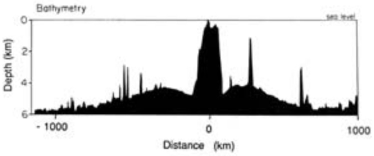


Gravitational constraints

Reading: Fowler p206 – 228

Gravity anomalies

Free-air anomaly:

$$g_F = g_{obs} - g(\lambda) + \delta g_F$$

Corrected for expected variations due to
- the spheroid
- elevation above the spheroid

Bouguer anomaly:

$$g_B = g_{obs} - g(\lambda) + \delta g_F - \delta g_B + \delta g_T$$

Corrected for expected variations due to
- the spheroid
- elevation above the spheroid
- Rock above sea level

Isostatic equilibrium

Can we use gravity anomalies to tell if a region is in isostatic equilibrium?

Isostatic equilibrium means no excess mass.
Does this mean no gravity anomaly?

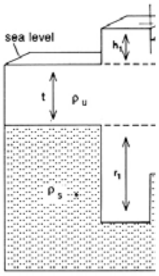
Not quite!

Consider the figure:

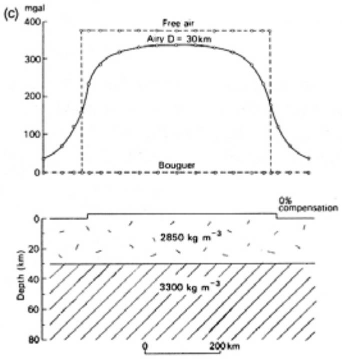
- Assume isostatic equilibrium
- Bouguer anomaly will be large and negative
- Free-air anomaly: small but positive

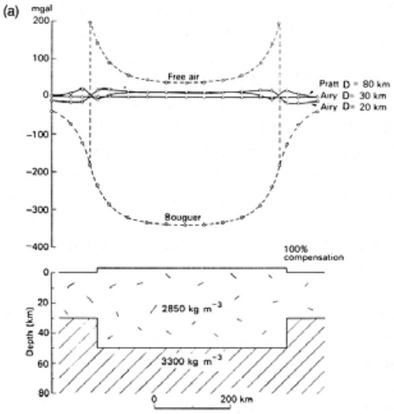
WHY?

WHY?



Isostatic equilibrium





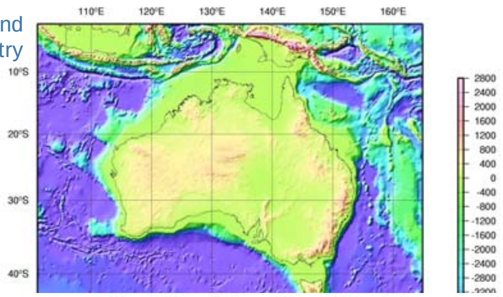
Uncompensated
- Large positive Free-air
- Zero Bouguer

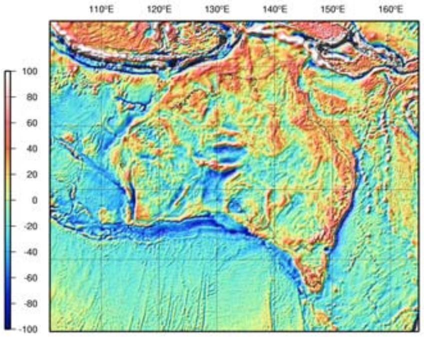
Compensated
- Small positive Free-air
- Large negative Bouguer

…away from the edges – why?

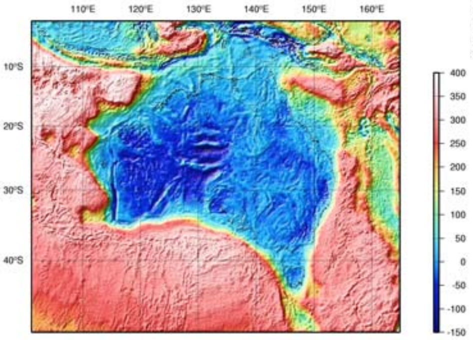
Australia

Topography and bathymetry





Free-air anomaly



Bouguer anomaly

Gravity anomalies

Analytical expressions for simple shapes: Buried sphere

Only the density contrast is important

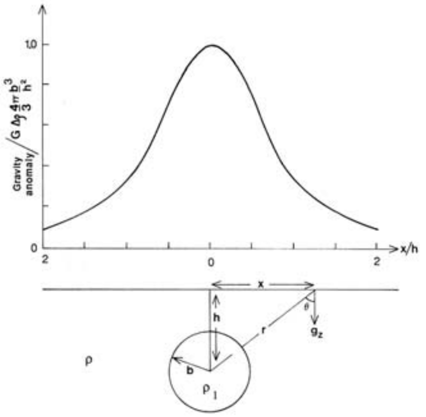
$$\Delta\rho = \rho_1 - \rho_2$$

Gravitational acceleration toward sphere

$$g = \frac{Gm}{r^2}$$

Gravimeters measure the vertical gravitational acceleration

$$g_z = \frac{Gm}{r^2}\cos\theta = \frac{Gmh}{r^3} = \frac{Gmh}{\left(x^2+h^2\right)^{3/2}}$$



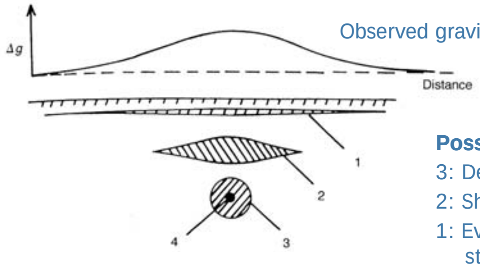
Mass excess of sphere $m = \frac{4}{3}\Delta\rho\pi b^3$

Gravity anomaly:

$$\delta g_z = \frac{4G\Delta\rho\pi b^3 h}{3\left(x^2+h^2\right)^{3/2}}$$

Ambiguity

An observed gravity anomaly can be explained by a variety of mass distributions at different depths



Observed gravity anomaly

Possible causal structures:

3: Deep sphere

2: Shallower elongated anomaly

1: Even shallower, more elongated structure

Ambiguity in formula for a sphere:

$$\delta g_z = \frac{4G\Delta\rho\pi b^3 h}{3\left(x^2 + h^2\right)^{3/2}}$$

- Trade-off between density and radius

Gravity observations

Mid-ocean ridge

Four density models adequately satisfy the observations

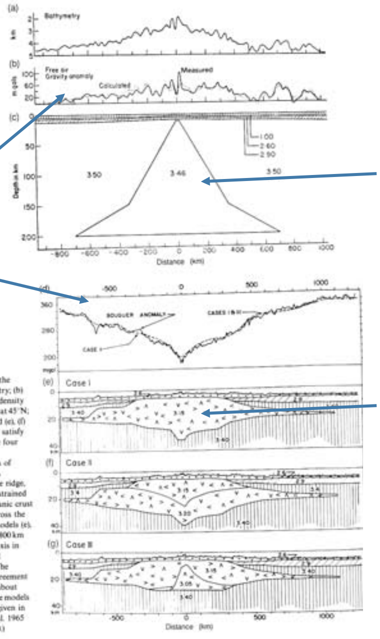


Figure 8.11. Gravity models for the Mid-Atlantic Ridge: (a) bathymetry; (b) free-air gravity anomaly and (c) density model for the Mid-Atlantic Ridge at 45°N; (d) Bouguer gravity anomaly and (e), (f) and (g) density models which all satisfy the anomaly shown in (d). These four density models – (c), (e), (f) and (g) – illustrate the nonuniqueness of models based on gravity data. A low-density zone lies beneath the ridge, but its dimensions need to be constrained by other methods also. The oceanic crust is assumed to be continuous across the ridge axis in model (c); but in models (e), (f) and (g) there is a zone some 800 km wide and centred on the ridge axis in which normal oceanic crust and uppermost mantle are absent. The density model in (c) is in better agreement with everything that is known about midocean ridge structure than the models in (e), (f) and (g). Densities are given in 10^3 kg m^{-3}. (From Talwani et al. 1965 and Keen and Tramontini 1970.)

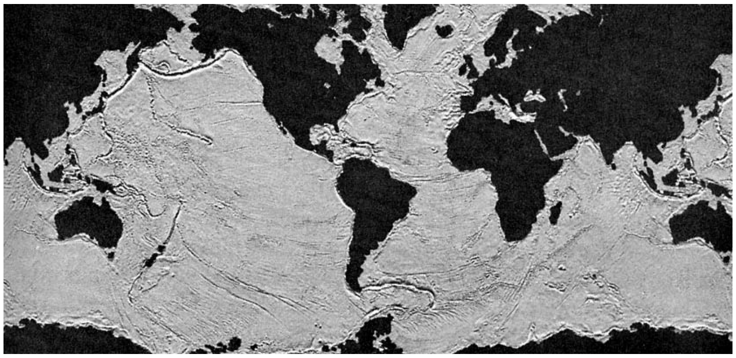
Small positive free-air anomaly, large negative bouguer anomaly: close to isostatic equilibrium

Deep model has small density contrasts

Shallow models have larger density contrasts

Gravity observations

Ocean trenches



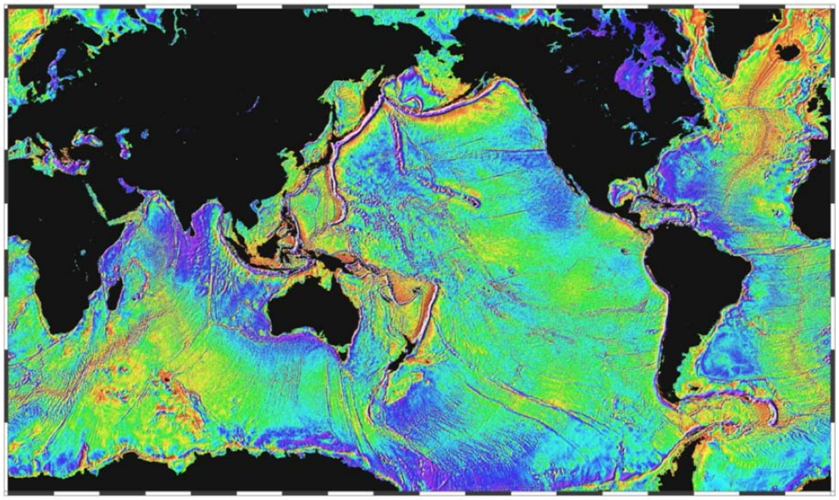
SEASAT Gravity Map

Largest anomalies are associated with the trenches

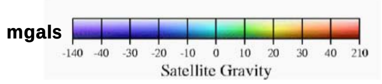
- 10 km deep and filled with water rather than rock
- Not compensated as they are being loaded down dip

Gravity observations

Ocean trenches

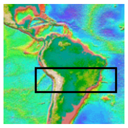


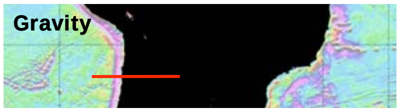
SEASAT Gravity Map

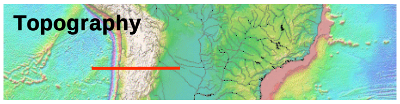


Subduction profiles

Across the Chile Trench

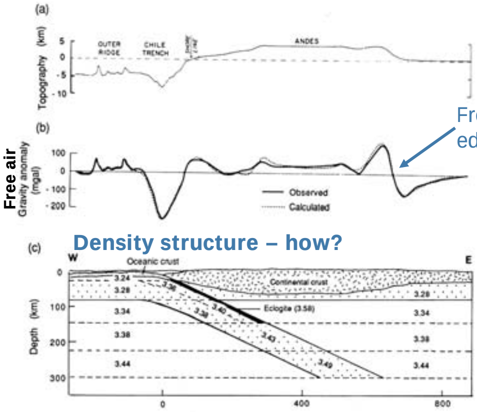






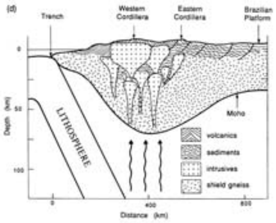
Classic low-high pair

- Low over trench
- High on ocean-ward side of the volcanic arc



Free air edge effect

Density structure – how?



60km thick Andean crust

- Believed to have been thickened from below by intrusive volcanism from slab

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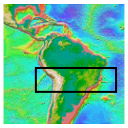
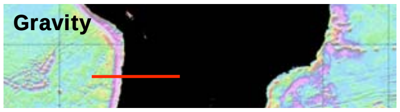
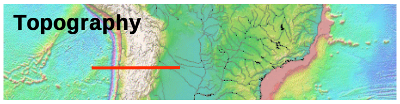
Subduction profiles

Across the Chile Trench



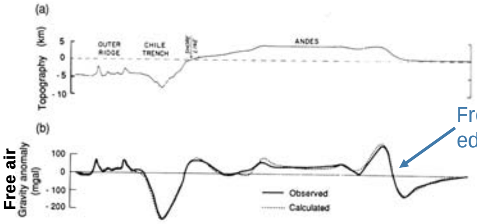




Estimate the expected gravity anomaly using infinite slab formula:

$$\Delta g = 2\pi G\rho h$$

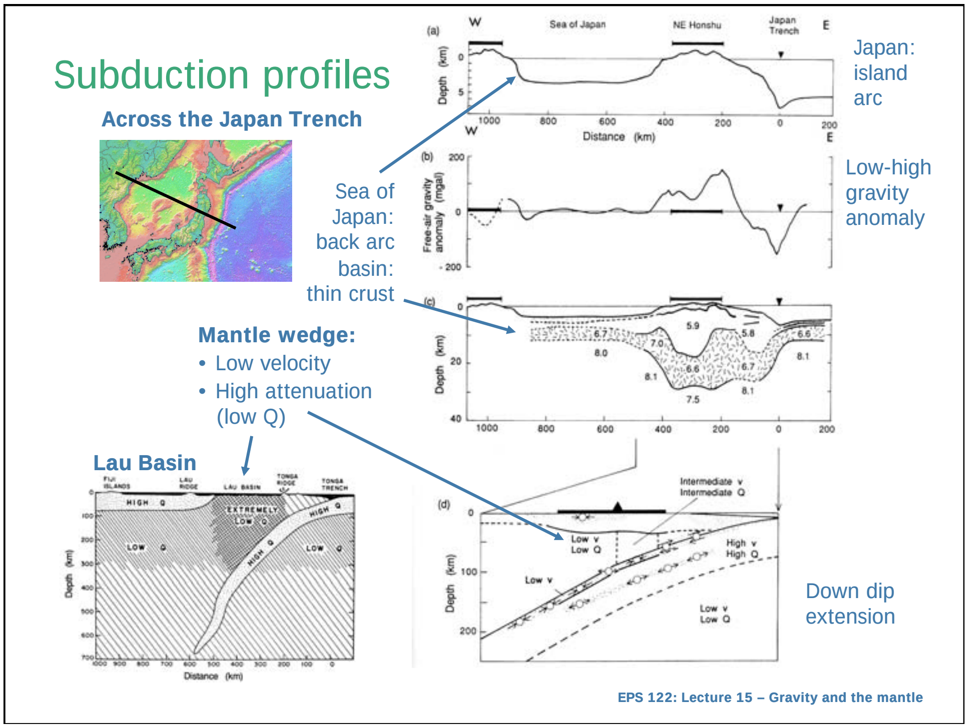
Universal gravitational constant
G=6.67 x 10^{-11} Nm^2/kg^2



Free air edge effect

1 km water (1000 kg/m^3) ➔ Δg = 42 mgal
1 km crust (2700 kg/m^3) ➔ Δg = 113 mgal

➔ 5 km crust vs. 5 km water: 565 mgal vs. 210 mgal



Subduction profiles

Across the Japan Trench

Japan: island arc

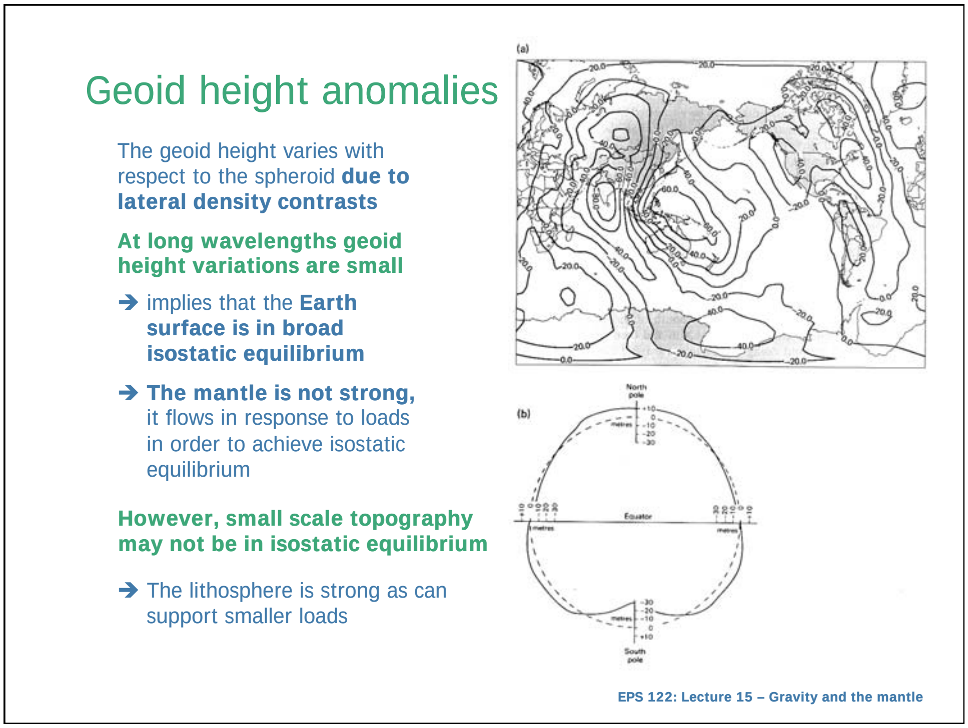
Low-high gravity anomaly

Sea of Japan: back arc basin: thin crust

Mantle wedge:
- Low velocity
- High attenuation (low Q)

Lau Basin

Down dip extension



Geoid height anomalies

The geoid height varies with respect to the spheroid **due to lateral density contrasts**

At long wavelengths geoid height variations are small

➔ implies that the **Earth surface is in broad isostatic equilibrium**

➔ **The mantle is not strong,** it flows in response to loads in order to achieve isostatic equilibrium

However, small scale topography may not be in isostatic equilibrium

➔ The lithosphere is strong as can support smaller loads

Gravity and convection

Our knowledge of the temperature and material properties of the mantle lead us to believe that there is convection

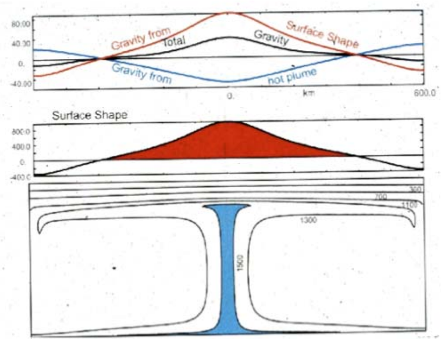
We would expect to see this in the global gravity field:

Upwelling of low density material

1. Gravity low due to low density
2. Gravity high caused by bulge due to upwelling

The winner: 2 > 1 so get bulge

See opposite effect above downwellings of cold dense material



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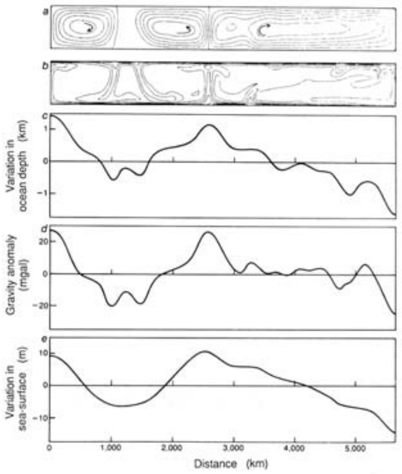


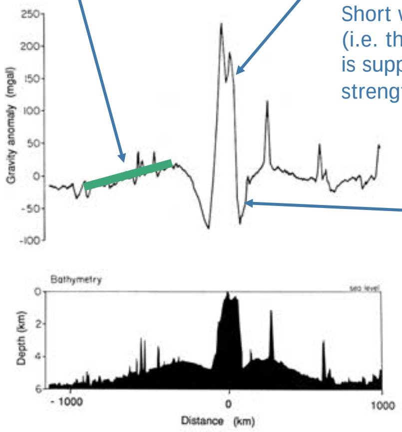
Gravity over Hawaii

Broad swell due to dynamic buoyancy

Lithosphere has no strength on this scale and is responding to dynamic forces

Gravity high due to thick islands

Short wavelength features (i.e. the islands: ~200 km) is supported by plate strength



Trough due to regional compensation

Lithospheric flexure

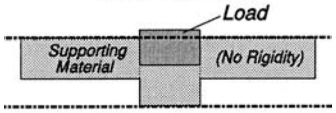
Local isostasy

- Pratt and Airy are about local isostasy
- Any load is perfectly compensated
- Therefore there is no rigidity

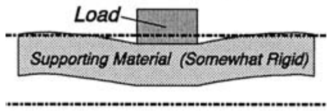
Regional isostasy

- Some of the load is supported by the strength of the lithosphere
- Isostatic compensation still occurs on a larger regional scale

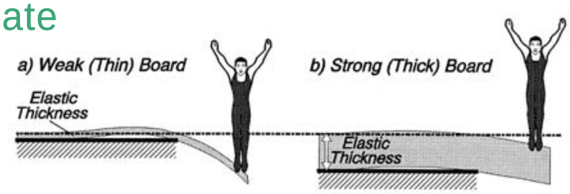
a) Local Isostasy



b) Regional Isostasy

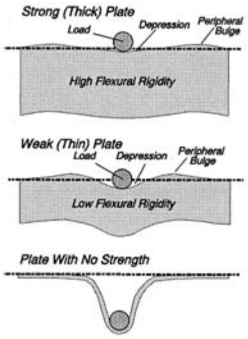


The elastic plate



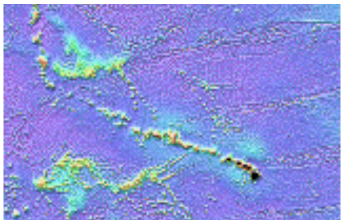
An elastic plate has strength and can be bent to support a load

- The flexural rigidity represents the strength of the plate and is dependent on the elastic thickness
- High flexural rigidity: small depression in response to a load and flexure on a long wavelength
- Low flexural rigidity: large depression and short wavelength response
- Peripheral (or flexural) bulge forms around the load
- Plates with no strength collapse into local isostatic equilibrium



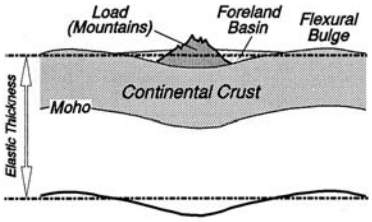
Examples

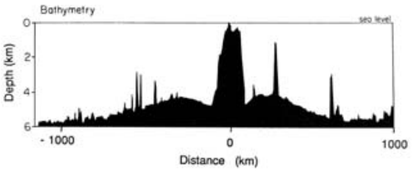
Ocean islands



Hawaii

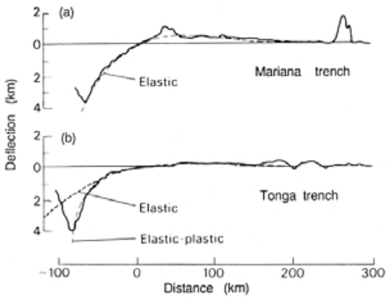
- Volcanoes load the oceanic plate causing flexure
- By modeling the shape of the flexure we can estimate the elastic thickness of the Pacific plate
- Note: there are two effects here (1) the flexure due to the island load, and (2) the bulge due to mantle upwelling





Examples

The elastic plate

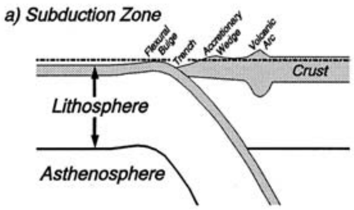


Subduction zones

- The accretionary wedge loads the end of the plate causing it to bend
- A flexural bulge is often observed adjacent to the trench

Mariana trench

- Topography matched with elastic plate, elastic thickness 28 km



Tonga trench

- Not all subduction zones can be modeled in this way eg the Tonga trench
- Instead this topography needs an elastic plate which deforms plastically once some critical yield stress is applied

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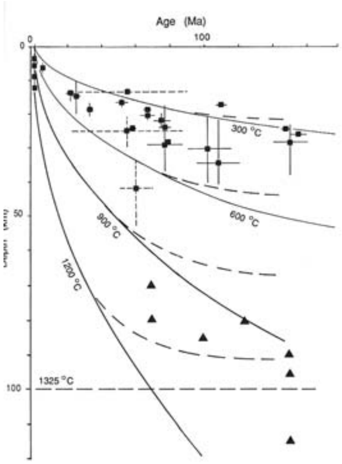
Elastic thickness of oceanic plates

and their thermal evolution



By modeling the flexure of the plates in response to loads we can estimate the elastic thickness (squares)

This shows an age dependency

- The elastic thickness increases with age and corresponds to the ~450°C isotherm
- Plate strength increases with age
- This is due to the gradual cooling of oceanic lithosphere

Elastic thickness

- **of oceans: 10-40 km**
- **of continents: typically 80-100 km**

Note: this is not the lithospheric thickness

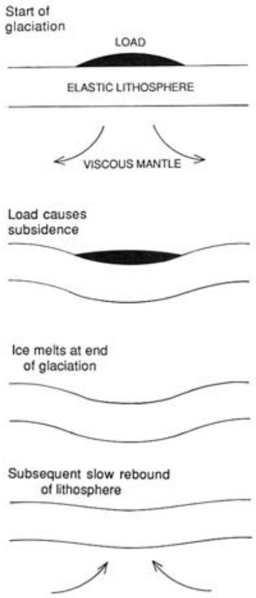
Isostatic rebound

The rate of deformation after a change in load is dependent on the flexural rigidity of the lithosphere and the viscosity of the mantle

Need a load large enough which is added or removed quickly enough to observe the viscous response of the mantle

1. Smaller loads: ~100 km diameter
tell us about uppermost mantle viscosity

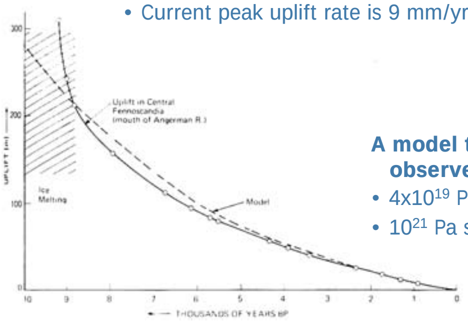
Lake Bonneville, Utah
- dried up 10,000 years ago: 300 m of water load removed
- Center of the lake has risen 65 m
- Viscosity: 10^{20} to $4x10^{19}$ Pa s for 250 to 75 km thick lithosphere



Isostatic rebound

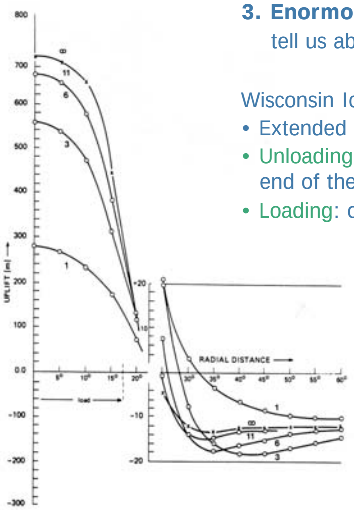
2. Larger loads: ~1000 km diameter
tell us about upper mantle viscosities

Scandinavia
- Removal of ice sheet at the end of the last ice age 10,000 years ago: ~2.5 km if ice removed
- Current peak uplift rate is 9 mm/yr



A model that satisfies the observed deformation:
- $4x10^{19}$ Pa s asthenosphere overlaying
- 10^{21} Pa s mantle

Isostatic rebound



3. Enormous loads: thousands km diameter
tell us about upper and lower mantle viscosities

Wisconsin Ice Sheet

- Extended over Northern US and Canada, up to 3.5 km thick
- Unloading: of continent due to removal of ice sheet at the end of the last ice age 10,000 years ago
- Loading: of the ocean basins due to additional water

Mantle viscosity estimates from isostatic rebound:

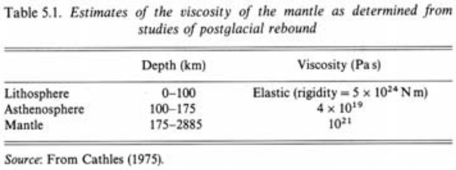


Table 5.1. *Estimates of the viscosity of the mantle as determined from studies of postglacial rebound*

	Depth (km)	Viscosity (Pa s)
Lithosphere	0–100	Elastic (rigidity = 5×10^{24} N m)
Asthenosphere	100–175	4×10^{19}
Mantle	175–2885	10^{21}

Source: From Cathles (1975).