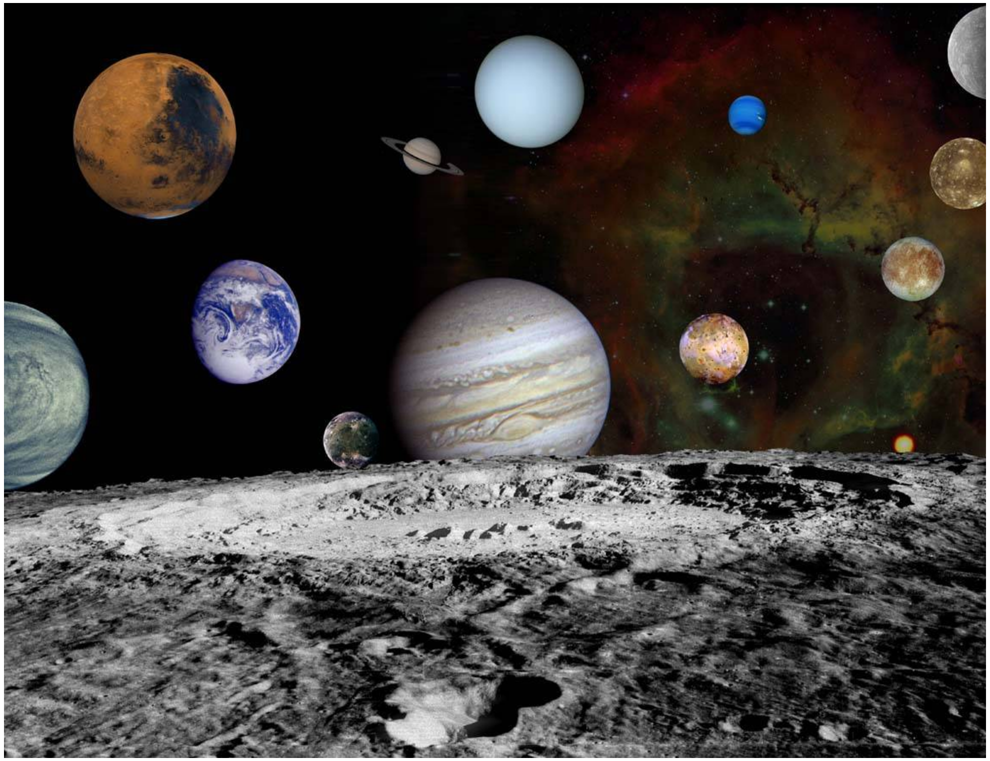
Surface features and thermal structure of a planet

Observations : 1- impact craters (definitions)

2- volcanism

3- lithospheric flexure and thermal structure



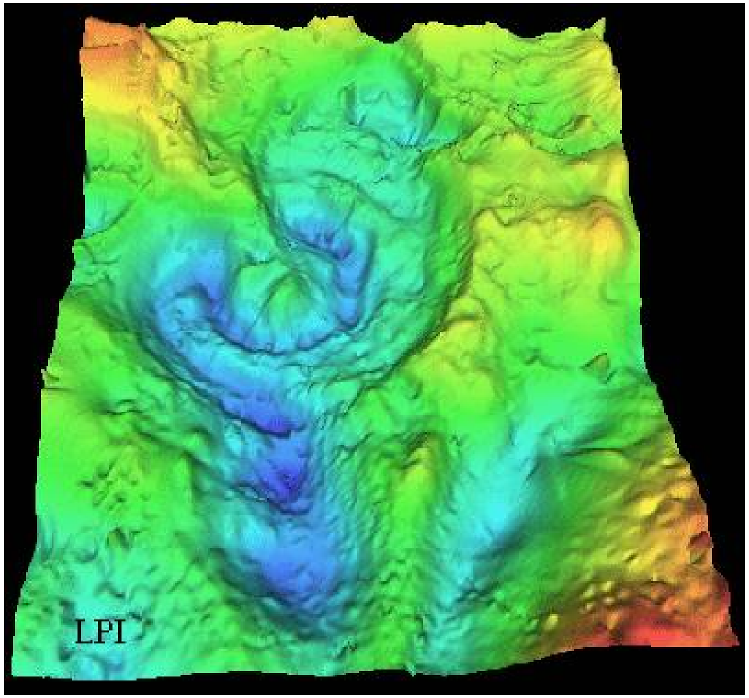
- What kinds of tectonic features are observed on planetary bodies, and what do they imply?
- How are loads on planetary bodies supported?

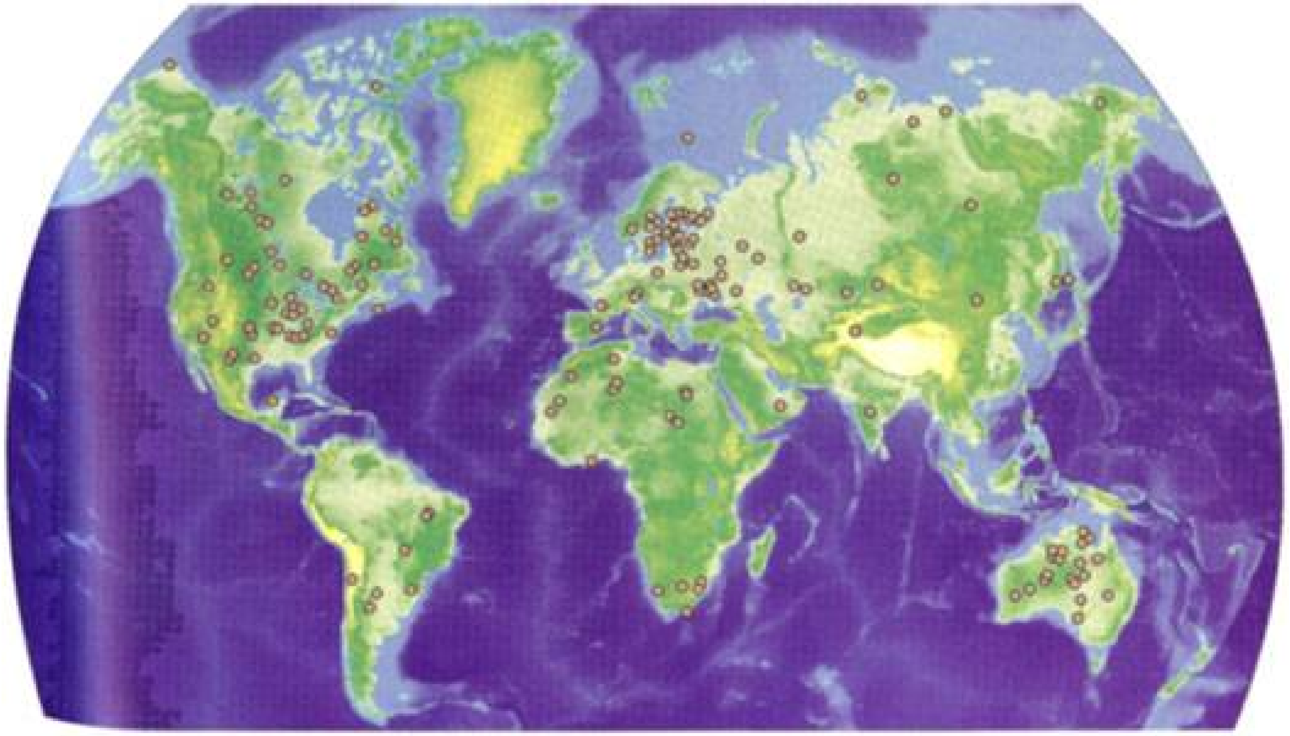
Aknowledgements:

Henrik Hargitai

- Crater morphologies
  - Morphologies of impacts rim, ejecta etc
  - Energies involved in the impact process
  - Simple vs. complex craters

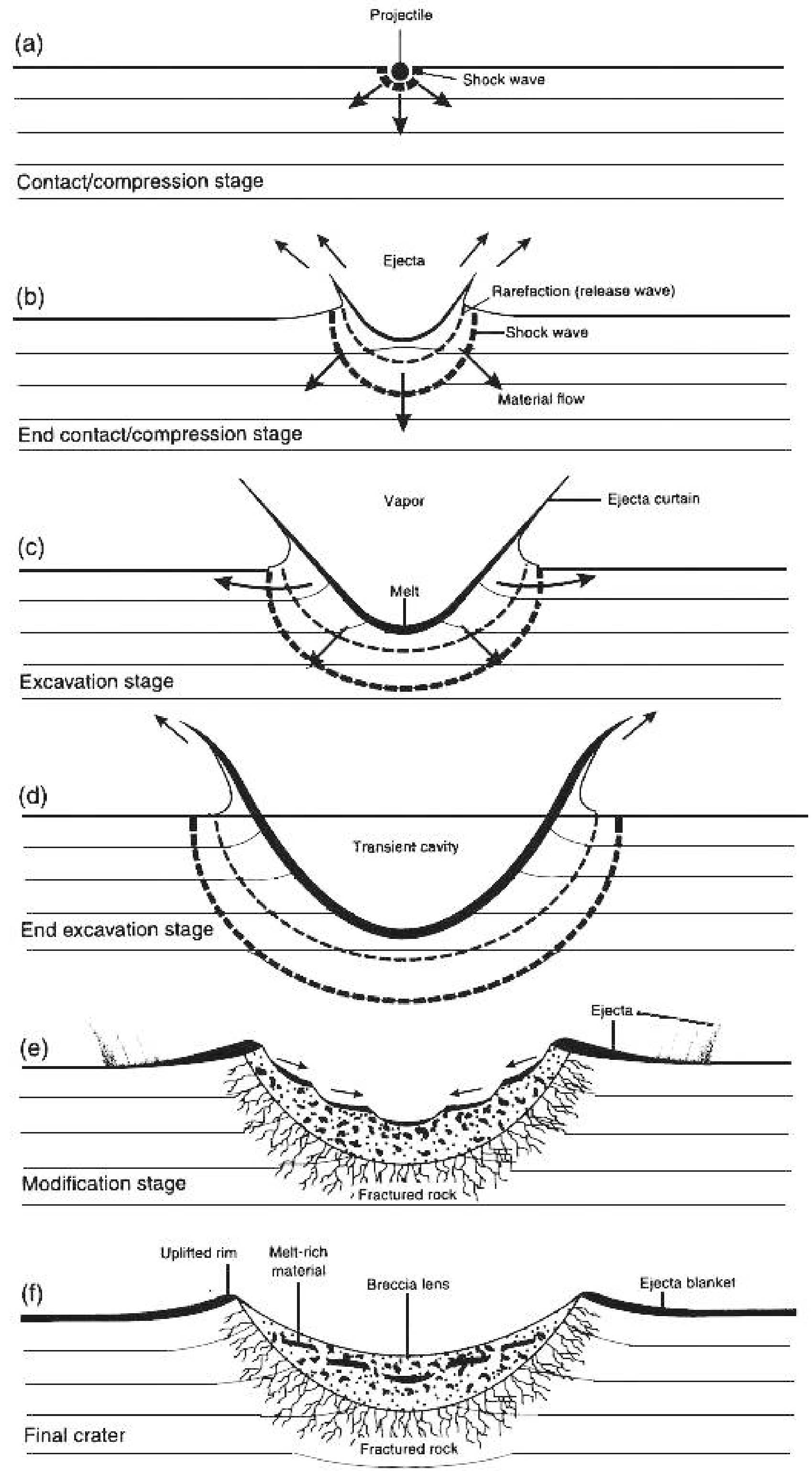
- Cratering mechanics
  - Contact and compression stage
  - Ejection and excavation stage
  - Collapse and modification stage





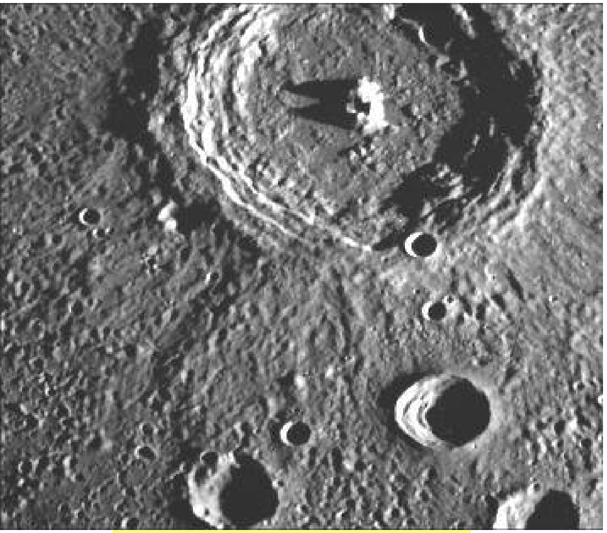
Formation stages

- Contact
- compression
- Excavation
- Modification

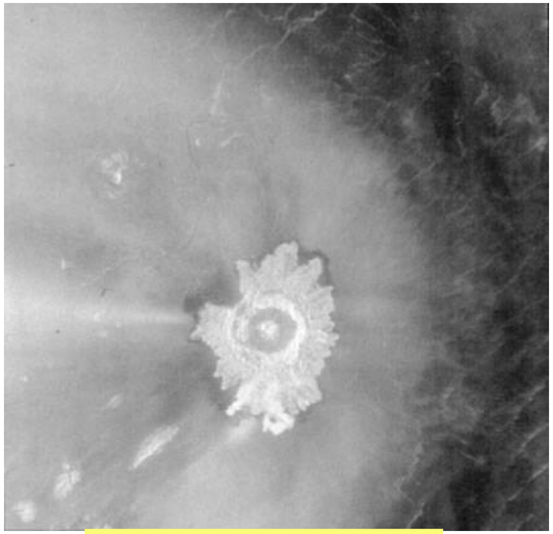


Where do we find craters? – Everywhere!

- Cratering is a geologic process that every solid solar system body experiences…



Mercury



Venus

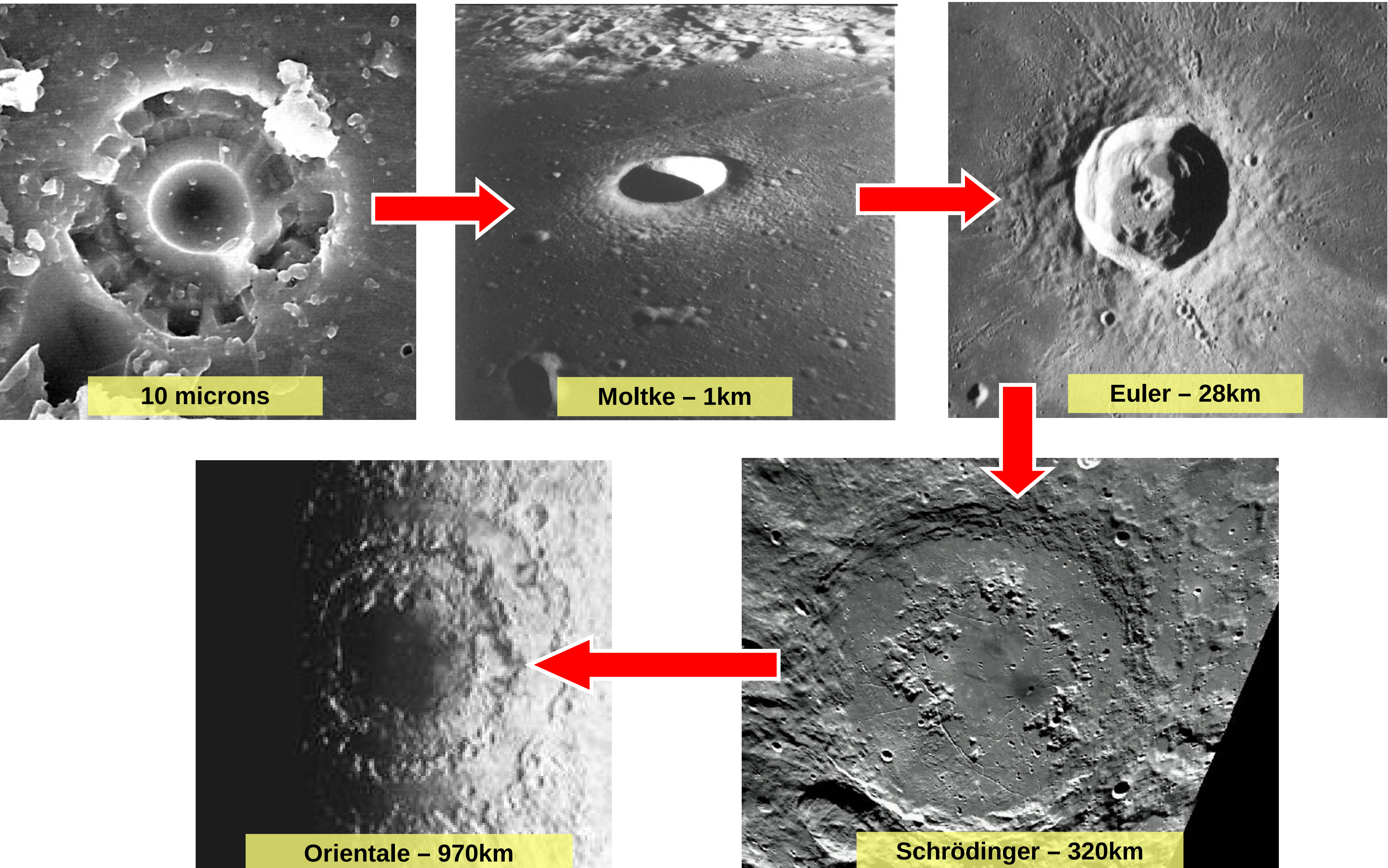
Moon

Earth

Mars

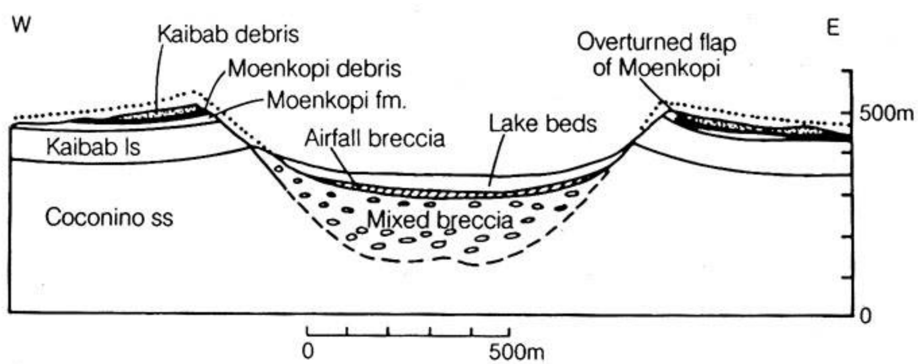
Asteroids

- Morphology changes as craters get bigger
  - Pit → Bowl Shape → Central Peak → Central Peak Ring → Multi-ring Basin



Craters are products of point-source explosions

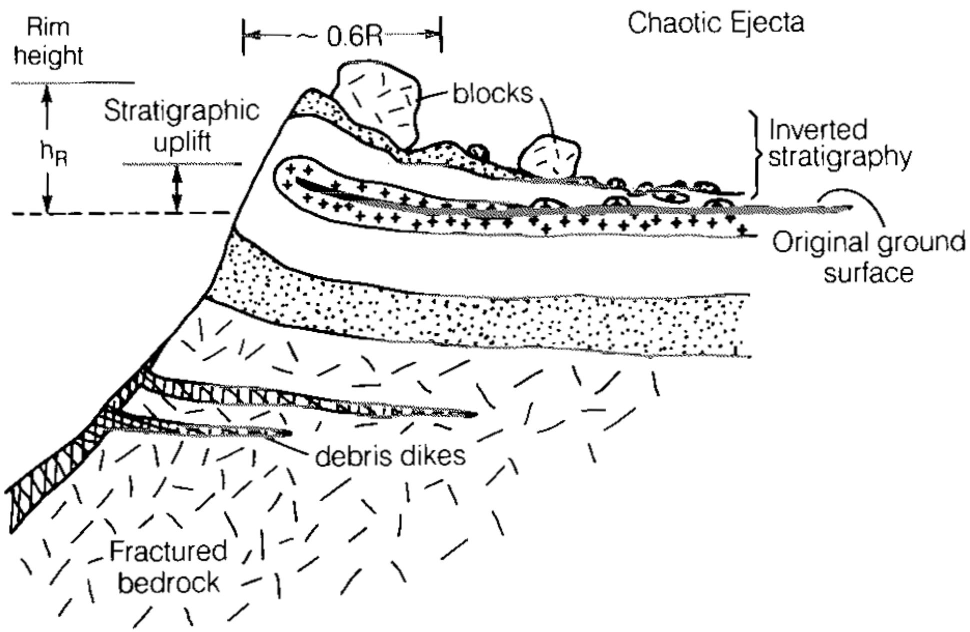
- Oblique impacts still make round craters



Meteor Crater – 1.2 km

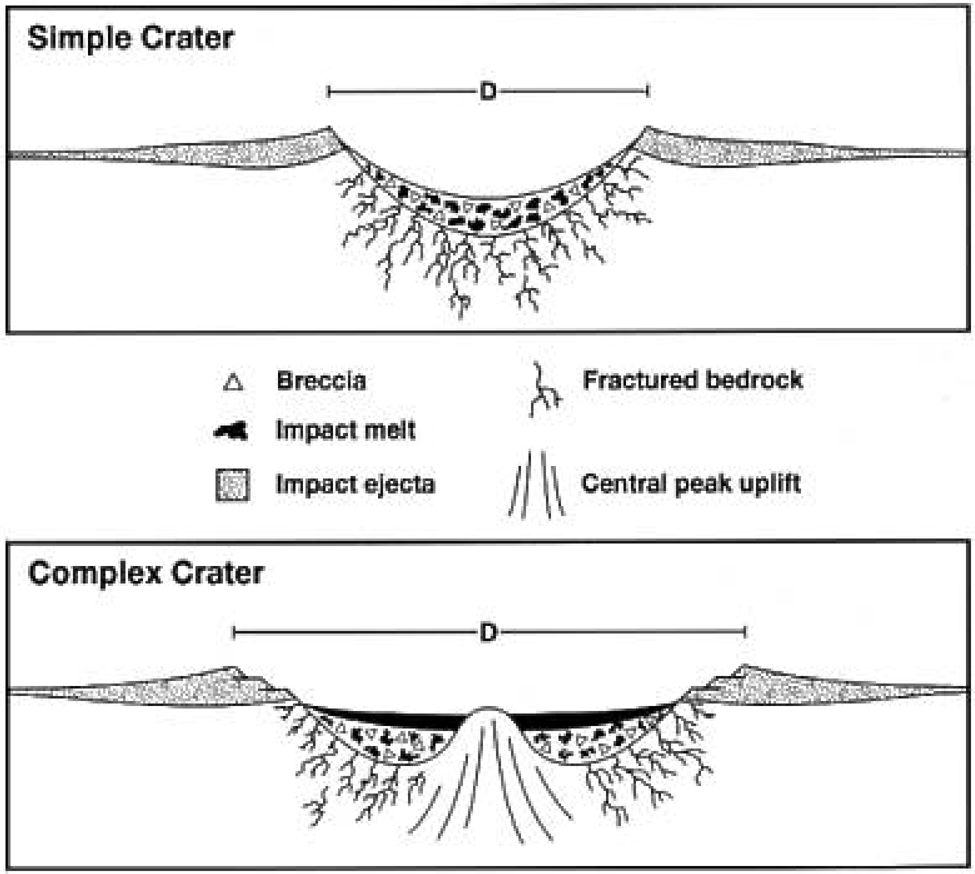
Sedan Crater – 0.3 km

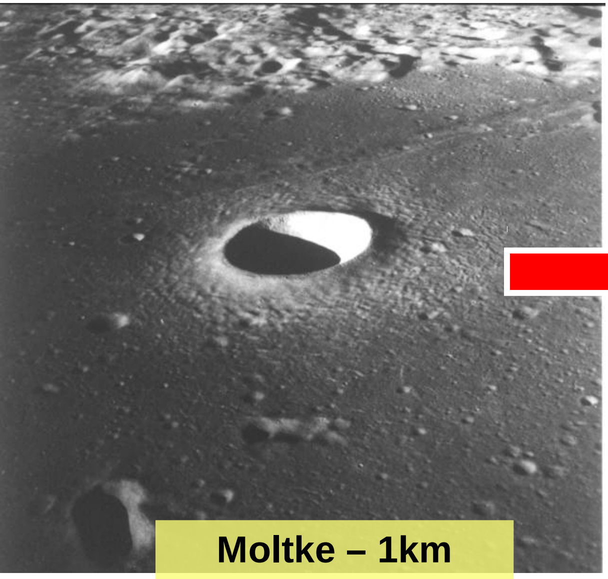
- **Overturned flap at edge**
  - **Gives the crater a raised rim**
  - **Reverses stratigraphy**
- **Eject blanket**
  - **Continuous for ~1 R_c**
- **Breccia**
  - **Pulverized rock on crater floor**
- **Shock metamorphosed minerals**
  - **Shistovite**
  - **Coesite**
- **Tektites**
  - **Small glassy blobs, widely distributed**

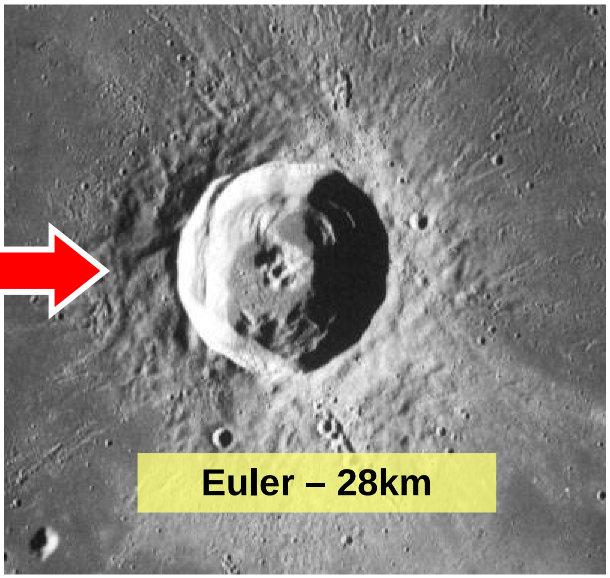


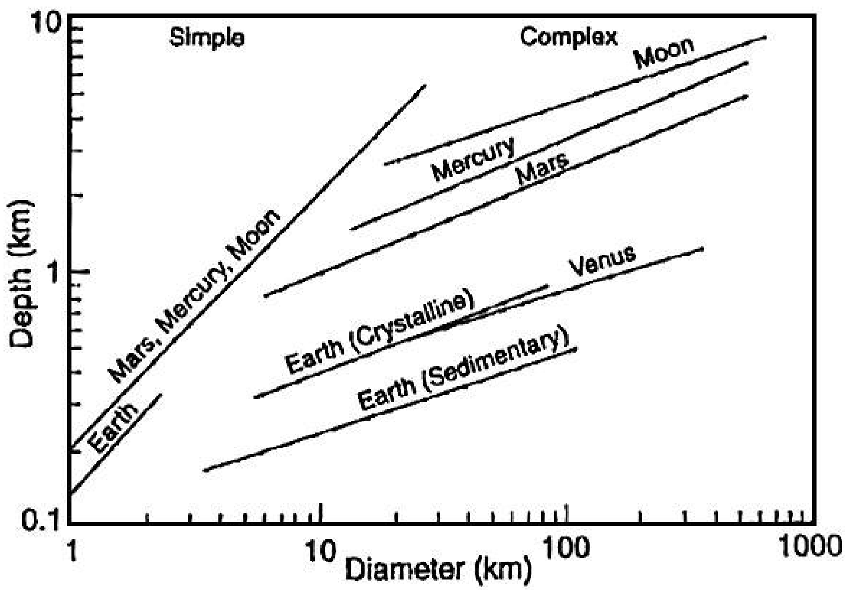
Differences in simple and complex morphologies

Simple	Complex
Bowl shaped	**Flat-floored** **Central peak** **Wall terraces**
Little melt	**Some Melt**
d/D ~ 0.2	**d/D much smaller** **Diameter dependent**
Small sizes	**Larger sizes**

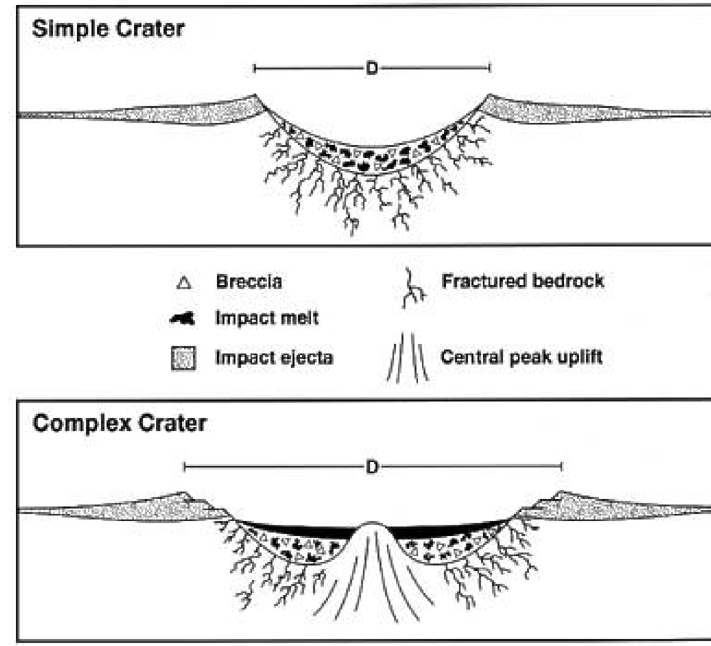


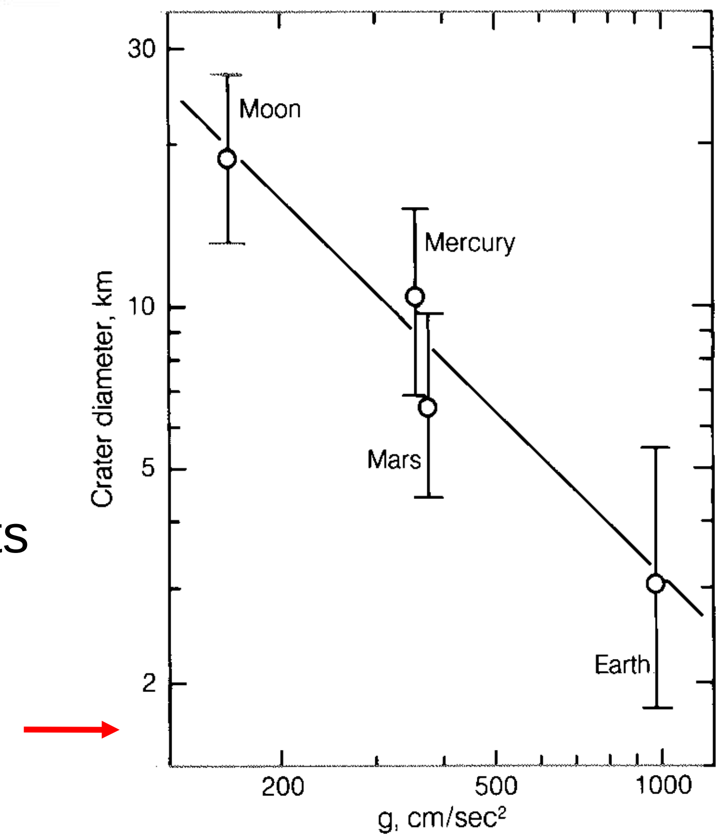




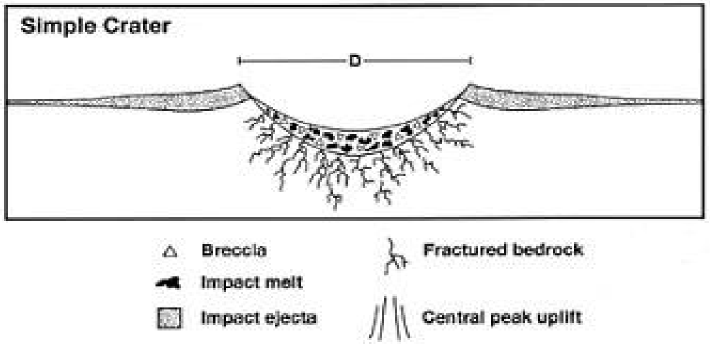


- craters start as a transient hemispheric cavity
- Simple craters
  - Most material pushed downwards
  - Size of crater limited by strength of rock
  - Energy ~ $\left(23\pi\, r^3\right) Y$
- Complex craters
  - Size of crater limited by gravity
  - Energy ~ $\left(23\pi\, r^3\right) \rho g\, D$
- Intermediate diameter
  - i.e. Energy ~ $\left(23\pi\, r_{T^3}\right) Y \sim \left(23\pi\, r_{T^3}\right) \rho g\, D_T$
  - So: $Y \approx \rho g\, D_T$ or $D_T \approx Y \rho g$
  - The transition diameter is higher when
    - The material strength is higher
    - The density is lower
    - <u>The gravity is lower</u>
  - Y ~ 100 MPa and ρ ~ 3×10^3 kg m^{-3} for rocky planets
  - D_T is ~3km for the Earth and ~18km for the Moon
    - Compares well to observations

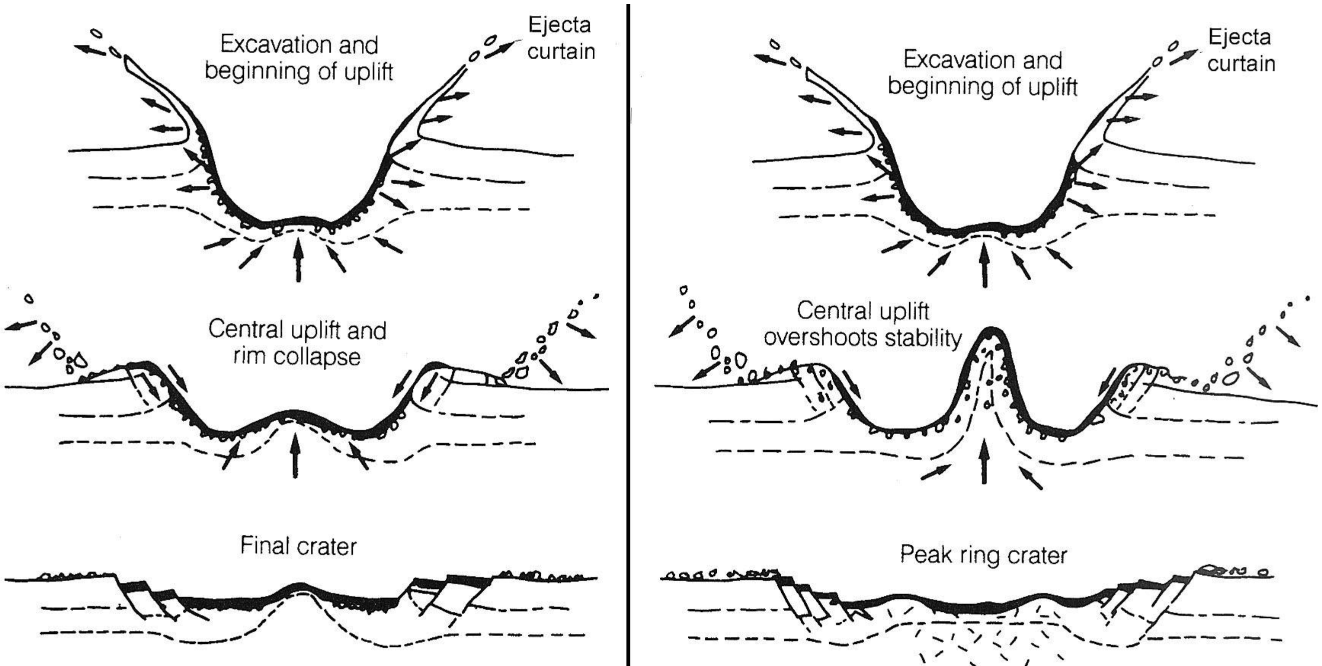




- Previous stages produces a hemispherical transient crater
- Simple craters collapse from d/D of ~0.5 to ~0.2
  - Bottom of crater filled with breccia
  - Extensive cracking to great depths



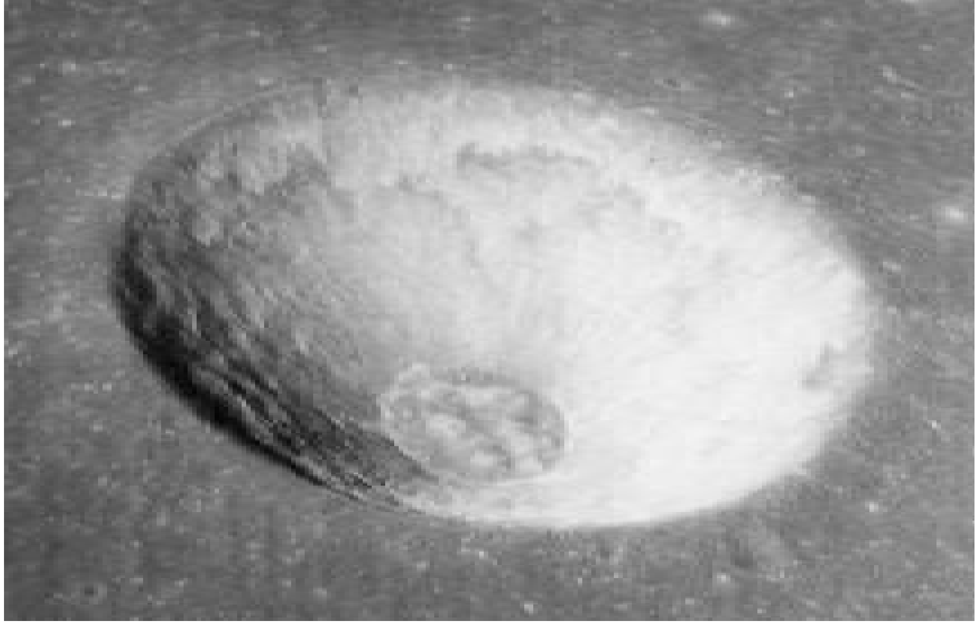
- Peak versus peak-ring in complex craters
  - Central peak rebounds in complex craters
  - Peak can overshoot and collapse forming a peak-ring
  - Rim collapses so final crater is wider than transient bowl
  - Final d/D < 0.1

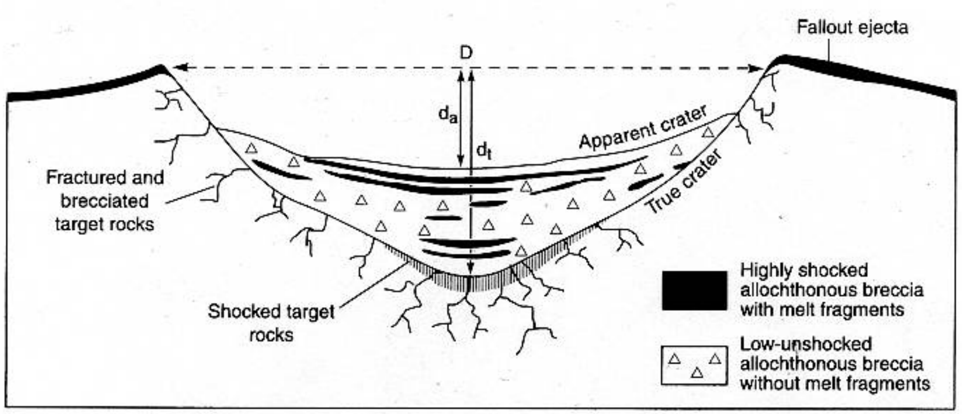


Melosh, 1989

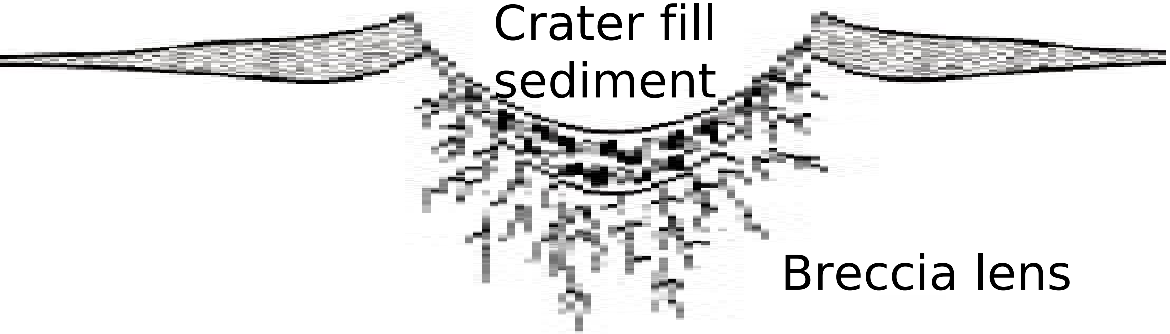
Simple Crater

- Small (3-10 km)
- Bowl-shaped
- Ejecta blanket extends one D from rim
- Rim height 4% of D





Ejecta blanket



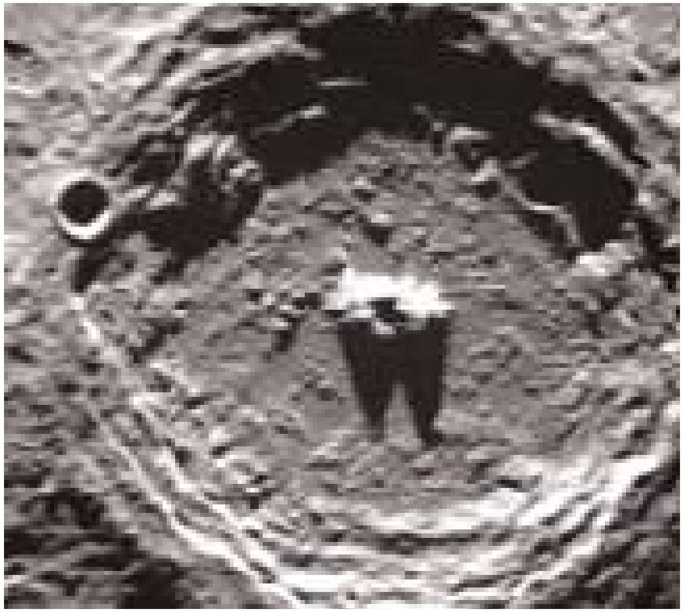
Simple Craters on Earth and Moon

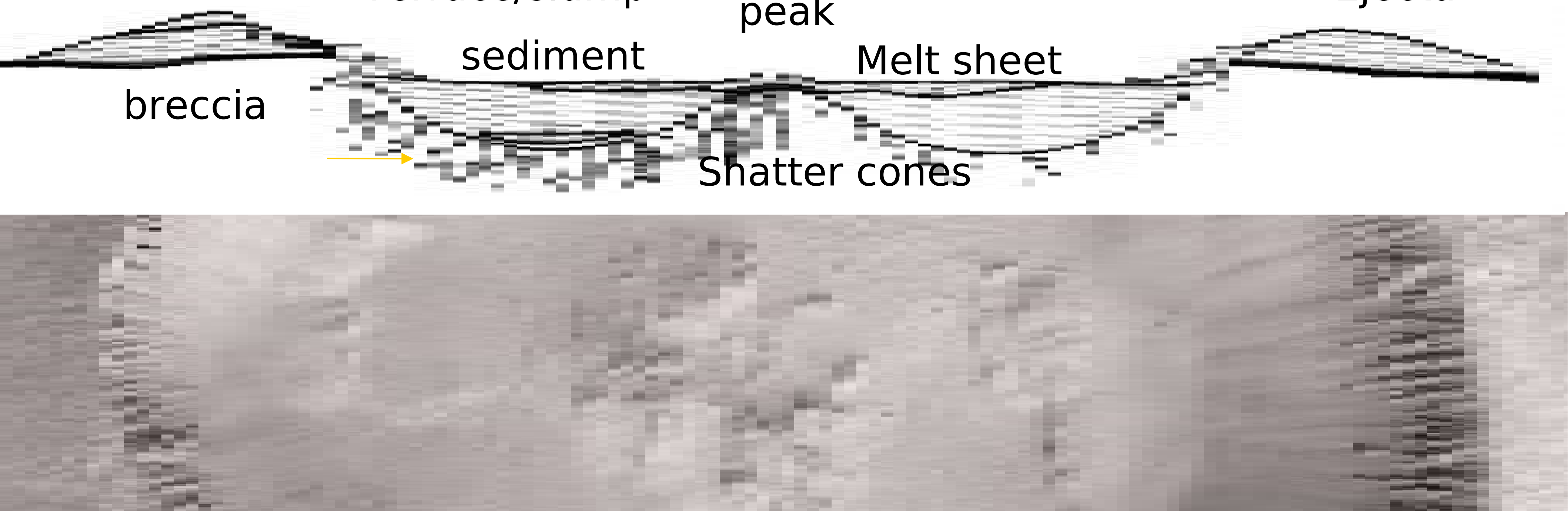
- First to be identified on Earth
- Not always completely circular
  - Faults
- Common at D= 3 km to 6 km

- Moltke crater, a simple crater, was photographed by Apollo 10 astronauts in 1969. The depression, about 7 km (4.3 miles) in diameter.
- Common up to D=15 km

Complex craters

- Elastic rebound
- Central peak (structural uplift [SU])
- Ring depression (flat floor/annular basin
- Depth increases slowly
  -d=3 - 6 km, D= 20 - 400 km



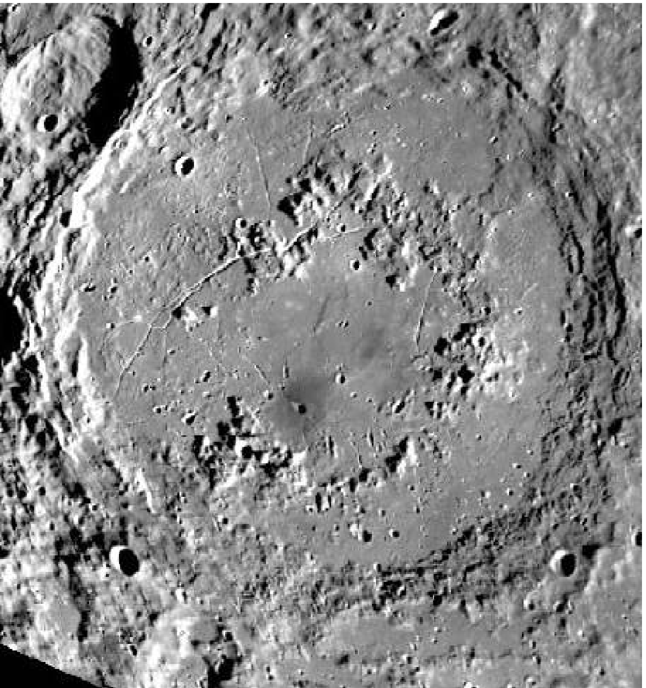


Central ring craters

- Complex crater with internal ring
- >4 km on Earth
- Transition to central ring at approx 140 km diameter on Moon
- Still follows the g^{-1} rule
- Central ring generally about 1/2 rim-to-rim D for terrestrial planets

W Clearwater, Québec, Canada

Schrödinger, Moon



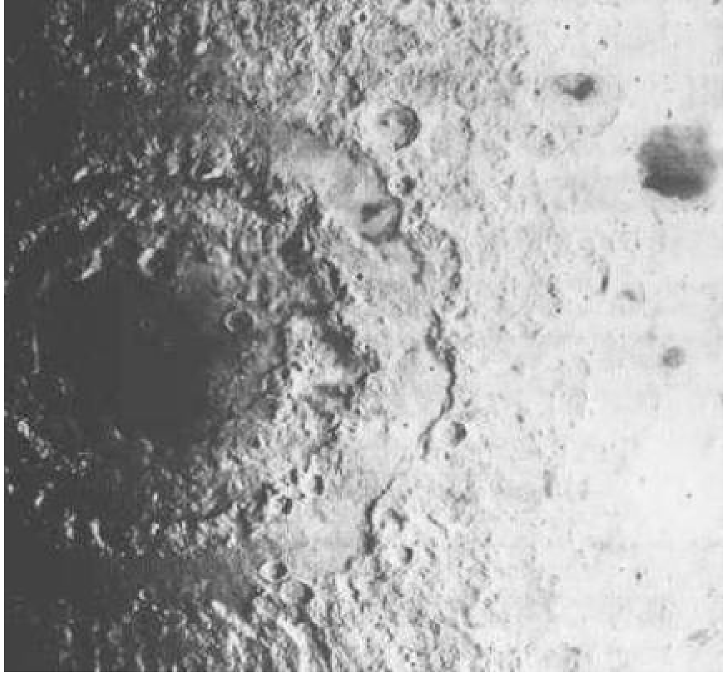
Lowell, Mars

Barton, Venus

Giant Multiringed Basins

- Impact-related inner,
- Tectonics related outer rings
- Lava-fill possible
- Mare Orientale Moon

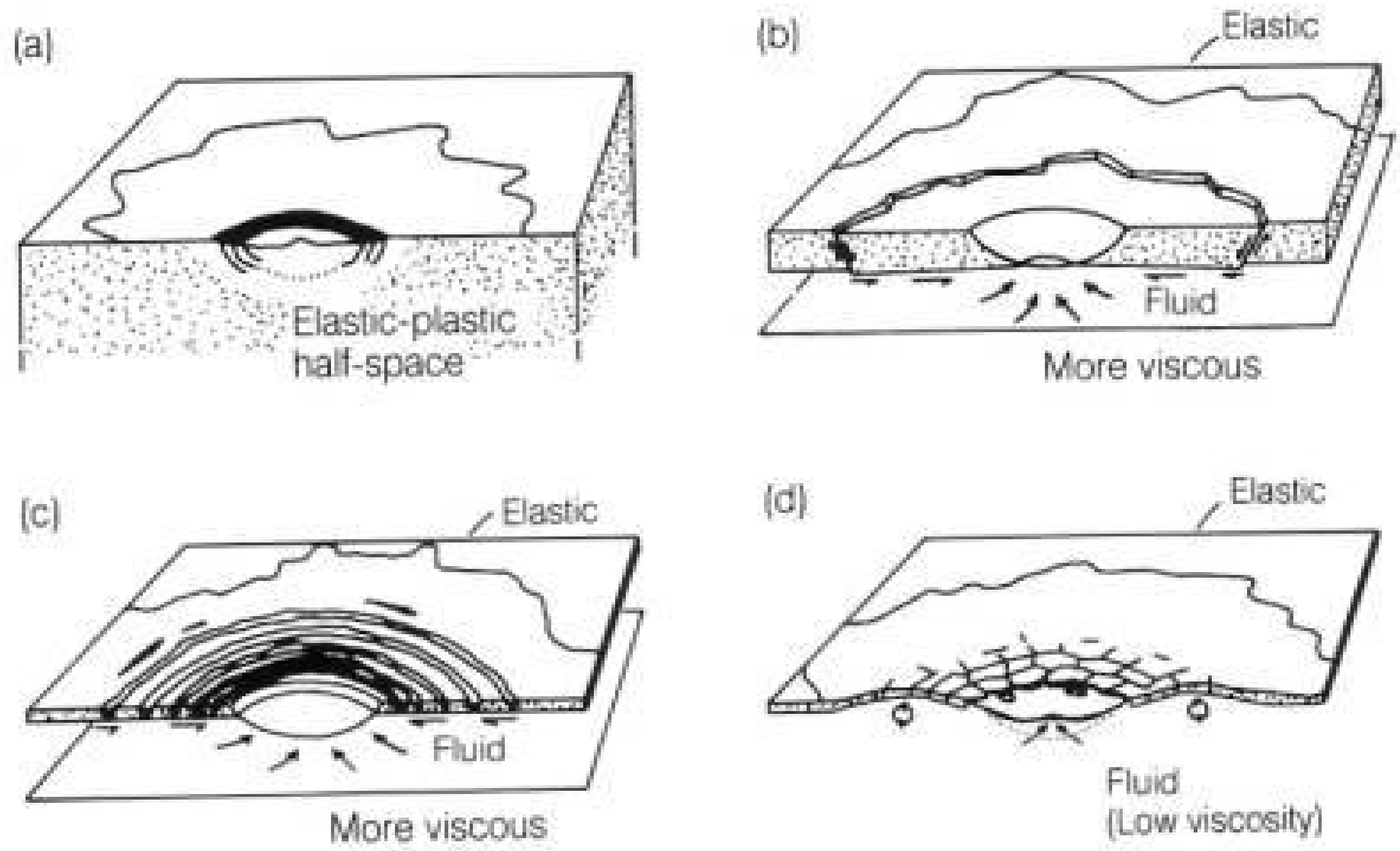
  - D~ 930km, d~2km
- Valhalla-type D~4000km
  - 20+ rings
  - Young elastic thin crust
  - Global effects



Mare Orientale, Moon

Valhalla, Callisto

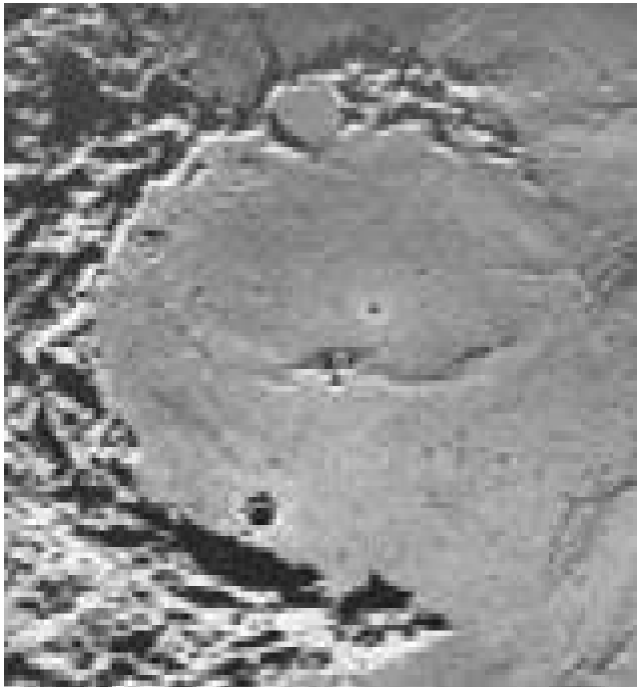
Multiring Schematic

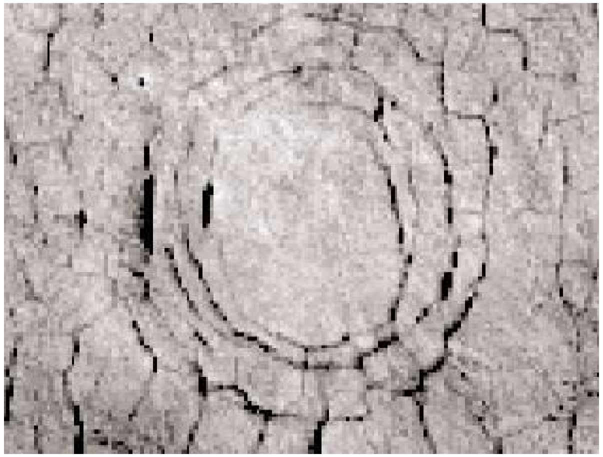


The ring tectonic theory suggests that in layered media in which the strength decreases with increasing depth, one or more ring fractures arise outside the rim of the original crater (figure 5) (Melosh and McKinnon, 1978). This suggests that for the formation of multiring basins to occur there must be a high brittle-ductile thickness ratio in the impacted material i.e. where thick crust exists over a deeper ductile layer (Allemand and Thomas, 1999). **www.spacechariots.biz/ creaters.htm**

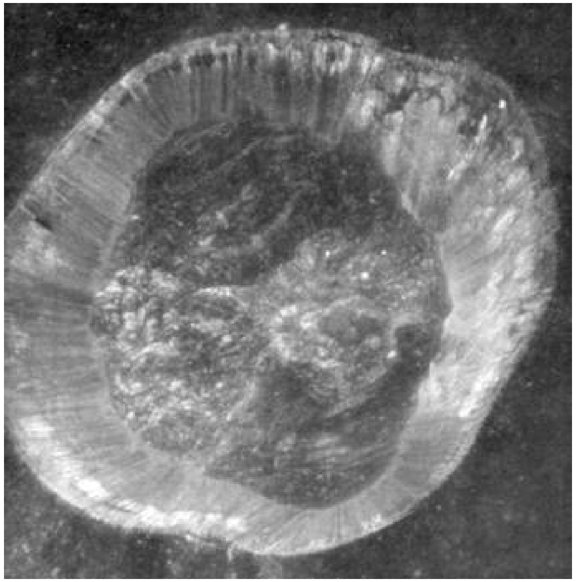
Flat floor and eroded craters

- „walled plains”
- Ghost or eroded
- Lava or sediment filled





Crater under
ice polygons (Mars)

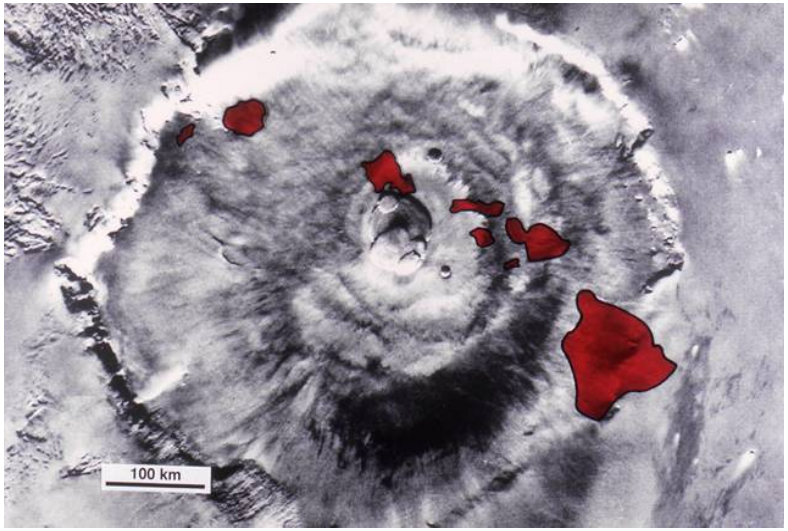


Dawes-type

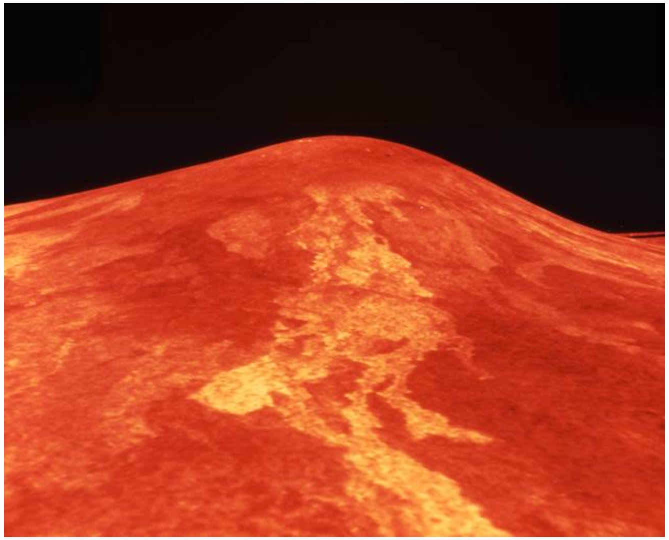
2- Volcanism

- Volcanism is an important process on most solar system bodies (either now or in the past)
- It gives information on the *thermal evolution* and *interior state* of the body

Hawaii

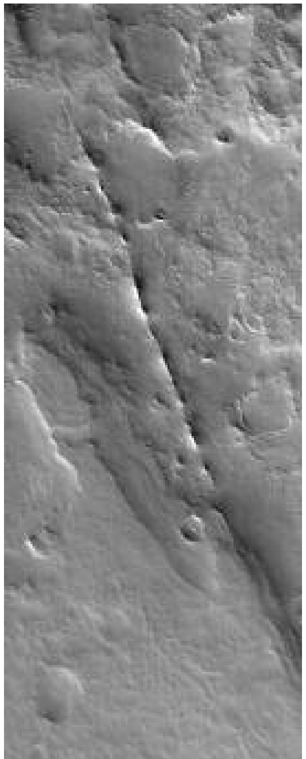


Olympus Mons,
Mars



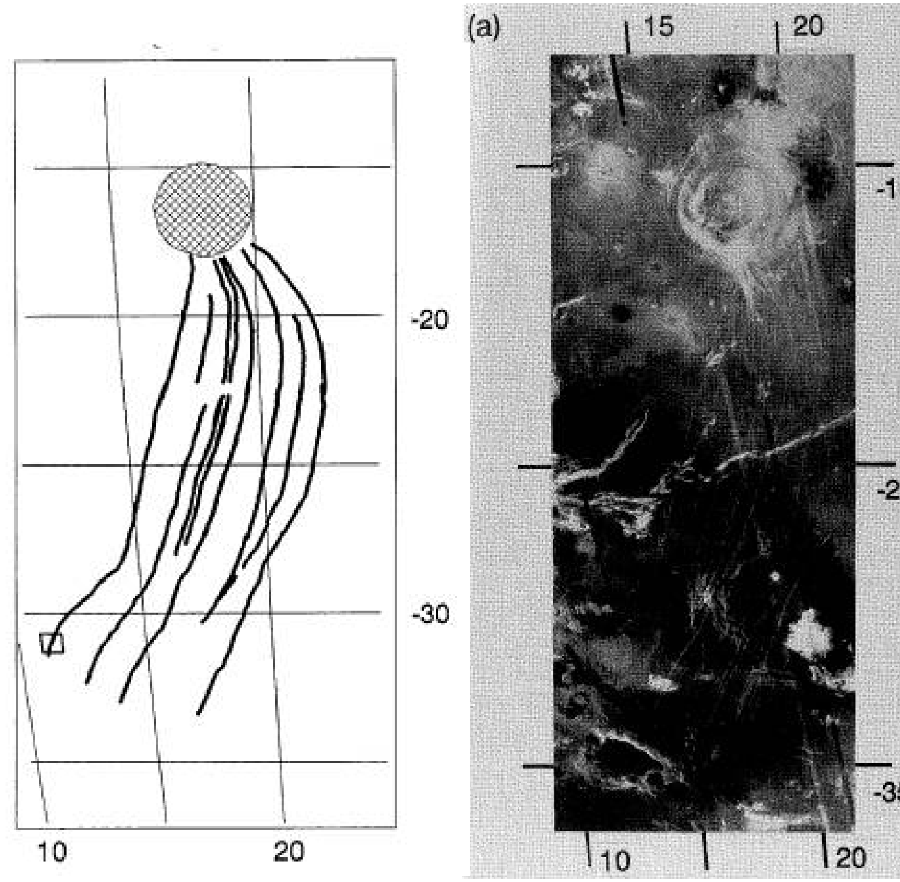
Sif Mons
(Venus)
2km x 300km

Dikes

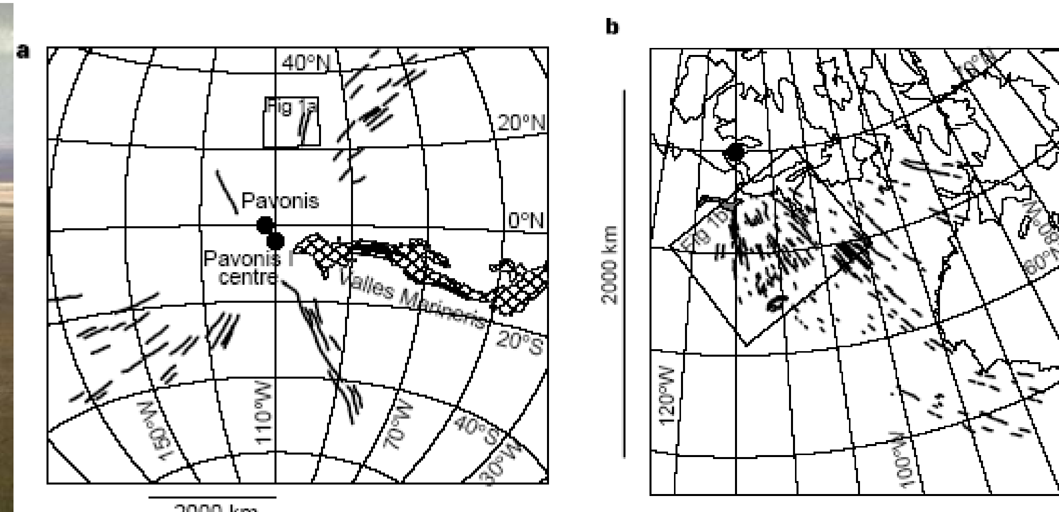


Mars image width 3km
MOC2-1249

Ship Rock, 0.5km high
New Mexico



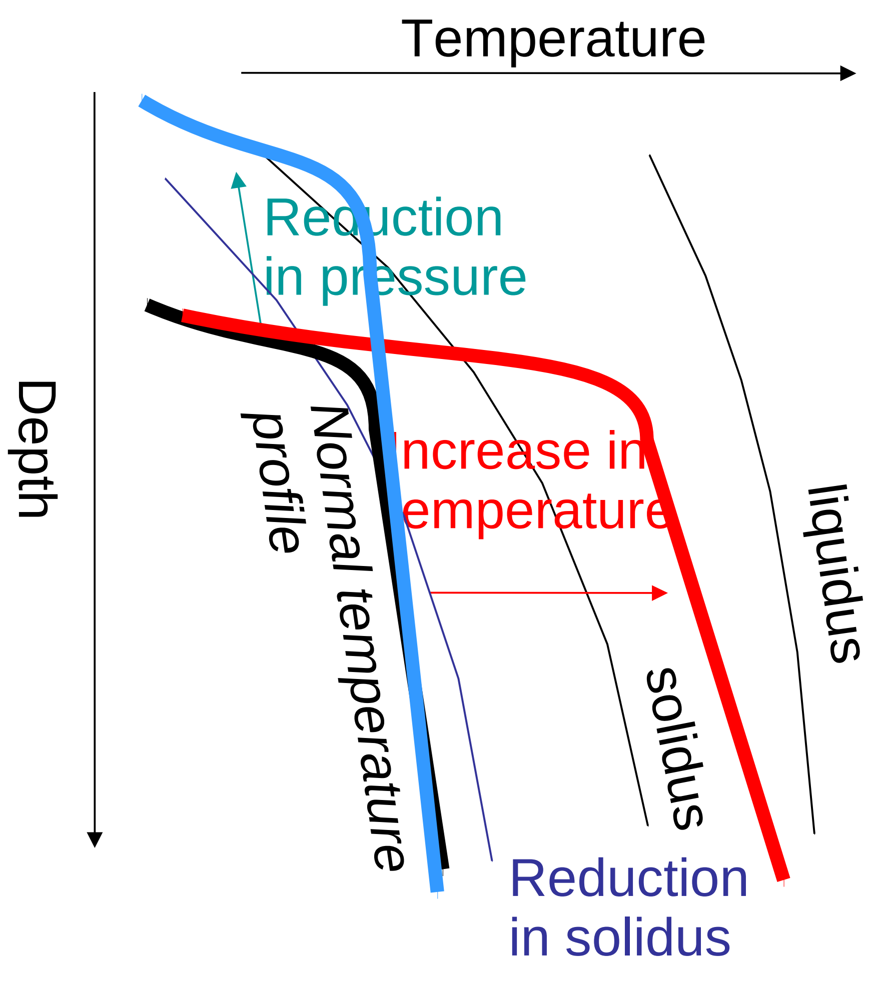
Radiating dike field, Venus



Dike Swarms, Mars and Earth

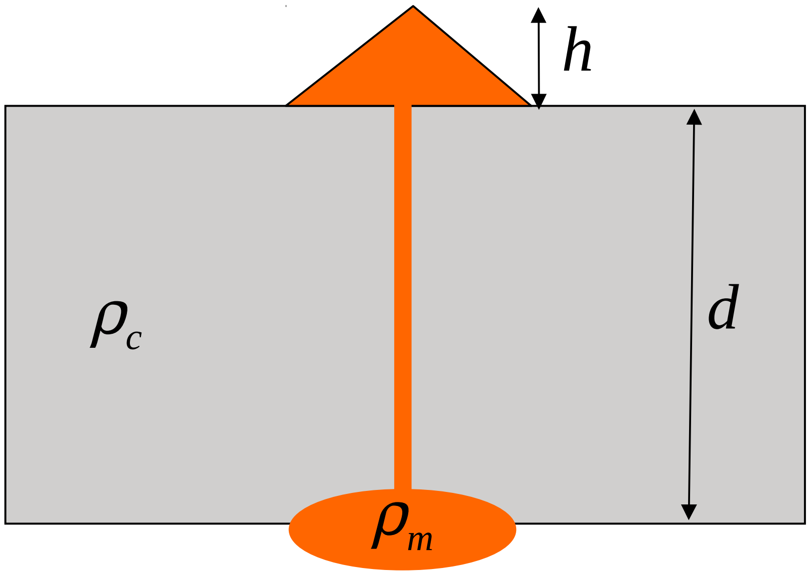
Why does it happen?

- Material raised above the melting temperature (solidus)
  - Increase in temperature (plume e.g. Hawaii)
  - Decrease in pressure (mid-ocean ridge)
  - Decrease in solidus temperature (hydration at island arcs)



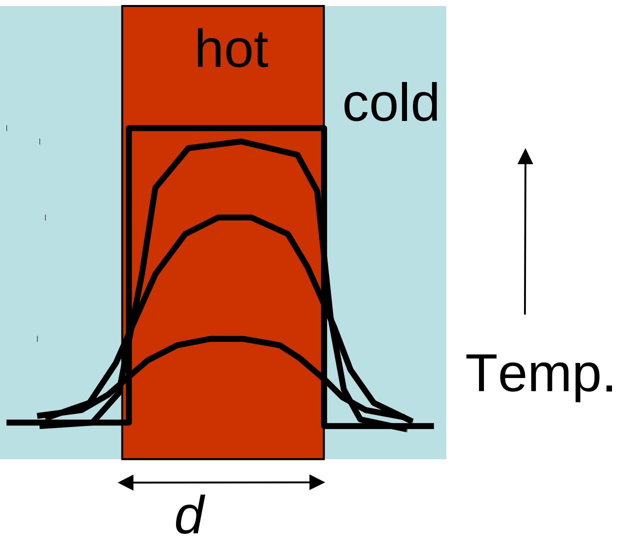
Eruptions

- Magma is often less dense than surrounding rock (why?)
- So it ascends (to the level of neutral buoyancy)
- For low-viscosity lavas, dissolved volatiles can escape as they exsolve; this results in gentle (effusive) eruptions
- More viscous lavas tend to erupt explosively
- We can determine maximum volcano height:



$$h = d(\rho_c - \rho_m)/\rho_m$$

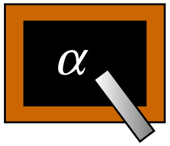
Cooling timescale



- Conductive cooling timescale depends on thickness of object and its *thermal diffusivity* κ

  Typical value for rock/ice is $10^{-6}\ m^2s^{-1}$

- How long does it take a meter thick lava flow to cool?
- How long does it take the Earth to cool?



- Characteristic cooling timescale $\boxed{t \sim d^2/\kappa}$
- *Demo:* $\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{z^2}$

* Si *T* change à une distance typique *L* - par exemple: *T = sin (z/L)*,
l’expression a droite est proportionelle à κ/L^2

* Si le changement de temps est à l’échelle de τ secondes, par exemple - *cos t/τ*,
on trouve à gauche un proportionalité avec 1/τ.

Ces oeufs de macareux ont deux fois le diamètre d’un oeuf de poule. Si j’aime les oeufs de poule cuits pendant 3 minutes, combien de temps m’avisez-vous de bouillir ceux-ci?

- $\sqrt{2} \times 3 = 4{,}2$ min
- $\sqrt{3} \times 3 = 5{,}2$ min
- $2 \times 3 = 6$ min
- $2^2 \times 3 = 12$ min

- La distance L est deux fois plus grande
- Le temps de cuisson est proportionel à L^2
- Le temps de cuisson est donc $2^2 = 4$ fois plus long: 12 min

3- Flexure and Elasticity

- The near-surface, cold part of a planet (lithosphere) behaves elastically
- This lithosphere can support loads (e.g. volcanoes,meteoritic impacts)
- We can use observations of how the lithosphere deforms under these loads to assess how thick it is
- The thickness of the lithosphere tells us about how rapidly temperature increases with depth i.e. it helps us to deduce the thermal structure,
- The deformation of the elastic lithosphere under loads is called flexure.

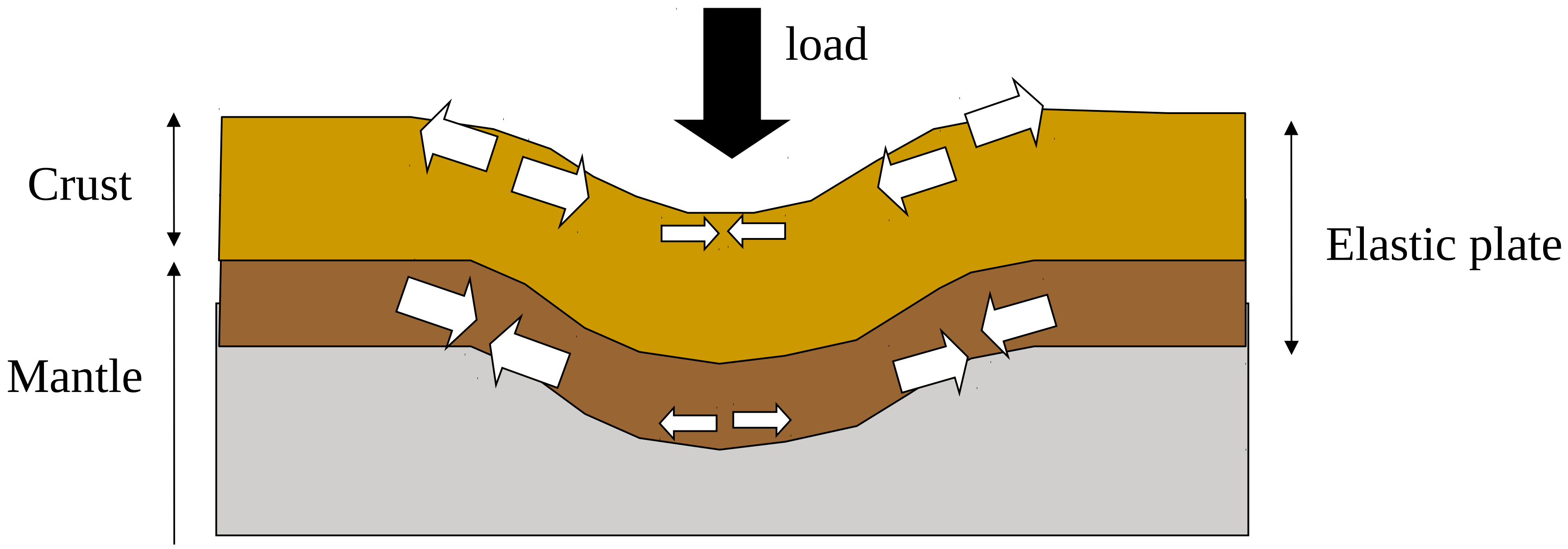
Stress and strain

- For many materials, stress is proportional to strain (Hooke's law); these materials are elastic
- Stress required to generate a certain amount of strain depends on Young's modulus E (large E means rigid)
- You can think of Young's modulus (units: Pa) as the stress σ required to cause a strain of 100%

$$\sigma = E\varepsilon$$

- Typical values for geological materials are 100 GPa (rocks) and 10 GPa (ice)
- Elastic deformation is reversible; but if strains get too large, material undergoes fracture (irreversible)

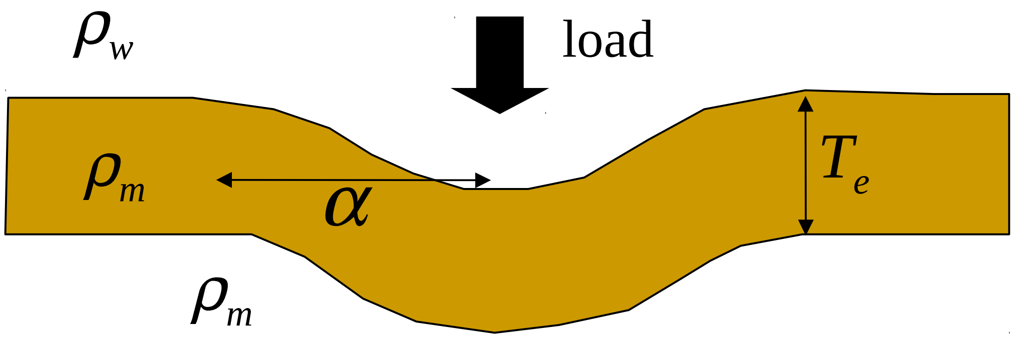
Flexural Stresses



- In general, a load will be supported by a combination of elastic stresses and buoyancy forces (due to the different density of crust and mantle)
- The elastic stresses will be both compressional and extensional

Flexural Parameter

- Consider a load acting on an elastic plate:



- The plate has a particular elastic thickness T_e
- If the load is narrow, then the width of deformation is controlled by the properties of the plate
- The approximate width of deformation α is called the flexural parameter and is given by

$$\boxed{\alpha = \left[\frac{E T_e^3}{3g(\rho_m - \rho_w)(1-\nu^2)}\right]^{\frac{1}{4}}}$$

Here E is Young's modulus, g is gravity and ν is Poisson's ratio (~0.3)

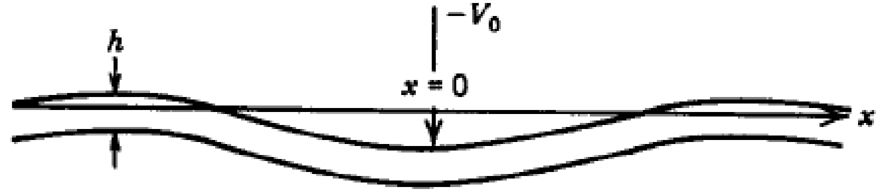


Figure 3-29 Deflection of the elastic lithosphere under a line load.

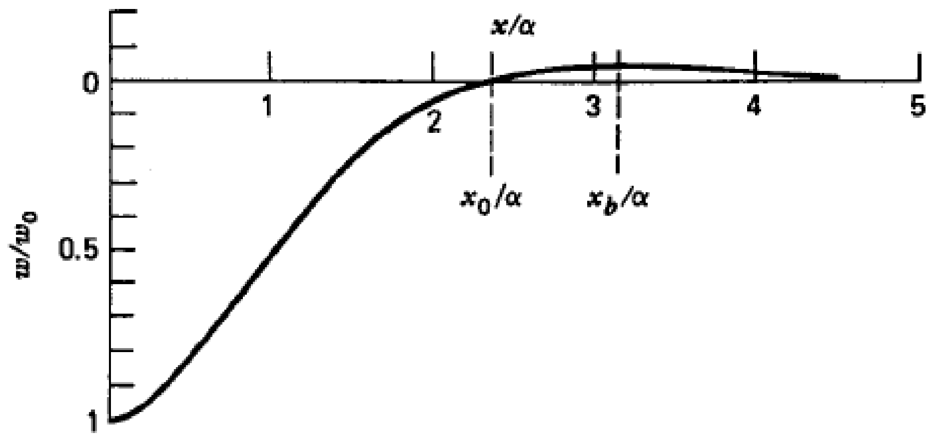
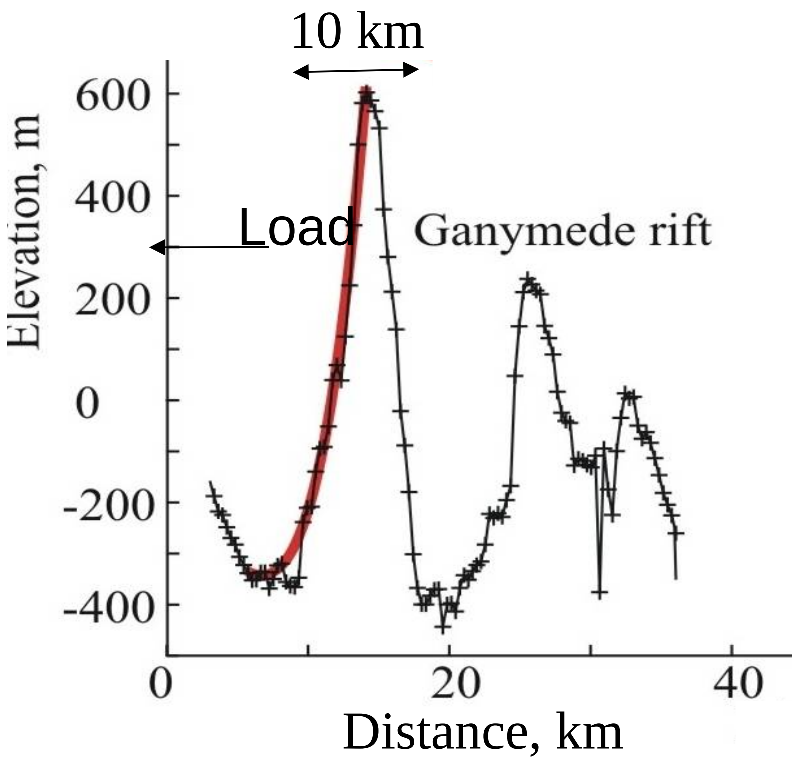


Figure 3-30 Half of the theoretical deflection profile for a floating elastic plate supporting a line load.

- Technically, the zero crossing $x_o = 3\pi\alpha/4$, and the forebulge maximum is at $x_b = \pi\alpha$.
- Therefore, α is less than the width of deformation.

- If the applied load is much wider than α, then the load cannot be supported elastically and must be supported by buoyancy (isostasy)
- If the applied load is much narrower than α, then the width of deformation is given by α
- If we can measure a flexural wavelength, that allows us to infer α and thus T_e directly.
- Inferring T_e (elastic thickness) is useful because T_e is controlled by a planet's temperature structure

Example



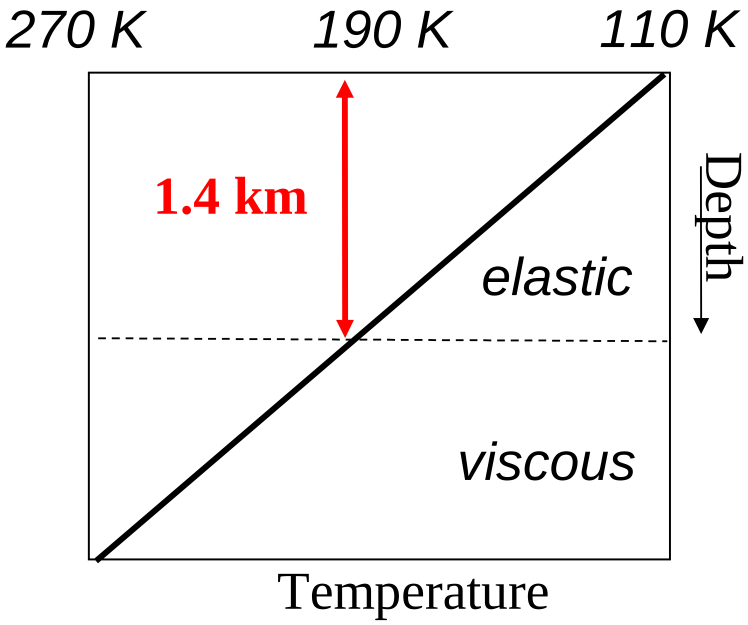
- Take this profile across a rift on Ganymede
- An eyeball estimate of $\alpha \sim 10$ km
- For ice, we take E=10 GPa, $\Delta\rho$=900 kg m^{-3}, g=1.3 ms^{-2}

- If α=10 km then T_e=1.5 km
- A numerical solution gives T_e=1.4 km
- This is useful since T_e is controlled by the temperature of the subsurface

T_e and temperature structure

- Cold materials behave elastically as opposed to Warm materials flowing in a viscous fashion.
- This means there is a *characteristic temperature* (roughly 70% of the melting temperature) which defines the base of the elastic layer

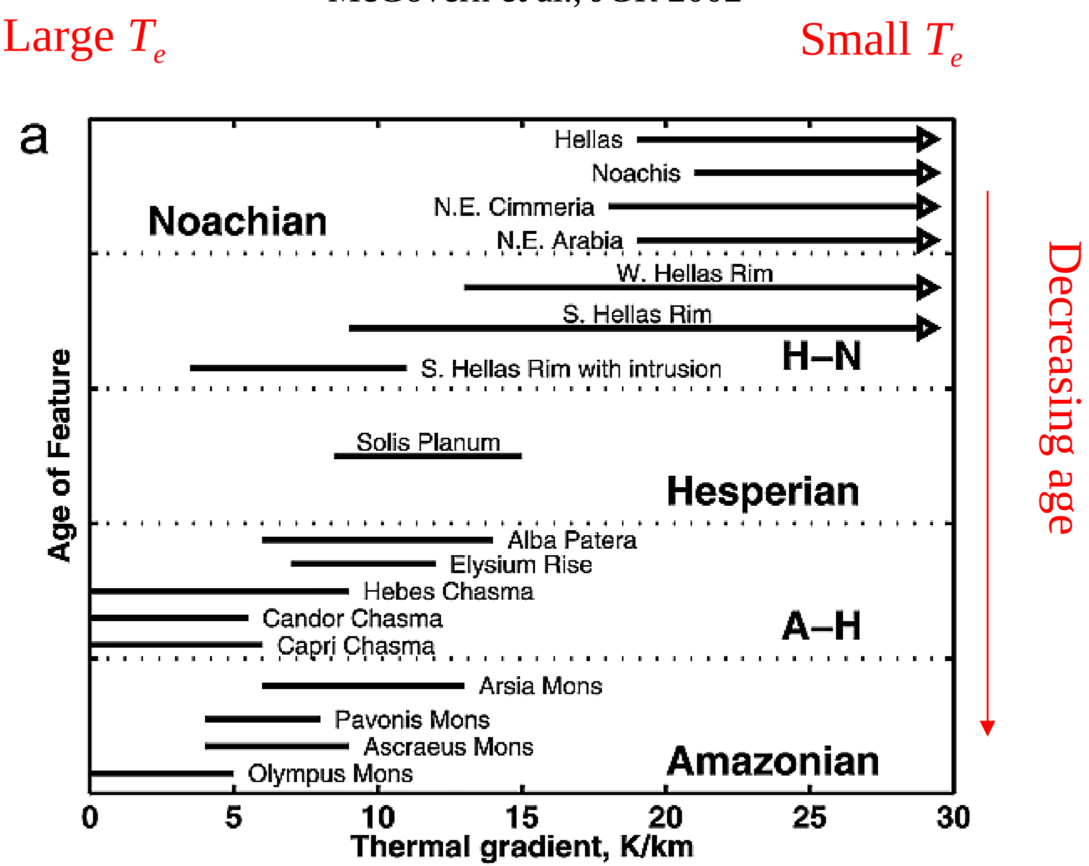
- E.g. for ice the base of the elastic layer is ~ 190 K
- The measured elastic layer thickness is 1.4 km
- So the thermal gradient is 60 K/km
- Thus the ice shell thickness is 2.7 km (!)



T_e and age

- The elastic thickness recorded is the *lowest since the episode of deformation*
- In general, elastic thicknesses get larger with time (why?)

McGovern et al., *JGR* 2002



- So by looking at features of different ages, we can potentially measure how T_e, and thus the temperature structure, have varied over time
- This is important for understanding planetary evolution

T_e in the solar system

- Remote sensing observations give us T_e
- T_e depends on the composition of the material (e.g. ice, rock) and the temperature structure
- If we can measure T_e, we can determine the temperature structure (or heat flux)
- Typical (approx.) values for solar system objects:

Body	T_e (km)	*dT/dz* (K/km)	Body	T_e	*dT/dz* (K/km)
Earth (cont.)	30	15	Venus (450^{o}C)	30	15
Mars (recent)	100	5	Moon (ancient)	15	30
Europa	2	40	Ganymede	2	40

Summary

- Volcanism happens because of higher temperatures, reduced pressure or lowered solidus
- Conductive cooling time $t = d^2/\kappa$
- Planetary cooling leads to compression
- Elastic materials $\sigma = E\,\varepsilon$
- *Flexural parameter* controls the length-scale of deformation of the elastic lithosphere

$$\alpha = \left[\frac{ET_e^3}{3g(\rho_m - \rho_w)(1-\nu^2)}\right]^{\frac{1}{4}}$$

- Lithospheric thickness tells us about thermal gradient