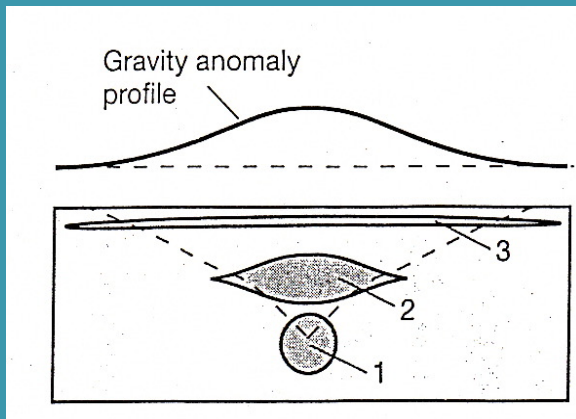
FREE-AIR GRAVITY, BASED ON GEOSAT AND ERS-1 SATELLITE RADAR ALTIMETRY (VERSION 7.2, D. SANDWELL & W. SMITH).

“Predicted” topography based on satellite radar altimetry measurements combined with shipboard bathymetry measurements



Use gravity to map
hidden structures

The interpretation is
ambiguous unless you
have an independent
way of mapping the
shape of the source



Three shapes, all
have same
anomaly

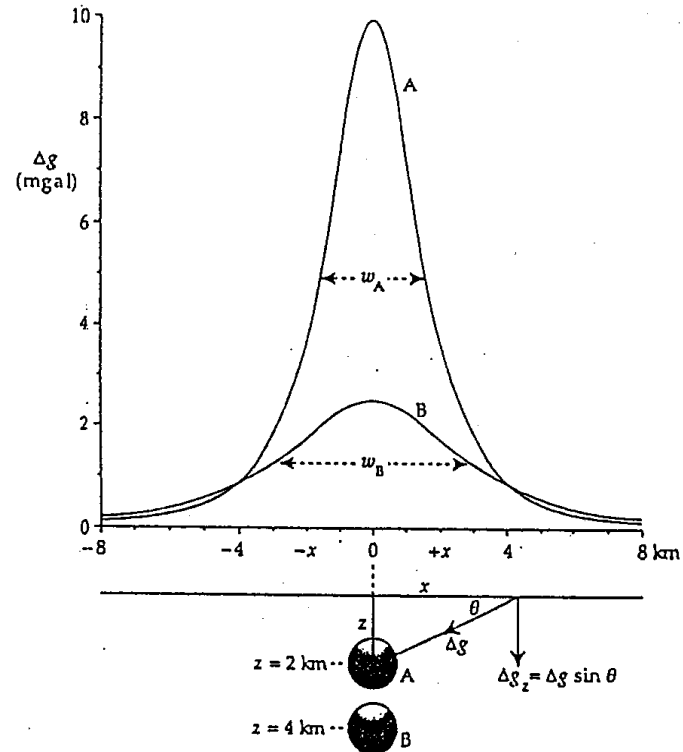


Fig. 2.47 Gravity anomalies for buried spheres with the same radius R and density contrast $\Delta \rho$ but with their centers at different depths z below the surface. The anomaly of the deeper sphere B is flatter and broader than the anomaly of the shallower sphere A.

Lowrie (1997)

gravity anomalies: compare observed to "expected" values
(g_o) (γ_o)



γ_o = gravity predicted on reference ellipsoid

We use gravity anomalies to study isostatic compensation

Free-air correction

Accounts for the $1/r^2$ decrease in gravity with distance from the center of the Earth. A given gravity measurement was made at an elevation h , not at sea level, recall:

$$g = \frac{GM_E}{R_E^2}$$

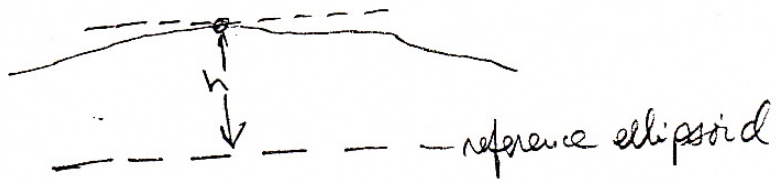
$$g = \frac{GM}{r^2} \quad \frac{dg}{dr} = -2gr = -.31 \text{ mgal/m}$$

$$\Delta g_{FA} = g_o + \underline{\underline{.31 h}} - \gamma_o \quad \text{free-air anomaly}$$

Free-air gravity anomalies

Over oceans, no free-air correction is needed

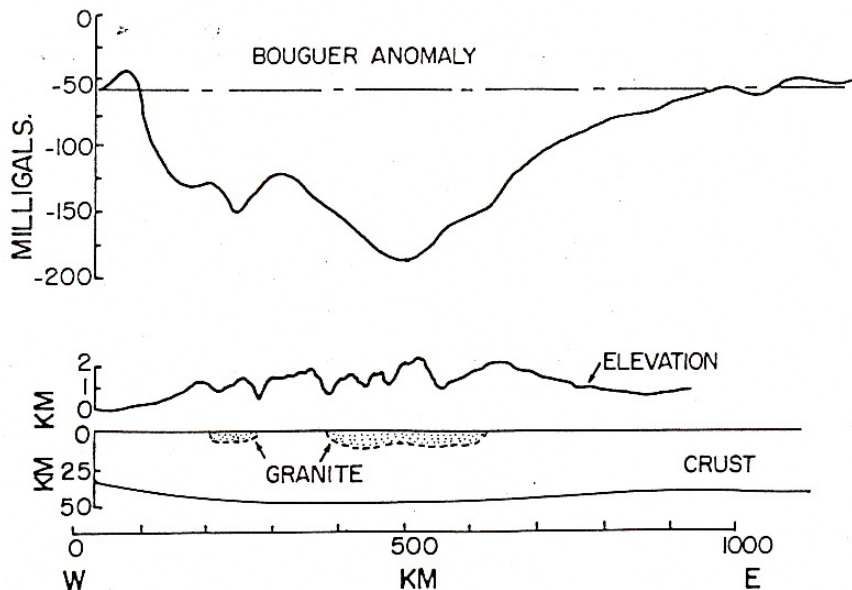
Correct for mass between observation point
and reference level



approximate effect with a slab of thickness h

$$\text{Bouguer correction} = 2\pi \Delta \rho G h = 0.04 \Delta \rho \text{ mgal/m} \\ \text{(or } \approx 0.1 \text{ mgal/m over continent)}$$

$$\text{Bouguer Anomaly} = \Delta g_B := \Delta g_{FA} - 2\pi \Delta \rho G h$$



Note: VE is not the
same above and
below sea level

Garland (1971)

Bouguer Anomaly

Pierre Bouguer was
one of the French
surveyors in the
Andes in the 1740's

The bouguer
anomaly is negative
over a mountain
range and reflects
the low-density
crustal root
associate with Airy
isostatic
compensation

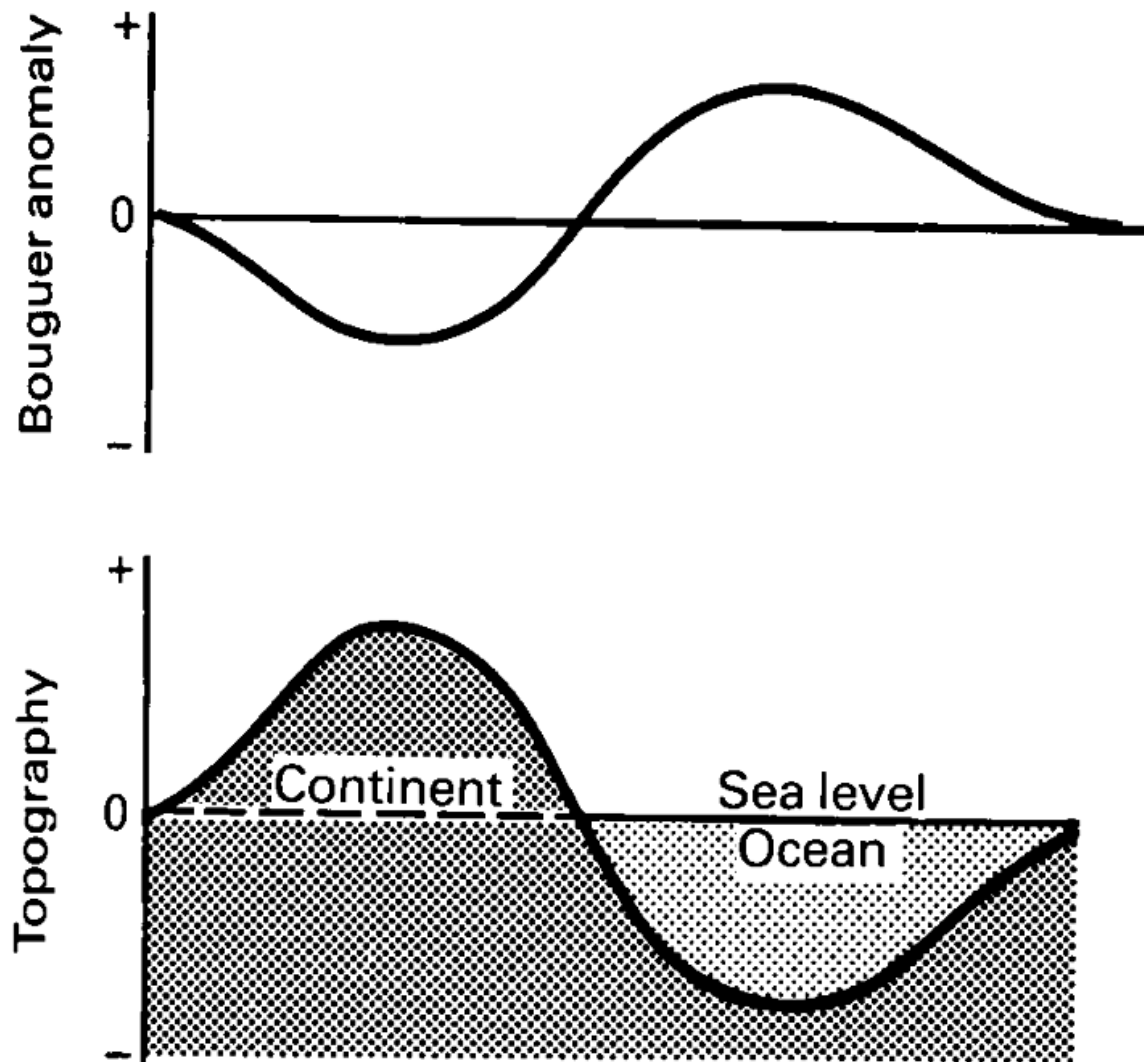
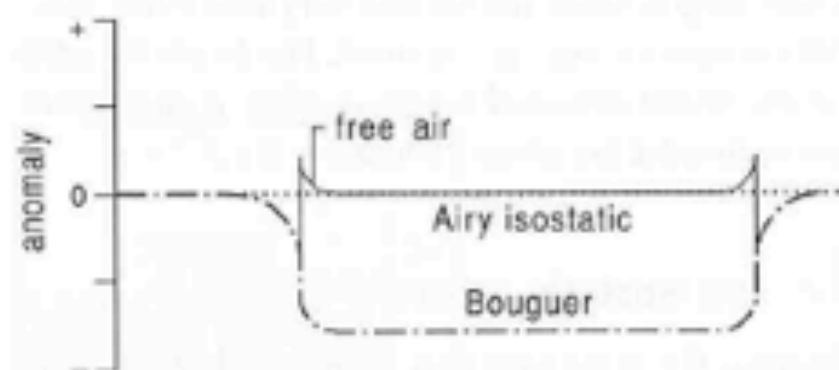


Fig. 2.28 Inverse correlation of Bouguer anomalies with topography indicating its isostatic compensation.

(a) full Airy compensation



(b) no compensation

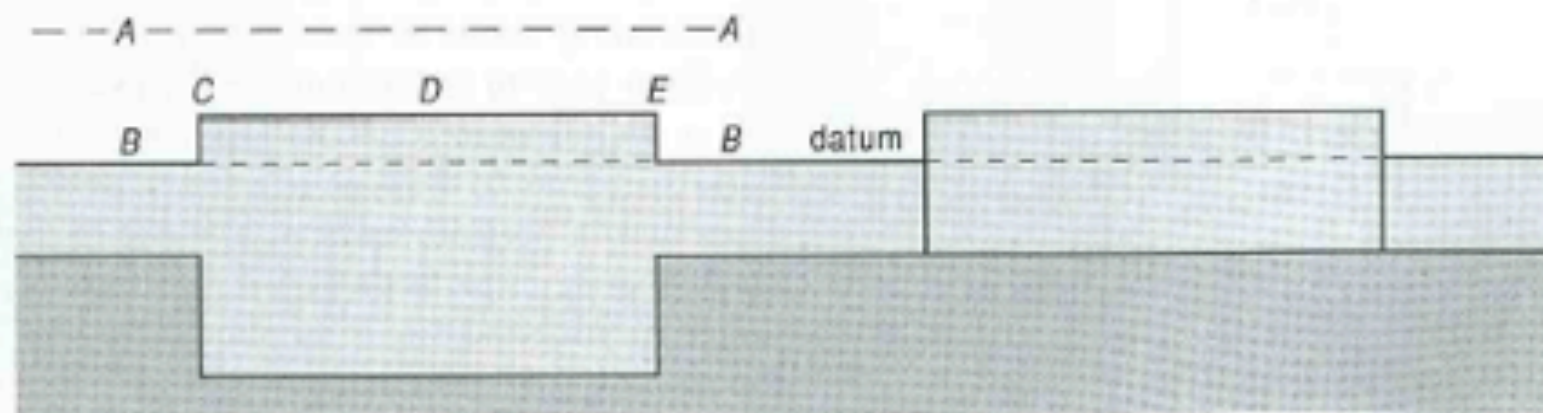
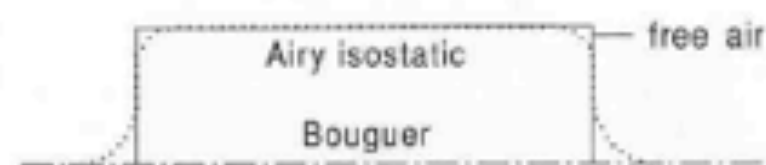
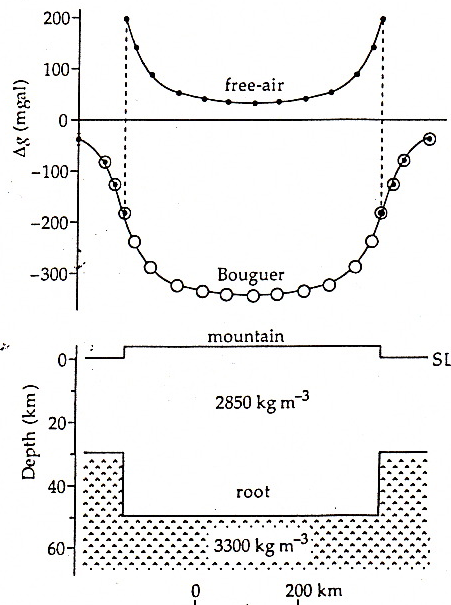
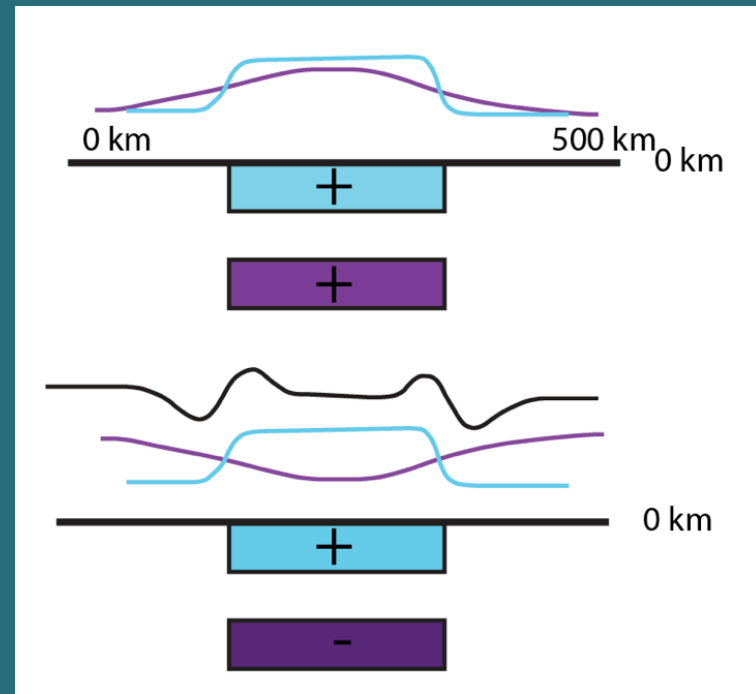


Figure 9.11 Anomalies with and without isostatic compensation.

Another way of thinking of it:
 With the Bouguer Anomaly you only
 see the effect of the root
 (think of example I showed earlier)



KK&V (2009)



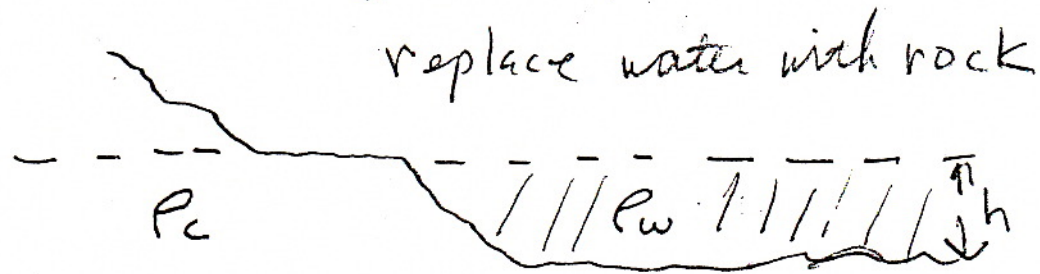
Simple model from earlier:

Let's move zero level to bottom of top body

Composite anomaly = free-air anomaly

Bouguer correction = effect of top body

Bouguer anom = composite – Bouguer correction

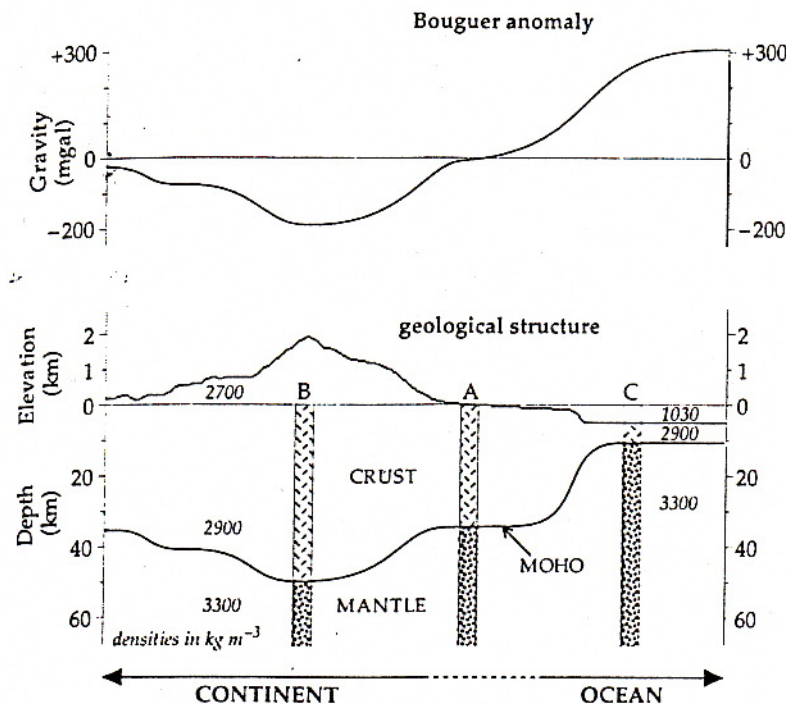


over water: $\Delta g_B = \Delta g_{FA} + 2\pi \Delta \rho G h$

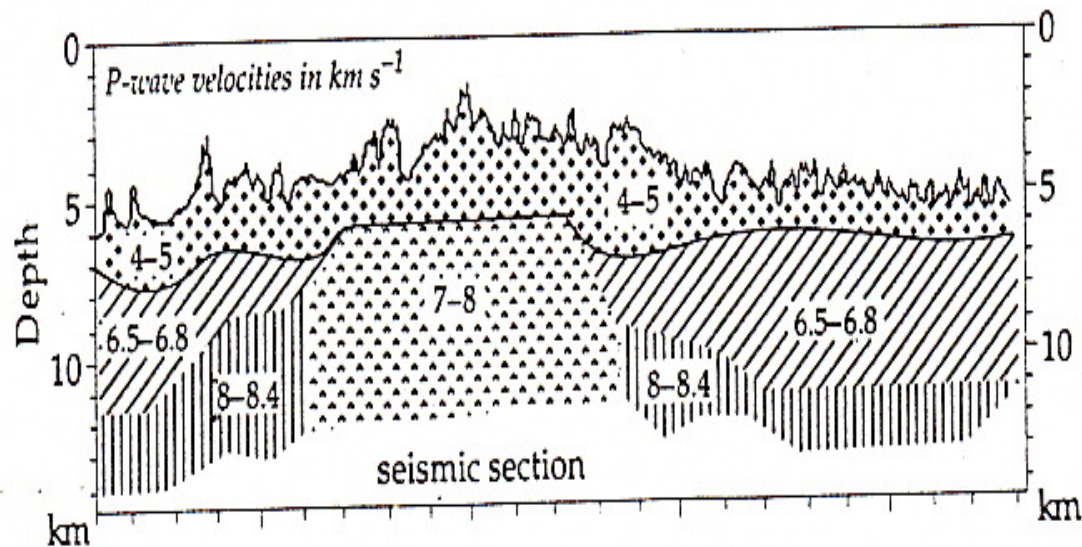
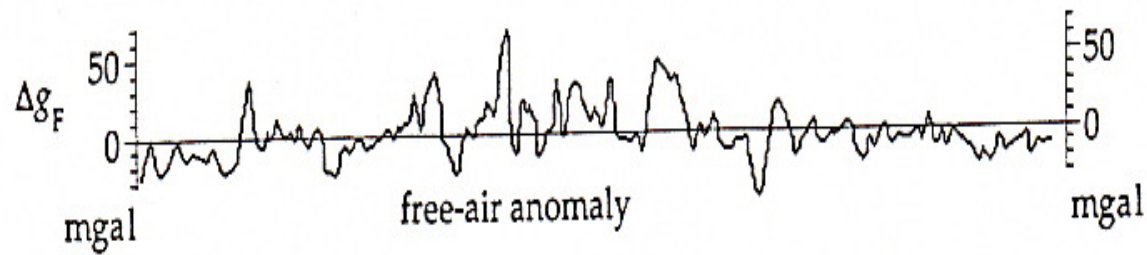
Bouguer anomaly over the oceans

Replace water with rock with same density as adjacent continent

Reveals “anti-root” (thin crust) beneath oceans



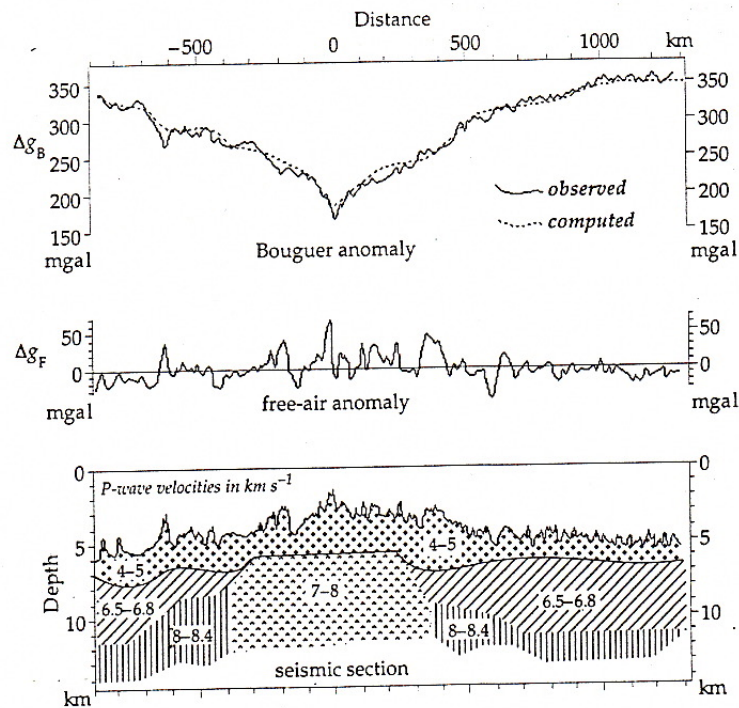
Lowrie (1997)



What does the Bouguer anomaly look like over a mid-ocean spreading center like the Mid-Atlantic ridge?

And what does that tell us?

Note: topography gets shallower but free-air anomaly stays near zero (on average)



Bouguer anomaly drops by 200 mgals over the mid-ocean ridge:

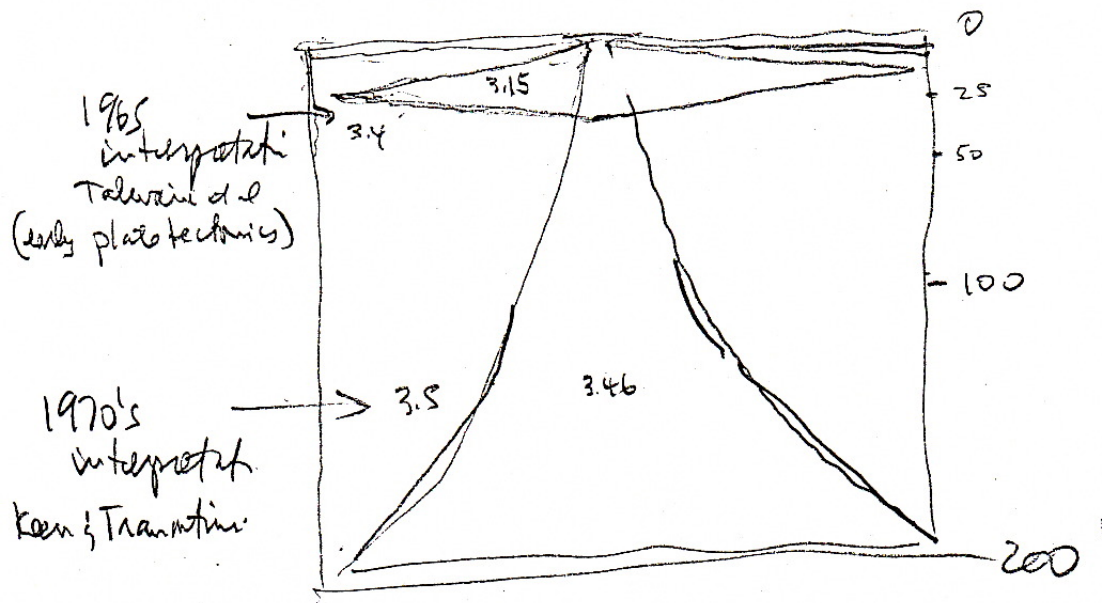
The mid-ocean ridges are generally isostatically compensated:

Free-air anomaly is roughly zero

Since the ridge axis is shallower than flanks, the Bouguer correction is smaller, so get a drop in the Bouguer anomaly centered on ridge axis

Reveals body of warm (less dense) asthenospheric material

Note ambiguity in size of body



Isostatic anomaly = Bouguer anomaly minus
the gravity anomaly of an ideal root

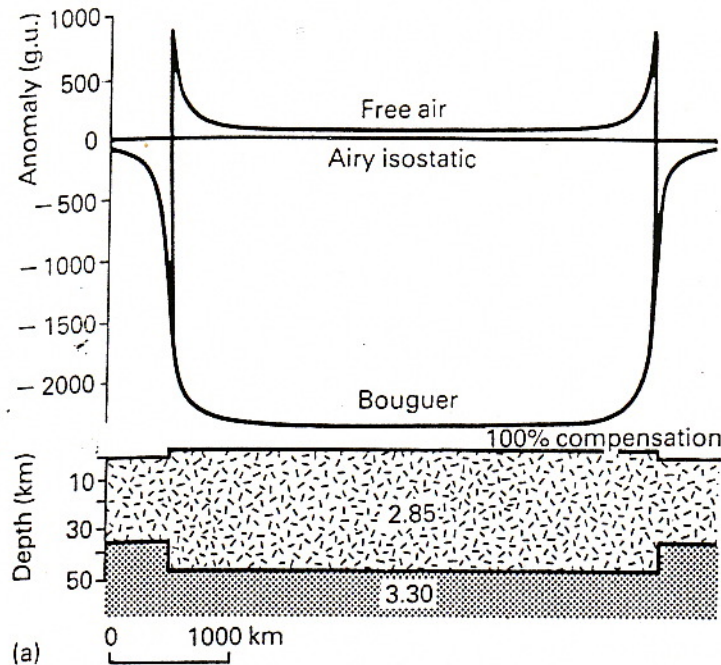
$$\Delta g_I = \Delta g_B - A_{\text{root}}$$

Isostatic anomaly (IA)

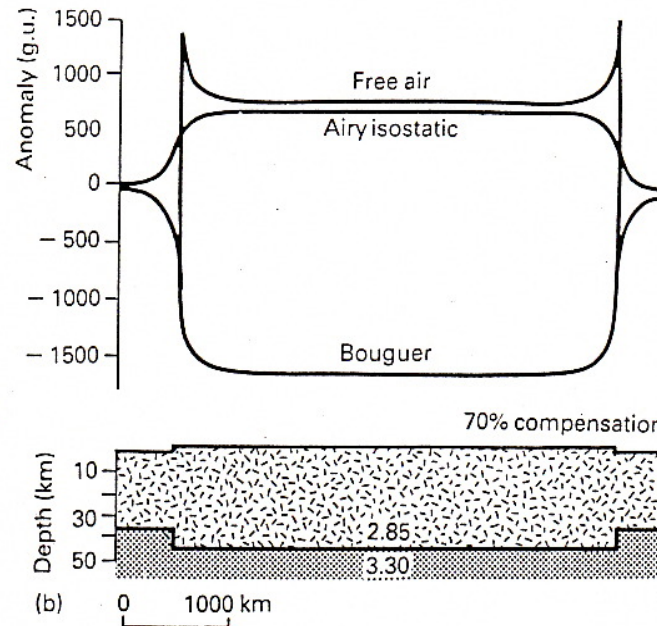
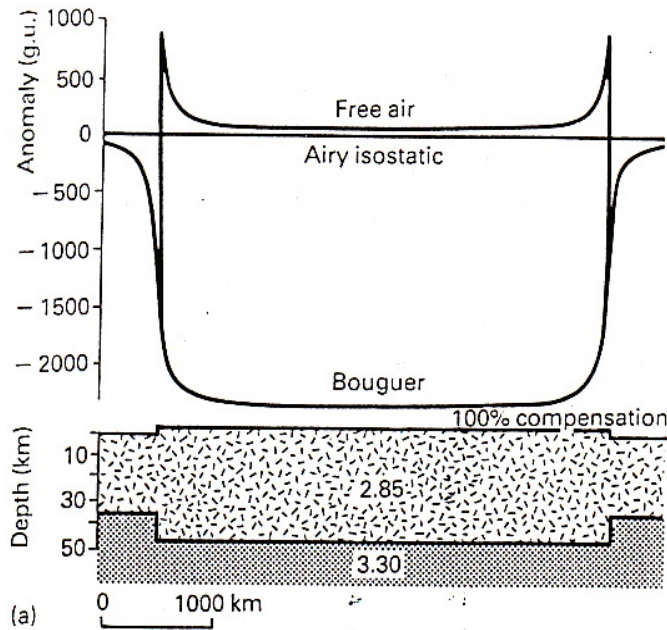
You calculate the
shape of root by
assuming Airy isostasy
and typical crustal
densities

Isostatic anomaly
should be zero if your
model is correct

Of course, you don't
really know the shape
of the root ...

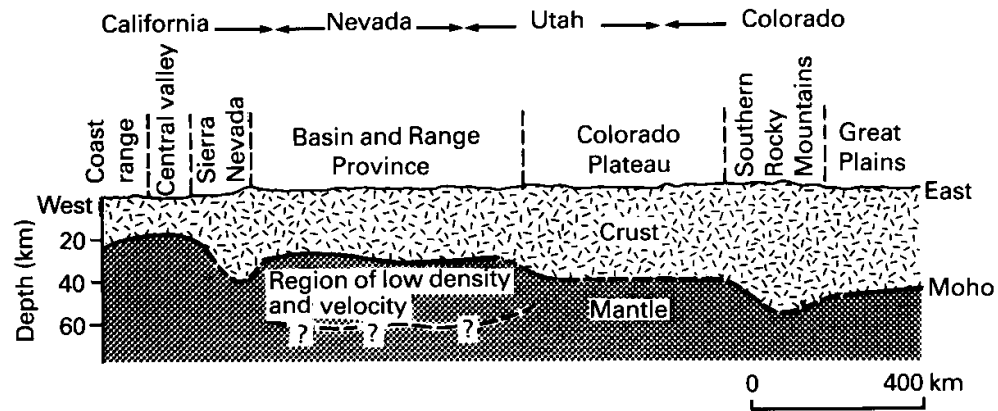
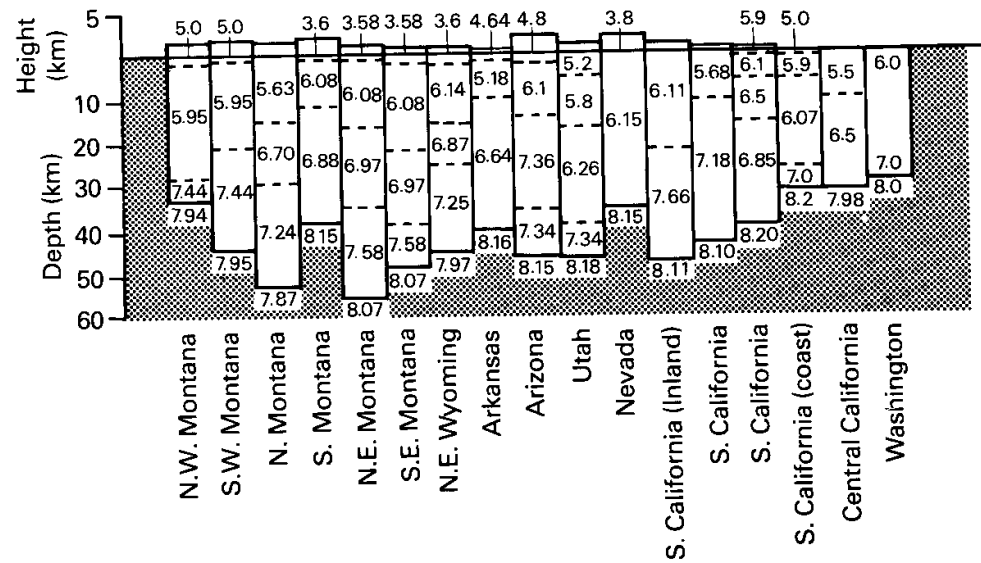
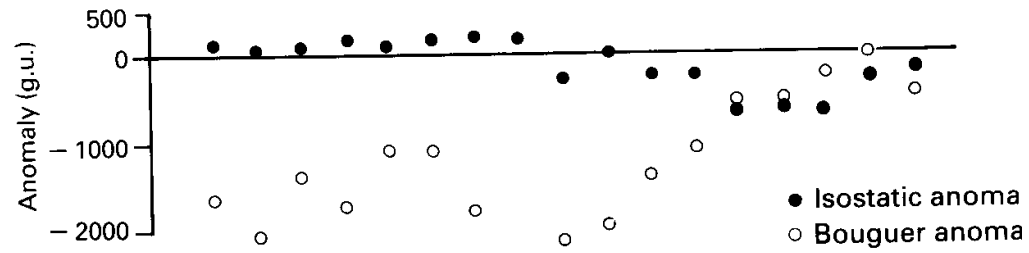


If the compensation is not 100%
 then the isostatic anomaly will not be 0
 which, of course, is the reason for
 calculating isostatic anomalies

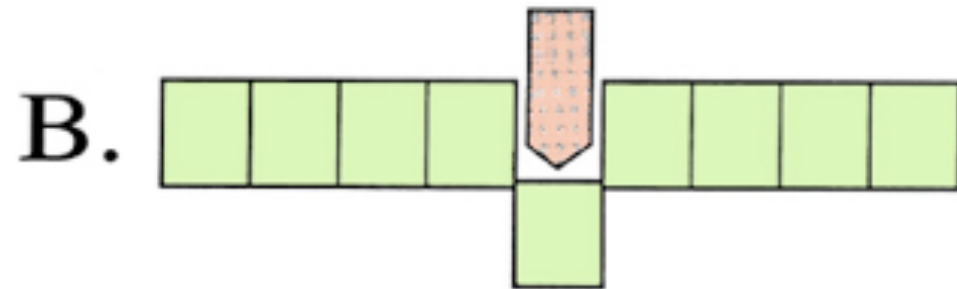
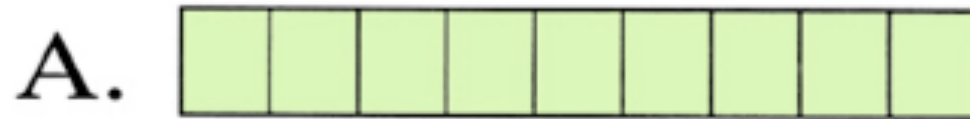


The root on the right is not as large as the one on the left but the “correction” was made for an “ideal” root so the isostatic anomaly is positive

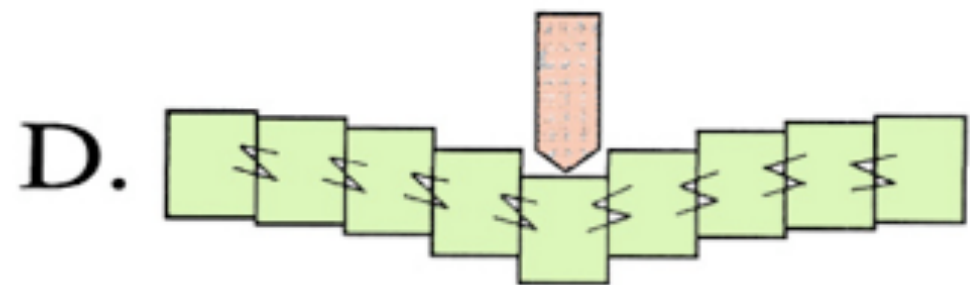
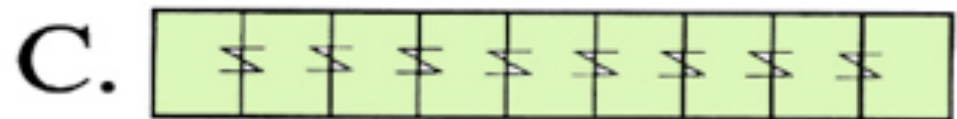
Why might the compensation not be 100% ?

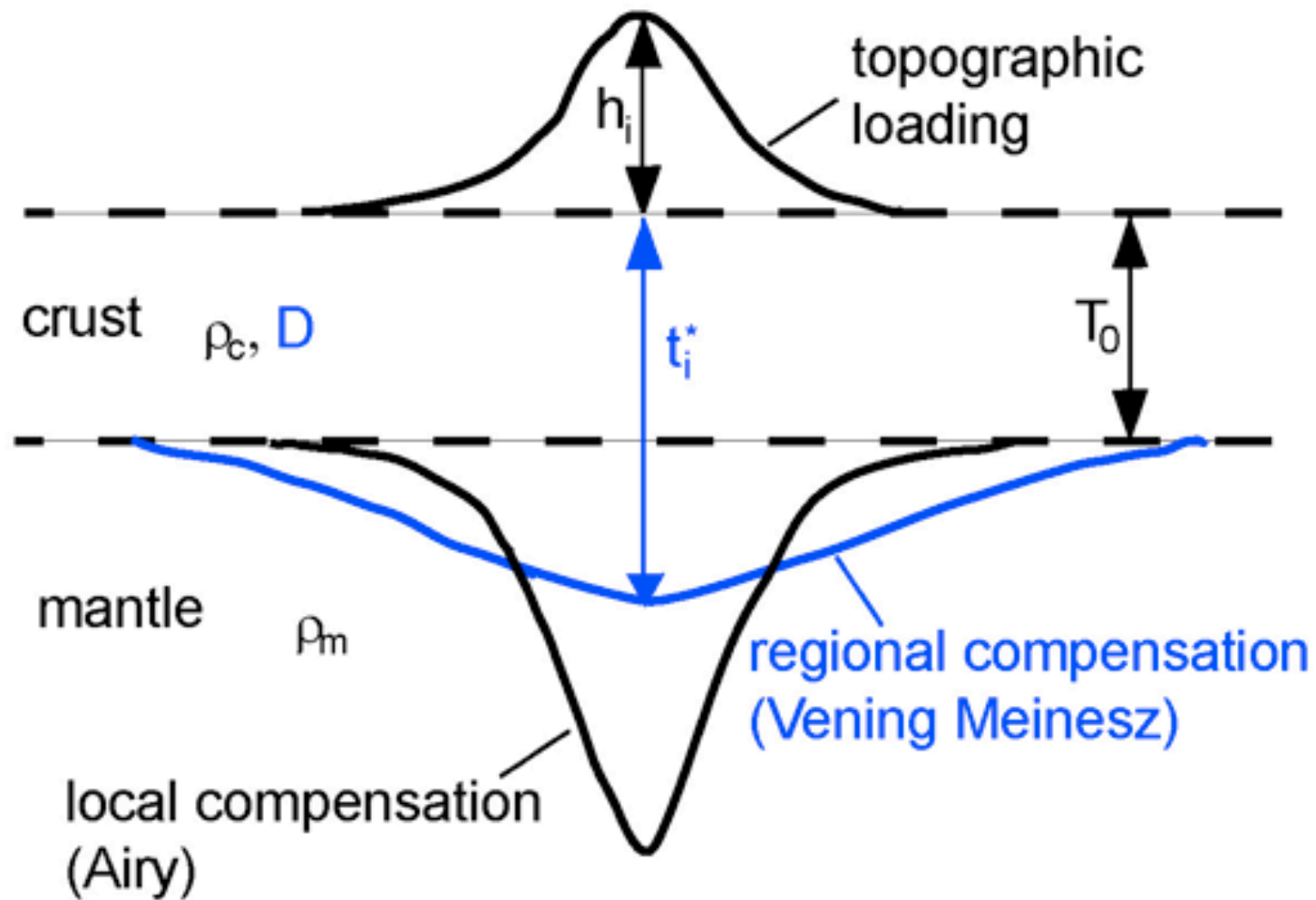


LOCAL ISOSTASY



FLEXURE (REGIONAL ISOSTASY)





Flexure

Earth's crust acts as a rigid plate of finite thickness and deforms "elastically" in response to a load.

Before plate tectonics, these concepts were used to analyze the effects of loading and unloading of glaciers, mountains, lakes, deltas.

The "plate" of plate tectonics

Flexure generates large gravity anomalies because the load is not compensated locally.

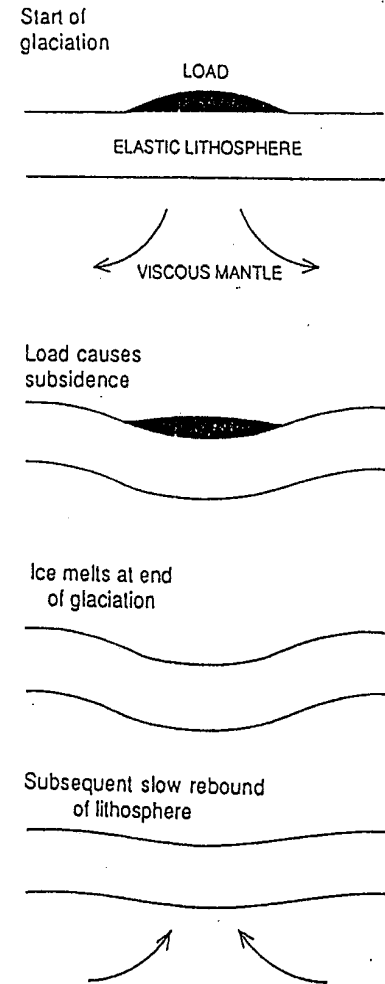


Figure 5.16. The deformation and uplift which occur as a result of loading and unloading of an elastic lithospheric plate overlying a viscous mantle.

Regional Isostatic Compensation

- The effect of the load can therefore be distributed over a wide area, depending on the flexural rigidity of the supporting material.
- A common model of regional isostatic compensation is that of an elastic plate that is bent by topographic and subsurface loads.
- The *flexural rigidity* of the plate (D) determines the degree to which the plate supports the load.
 - A thin, weak diving board bends greatly, especially near the diver.
 - A thicker board of the same material behaves more rigidly – the diver causes a smaller deflection.
 - The *flexural rigidity* (resistance to bending) thus depends on the *elastic thickness* of each board



From Lillie, 1999

Regional Isostatic Compensation

The deflection of a 2-D plate due to a linear load depressing the plates surface (assuming a fluid material below) is given by:

$$D(d^4w/d^4x) + (\rho_b - \rho_a)gw = q(x)$$

Where:

D = Flexural rigidity

w = vertical deflection of plate at x

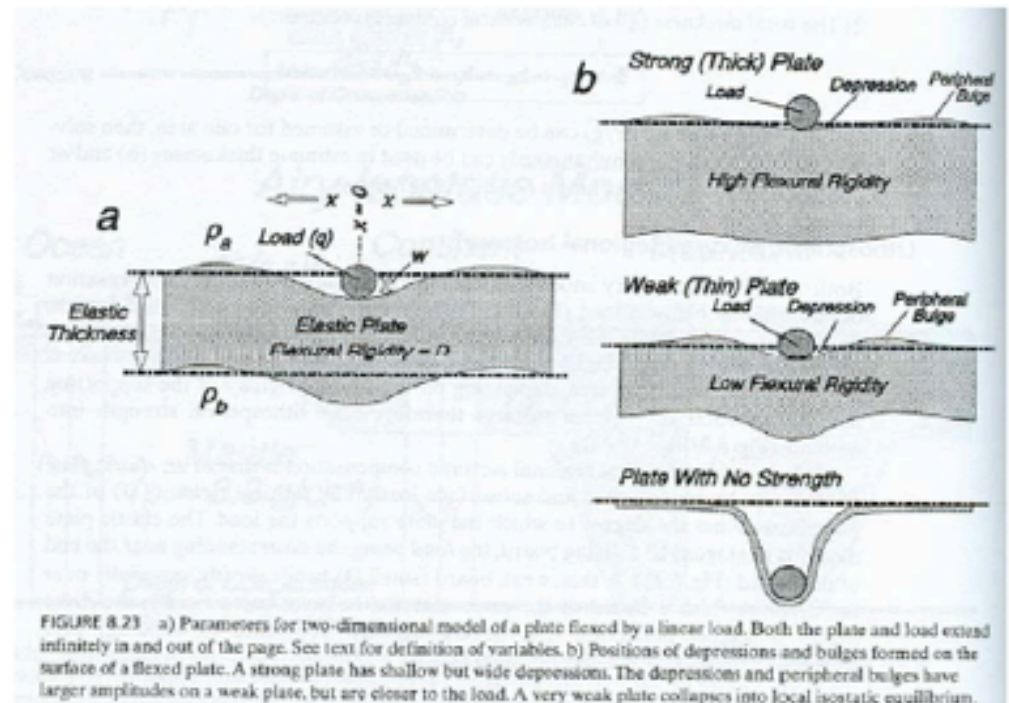
x = horizontal distance from load to a point on the plates surface

ρ_a = density of material above plate

ρ_b = density of material below plate

g = gravitational acceleration

$q(x)$ = load applied to the plate at x

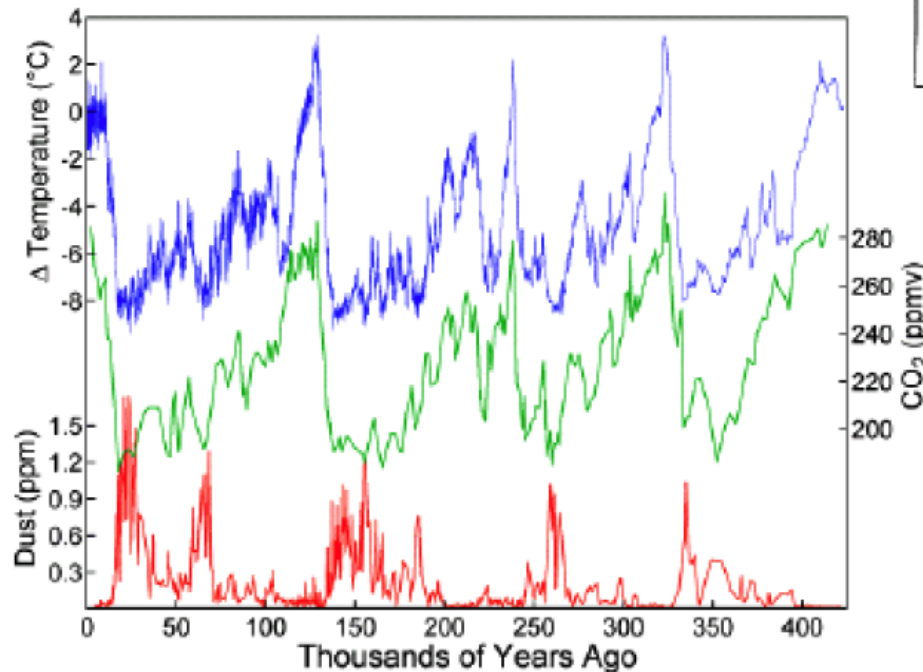


Strong plate: Long wavelength small amplitude (w) deflection.

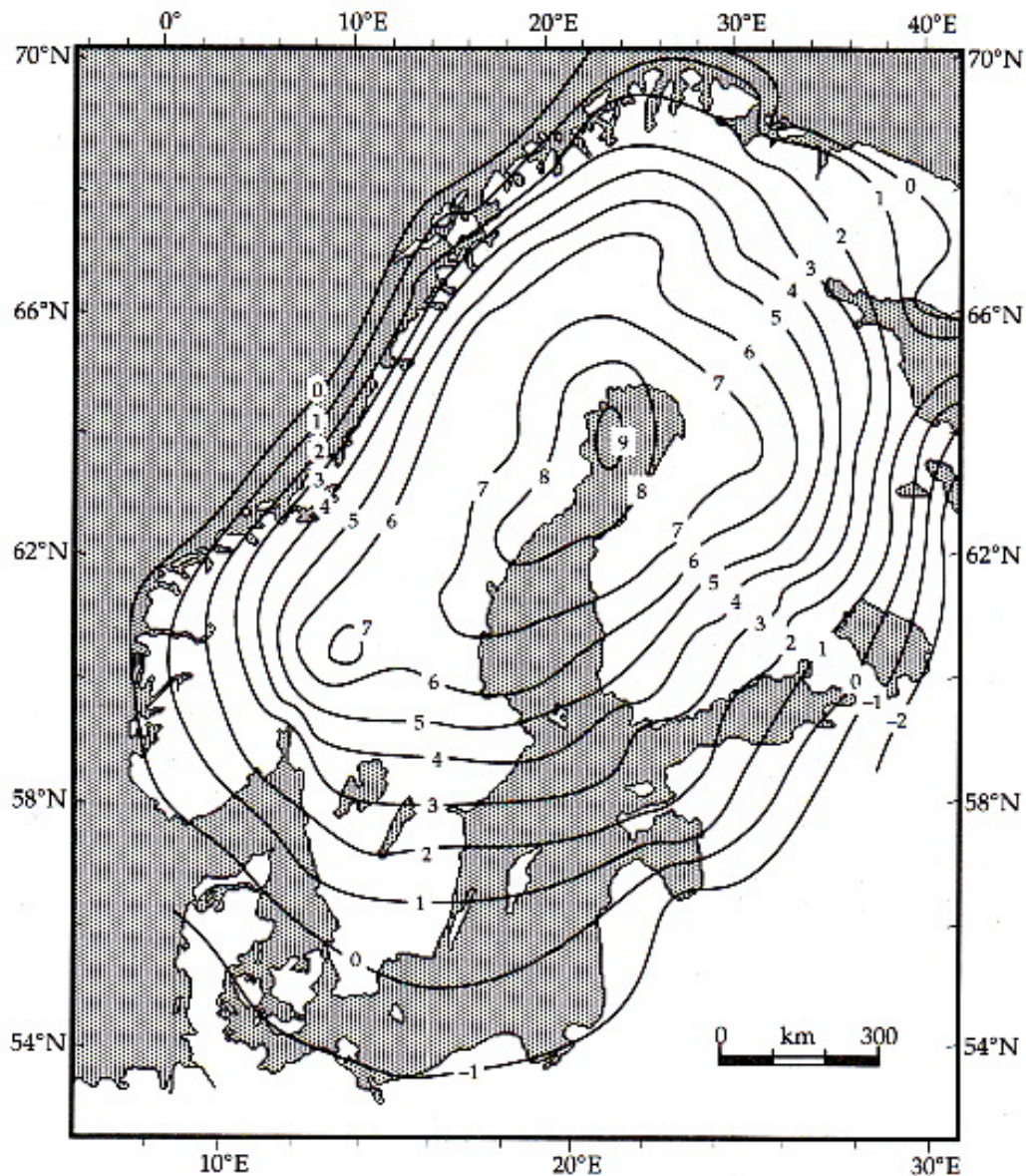
Weak plate: Short wavelength large amplitude deflection.

A special case of flexure and isostasy...

11,000 years ago, large parts of N. Europe and N. America were covered by ice sheets up to 3 km thick.



Ice sheets melted rapidly ~10,000 years ago as a result of global climate change.

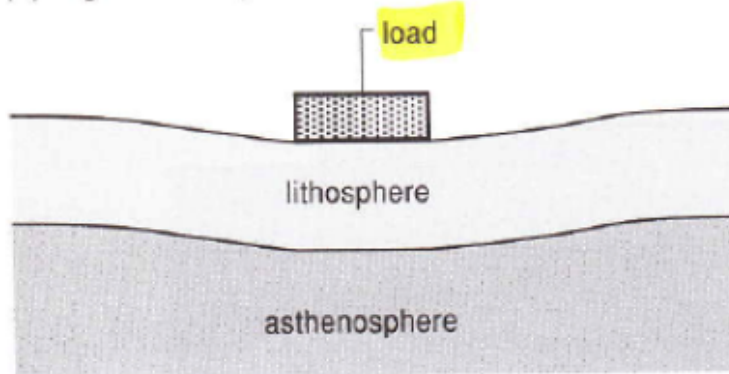


Glacial rebound: the best constraint on the viscosity of the upper mantle:

This figure shows the rate of vertical crustal movement relative to mean sea-level in mm/yr

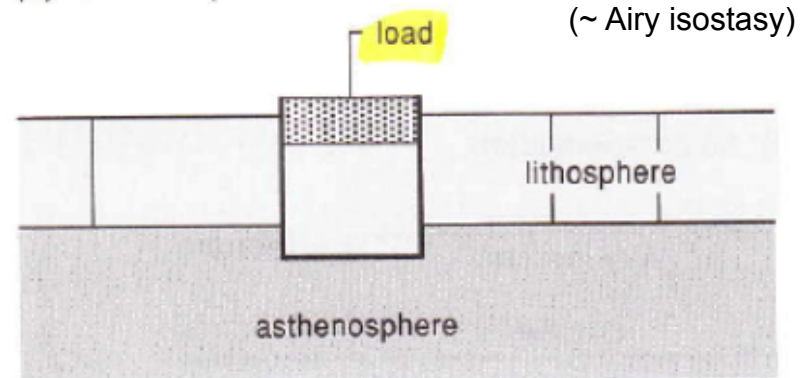
Local versus regional compensation

(b) regional compensation



The Lithosphere (crust) is strong enough to support the load (weight) of the mountain. But, the lithospheric strength is finite and the surface of the lithosphere is bowed down 'regionally' to support the load.

(a) local compensation



The Lithosphere (crust) is **NOT** strong enough to support the load (weight) of the mountain. In fact, in the limit, the lithosphere is broken on either side of the load and has a near zero strength. The load is supported by the hydrostatic pressure of the asthenosphere pushing on the bottom of the loaded block.

Regional compensation of Hawaiian Island Chain

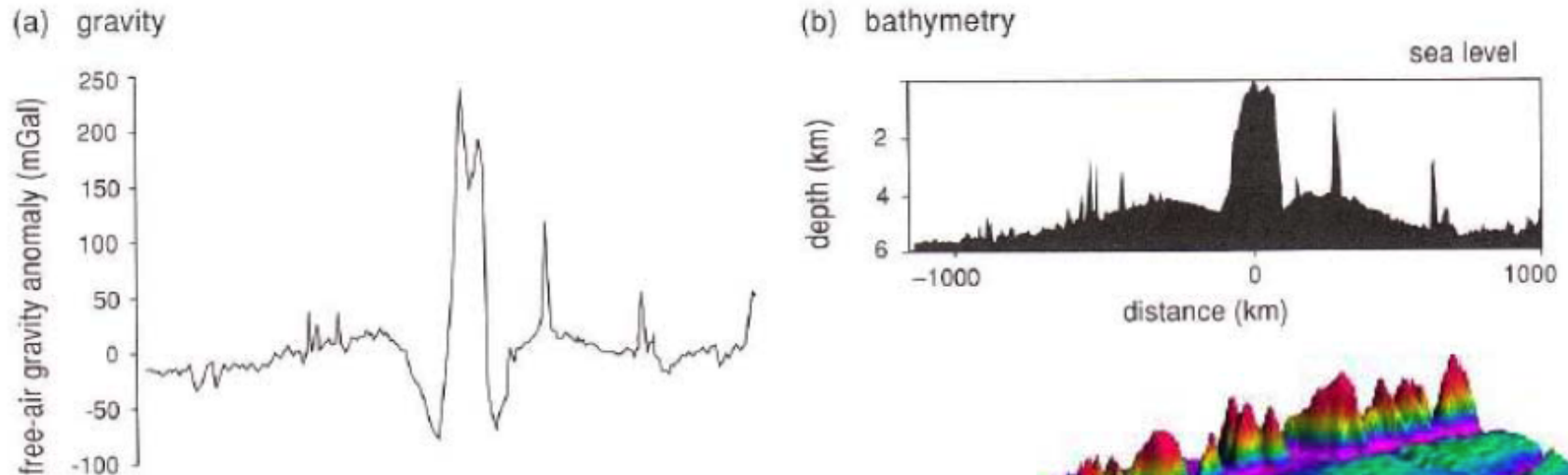
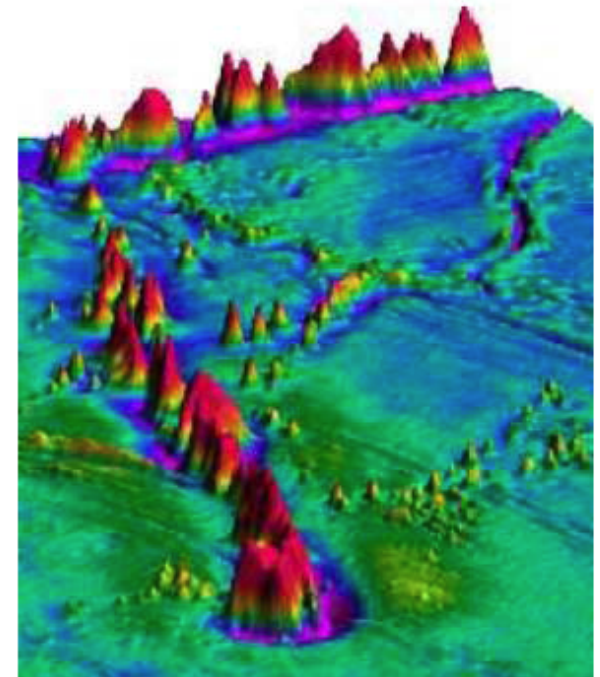
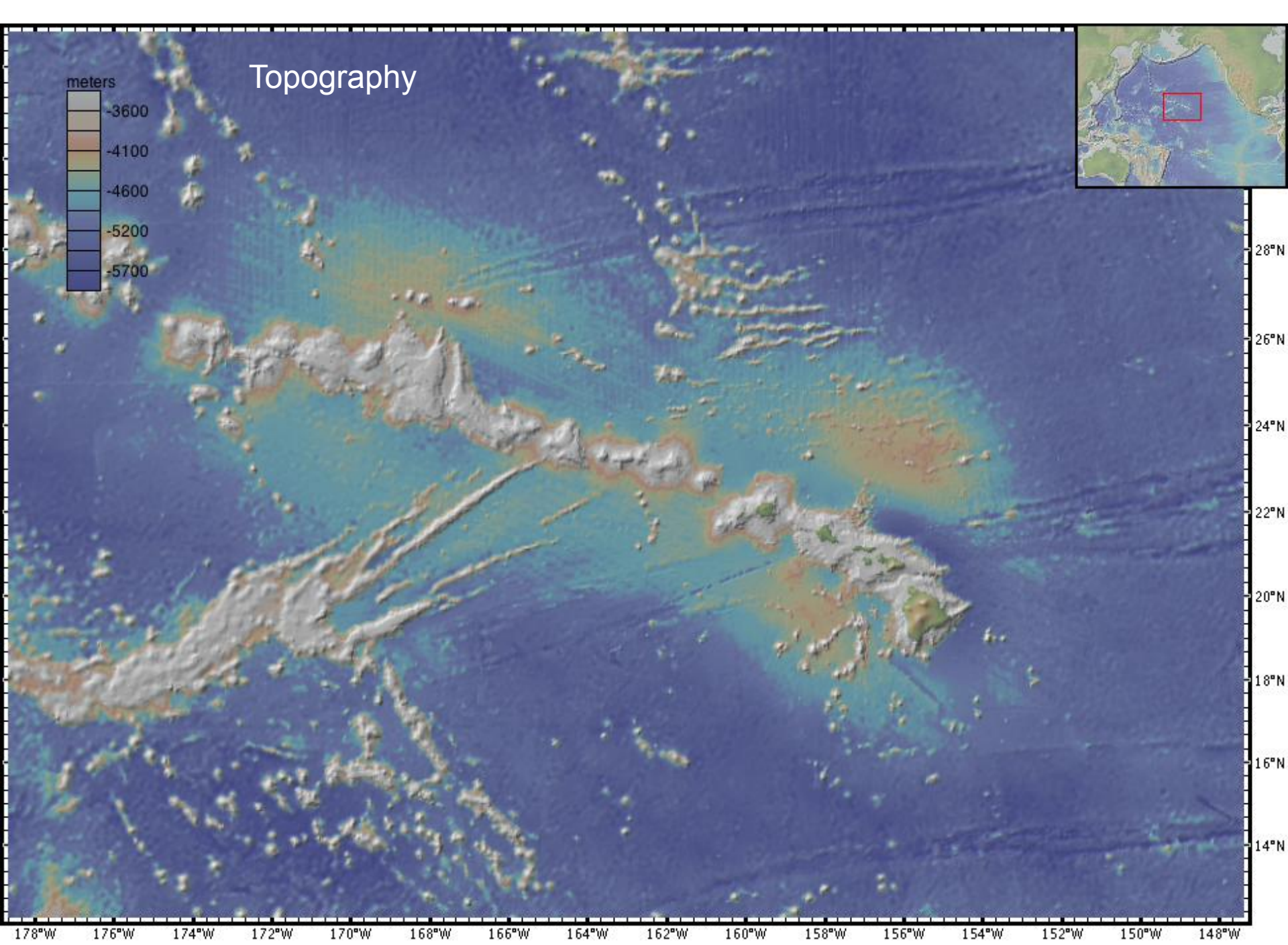


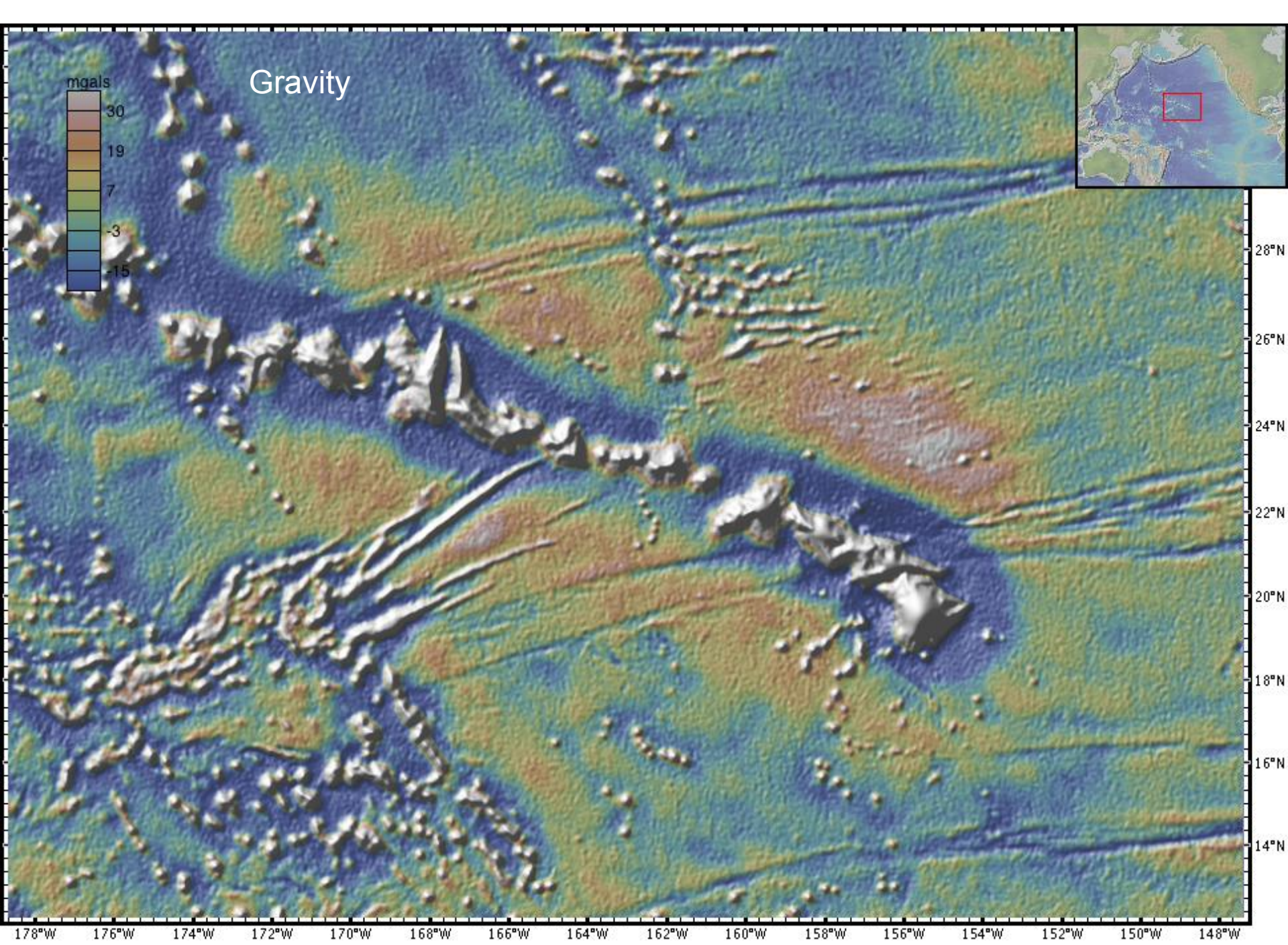
Figure 9.12 Bathymetry and gravity anomaly across Hawaiian Islands.

Note the strong correlation between the bathymetry and the free air gravity profile. This is because this topography is **NOT** isostatically compensated but is 'regional' compensated by the strength of the lithosphere. Also, note the downwarps to either side of the big island caused by the load that downflexes the lithosphere (see Fig. 9.10).

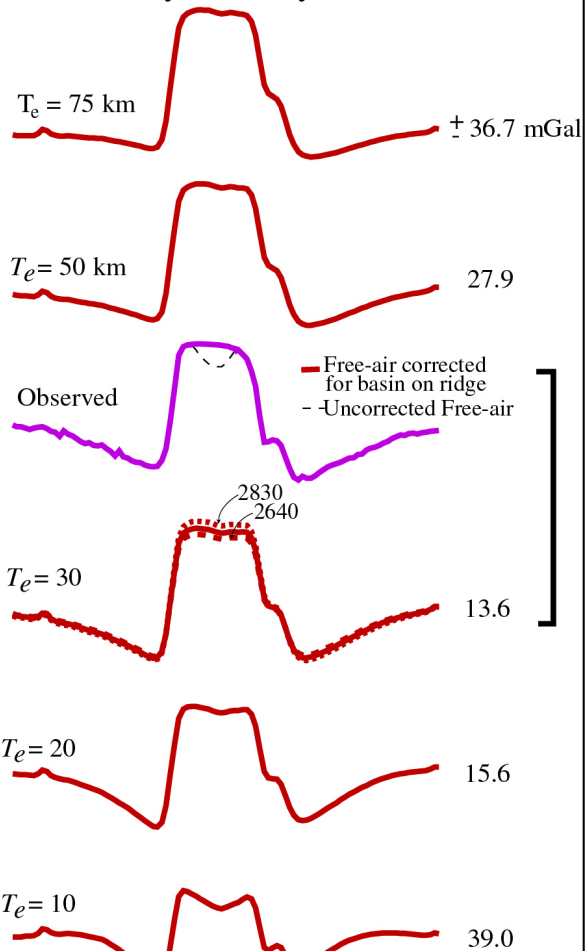


(Ken Dueker, U. Wyoming)

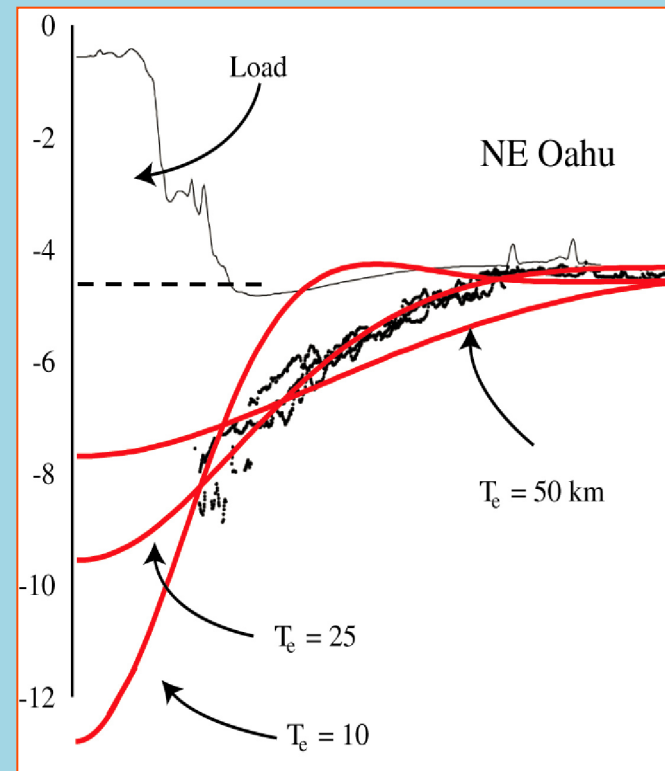
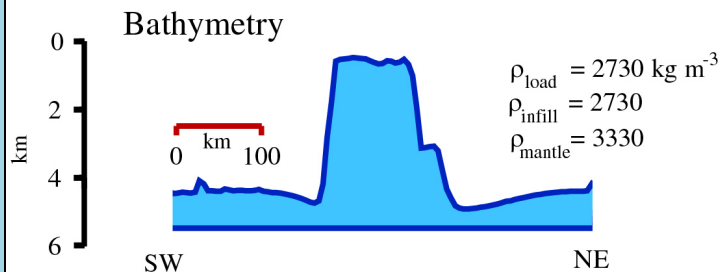




Free-air Gravity Anomaly

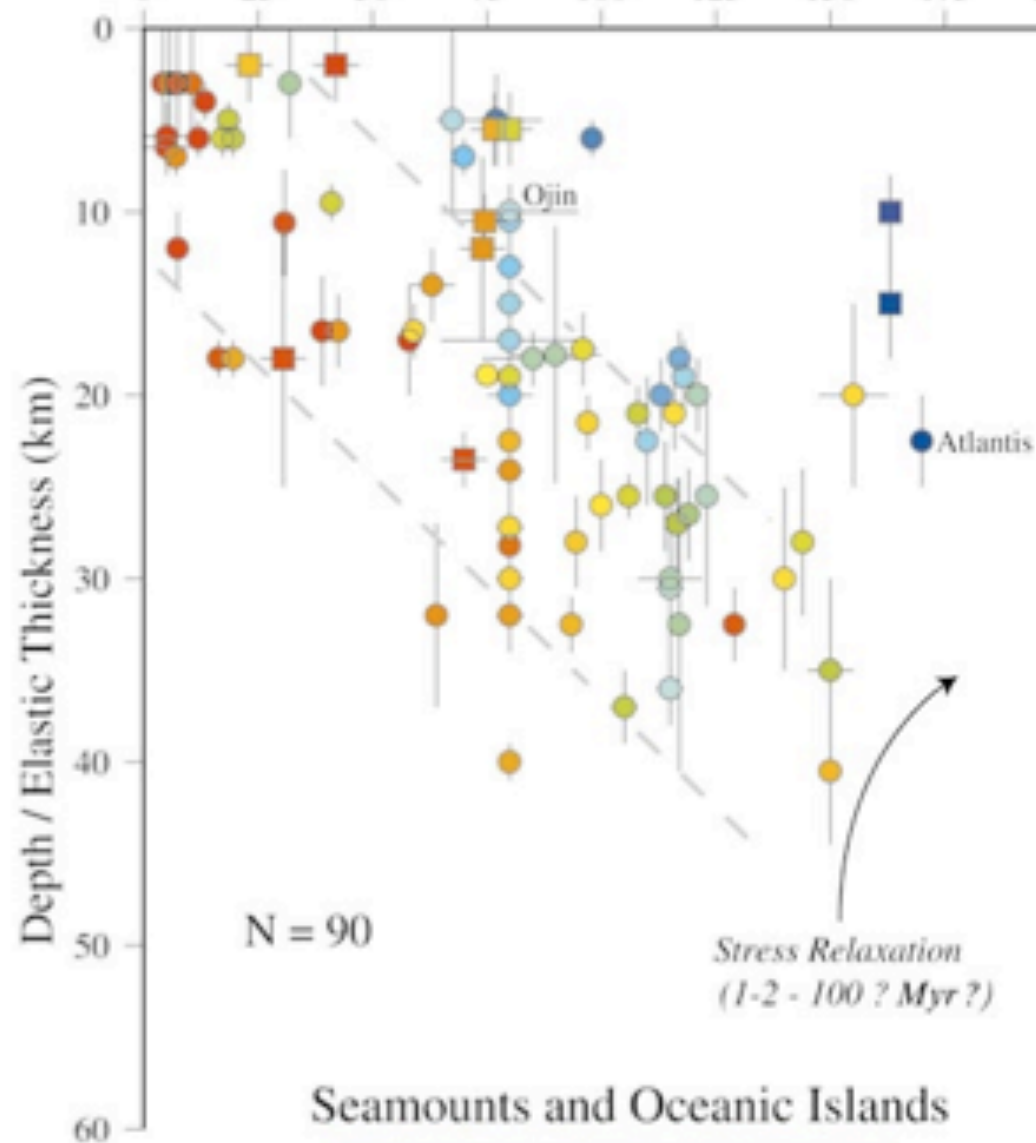


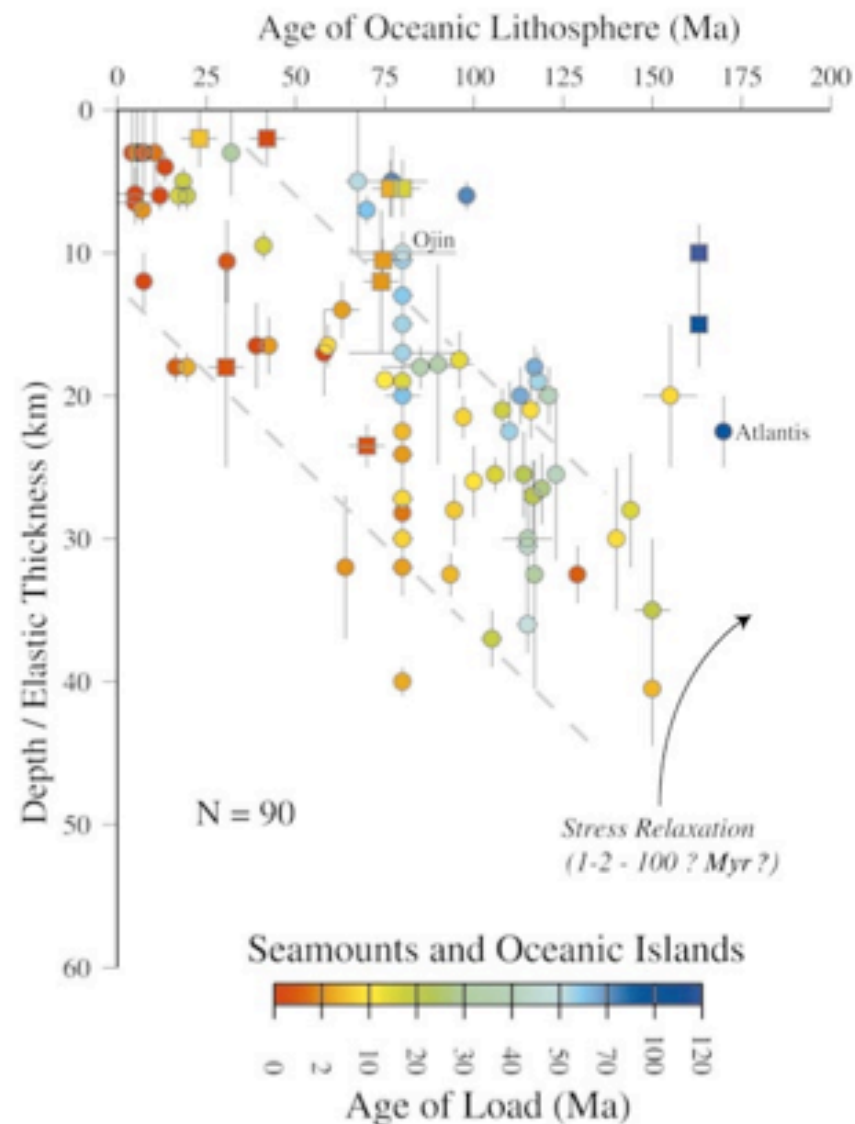
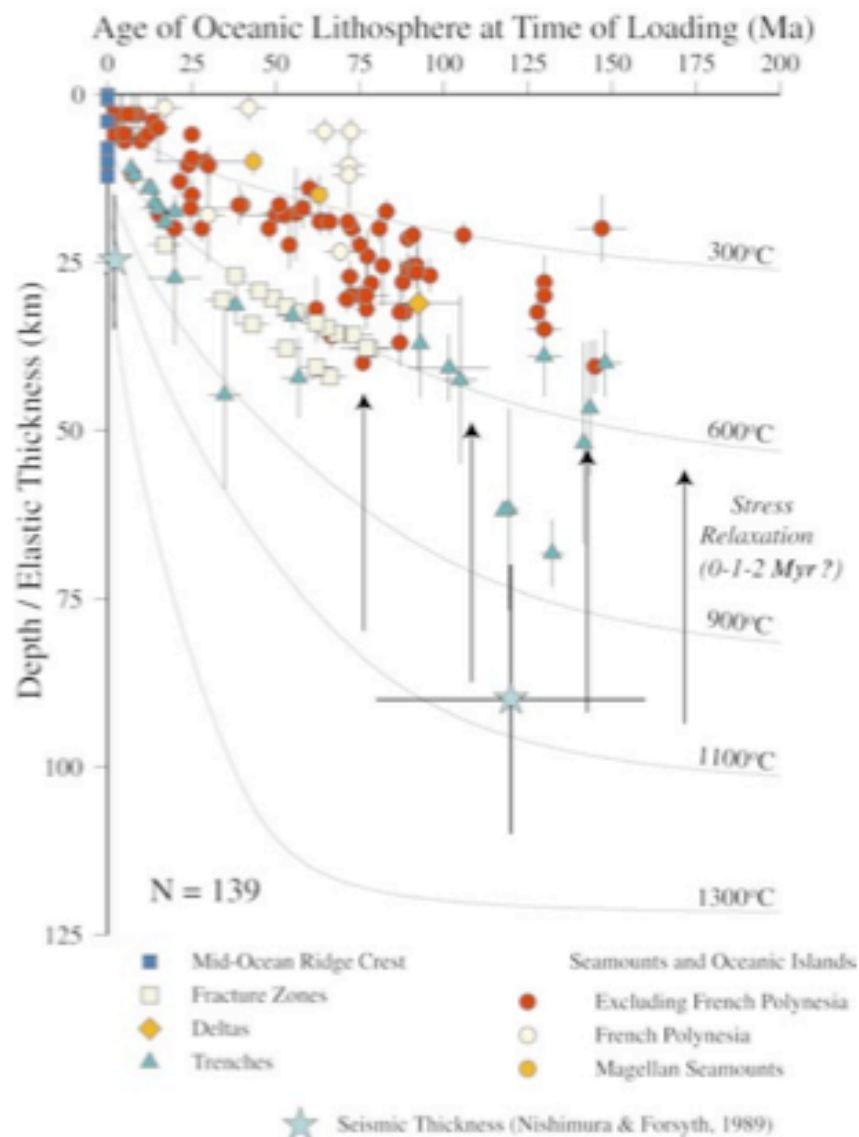
T_e can be estimated by comparing the amplitude and wavelength of the observed gravity anomaly to the predicted anomaly based on an elastic plate model.

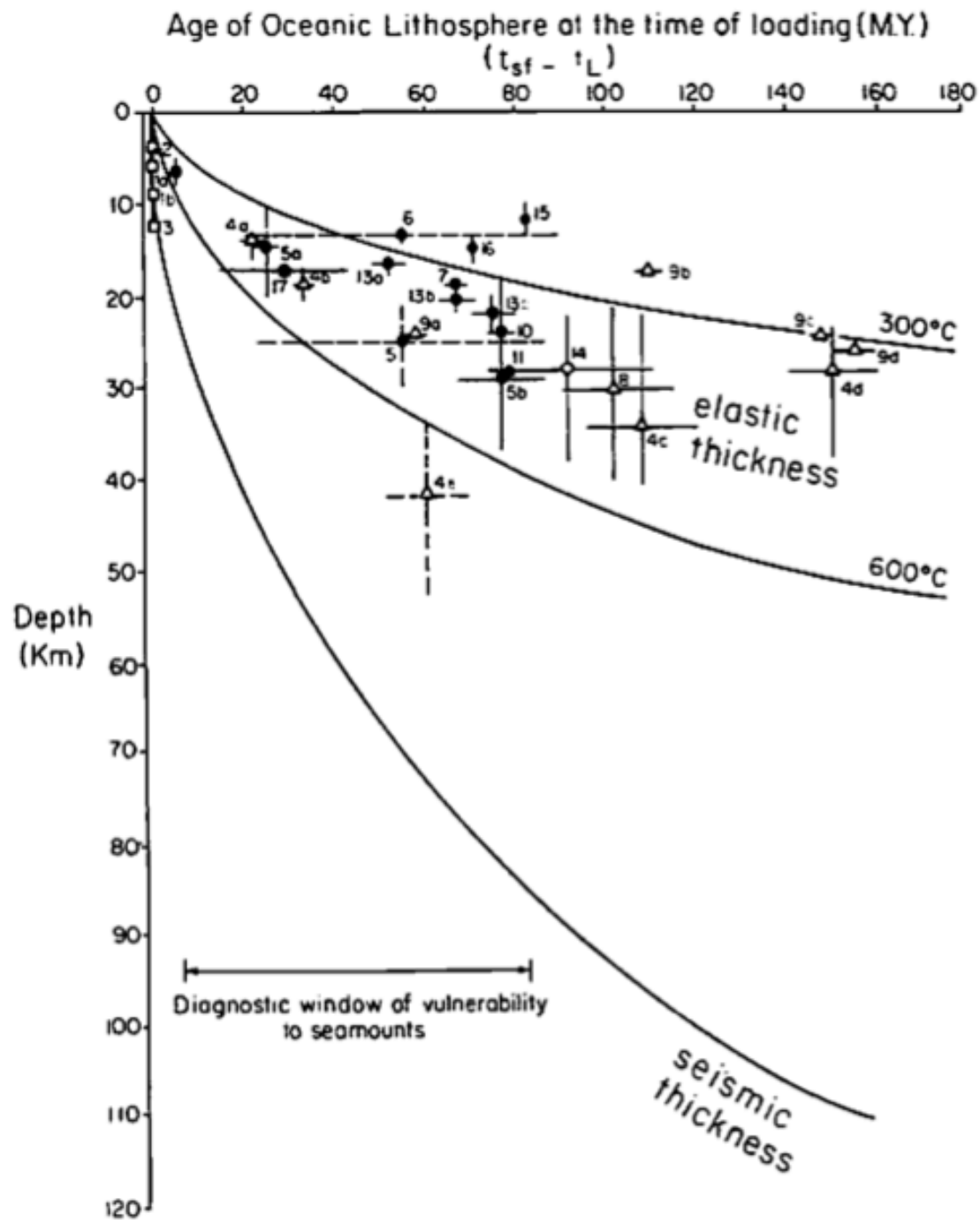


Age of Oceanic Lithosphere (Ma)

0 25 50 75 100 125 150 175 200

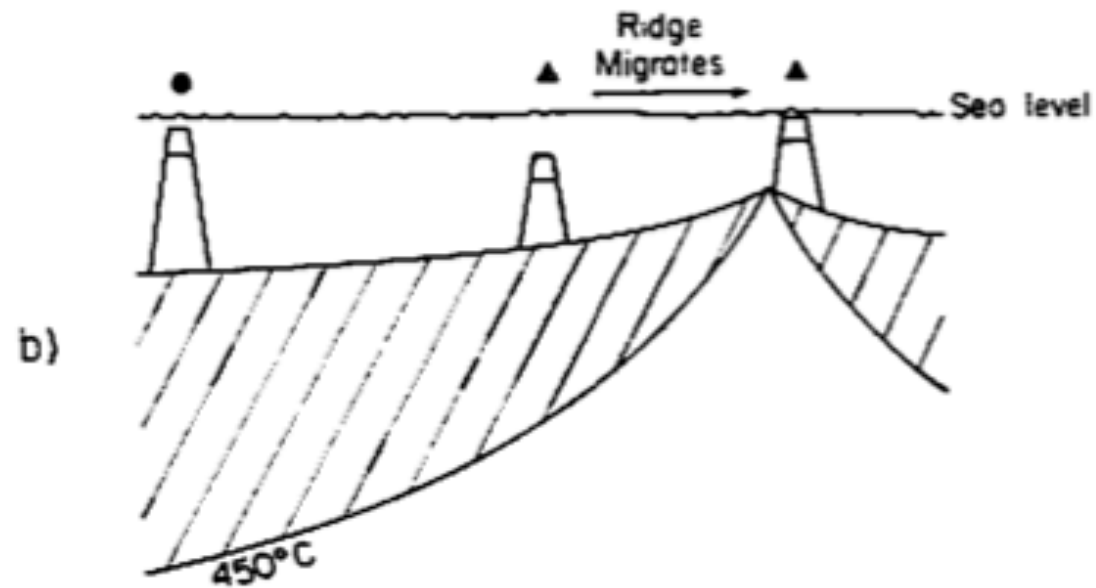
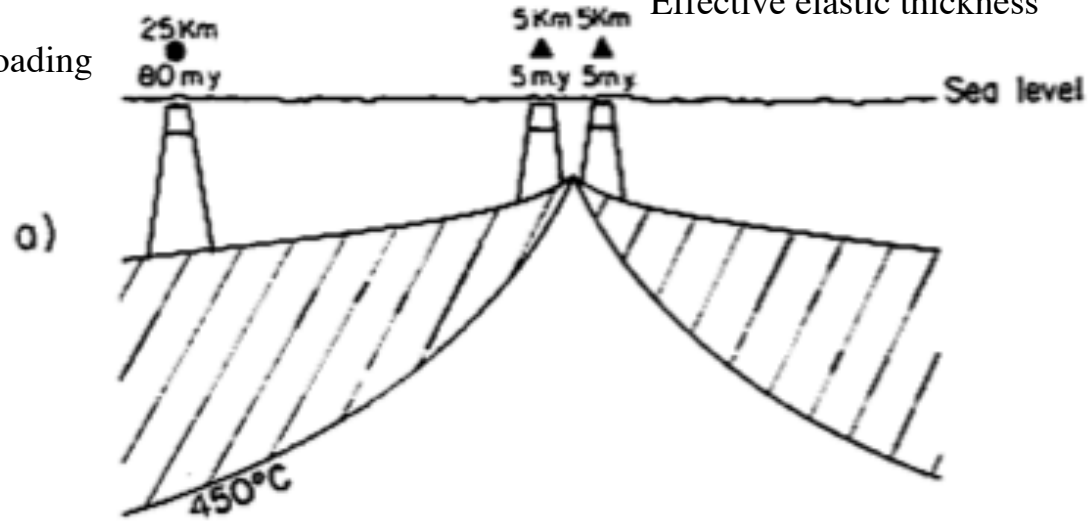




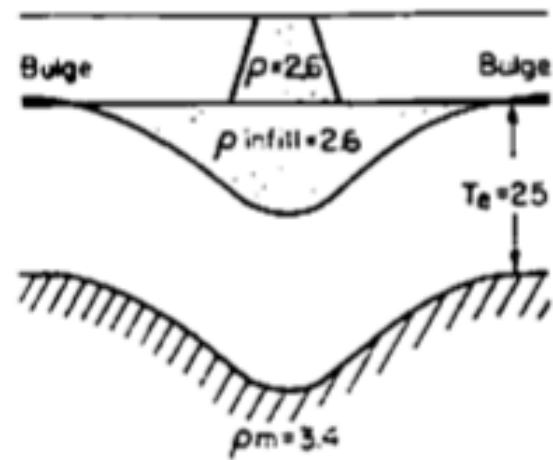


age of loading

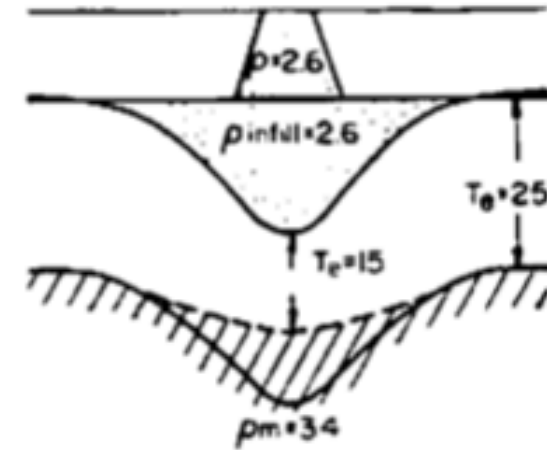
Effective elastic thickness



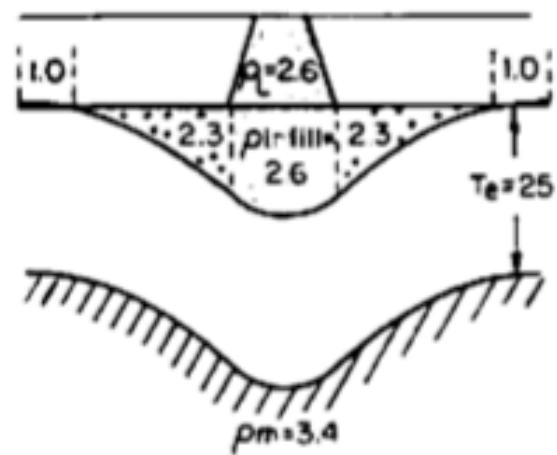
a) Continuous plate



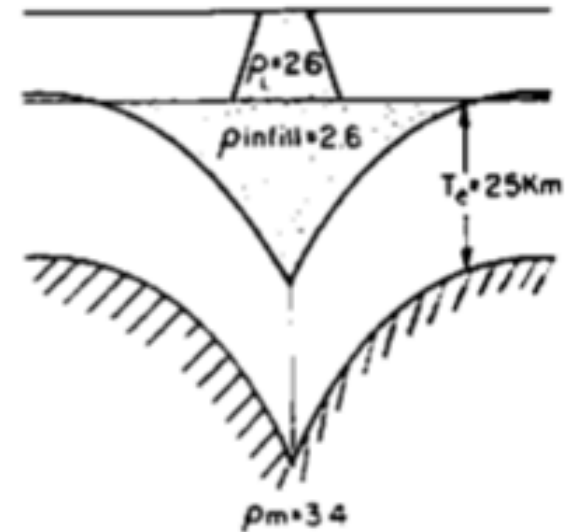
c) Variable T_e



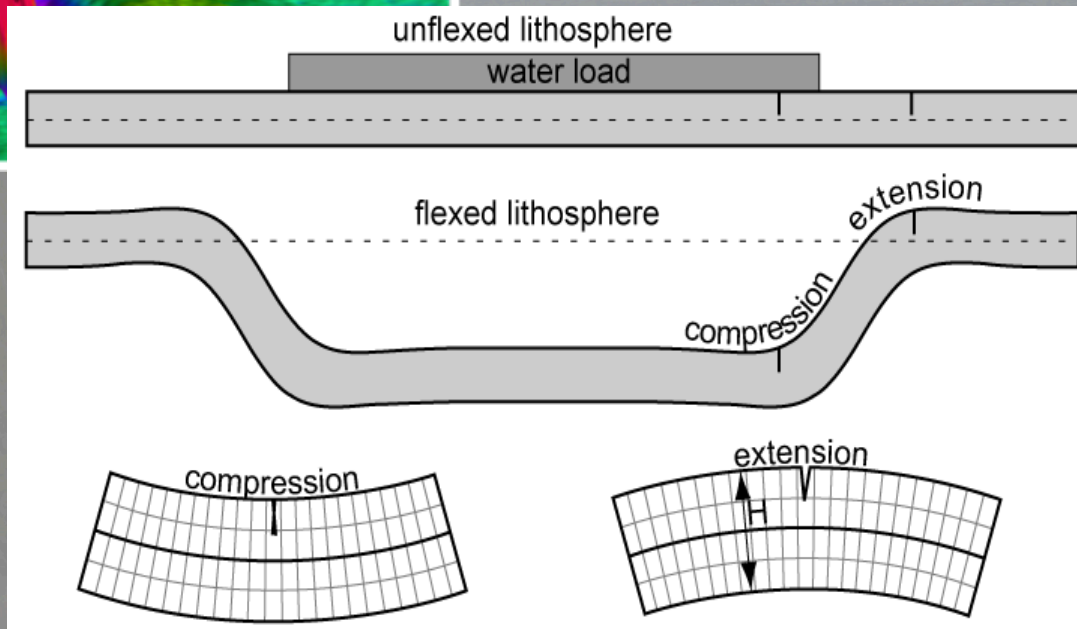
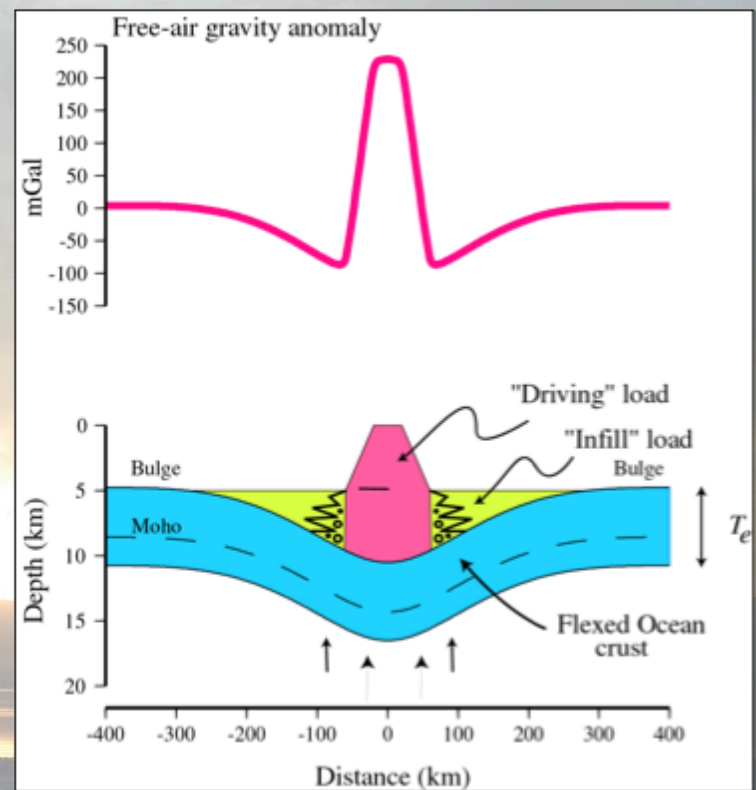
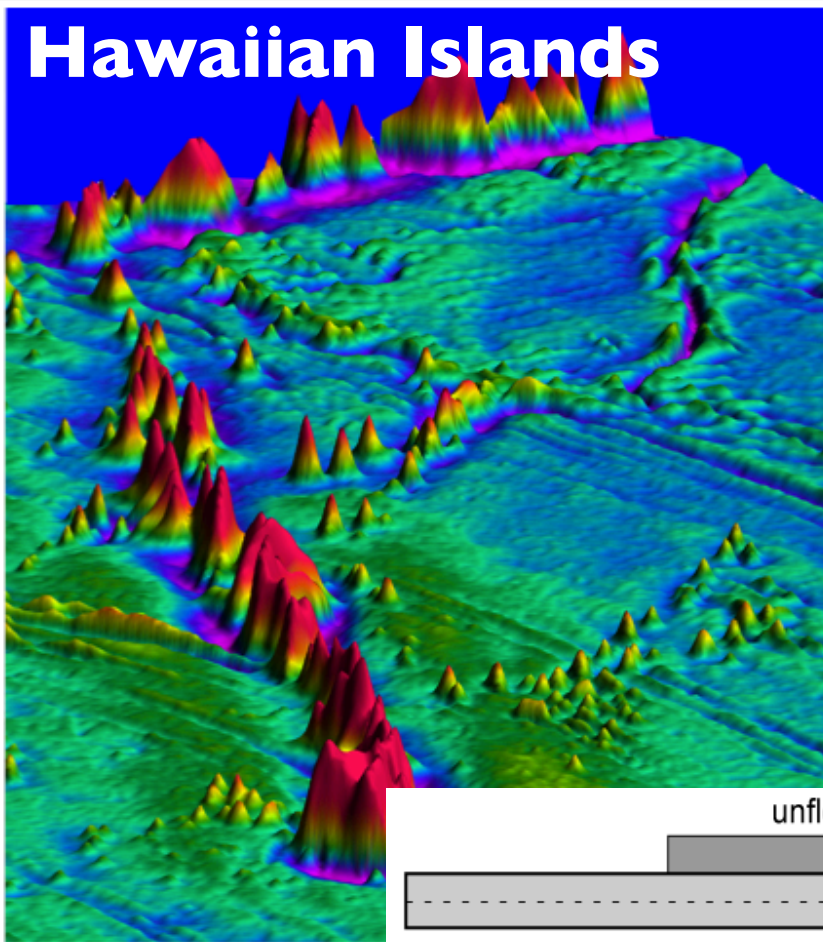
b) Variable infill



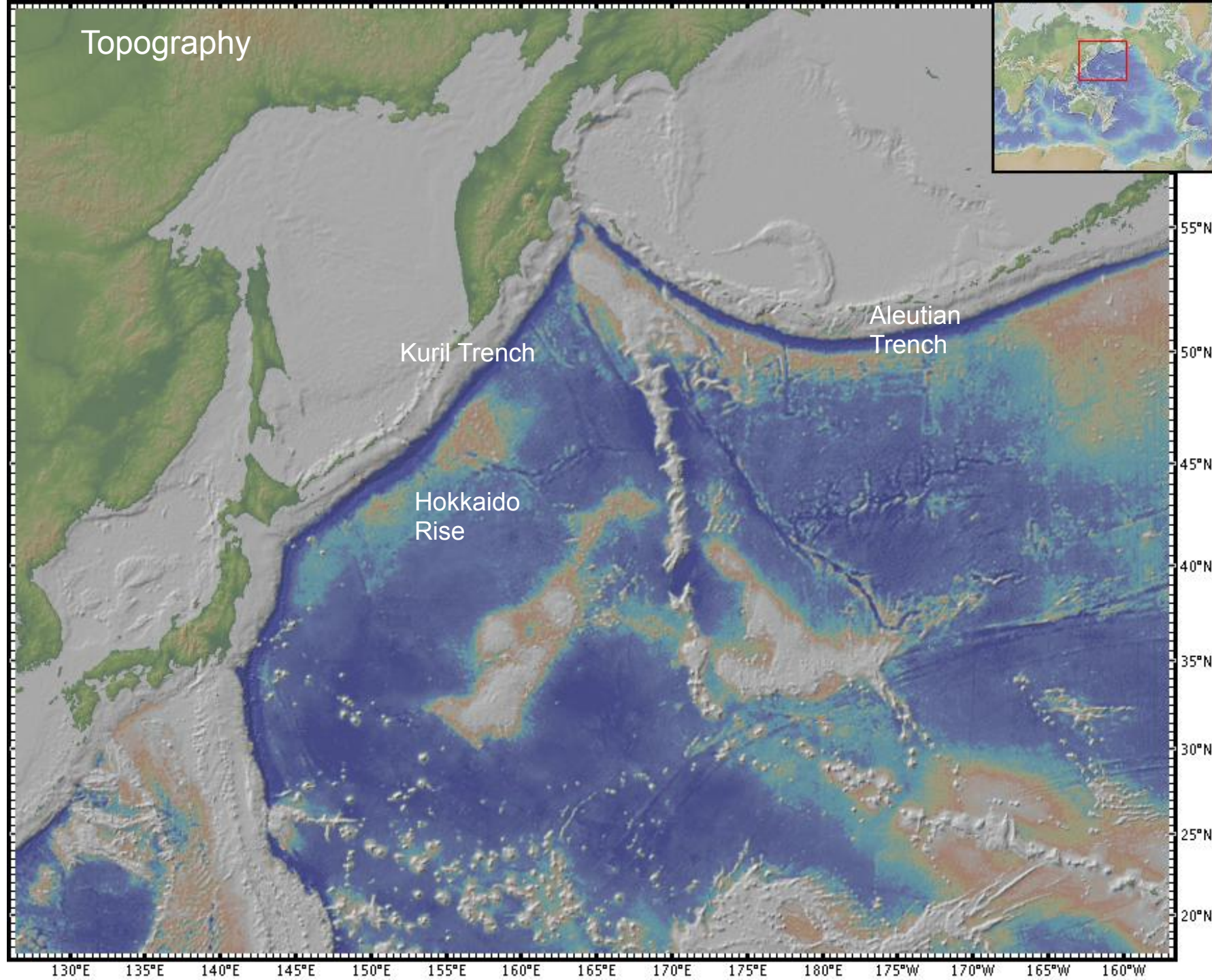
d) Fractured plate



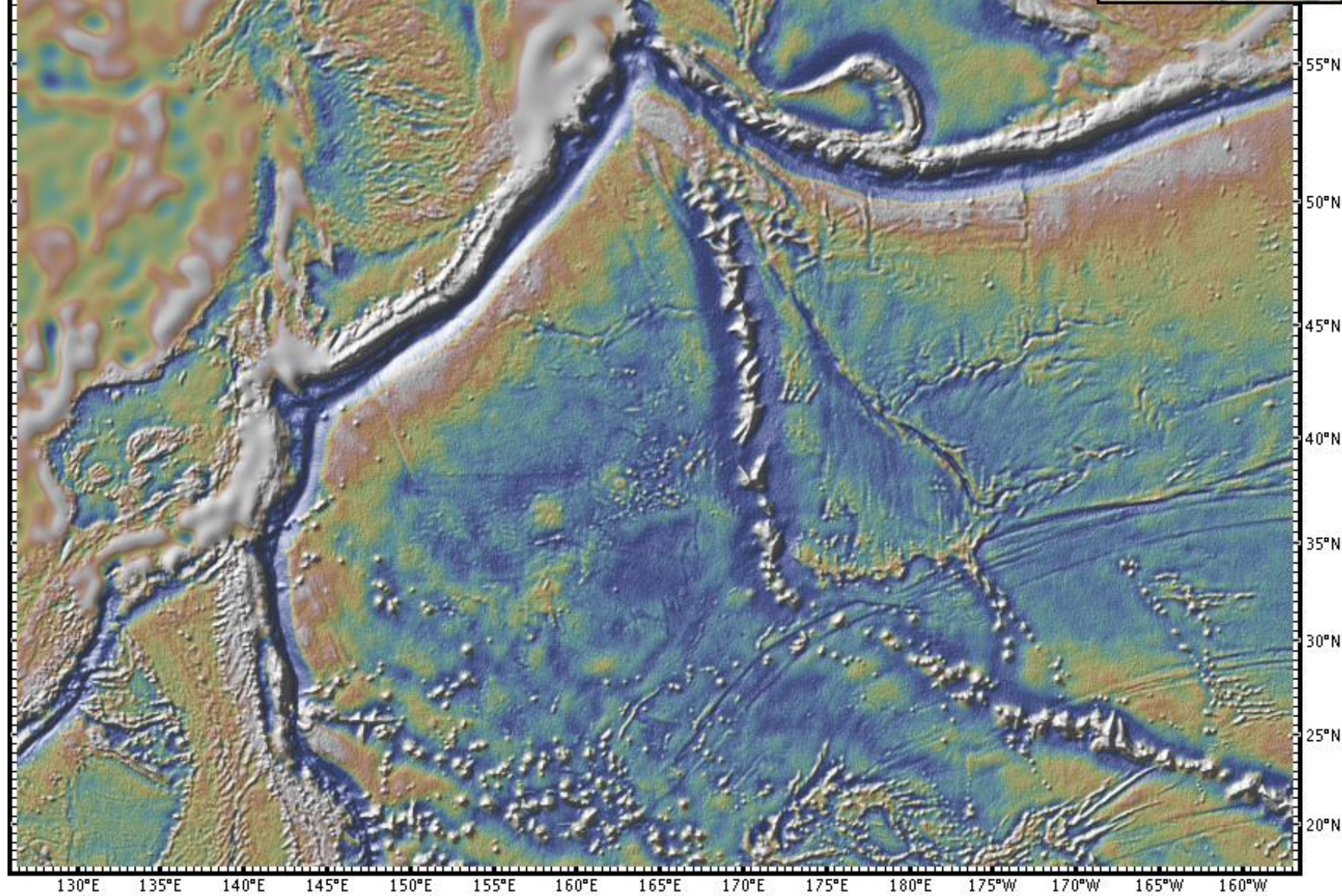
Hawaiian Islands



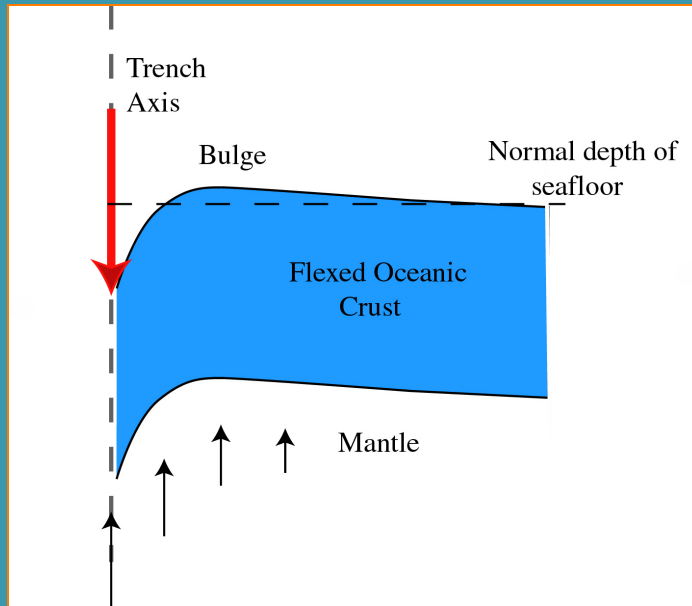
Topography



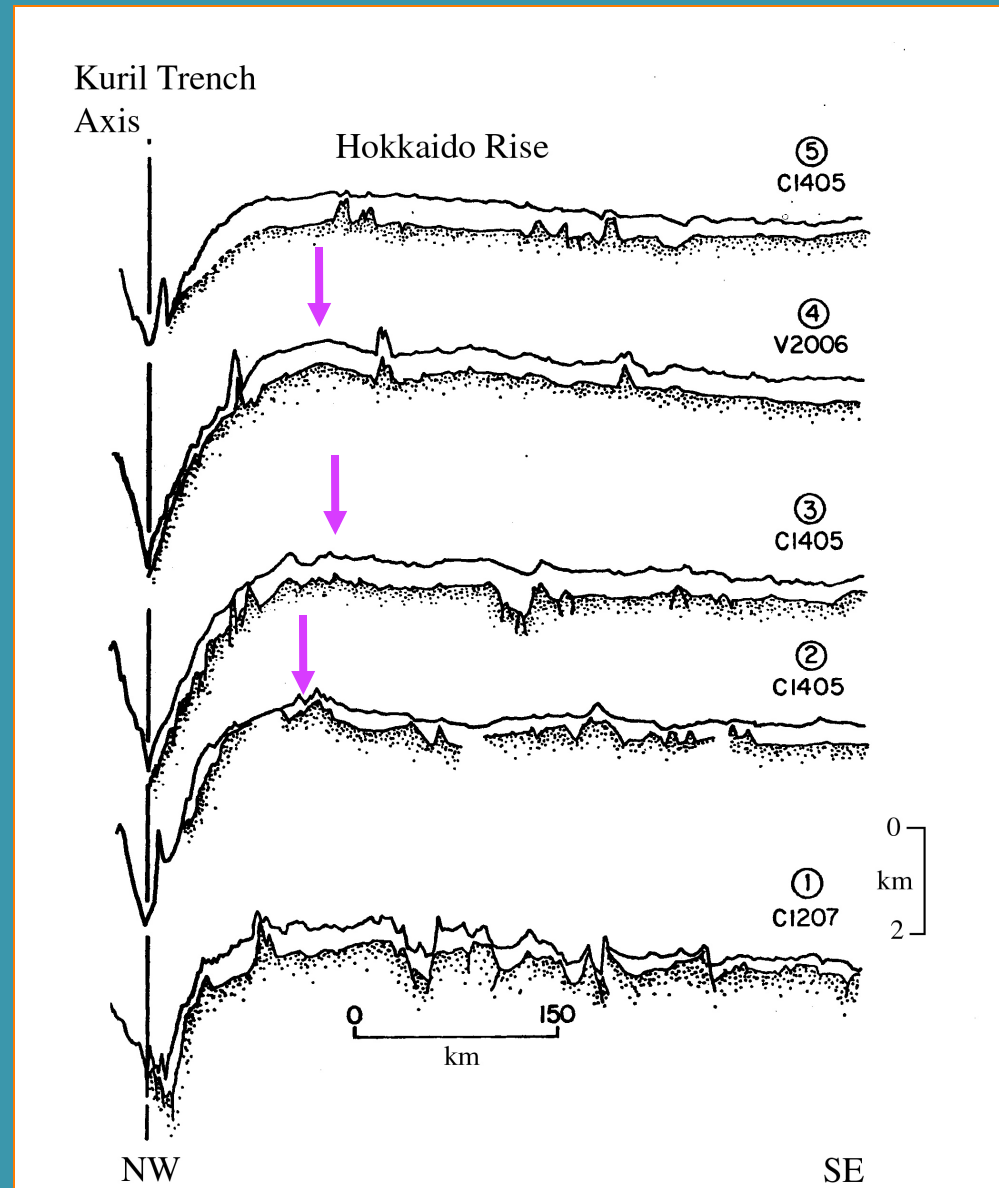
Gravity

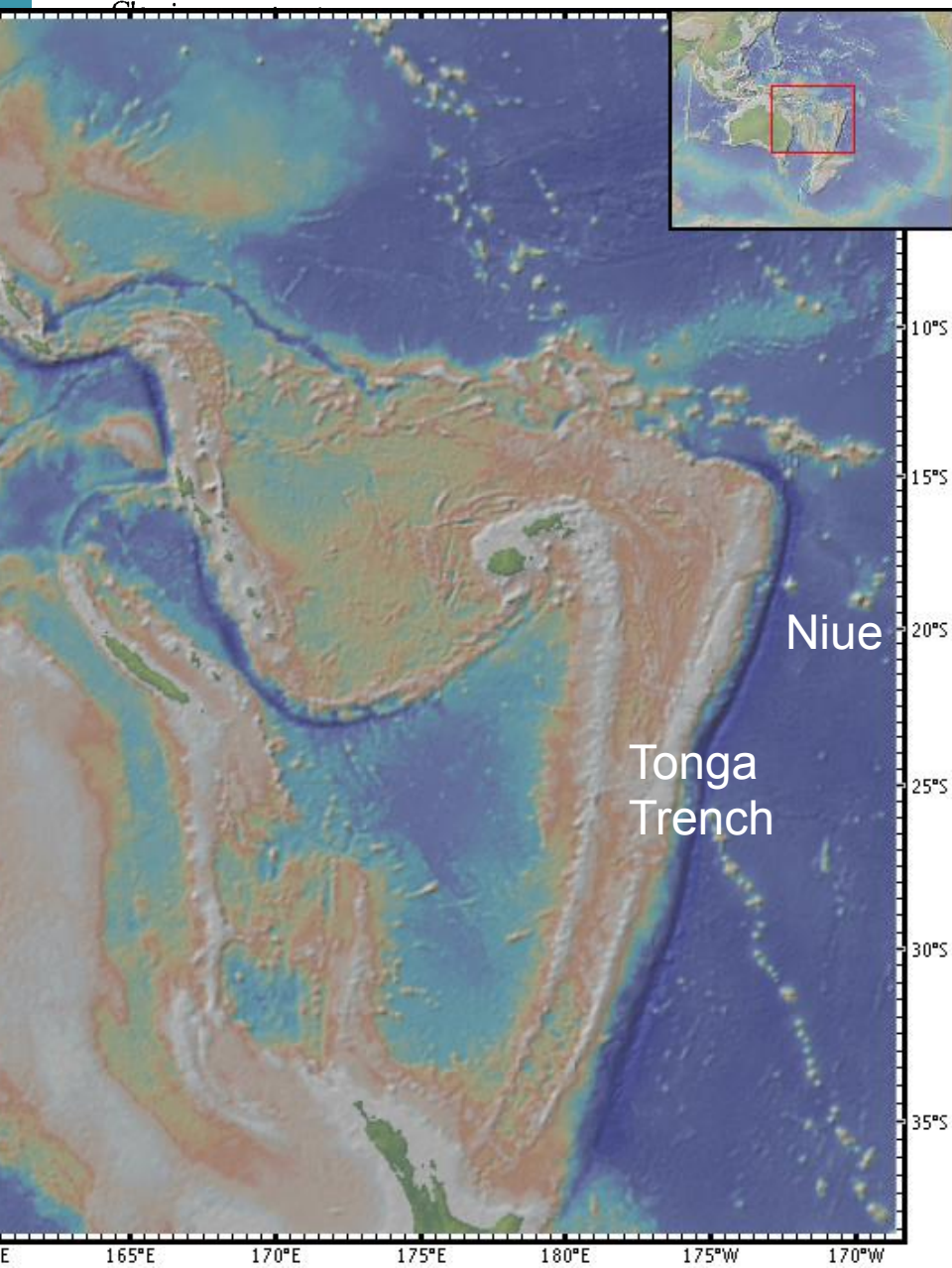


Topography seaward of the Kuril Trench

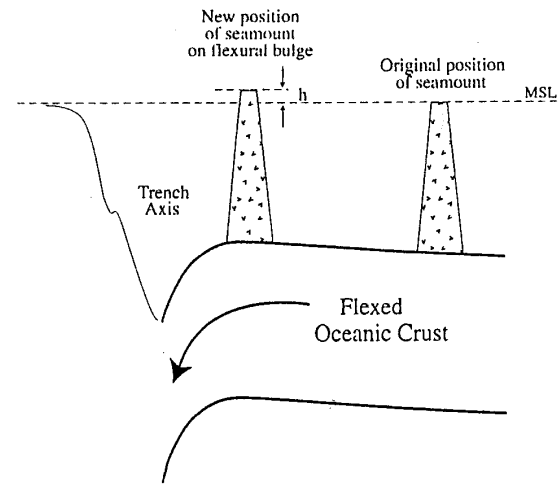


Distance to bulge ~ 120-140 km
 $T_e \sim 30$ km





Uplift (70m) of Niue where Pacific plate bends as it enters Tonga trench



Simple model for the uplift of a seamount as it approaches the trench-outer rise system.

Watts (2001)

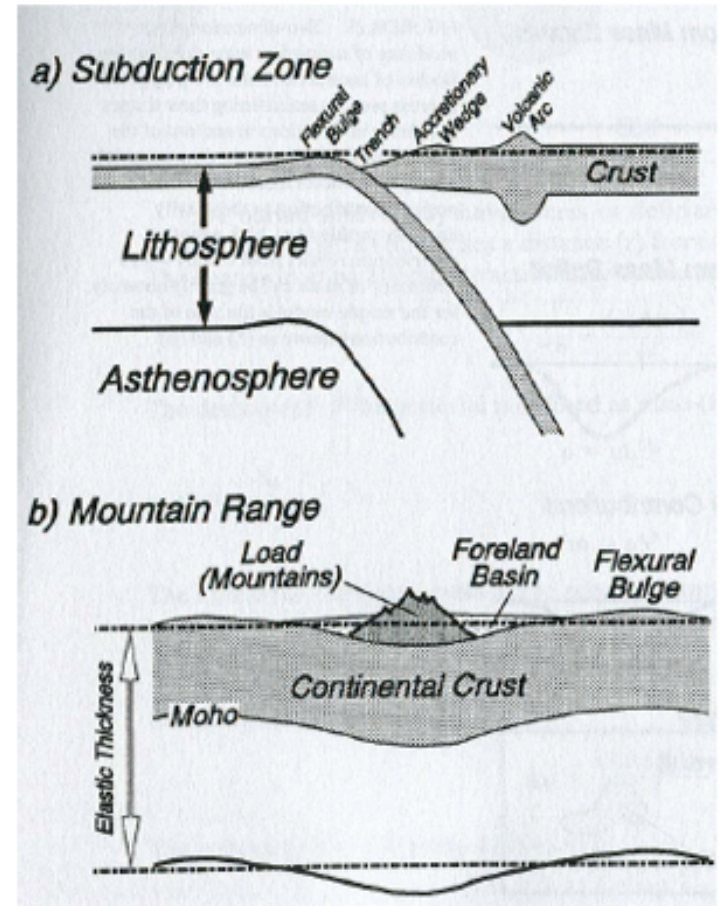
Regional Isostatic Compensation

In (a) the subducting plate is analogous to the diving board – the load is primarily the downgoing plate/slab

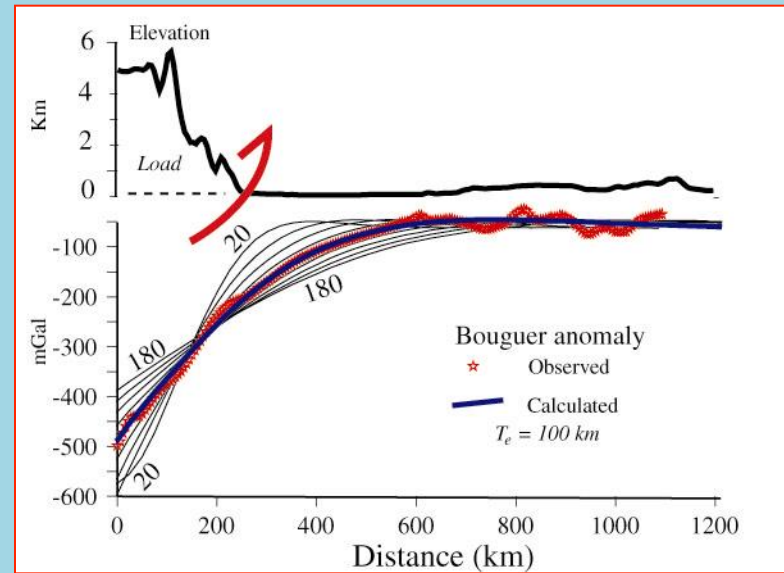
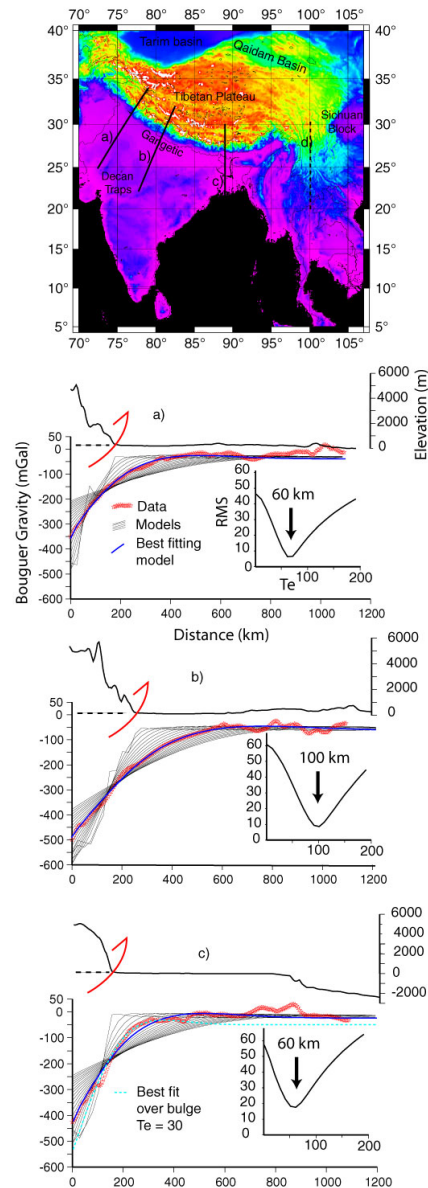
- Flexure of the downgoing plate results in a depression (trench) and, farther out to sea, a bulge on the oceanic crust.

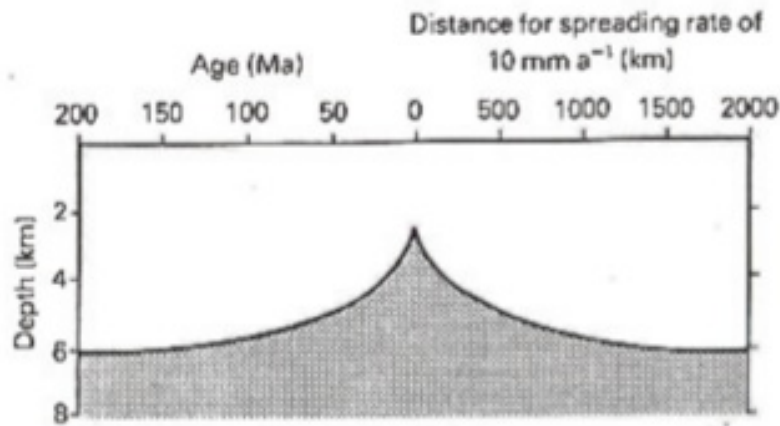
In (b) the mass of the mountains puts a load on the plate that can be expressed in both directions.

- Depressions between the mountains and the flexural bulges (foreland basins) can fill with sediments to considerable thickness.



Example: Himalaya foreland basin: Continent – continent collisions



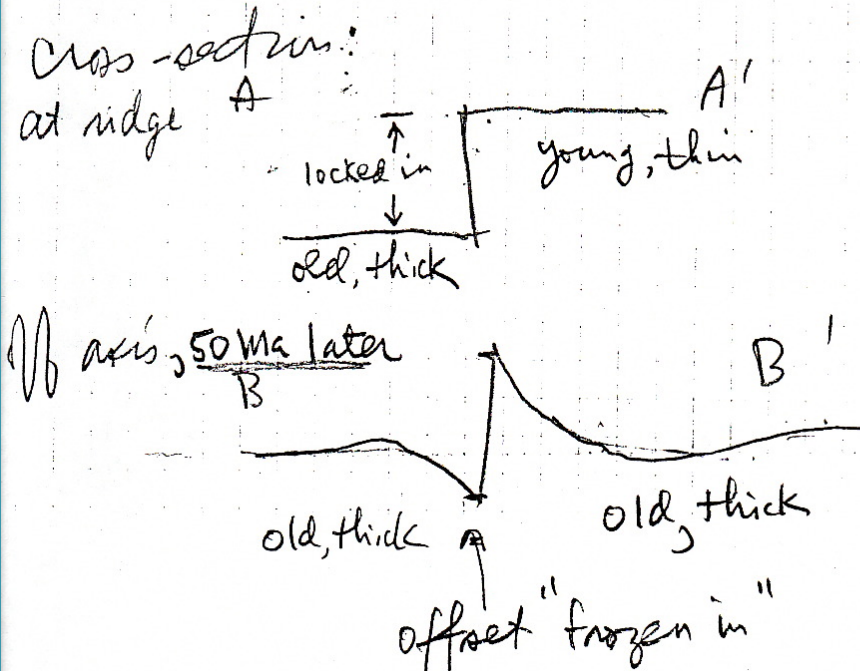
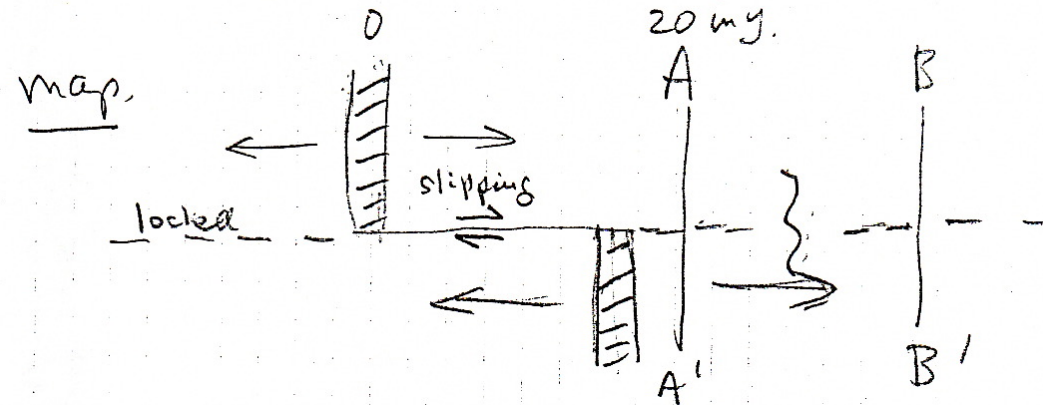


Another location where flexure plays an important role: oceanic fracture zones

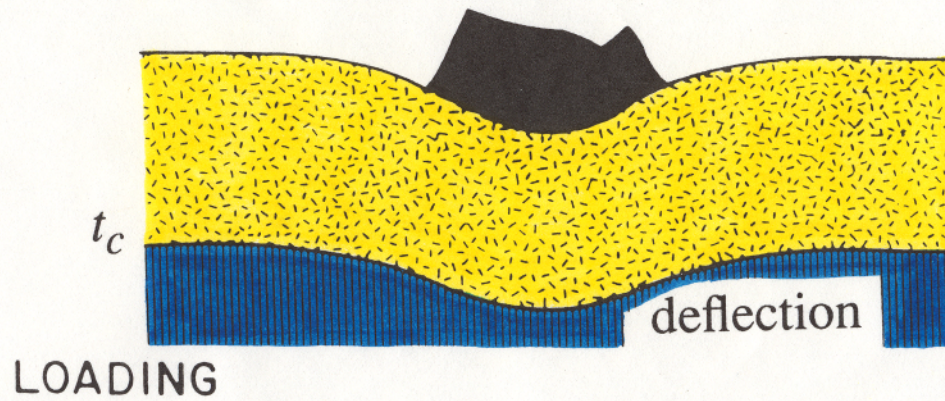
Young oceanic crust is shallower than older oceanic crust; topographic step is frozen in at the ridge axis.

Young oceanic crust subsides faster and difference in depth across fracture zone gradually decreases

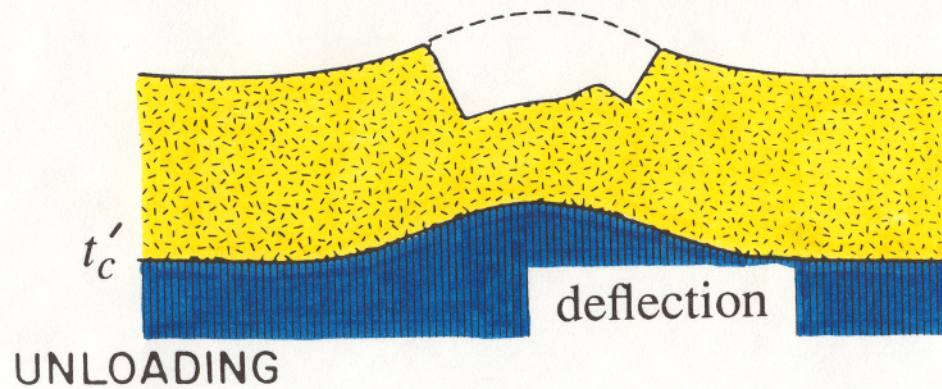
Plate flexes to accommodate decreasing difference in depth

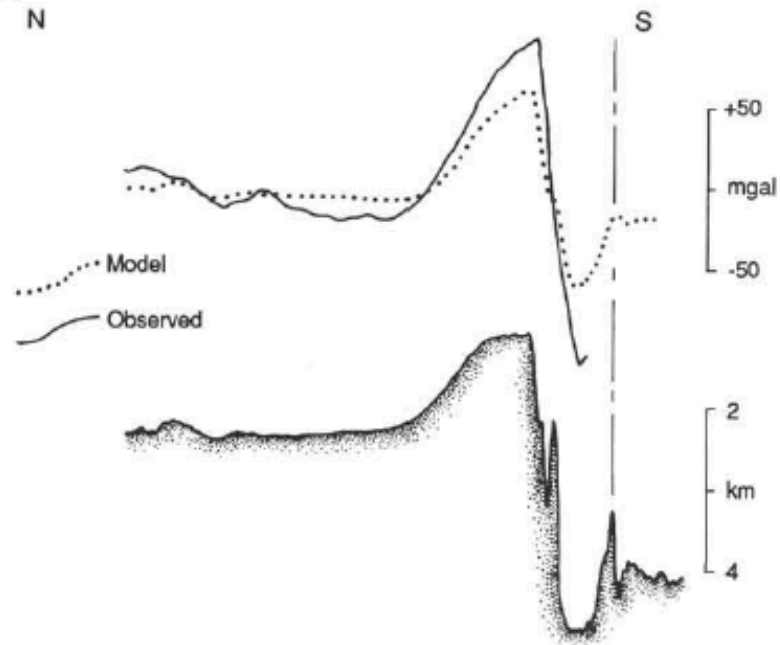
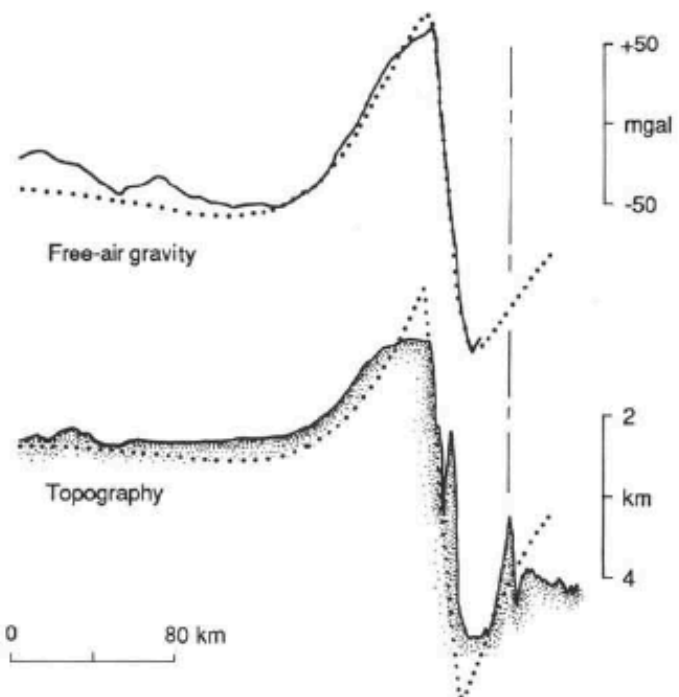


Seamount load



Region over which material has been removed laterally by extension



A
N**B**

Classic settings where plate flexure is observed:

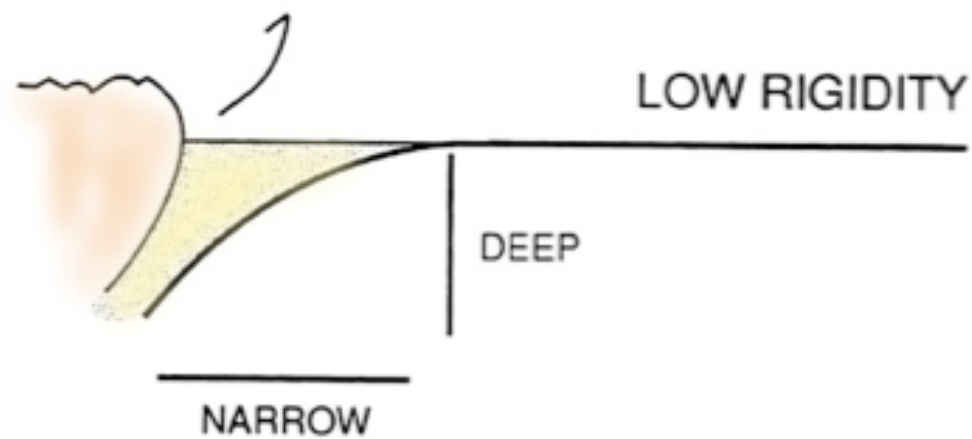
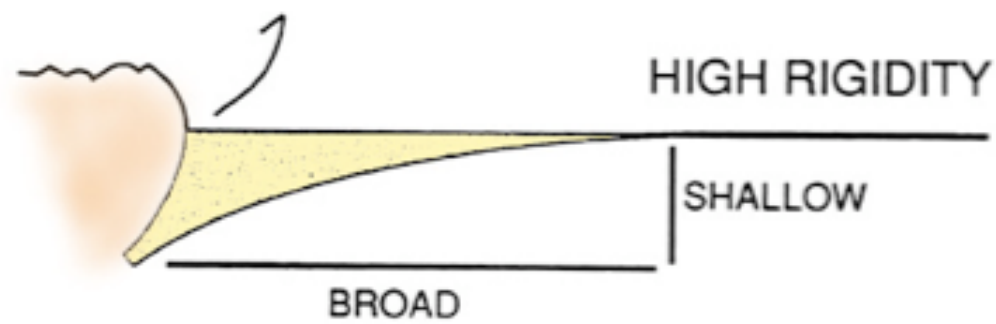
Post-glacial rebound in Scandinavia and Canada

Large seamounts, seamount chains (Hawaii)

Incoming oceanic plates at trenches (Kuril trench)

Fracture zones with large age offsets (Udintsev FZ)

Foreland basins in continent-continent collisions (India)



A complex body can be represented by a simple source.

In this case, upper and lower layers have no effect.

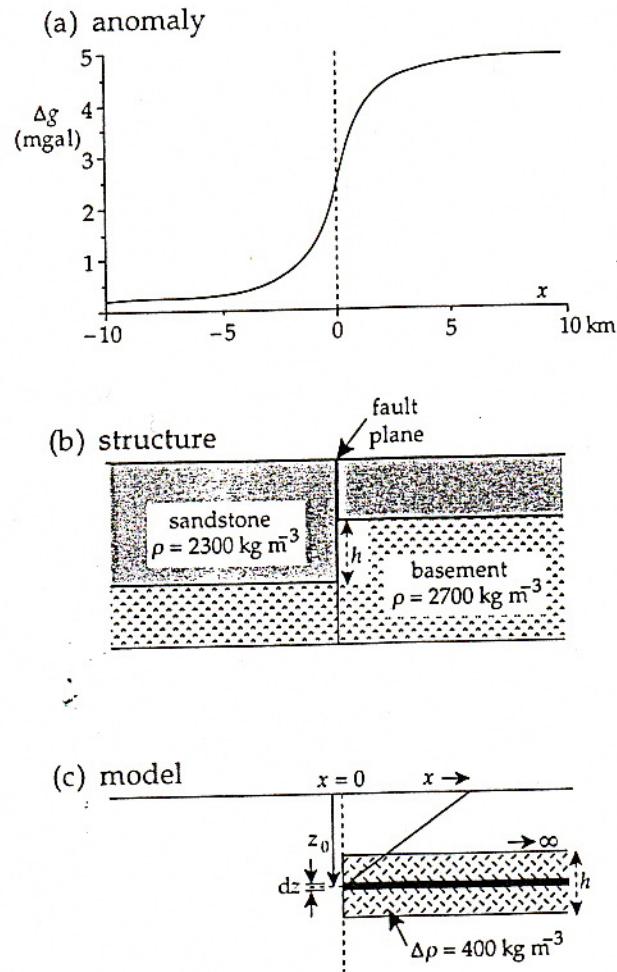
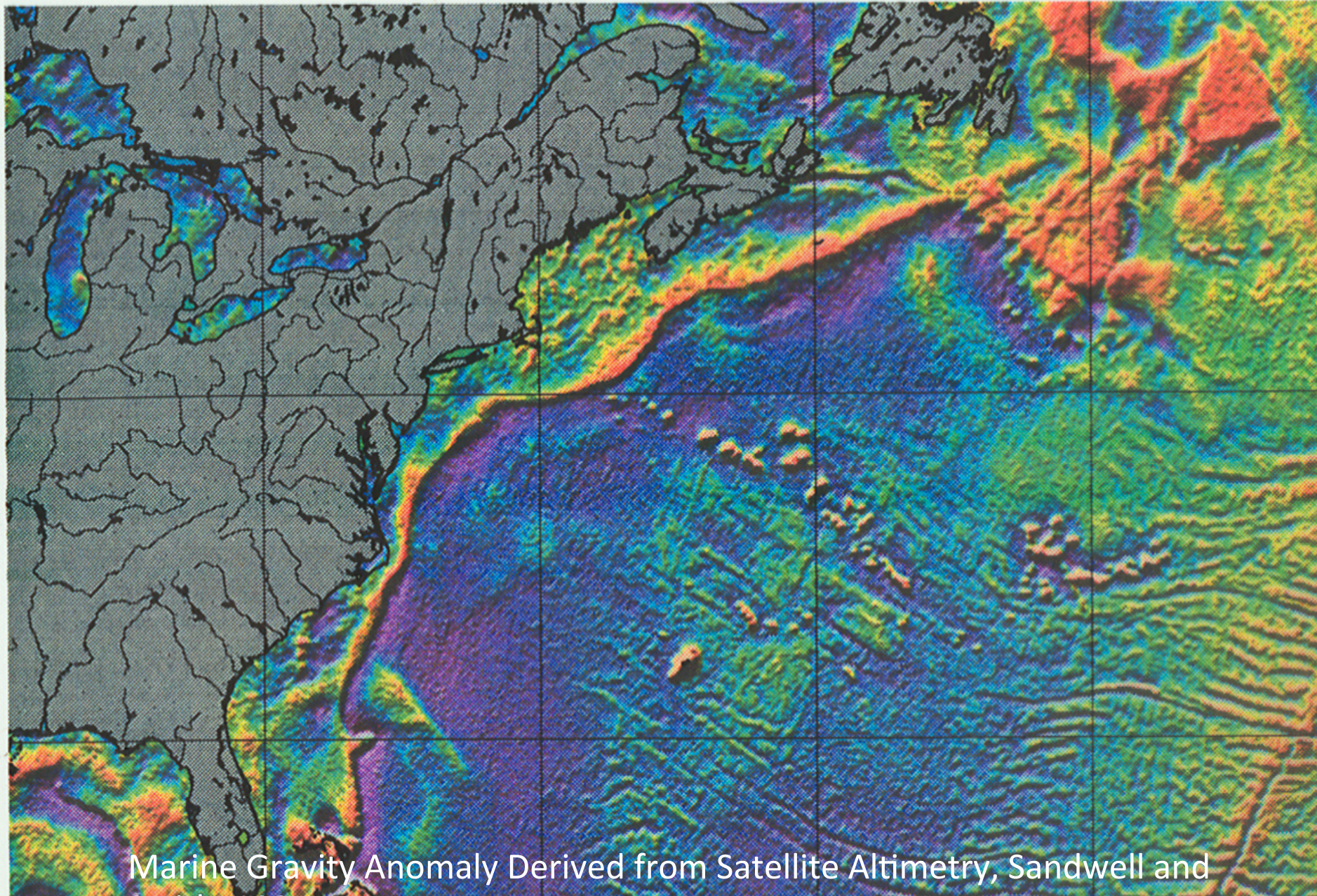


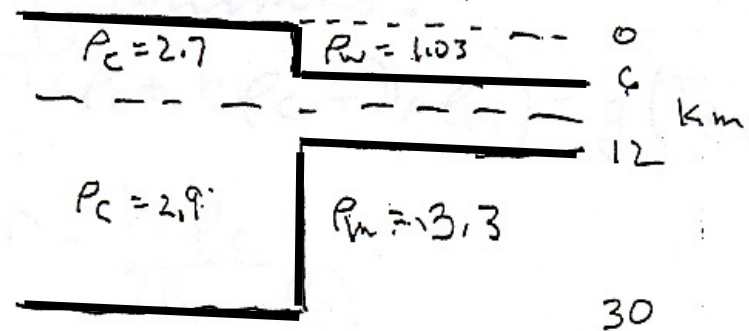
Fig. 2.52 (a) The gravity anomaly across a vertical fault; (b) structure of a fault with vertical displacement h , and (c) model of the anomalous body as a semi-infinite horizontal slab of height h .

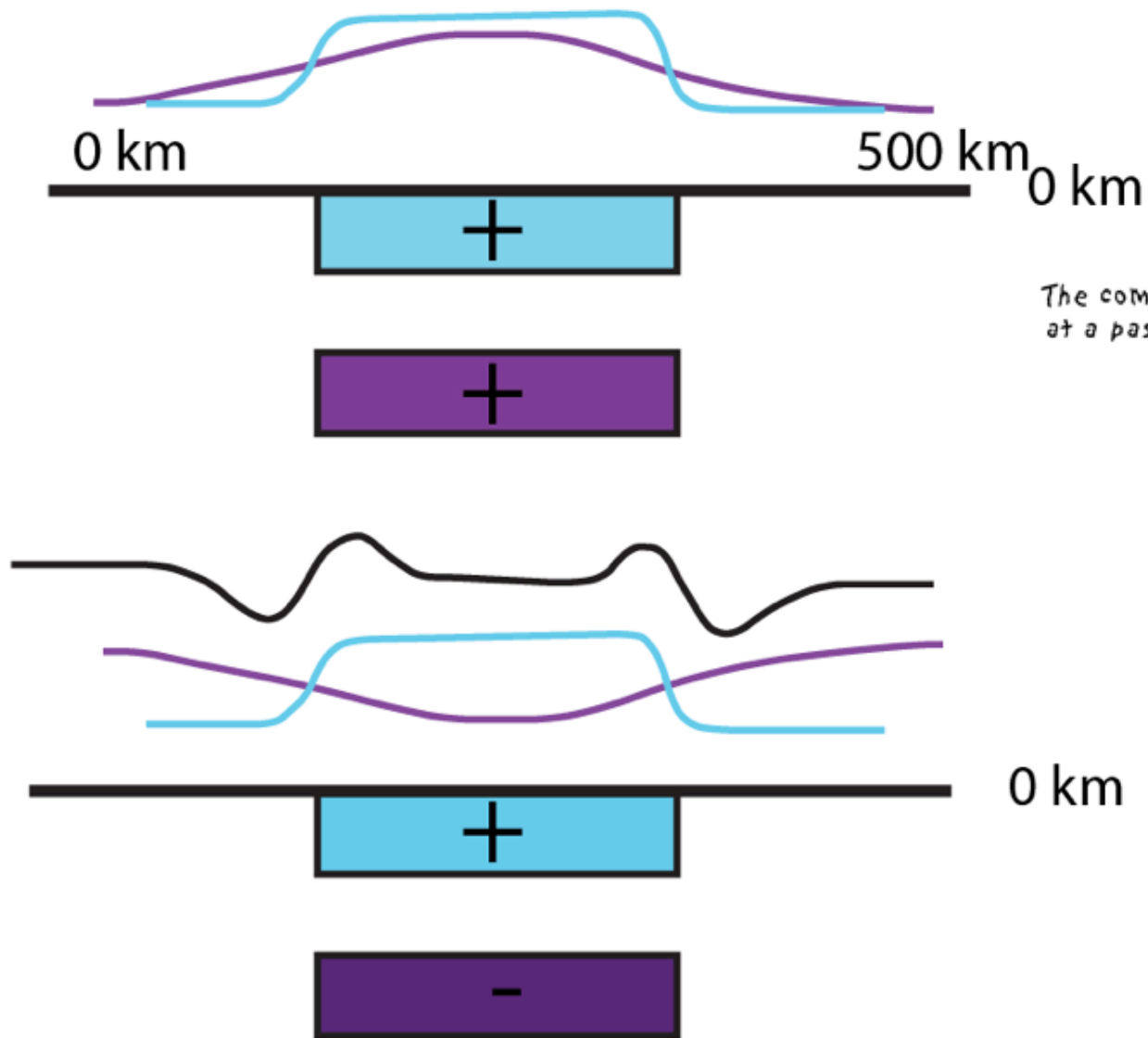
Why is there a gravity high over the continental shelf break?



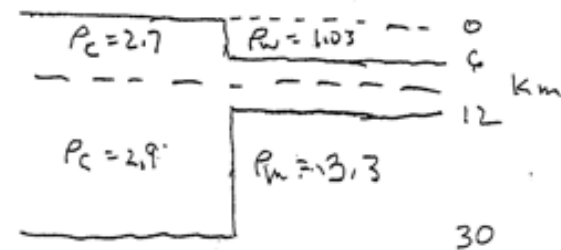
Marine Gravity Anomaly Derived from Satellite Altimetry, Sandwell and

Topography



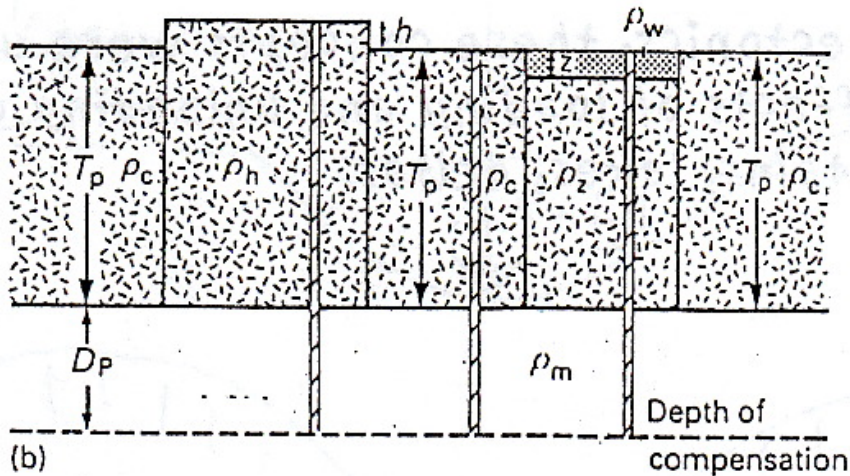


The composite anomaly is similar to what you see at a passive continental margin.

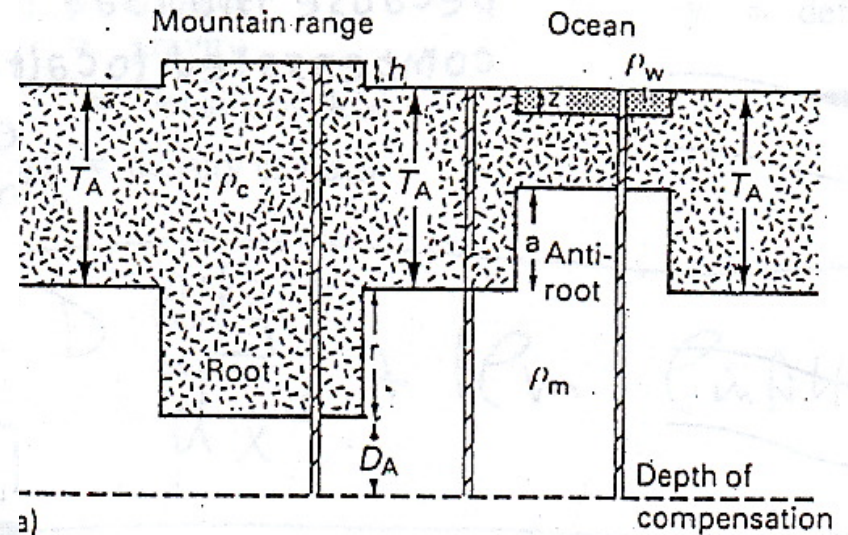


Two basic models to explain isostatic compensation

Pratt's hypothesis:
horizontal variations in density



Airy's hypothesis:
Horizontal variations in crustal thickness



Formula for Airy isostasy:

equating weights of columns:

$$g(h\rho_c + T_A\rho_c + r\rho_c + D_A\rho_m) = g(T_A\rho_c + r\rho_m + D_A\rho_m)$$

HW: find: $r = \frac{h\rho_c}{(\rho_m - \rho_c)}$

Do same over ocean and find: $a = \frac{z(\rho_c - \rho_w)}{(\rho_m - \rho_c)}$

For typical densities of
water (1.03), crust (2.8) and
mantle (3.3) (gm/cc):

$$r = 5.6 h \quad \text{and} \quad a = 3.5 z$$