Geothermal Field

Prof. Dr. Emad A. M.Al-Heety

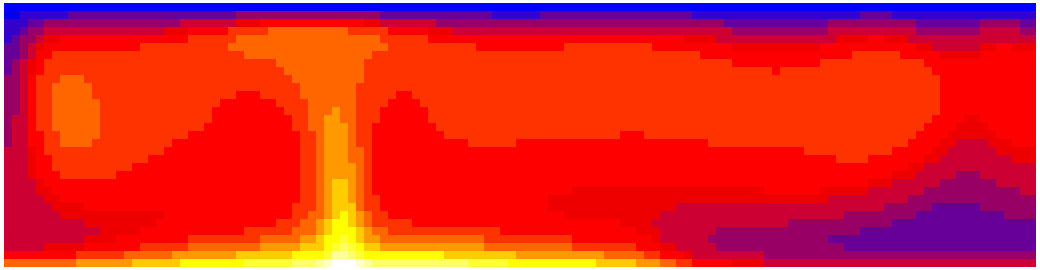
Department of Applied Geology

College of Science

University of Anbar
Email: emadsalah@uoanbar.edu.iq

Geothermics

- The earth is hot at the core-mantle boundary (about 4000° C) and cold at its surface (19 ° C) and very cold in space (-270 ° C). This large temperature contrast means the earth is in thermal dis-equilibrium and must be transferring heat into space.
- The internal earth is cooled by convection that delivers heat to the base of the lithosphere where the heat is then conducted through the lithosphere. This heat flux warms the atmosphere and finally the atmosphere is always radiating heat into space.
- Note: the Earth primarily cools itself by creating new lithosphere at ridges that is pulled across the ocean basin and finally subducted deep into the mantle to cool it.
- The solid earth is a heat engine because work is done moving the plates and raising the mountains and pulling apart the ridges/rifts. The ocean-atmosphere is also a heat engine.
- In the last 4 Ga, the earth's cooling has been about 100 ° C per Billion years.
- The earth is predicted to reach thermal equilibrium with space in about 10 Ga.



Three heat transport processes

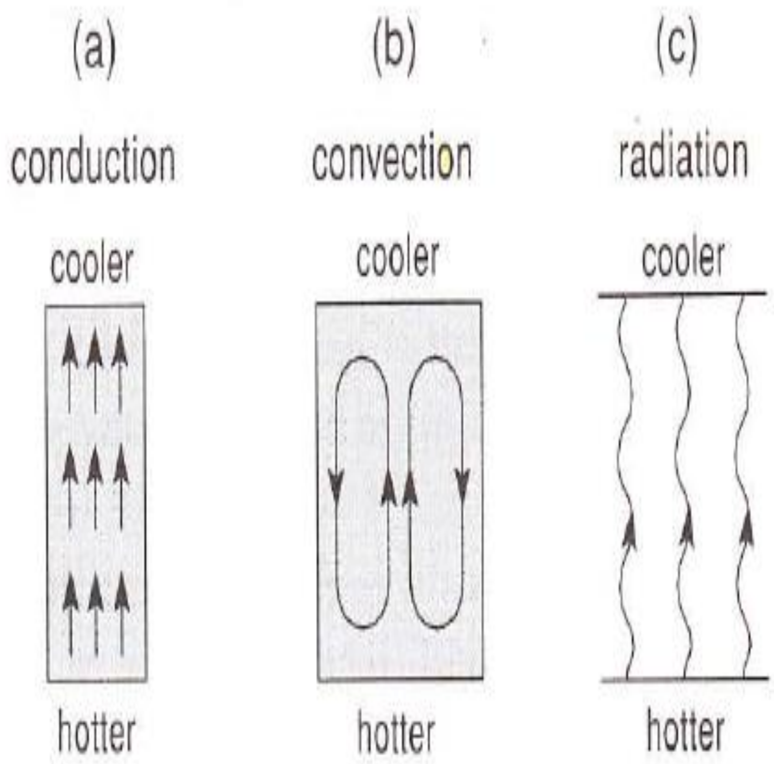


Figure 17.1 Ways that heat travels.

There are three ways to transfer heat

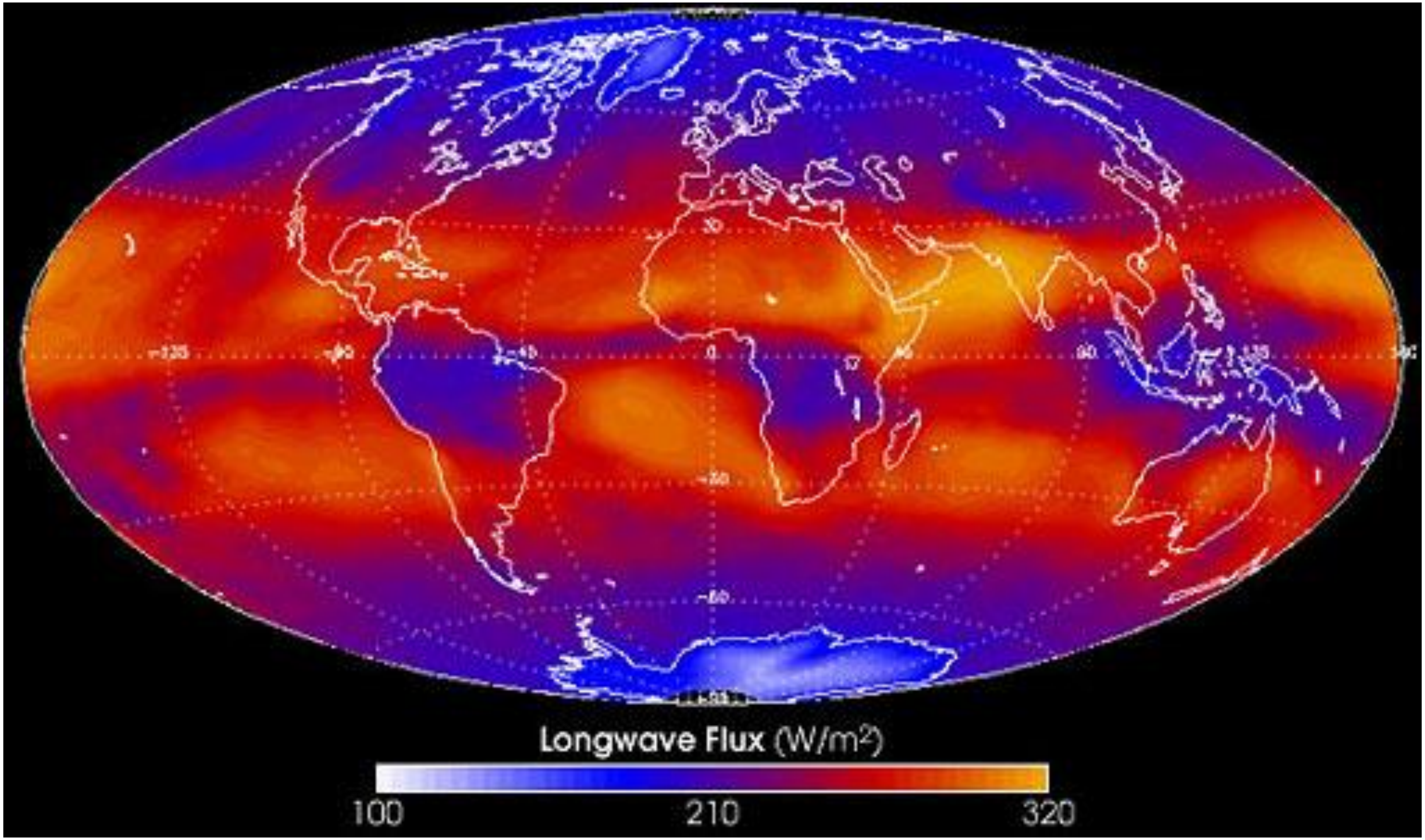
Conduction: heat in a solid is transferred via diffusion of crystal lattice vibrations. Conduction can never be stopped if a temperature difference exists between two regions. Examples?

Convection: heat is moved (advected) by different density parcels flowing up and down in the earth's gravity field. This only occurs if a parcels buoyancy force (Archimedes principle) is sufficiently large with respect to the strength (viscosity) of the mantle. Examples?

Radiation: heat moves as infrared electromagnetic waves at speed of light. All matter at non-zero temperatures radiates energy. Examples?

Radiative heat flux from Planet earth

Energy (N*m or J)   Power=energy/time (J/s or W)   Flux = Power/area (J/s-m^2 or W/m^2)



Conductive heat transport in lithosphere and convective heat transport in mantle and fluid outer core

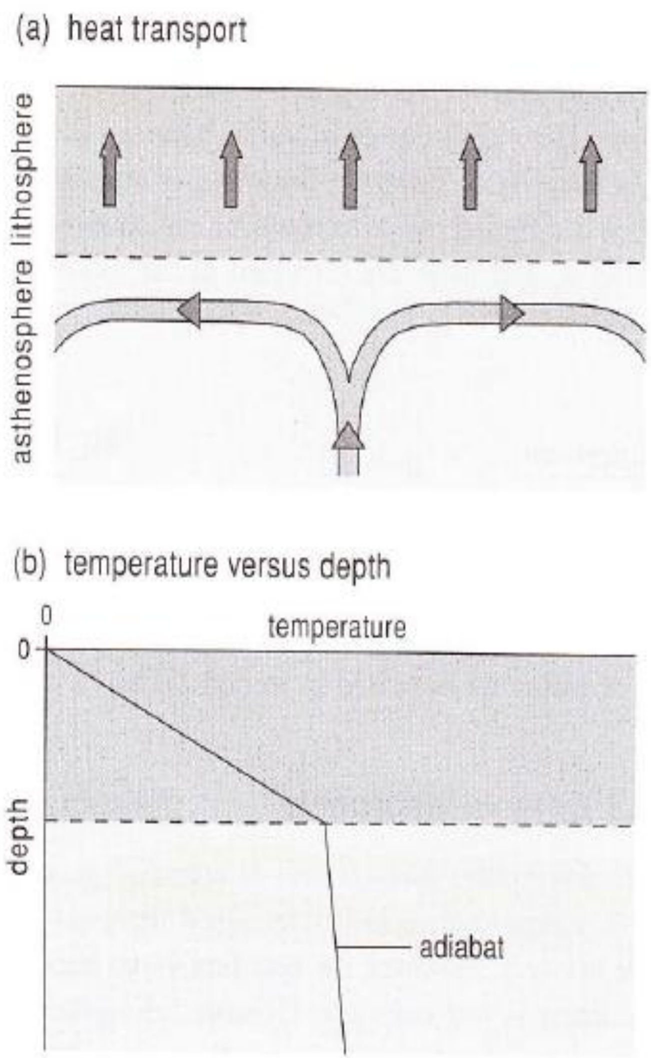


Figure 17.2 Heat transport within the Earth (schematic).

The lithosphere is strong and hence translates as a quasi-rigid block over the asthenosphere, therefore heat cannot be convected across the lithosphere and instead heat is conducted across the lithosphere.

The mantle dominantly moving heat via convection, although heat conduction is always occurring. But the 1-10 cm/yr mantle flow rates moves (advects) the heat much faster than heat conduction!

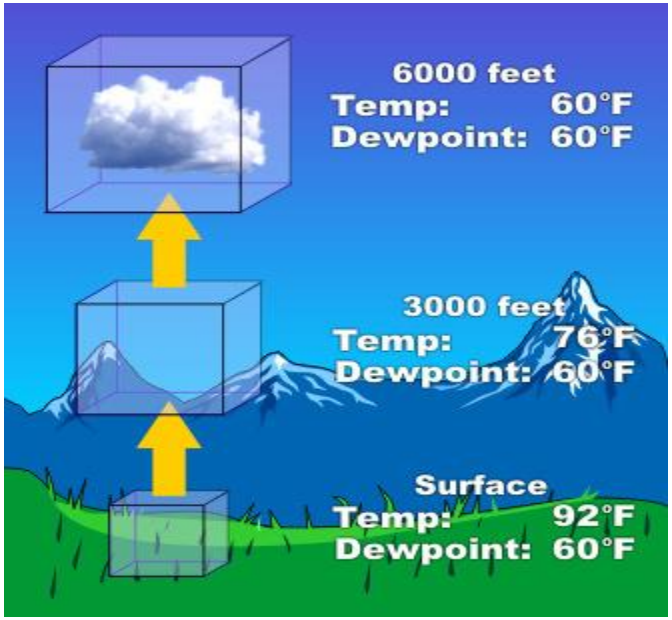
The temperature with respect to depth (geotherm) in the lithosphere is near a straight line which is called the conductive geotherm: $dT(z)/dz = 30°C/km$. The lithospheric geotherm will have curvature when there is significant heat production (e.g., in continental crust).

The adiabat geotherm is about $dT/dz = 0.3\text{-}0.5°C/km$; this temperature rise with depth is because each parcel of heat is being compressed into a smaller volume due to the pressure increase with depth.

Adiabatic thermal gradients

•*adiabatic* - a process where a parcels temperature changes due to an expansion or compression, but no heat is added or taken away from the parcel.

Adiabatic temperature gradients do NOT drive heat flow.



$$\left(\frac{\partial T}{\partial P}\right)_S = \frac{T\alpha}{\rho c_P}$$

...adiabatic gradient as a function of pressure

...but we want it as a function of depth

For the Earth $$\frac{dP}{dr} = -g\rho$$

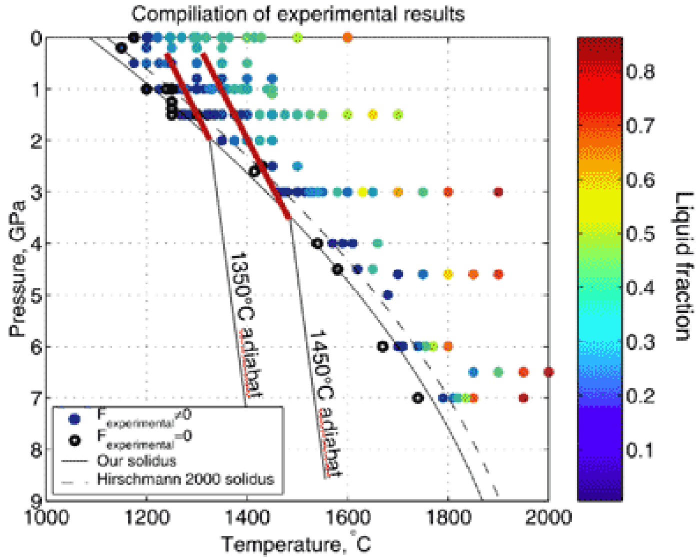
Substitute... $$\left(\frac{\partial T}{\partial r}\right)_S = \frac{T\alpha g}{c_P}$$

...adiabatic gradient as a function of radius

Temperature gradient for the uppermost mantle 0.4 $^oC\ km^{-1}$ using $T = 1700$ K, $\alpha = 3 \times 10^{-5}\ ^oC^{-1}$, $g = 9.8\ m\ s^{-2}$, $cp = 1.25 \times 103\ J\ kg\ ^oC^{-1}$

at greater depth 0.3 $^oC\ km^{-1}$ due to reduced α

Mantle Potential temperature



Heat flux and amount of heat power

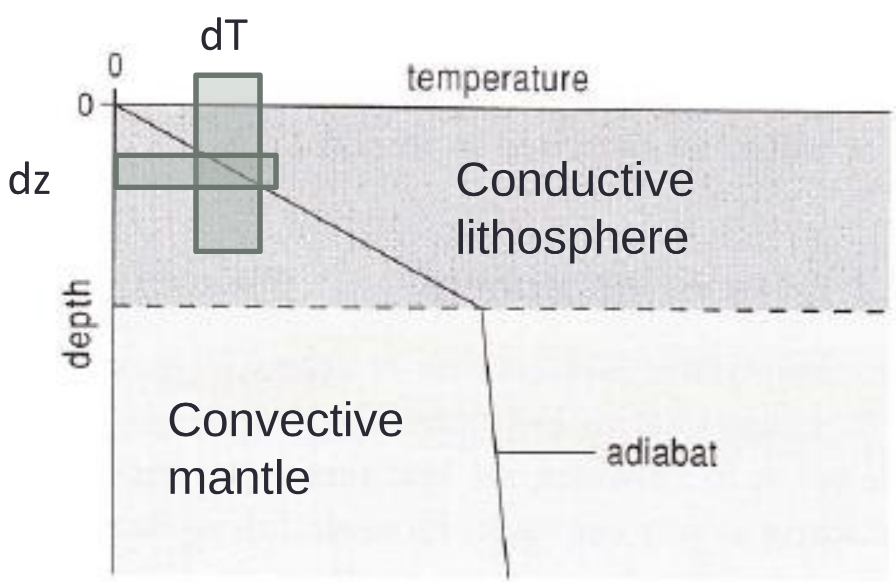
Little-q Heat flux: W/m^2
Power (W) per area (m^2)
Energy (J) per time (s) per area (m^2)

$$q\left(\frac{W}{m^2}\right) = K\left(\frac{W}{m^\circ C}\right) * \frac{\Delta T}{\Delta z}\left(\frac{^\circ C}{m}\right)$$

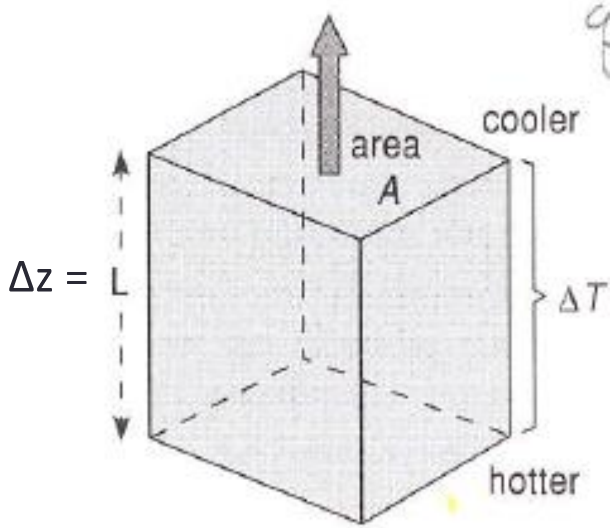
Big-Q Heat Power: W
Energy (J) per time (s)

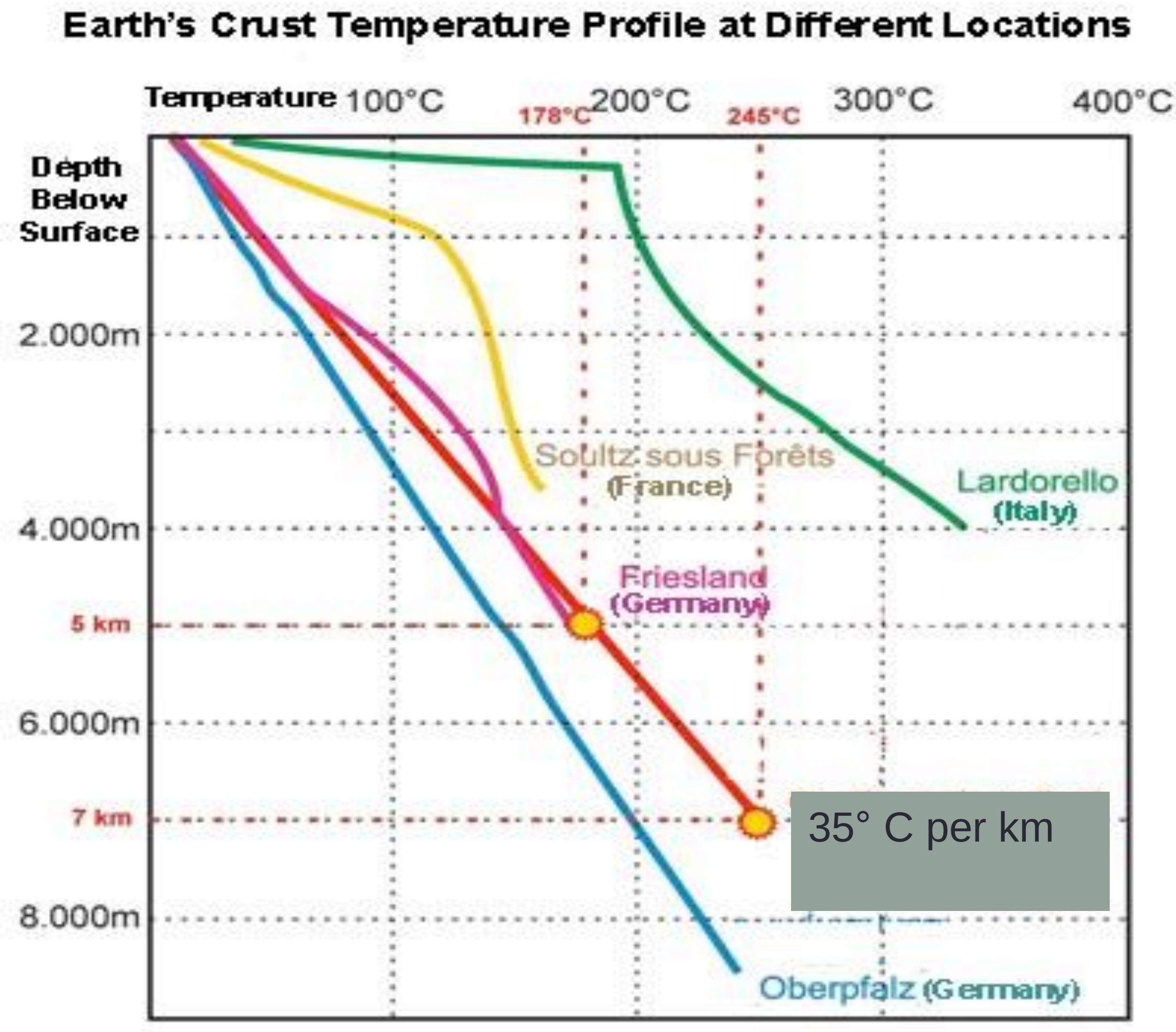
$$Q(W) = K\left(\frac{W}{m^\circ C}\right) * \frac{\Delta T}{\Delta z}\left(\frac{^\circ C}{m}\right) * A\left(m^2\right)$$

Geotherm with 0° C at surface (z=0 m)



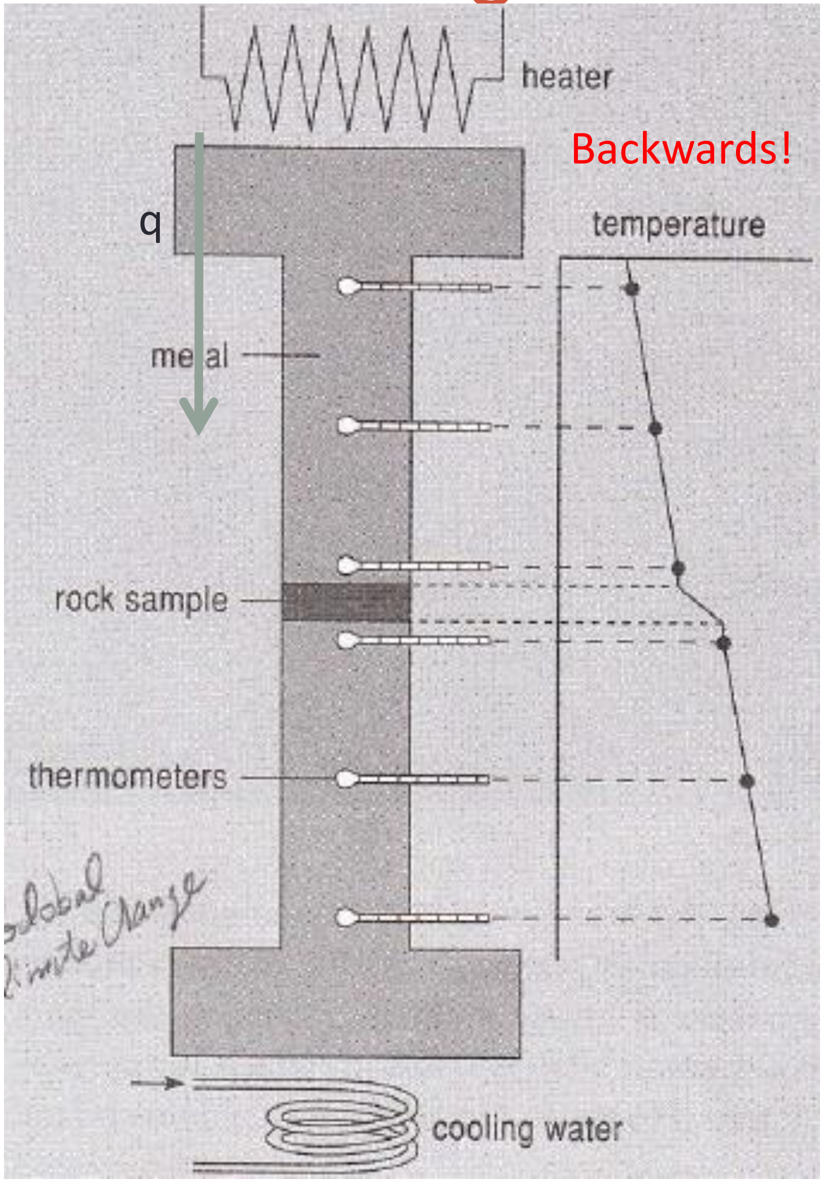
Heat flux through area A




Earth's Crust Temperature Profile at Different Locations
Temperature 100°C 200°C 300°C 400°C
178°C
245°C
Depth Below Surface
2.000m
4.000m
5 km
6.000m
7 km
8.000m
Soultz sous Forêts (France)
Lardorello (Italy)
Friesland (Germany)
Oberpfalz (Germany)

35° C per km

Measuring thermal conductivity



Because it is very hard to measure mW scale heat fluxes at earth's surface, a tactic is to measure the temperature gradient in a borehole and the thermal conductivity of the rocks in the borehole.

Heat flux is calculated: $q = k*dT/dz$ (W/m^2)

The thermal conductivity of a rock can be measured by heating one side of the metal rod and cooling the other side and measuring the thermal gradients in the metal and rock-sample regions.

Using the known conductivity of the metal, the heat flux is calculated.

Thus, the conductivity of the rock is:
$K = q / (dT/dz)$.

Thermal conductivity values

Good heat conductors:
Diamond, Metals

Poor heat conductors
Wood, Wool, Fiberglass

Conductivity is largely determined by ability of electrons to wandered through a substance. Good heat conductors are also good electrical conductors.

Material	Thermal conductivity (cal/sec)/(cm^2 C/cm)	Thermal conductivity (W/m K)*
Diamond	...	1000
Silver	1.01	406.0
Copper	0.99	385.0
Gold	...	314
Brass	...	109.0
Aluminum	0.50	205.0
Iron	0.163	79.5
Steel	...	50.2
Lead	0.083	34.7
Mercury	...	8.3
Ice	0.005	1.6
Glass,ordinary	0.0025	0.8
Concrete	0.002	0.8
Water at 20° C	0.0014	0.6
Asbestos	0.0004	0.08
Snow (dry)	0.00026	...
Fiberglass	0.00015	0.04
Brick,insulating	...	0.15
Brick, red	...	0.6
Cork board	0.00011	0.04
Wool felt	0.0001	0.04
Rock wool	...	0.04
Polystyrene (styrofoam)	...	0.033
Polyurethane	...	0.02
Wood	0.0001	0.12-0.04

Heat flow as function of oceanic plate age

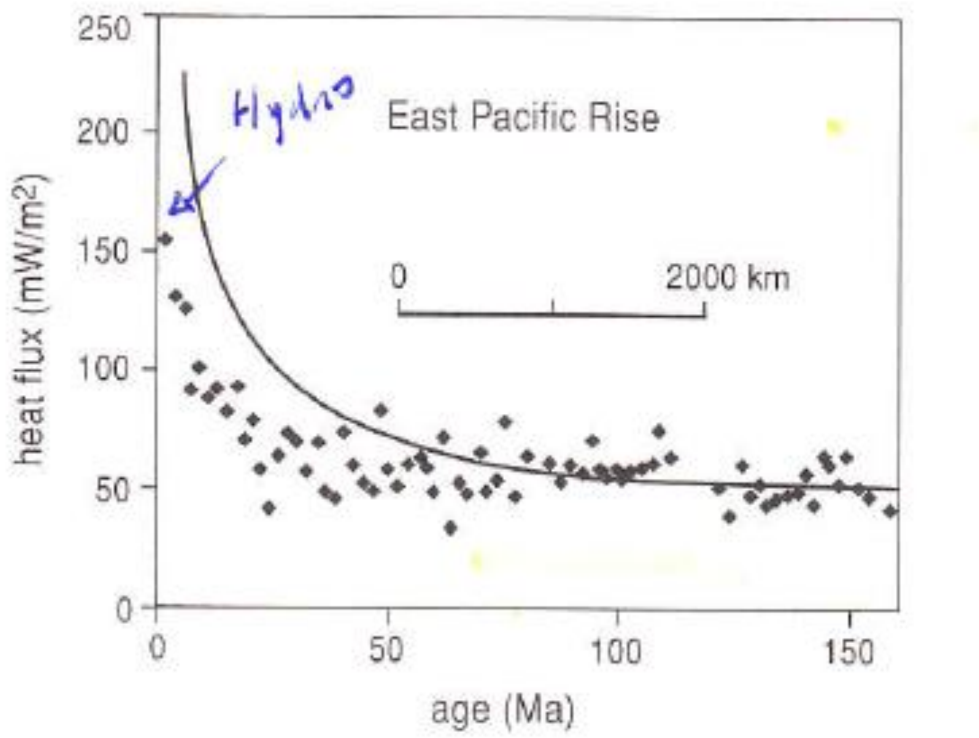
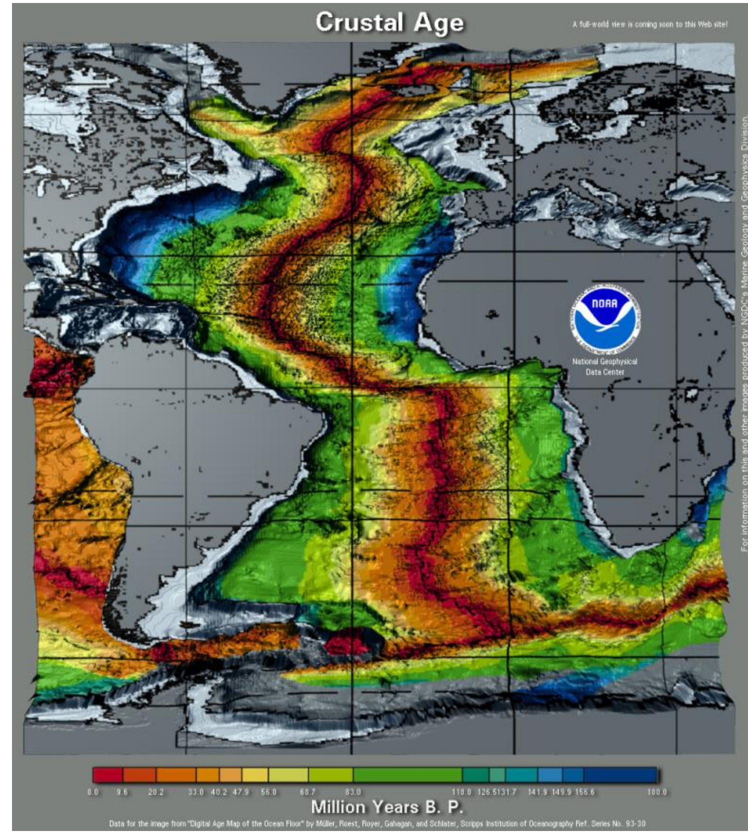


Figure 17.5 Heat flux near the East Pacific Rise (distance scale is approximate).



The plate moves horizontally away from the mid-ocean ridge and cools as heat is conducted into the ocean bottom over 150 ma. This cooling is manifest as the cool lithosphere getting thicker with time.

Heat production: heat flux increases in direction of heat flow

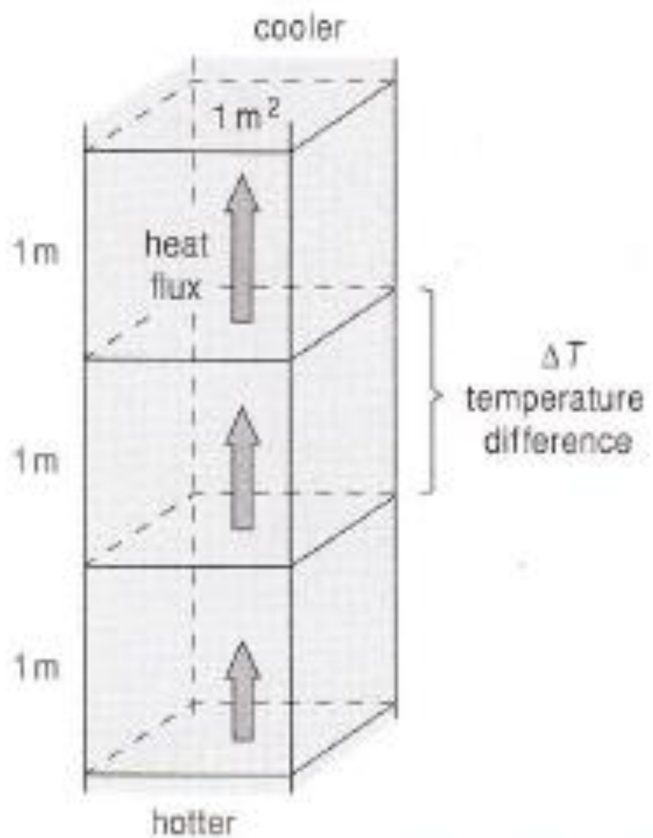


Figure 17.8 Heat flux with internal heat production.

Note that the heat flux vectors becomes longer (bigger) towards the cooler side of the column.

This is because a constant heat production ($\mu W/m^3$) throughout the column is assumed.

This creation of heat increase the heat flow and the thermal gradient towards the cooler end (i.e., the earth's surface).

Thus, the distribution of heat production in the crust changes the geotherm to make it a curve and not a straight line for a steady state thermal equilibrium.

Crustal heat productivity vs. heat flux relationship

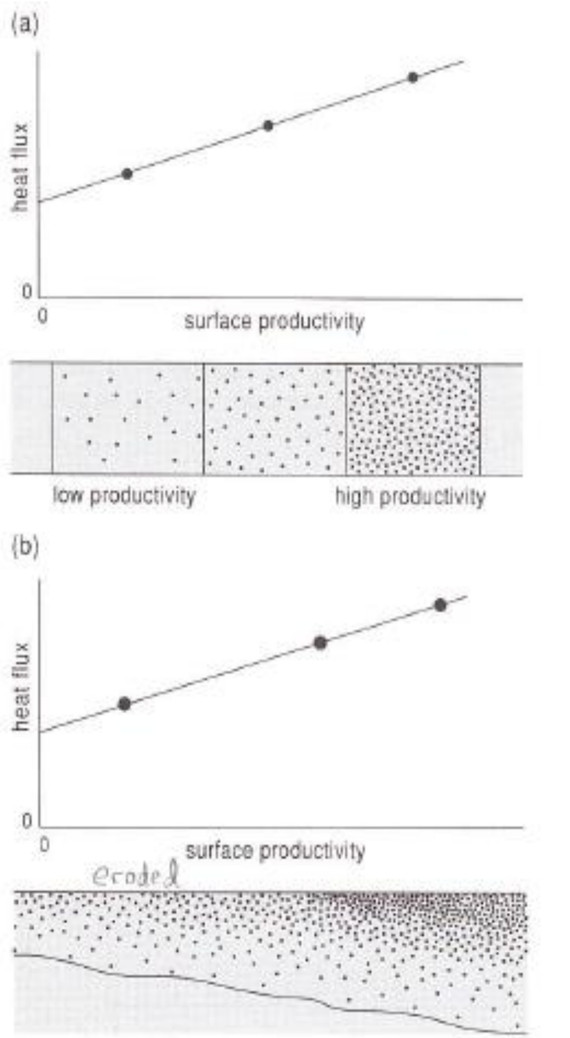
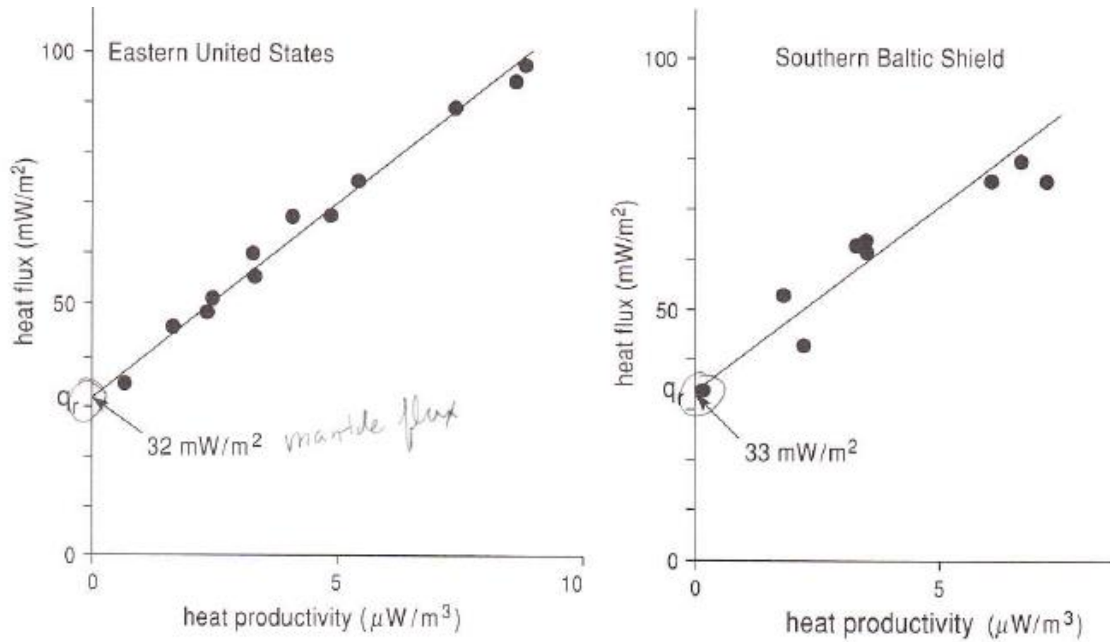


Figure 17.7 Heat fluxes and schematic thermal structures of a heat flow province.



The continental crust is highly concentrated in radioactive elements such as Uranium. As Uranium decays, this is termed heat productivity in micro-Watts per unit volume.

By measuring the surface rock heat productivity, a linear relation between heat productivity and heat flux is often found. This means crustal heat production controls the heat flux.

The y-intercept value is the reduced heat flow (q_r) which is heat flux from mantle convection.

Oceanic and continental heat flux and geotherms

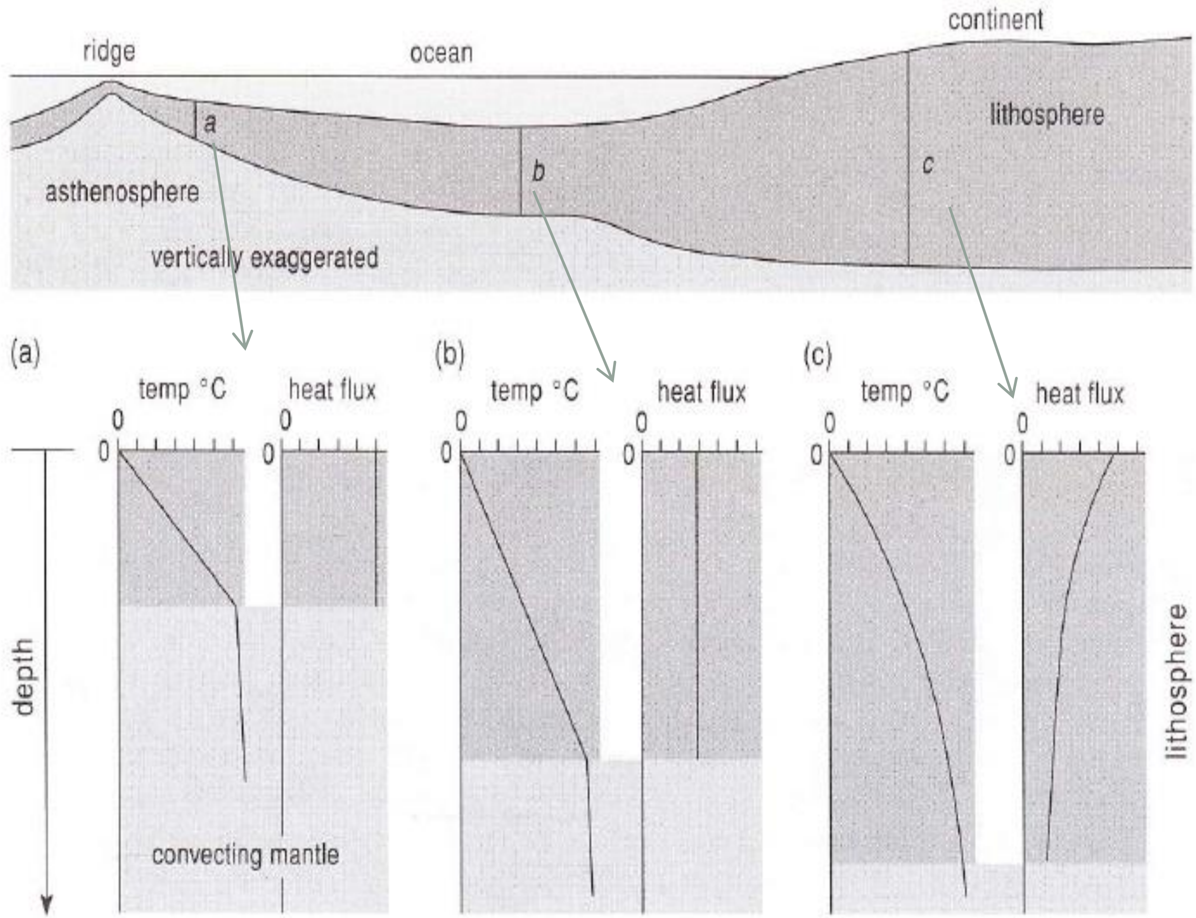


Figure 17.4 Temperatures and heat fluxes through oceanic and continental lithosphere.

The geotherm in the lithosphere is conductive and the geotherm in the convecting mantle is the adiabat.

Note that the conductive heat flux with respect to depth in the oceanic lithosphere is constant and hence the geotherm is a straight line.

But, the conductive heat flux in the continental lithosphere increases towards the surface and the geotherm is a curved line.

The base of the lithosphere has the same temperature at all three places of about 1300°C.

Radioactive heat generation

Table 17.1 Radioactive heat generation

Part of Earth	% of Earth's volume	Rock types	Thermal productivity (μW/m^3)	% of Earth's total heat outflow
continental crust	0.7	granites, etc.	0.55	c. 10%
oceanic crust	0.2	gabbro	0.03	c. 0.15%
mantle	84	peridotite	0.014	c. 30%
core	16	mainly iron	negligible	negligible
Total				c. 40%

Based on Table 8.1 of Brown and Mussett (1993).

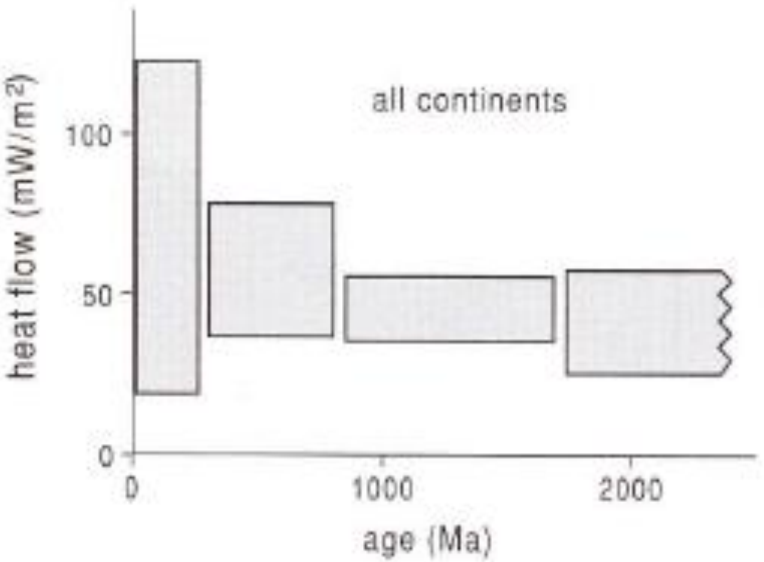


Figure 17.11 Decrease of continental heat flux with age.

Note that 40% of the Earth's total heat flow comes from radioactive heat generation!

The other 60% of heat comes from 'secular cooling' of the original accretional heat when the Earth formed 4.5 Ga ago.

Also, note that 10% of the heat comes from continental crust, but only 0.15% from ocean crust.

In general, there is a decrease in heat flow with the age of continents. Why ?

Planetary heat budget

46 +/- 3 TW (Terra-Watts)

Terra is 10^{12}

World power consumption is 15 TW.

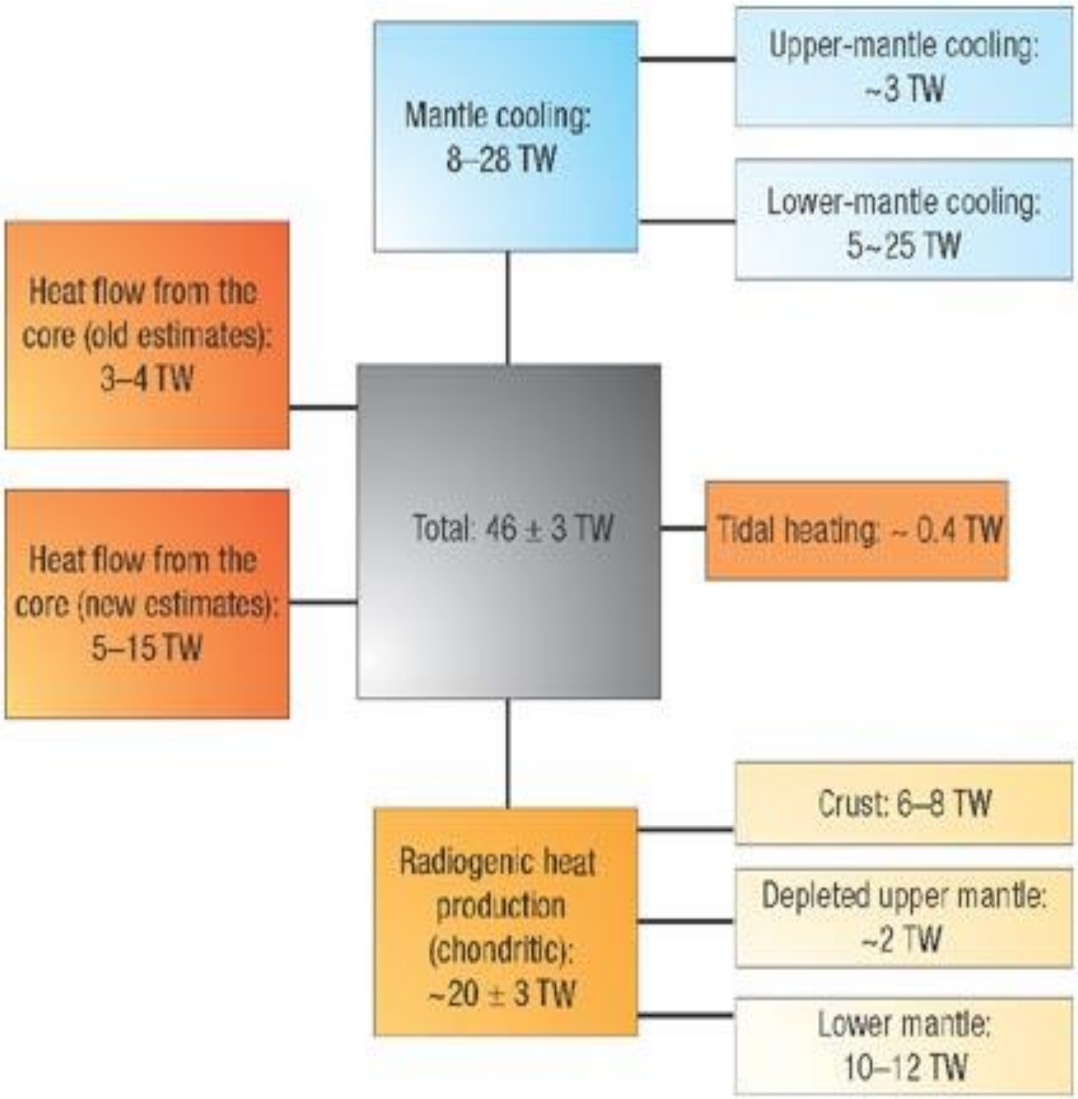
The primary contributions to observed total surface heat are shown.

Radiogenic heat production

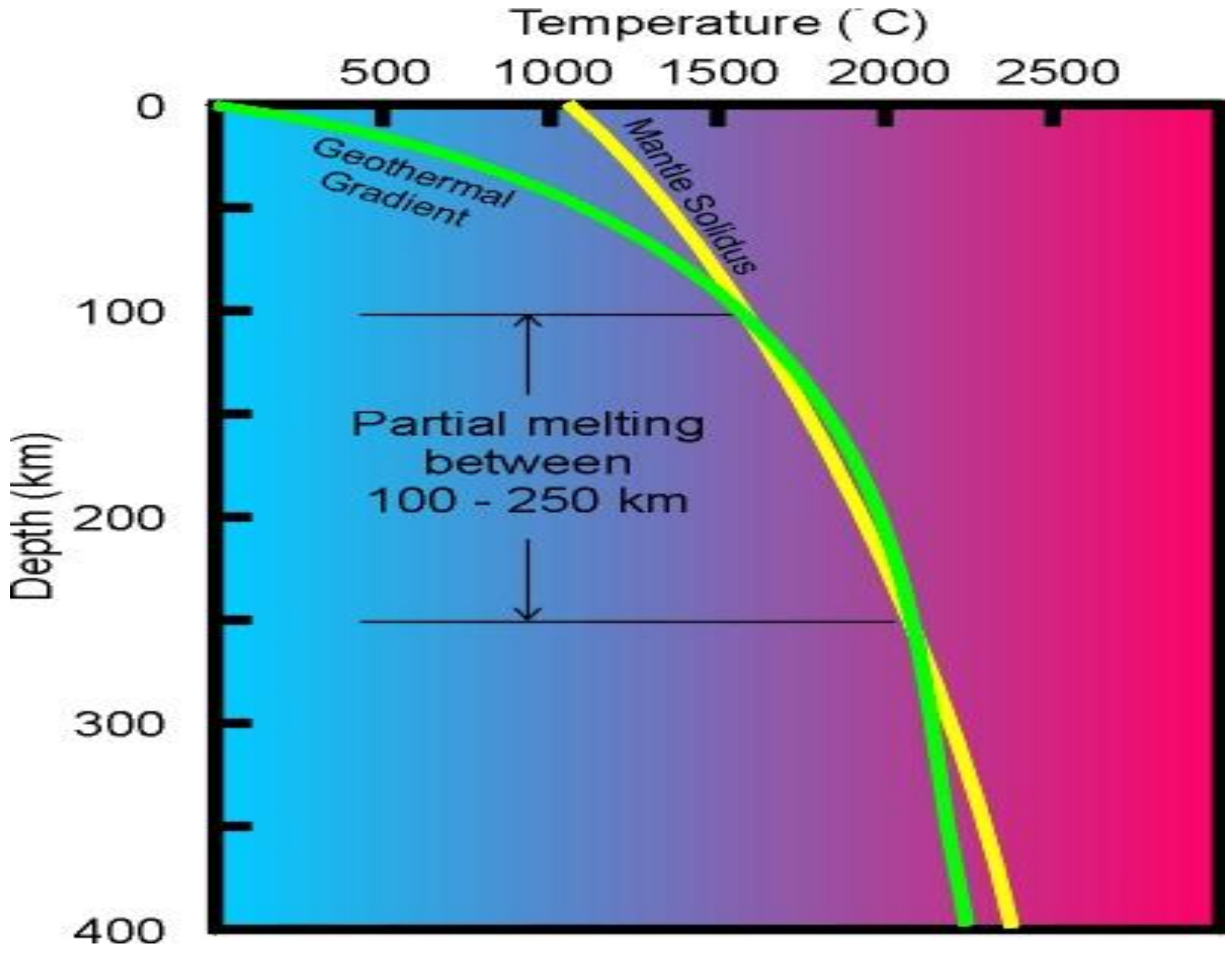
mantle cooling

heat flow from the core

These three heat sources dominate the mantle energy budget, but there are substantial uncertainties in the latter two contributions.

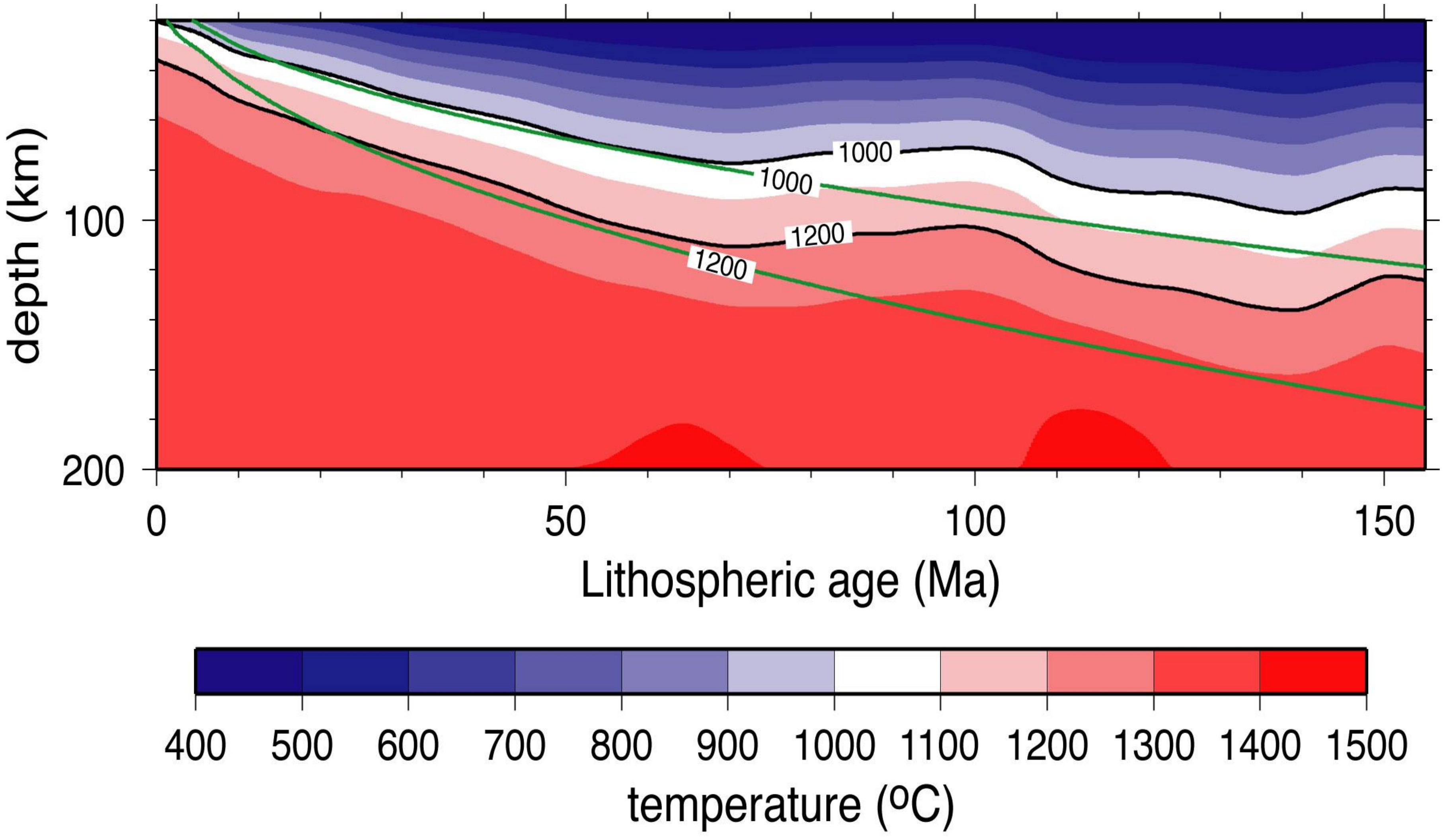


Mantle temperature and melting



By knowing the approximate temperature at which mantle rocks melt, the geotherm can be compared to the mantle solidus to see if melt should form.

Seismic velocity mapped to temperature for Oceanic lithosphere



Temperature re-equilibration

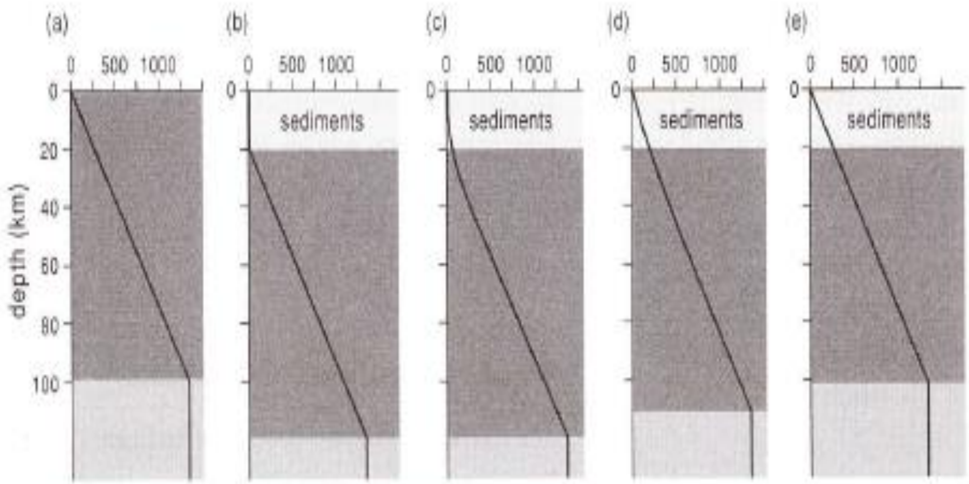


Figure 17.9 Temperature reequilibration for a sedimentary basin (schematic).

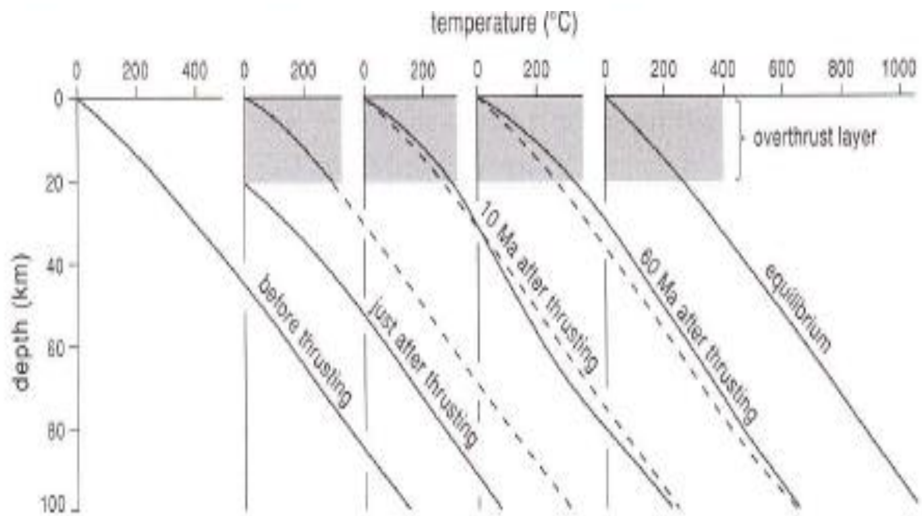


Figure 17.10 Temperature reequilibration after overthrusting (schematic).

So far, we have been assuming that the temperature does NOT vary in time. This is called thermal steady-state or dynamic equilibrium.

So, what if the temperature structure changes in time due to basins filling or tectonic overthrusts?

Adding sediments quickly to the top of the crust, causes a thermal disequilibrium that will be brought to thermal equilibrium over Ma time-scales.

Thrusting 20 km of the upper crust over the surrounding surface quickly, creates a disequilibrium that takes greater than 50 Ma to reach thermal equilibrium.

Geothermal heat extraction and energy production

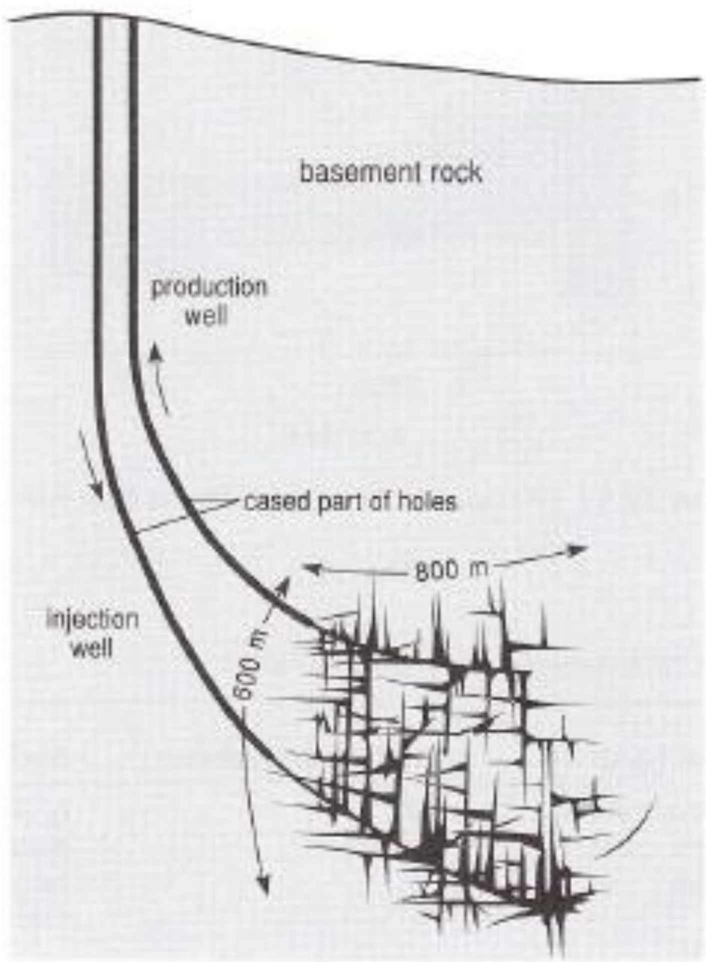


Figure 17.12 Hot, dry rock heat-extraction system.

Where high temperature rock is near the surface, the heat contained in the hot-rock can be mined to drive steam-turbines to make electricity.

This is done by pumping water down the injection well where it flows through the cracks in the rocks and heats up and corresponding cools down the rocks over time. The hot water is then extracted via a production well.

Note: geothermal energy is technically not a renewable resource as the water circulation extracts the heat much faster than the earth's mW scale heat flux can replace the mined heat (energy).

Atmospheric temperature changes effects on shallow geotherms

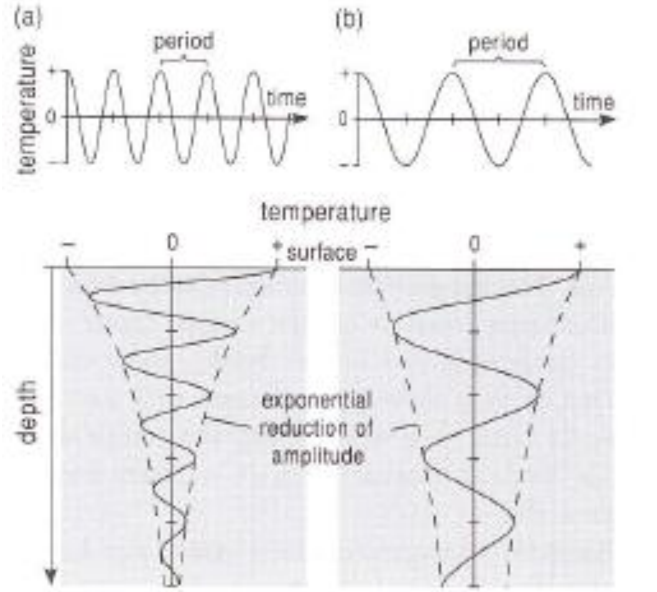


Figure 17.13 Propagation of alternating temperature changes at two frequencies.

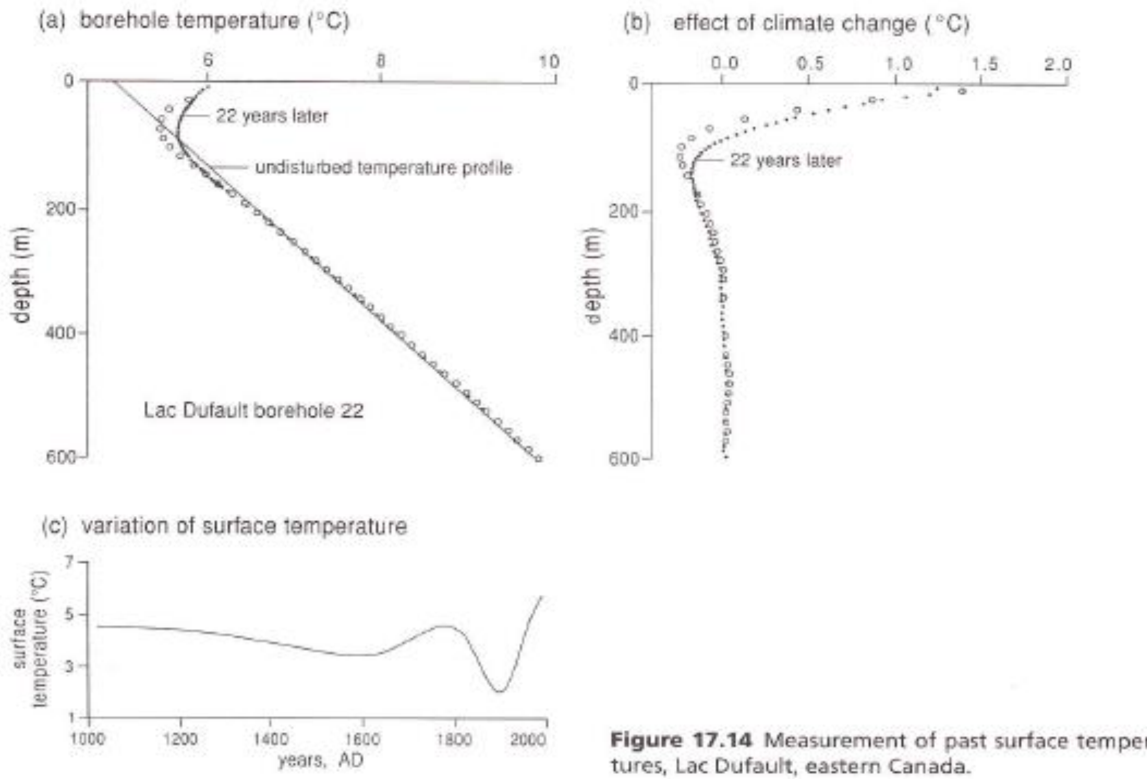
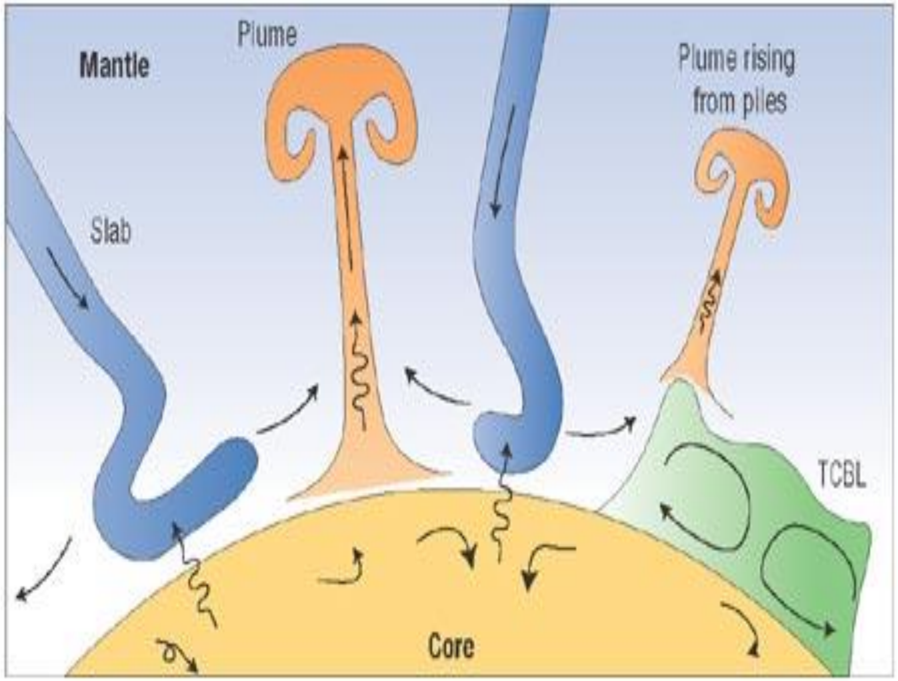


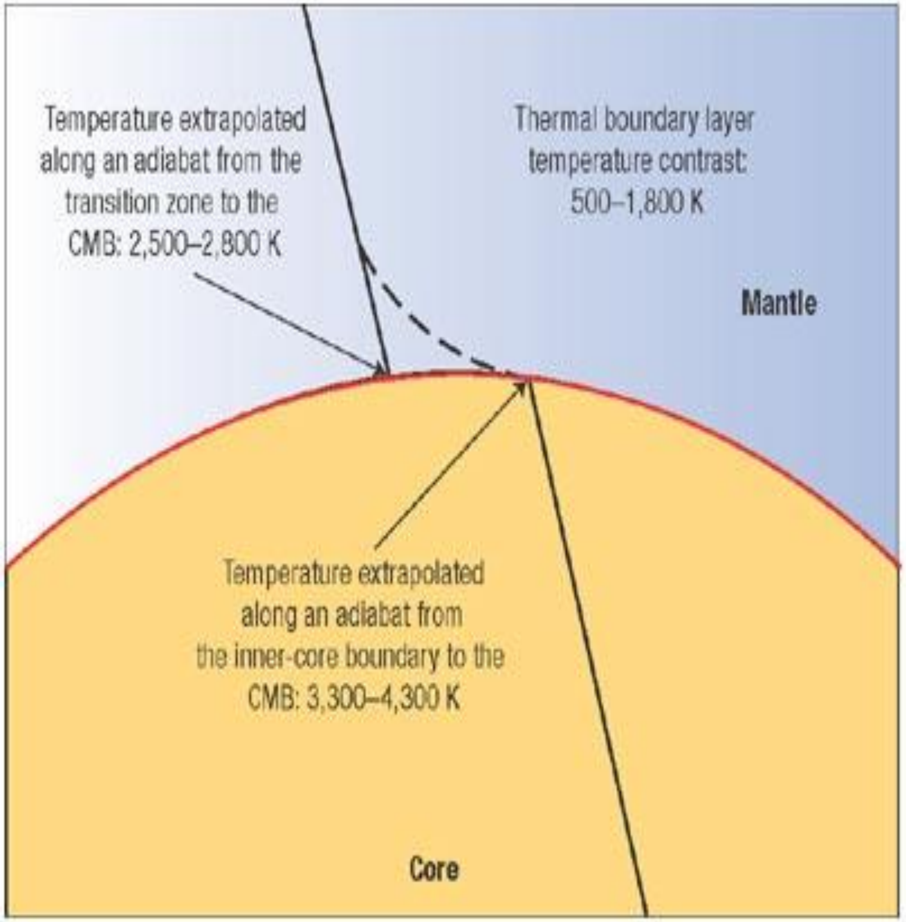
Figure 17.14 Measurement of past surface temperatures, Lac Dufault, eastern Canada.

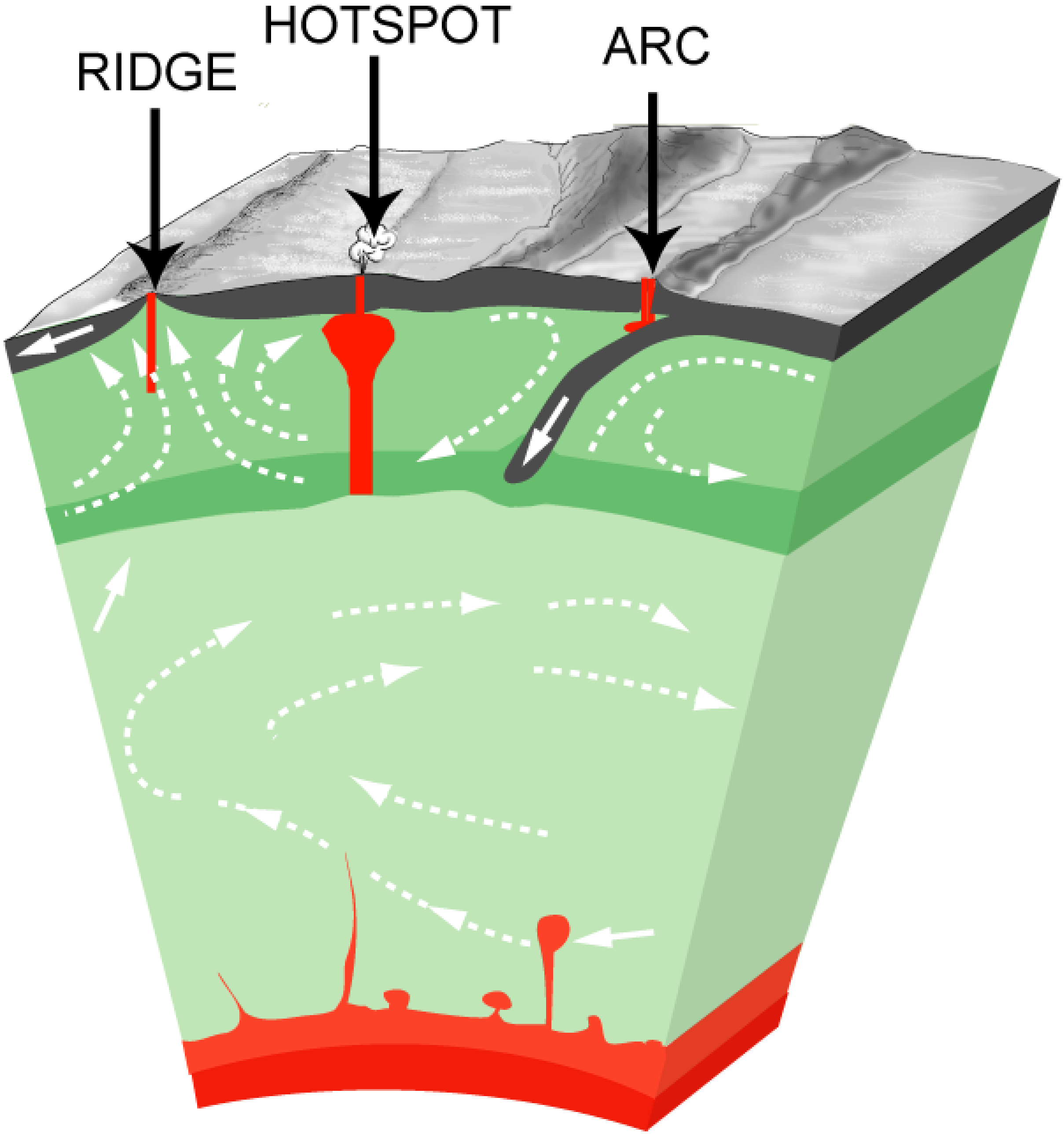
Periodic (daily, seasonal) changes in atmospheric temperature makes 'temperature waves' that propagate downward into the crust, but the attenuate exponentially with depth.

A recent 2 ° C increase in surface temperature between 1900-2000 can be measured in the borehole temperature profile and well modeled using the heat flow equations.

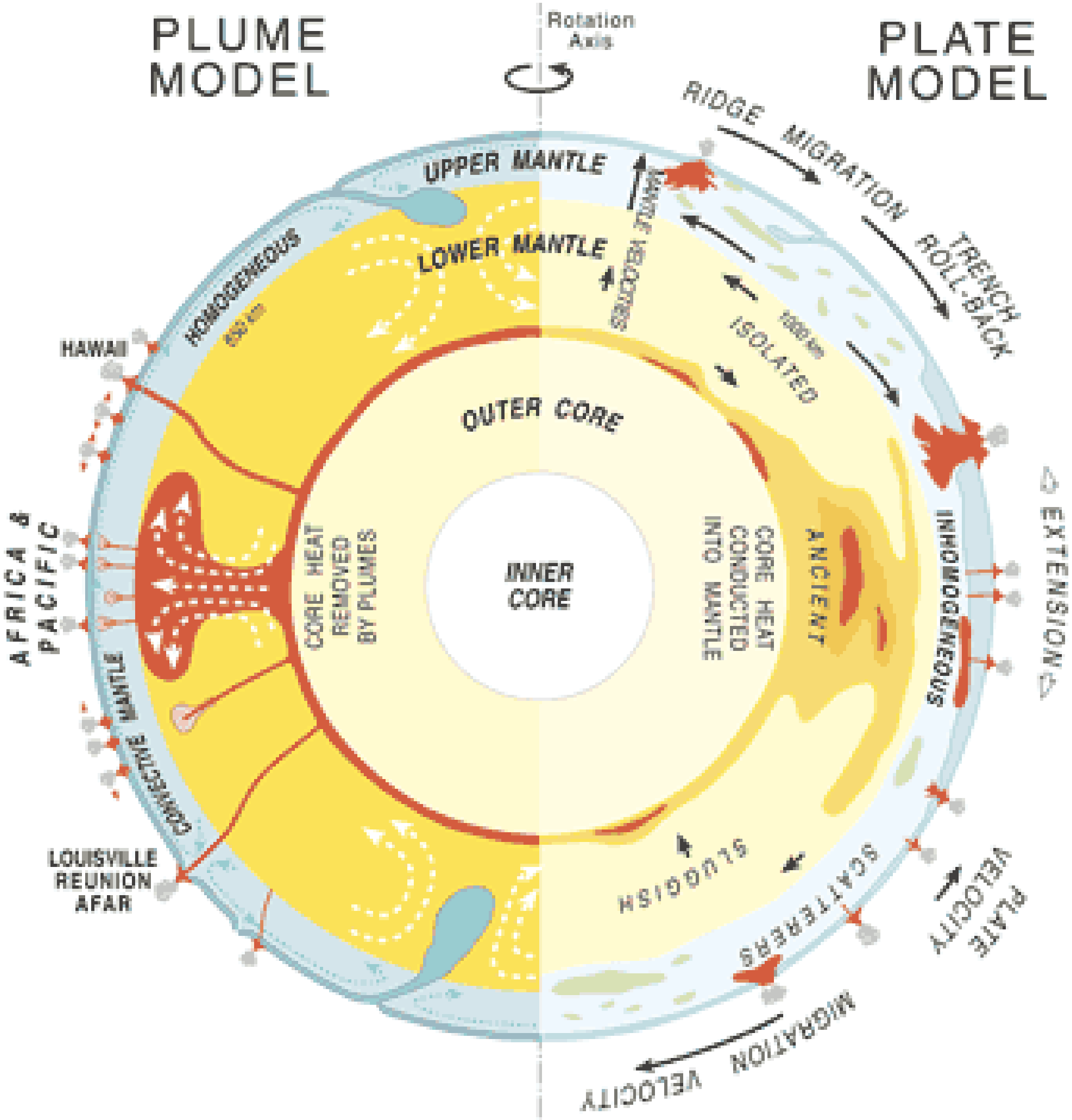
Thermal boundary layer and at core-mantle boundary and heat exchange



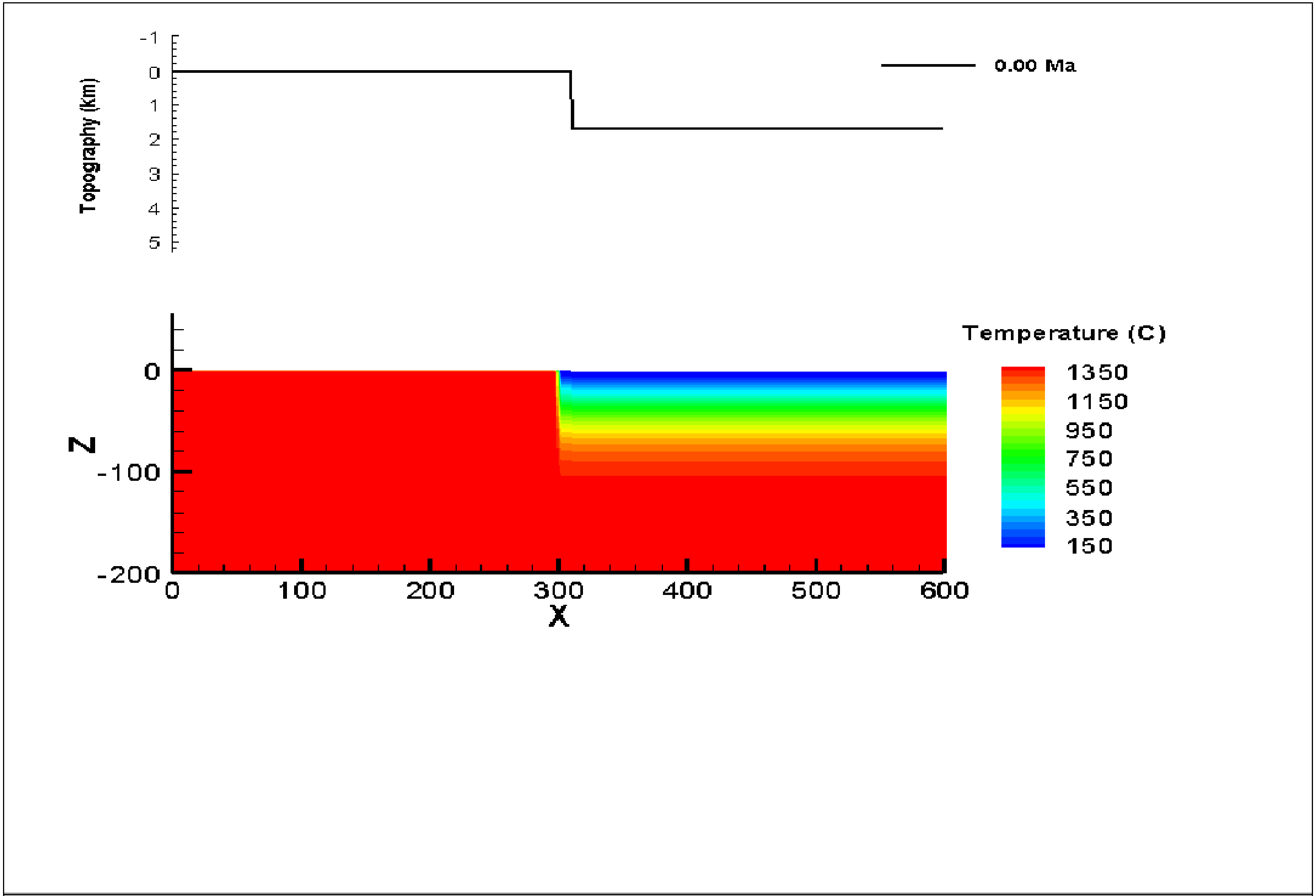



HOTSPOT
RIDGE
ARC

Plume vs. Plate convection models



Starting subduction convection



subduction start.avi

References

-Stacey F. and Davis P., Physics of the Earth, Cambridge University Press, 2008.

- Lowrie, W., Fundamentals of Geophysics, Cambridge University Press, 2007