Heat flow at the $\oplus$'s surface

Heat is a form of energy. It is the macroscopic manifestation of the energy of molecular motion.

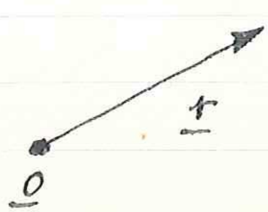
Heat tends to move from hot places to cold places.

Three mechanisms of heat transfer: conduction, convection and radiation.

Conduction dominant in lithosphere although as we shall see heat transfer by hydrothermal convection is also important in the oceans (and in hot spring areas on land).

We shall however study heat flow by conduction first: same mechanism that makes handle of a cast iron skillet get hot.

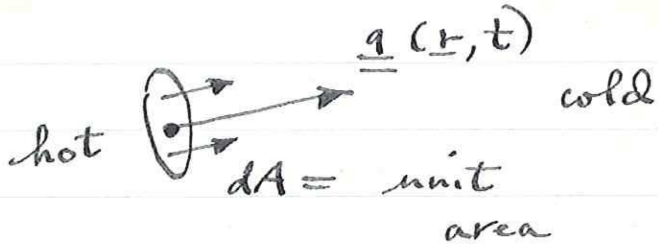
Consider a body: $\underline{r}$ = position vector



If some parts of the body are hotter than others, heat flow will occur to equalize temp.

Heat flow or flux is a vector field in the body, call it $\underline{q}(\underline{r})$.
To define $\underline{q}(r)$ must specify direction and magnitude.

1. direction: direction heat is flowing
2. amount of energy flowing through a unit area per second: magnitude



Units of $\underline{q}(\underline{r},t)$: SI units $\frac{J}{m^2\,sec}$

or $\frac{W}{m^2}$.

An older unit for heat (used before it was realized that heat was just a form of energy) the calorie = 4.184 J

A unit frequently used in geophysics:

$$\boxed{\begin{aligned} &1 \text{ HFU (heat flow unit)} \\ &= 1 \ \mu\text{cal}/\text{cm}^2\text{-sec} \\ &= 41.84 \ \text{mW}/\text{m}^2 \end{aligned}}$$

This a convenient unit because a typical heat flux thru ⊕'s surface is a few HFU.

The average heat flow escaping from the ⊕ is 80 ± 8 mW/m² ~ 2 HFU (Davies, 1980 RGSP). Sclater et al. say 70 mW/m². This cited in Turcotte & Schubert

[q] = 80mw/m2

Fourier's "law" of heat conduction relates the heat flux at a point to the temp. gradient

$$\boxed{\underline{q}(\underline{r}, t) = -k(\underline{r})\, \underline{\nabla T}(\underline{r}, t)}$$

Says heat flow is in direction of temp. grad. and $\propto$ to it. Not really a law like, say, $E = mc^2$, may break down in extreme situations, e.g.

during an explosion.

$-$ sign so that flow from hot to cold

$$\text{hot} \quad \xrightarrow{\ \underline{q}\ } \quad \text{cold}$$

$\longleftarrow$ direction of $\underline{\nabla T}$

κ is positive

κ, proportionality const. called <u>thermal conductivity</u>, units cal / cm - sec - °C. or W / m - °C.

<u>Conductivity</u> a <u>material</u> <u>property</u>, good conductors (Cu, Fe) have large κ, poor conductors (home insulation, e.g.) have low conductivity. Also κ can be temperature dependent.

One measures κ by applying heat (electrical resistance heating) to one side of a small sample and detecting the thermal response on the other side.

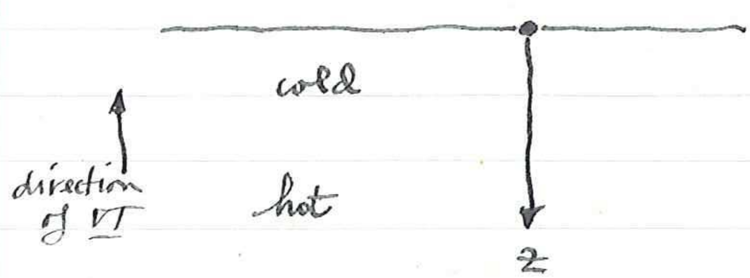
Example: high T conductivity of <u>olivine</u> (dominant constituent of oceanic lithosphere) from Schatz + Simmons <u>JGR</u> 1972 shown.

Typical value is $\kappa \sim 0.01$ cal/°C.-cm-s
Parsons and Sclater adopt a somewhat lower value as an average of for upper ~100 km of mantle, viz.

$$\boxed{\kappa \approx 0.0075 \text{ cal/°C.-cm-sec}}$$

This is about 20 – 30 times lower than a good conductor such as Cu or Fe. It is a common observation that rocks are better insulators than steel plates.

The interior of the ⊕ is hotter than the surface and as a result heat is flowing out. The heat flow at ⊕'s surface is given by



$$\boxed{q = \kappa\,(dT/dz)}$$

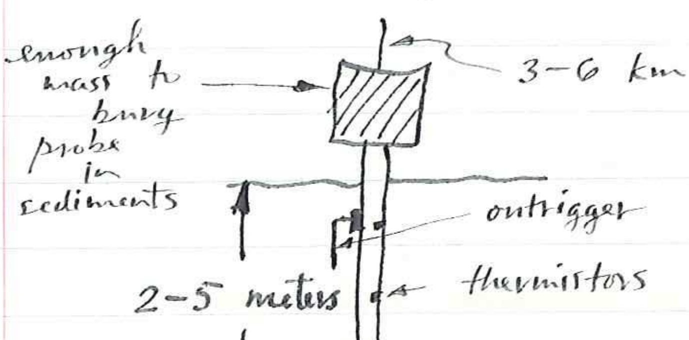
To measure the heat flow we measure dT/dz and multiply by κ.

First measurements were made on land down boreholes. Preferably these should be as deep as possible to avoid surface disturbances. Also heat transfer by ground water convection can and does plague many measurements. Perturbations associated with last ice age also important.

First measurements at sea 1956 Bullard, Maxwell and Revelle. Previously thought to be very difficult. Now commonplace, easier and more reliable and cheaper than land measurements which require deep ($\gg 100$ m) boreholes. Ocean floor very stable thermally.

on land must be careful to avoid circulation of drilling fluid.

Seafloor almost everywhere covered by soft sediments. Method, oceanic heat flow probes



To measure K of sediments can core or add an electrically heated outrigger

Or can core seds and bring back to lab.

κ of ocean-floor muds depends on H_2O content (must not allow to dry out before measuring κ).

or, equiv., $\sim 2 \frac{W}{m\,°C}$

Typically $\kappa_{mud} \sim 0.004$ cal / °C . cm s which means $dT/dz \sim 0.05°$ per meter or $0.1°$ for a 2 m probe. This not difficult to measure because of thermal stability of ocean floor.

[redacted] First oceanic heat flow measurement eagerly awaited. Expected to be much lower than on land because radioactive heat productivity of basalts is $\sim$ 30 – 40 times less than granites.

Famous story "I put it in the fixer first".
Result: average oceanic and continental heat flows roughly equal. Now thought that average oceanic heat flux about 70% greater (including hydrothermal fluid circulation) than average continental.

Tendency for higher heat flow over and near mid-ocean ridges an early discovery. Early evidence for seafloor spreading.

Now many 1000's of oceanic heat flow measurements. General characteristics:

1. very scattered in young crust less than, say, 30–40 m.y. includes high values up to 5–6 HFU.
2. less scatter beyond 30–40 m.y. typically about 1 HFU.

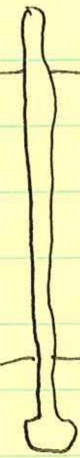
Example: compilation of data from Indian Ocean by Anderson et al. JGR (1977). Age determined from identification of magnetic anomalies, locations shown in map. Heat flow in HFU plotted as a function of age in Fig. 9A.

When sorted according to "reliability" using various criteria (e.g. uniformity of sediment cover) and averaged and smoothed the general trend shown in Fig. 9B emerges. Much local variability (between nearby stations) in region < 30–40 m.y., heat flow about 1 HFU in old basins.

We shall now investigate the explanation of heat flow variation + subsidence of topog. + geoid.

Celsius Fahrenheit

Temperature :



100°C — 212°F water boils at sea level

0°C — 32°F water freezes at sea level

Kelvins

also: $T(K) = T(°C) + 273$

temperature above absolute zero

$$\boxed{T(°F) = \frac{9}{5}\,T(°C) + 32}$$

We will always use °C. (or K)

Heat is a form of energy — measured in Joules

$$1 \text{ Joule} = 1 \, \frac{kg \, m^2}{s^2} \qquad (E = mc^2)$$

Power is a measure of the rate at which energy is produced, or consumed, or used, or expended

$$1 \text{ watt} = 1 \, \frac{J}{s}$$

A 60-watt light bulb consumes 60 Joules of energy every second that it is turned on.

There are also antiquated units for energy, which date to the time when heat was thought to be a separate substance — phlogiston

1 calorie is by definition the amount of heat required to raise the temperature of one gm of H_2O by $1°C$.

$$\boxed{1 \text{ cal} = 4.184 \text{ J}}$$

The energy content of food is measured in Calories

cap. C

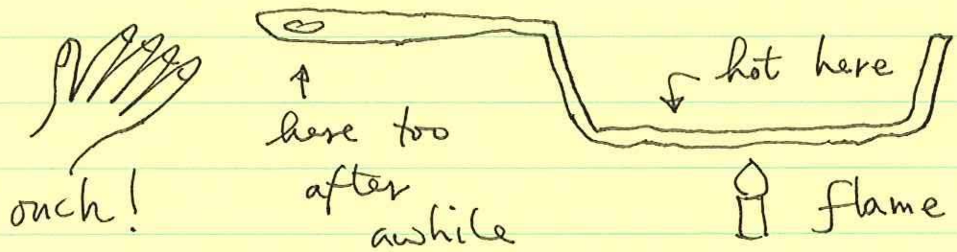
$$1 \text{ Calorie} = 1 \text{ kcal} = 1000 \text{ cal}$$
$$= 4184 \text{ J}$$

There are three modes of heat transfer:

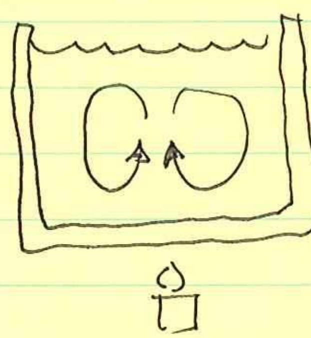
we will study this in 4^{th} quarter of course

1. radiation — this is how we get heat from the sun — in the form of photons — not significant within the $\oplus$ (or any opaque solid material)

2. conduction — this is how a skillet on a stove gets hot



3. convection — put a pot of H_2O on stove

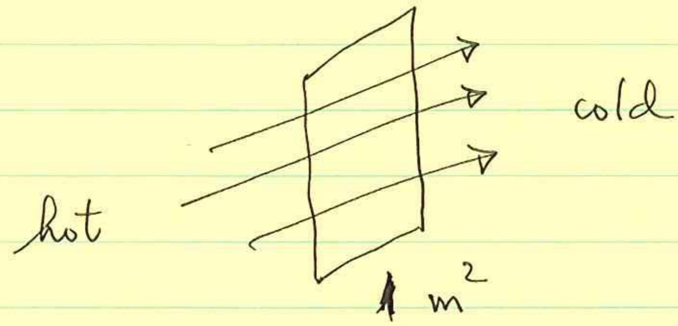


before it boils — by conduction

after it boils — heat is <u>conv</u>ected to top — physical exchange of hot & cold water.

Both conduction & convection are important in the Earth — conduction dominates in the uppermost ~ 100 km (except in geothermal areas such as Yellowstone and Iceland)

Heat conduction — consider the amount of heat flowing through an area in a given time

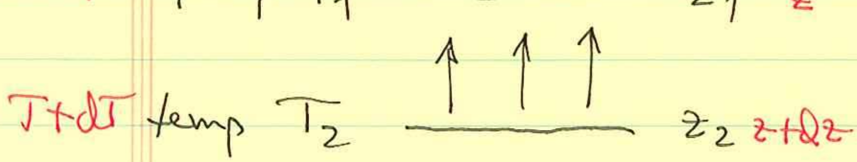


units of heat flow

$J/m^2\text{-}s = W/m^2$

Heat flow within the ⊕ is small (usually expressed as mW/m^2).

Consider the skillet again



$T_2 > T_1$

$$\underline{q} = k\frac{T+dT-T}{z+dz-z} = k\frac{dT}{dz}$$

heat flow $\underline{q} = k \dfrac{T_2 - T_1}{z_2 - z_1}$ or

$$\underline{q} = k\frac{dT}{dz}$$

k = kappa

thermal conductivity

temperature gradient °C/m

The thermal conductivity κ is a <u>material</u> property

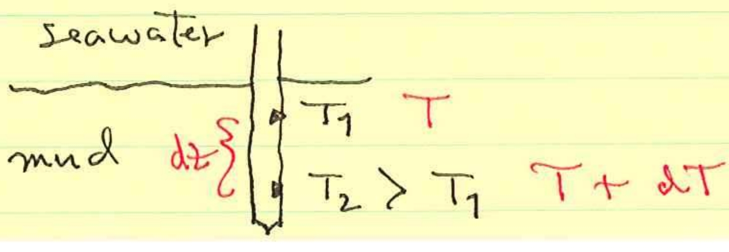
units : $\dfrac{W/m^2}{{}^\circ C/m} = \dfrac{W}{m\,{}^\circ C}$

material	κ ($W/m\,{}^\circ C$)	
Cu	400	good skillets
Al	240	← forget this one
Fe	80	
dry upper mantle rocks (high T olivine)	3	— typical of hard dry rocks
water	0.6	
soil	about the same	
styrofoam	0.03	— good insulation

↑ factor of 10^5 variation ↓

<u>conductors</u> vs. <u>insulators</u>

Heat flow from the surface of the ⊕ is most easily measured in the <u>oceans</u>. Probe 3-10 m in length



$$q = \kappa \frac{dT}{dz}$$

Typically $\kappa_{mud} \sim 0.5 \dfrac{W}{m\,^\circ C}$

$dT/dz \sim 0.1\ ^\circ/m$ — easily measured — very stable thermal environment: no diurnal fluctuations as on land or annual

on land need much deeper holes to get below seasonal cycle

Mean surface heat flow

$$\boxed{\bar{q} = 60 \frac{mW}{m^2}}$$

— heat flow through $1000\ m^2$ ↓ $\frac{1}{3}$ football field $= 30m \times 30m$ would heat one 60W bulb

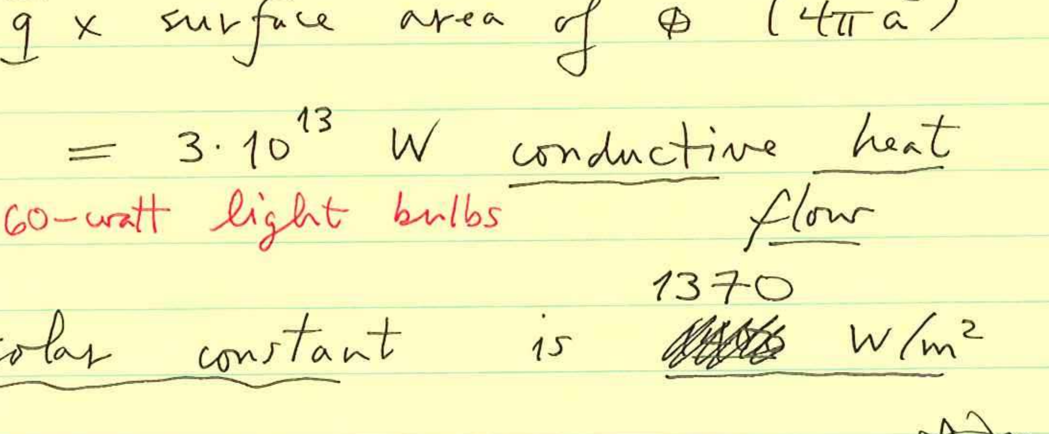
In comparison: $5 \cdot 10^{14}\ m^2$

$\bar{q} \times$ surface area of ⊕ $(4\pi a^2)$

$= 3 \cdot 10^{13}$ W <u>conductive</u> <u>heat flow</u>

enough to light $5 \cdot 10^{11}$ 60-watt light bulbs

The <u>solar constant</u> is 1370 W/m^2



Averaged over $\oplus$ surface: $\frac{1370}{4} = 340\ \mathrm{W/m^2}$

Total solar energy striking the $\oplus$

$1.7 \cdot 10^{17}$ W

another comparison — total world fossil fuel consumption $\sim 10^{13}$ W

of this $\sim 70\%$ absorbed by atmosphere and surface, $\sim 30\%$ reflected back into space

4000 times more energy absorbed from sun than escaping from interior.

fossil fuel consumption $\sim 1/3 \times$ total terrestrial heat flow

The thermal conductivity of typical crustal rock (not porous mud) is $\kappa_{rock} = 3\ \mathrm{W/m\,^\circ C}$.

so that $3 \cdot 20 \cdot 10^{-3} = 60\ \frac{\mathrm{mW}}{\mathrm{m^2}}$

$$\boxed{\text{Typical } \underline{\text{geotherm}} \quad \cancel{25}\,^\circ\mathrm{C/km} \quad 20\,^\circ\mathrm{C/km}}$$

gets hotter by this amount every km we drill down — on land or in basalt beneath seafloor seds

How hot would it be at center of $\oplus$ if this gradient were maintained all the way down?

$\cancel{25}\ 20\ ^\circ C / km \times 6371\ km$

$= \cancel{160{,}000\ ^\circ C}\ 130{,}000\ ^\circ C$!

in fact this was widely believed until late 19th century (Kelvin & Darwin — fortnightly tides)

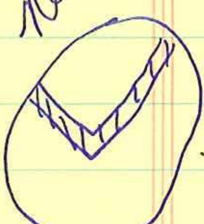
Almost entire interior of ⊕ would be molten if this were true

<u>solidus</u> — T (°C) at which the lowest-melting component of a multi-component system melts.

<u>liquidus</u> — T(°C) at which all components have melted

Discuss ellipsoidal shape of ⊕: Newton

Seismic evidence for solidity of mantle ⟶ temperature gradient must decrease with depth



<u>Roughly speaking</u>, the temperature increases by ~ ~~25~~ 20°C / km down to about ~~[illegible]~~ 50 km depth.

The uppermost ~ 100 km of the ⊕ surface is the region where the heat transport from the interior is <u>dominated</u> by conduction lithosphere : mechanically strong; cannot flow in response to stresses; heat transport dominated by conduction

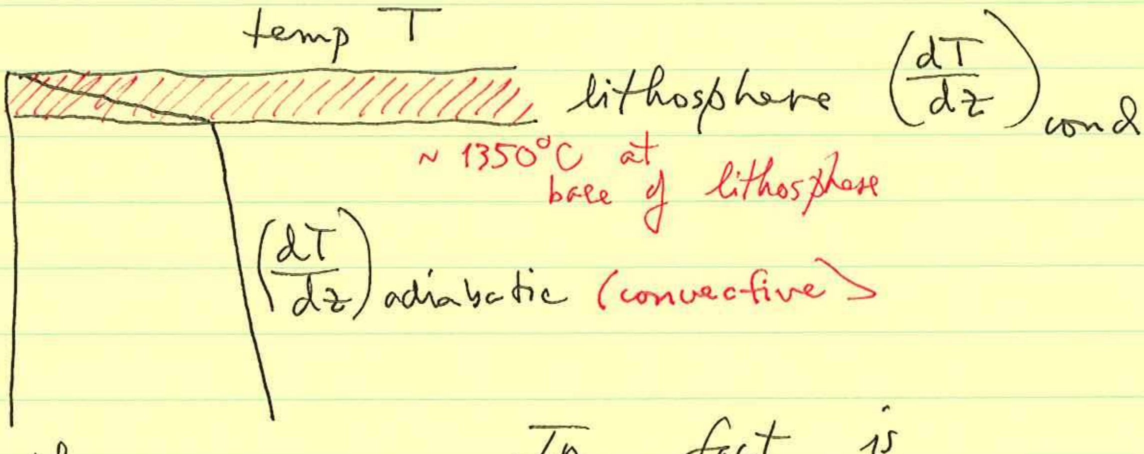
asthenosphere (and underlying mantle) in a state of active convection this is the dominant mode of heat transfer

then: 1350° + (0.6)(2800 km) ≈ 3000°C at CMB.

$$\boxed{dT/dz \sim 0.6 \ \mathrm{C/km}}$$

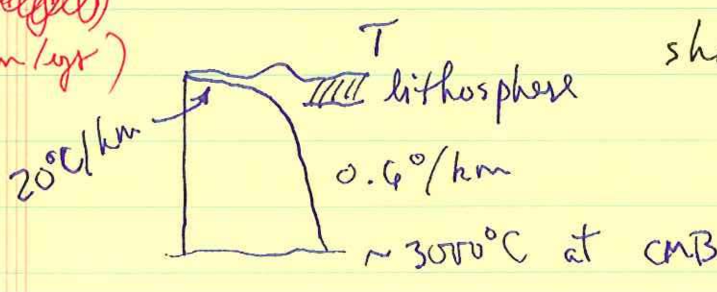
↑ adiabatic gradient

10 – 100 times less than conductive gradient



note – the (subsolidus) convective velocities are very slow (cm/yr)

In fact is rounded off as shown in J&R Figure 8.12



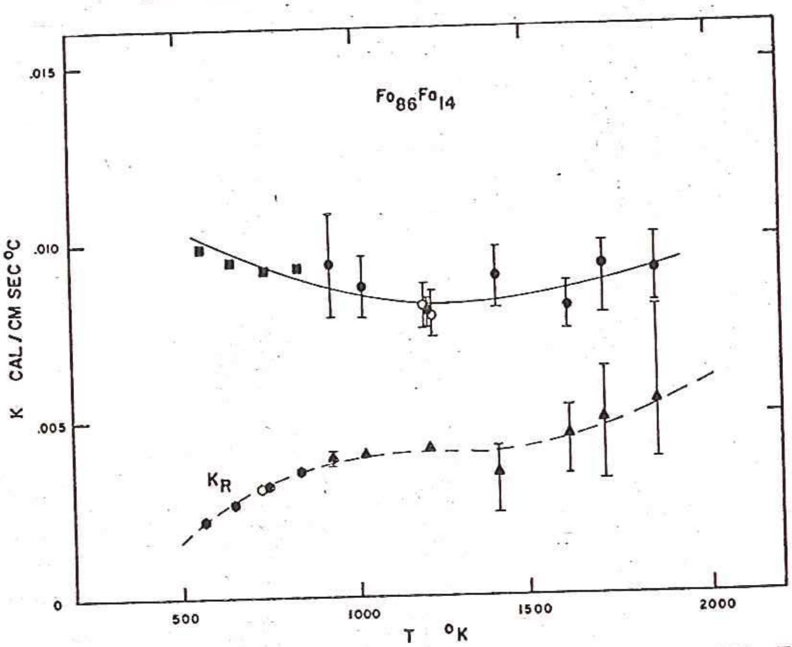


Fig. 5*b*. Total and radiative thermal conductivities in olivine single crystal $Fo_{86}Fa_{14}$. Open circles are points measured at descending temperature. Solution below 900°K are obtained by using assumed values of κ_L.

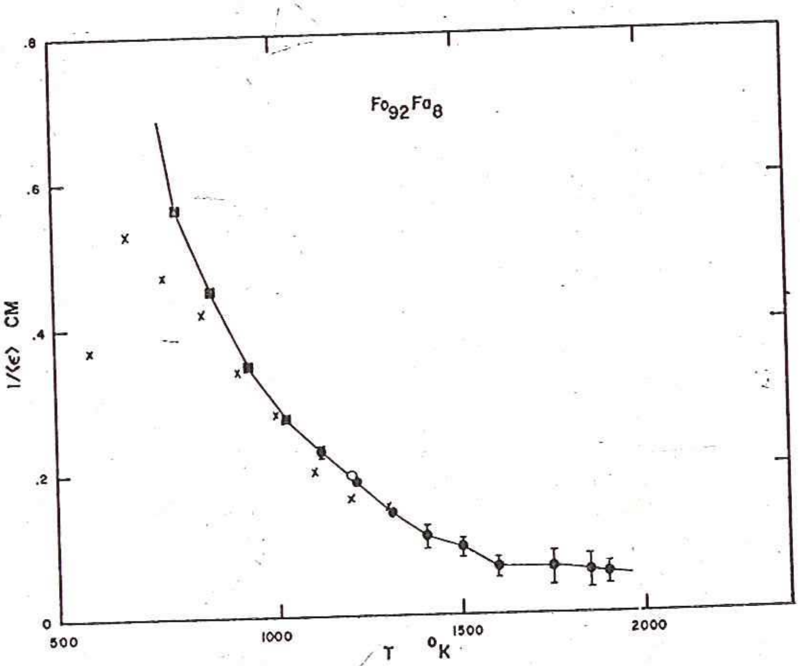
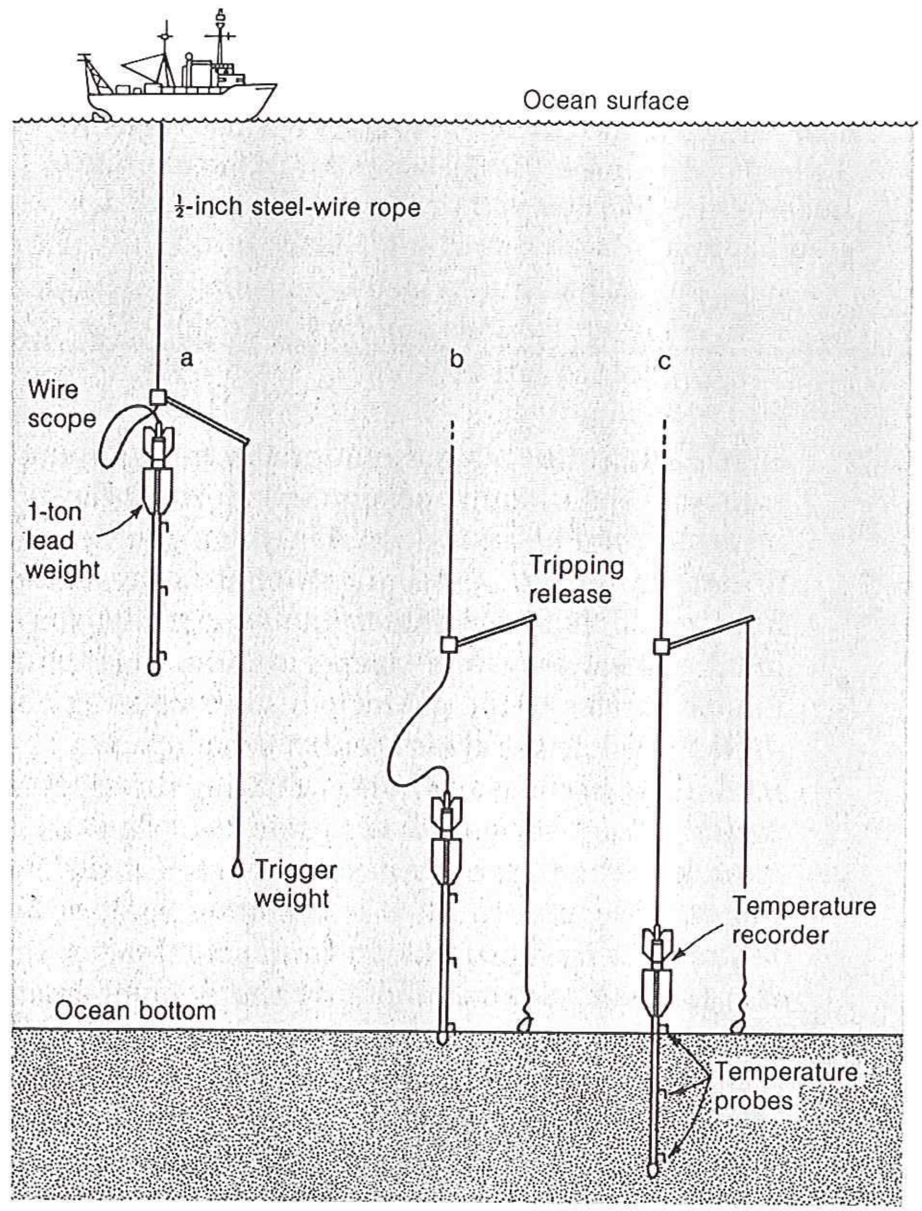


Fig. 6*a*. Photon mean free path in olivine single crystal $Fo_{92}Fa_8$. The open circle is a point measured at descending temperature. Solutions below 1100°K are obtained by using assumed values of κ_L. The crosses are data of *Fukao et al.* [1968] for a crystal of olivine composition $Fo_{88}Fa_{12}$.



'igure 13-6
Ieat flowing out of the sea is measured by plunging . core tube about 10 meters long into the sediments. 'hermometers on the side of the tube record the emperature increase with depth, and the thermal onductivity of the sediments is measured when the ore is retrieved. The product gives the heat flow.

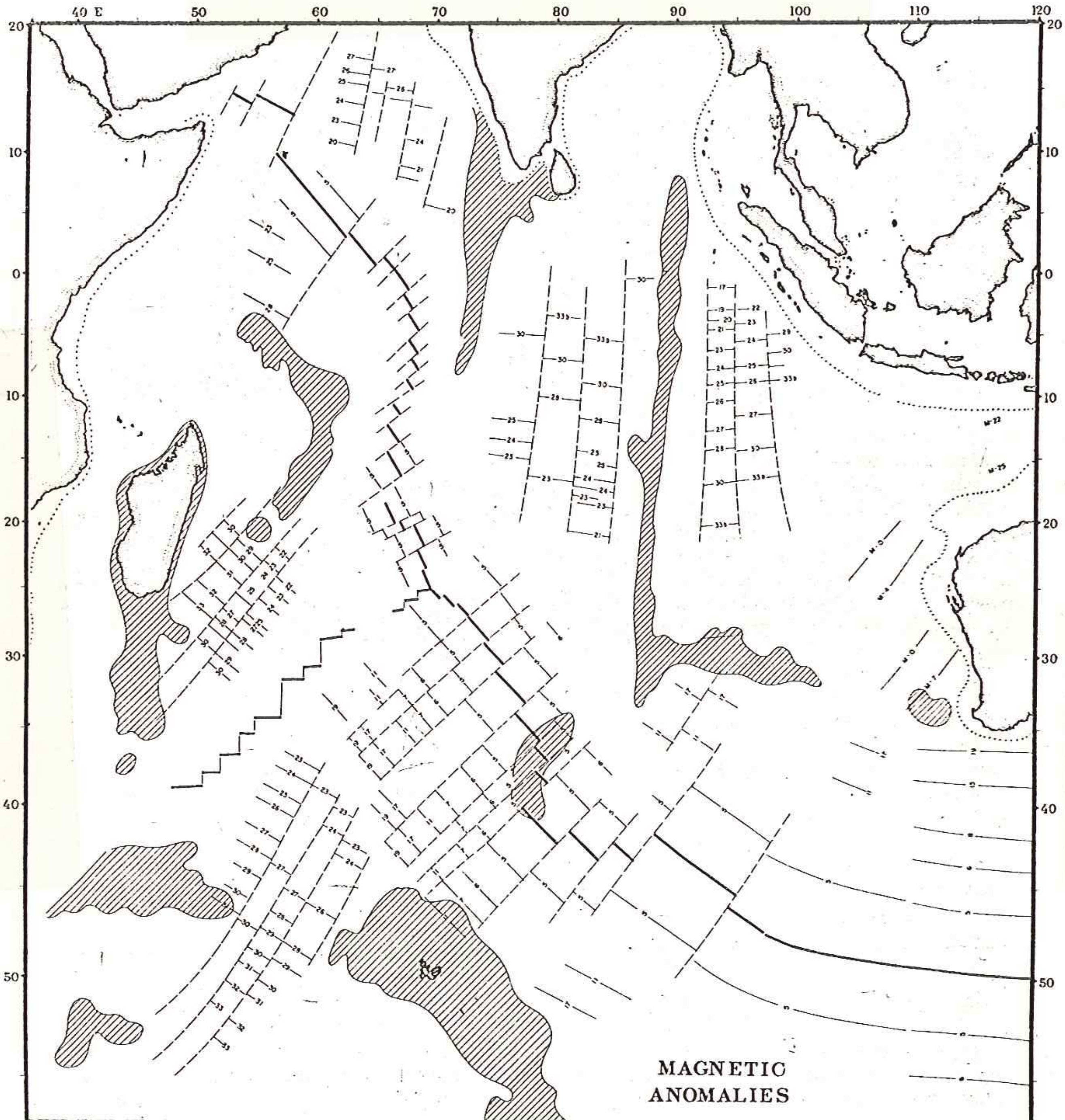


Fig. 6. Identifiable magnetic anomalies in Indian Ocean. For sources, see text. Hachures represent shoal regions of Figure 1.

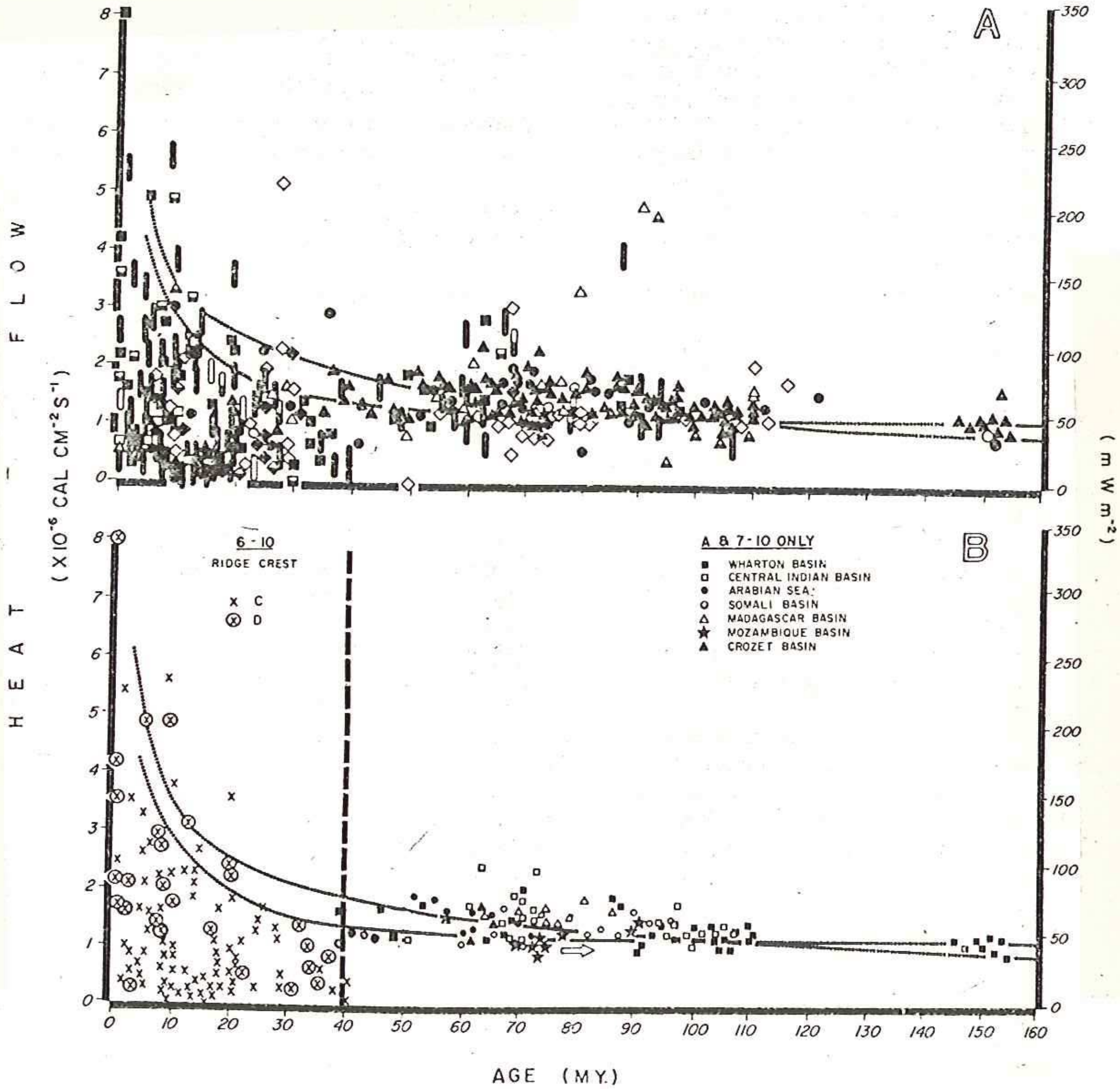


Fig. 9. Heat flow versus age in the Indian Ocean. (*a*) All the data. (*b*) Filtered data. Notice that the filtering removes much scatter but does not change the mean of values older than 50 m.y. B.P. The arrow refers to the Mozambique Basin values that appear to be from older sea floor than that indicated by DSDP hole ages in the basin. Solid curves are from *Sclater and Francheteau* [1970] and *Parker and Oldenburg* [1973].

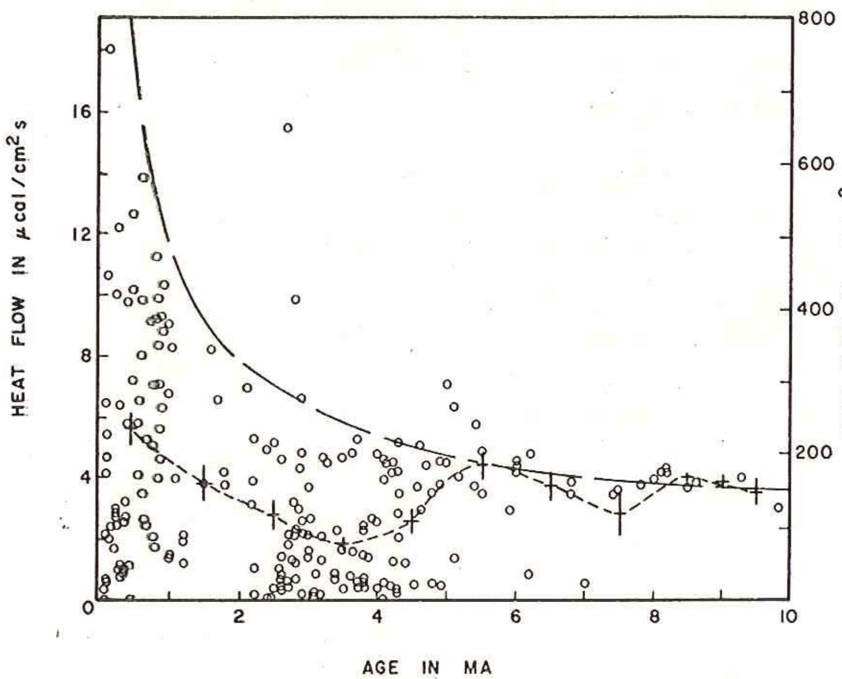


Fig. 5. Heat flow values plotted as a function of age on the Galapagos spreading center. Only those values which are on oceanic crust of well-defined age were used for this plot. Circles represent heat flow values. Pluses are 1-m.y. means. The long-dashed curve is the heat flow expected from the thermal model of *Parsons and Sclater* [1977], and the short-dashed curve connects the mean of the observed data [after *Anderson and Hobart*, 1976].

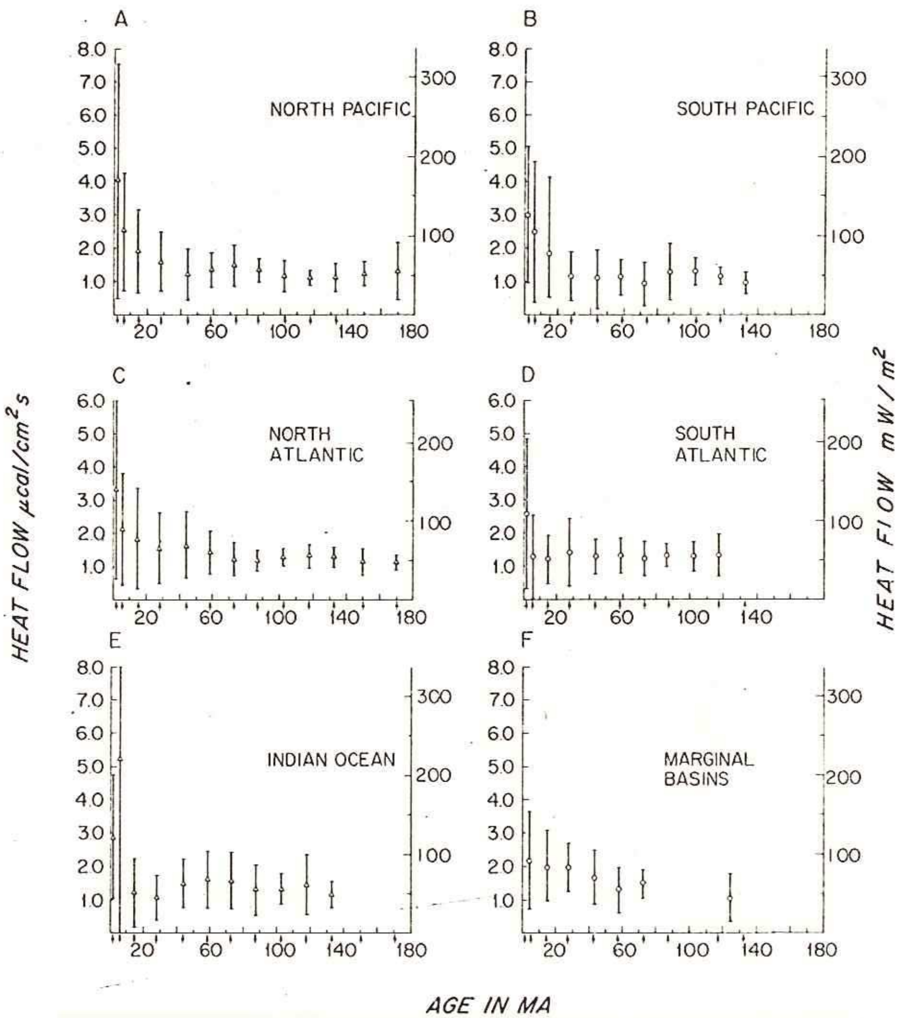
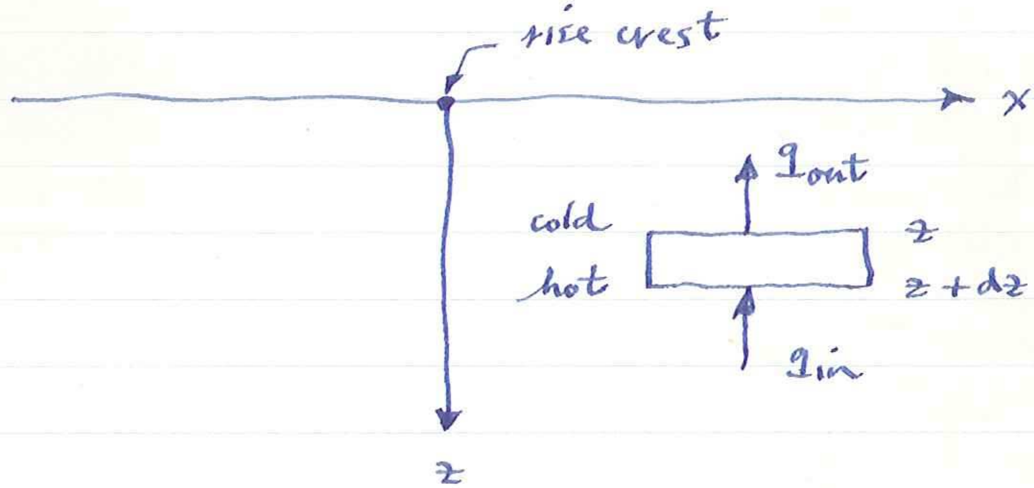


Fig. 1. Mean heat flow and standard deviation as a function of age for five major oceans and the marginal basins.

Thermal evolution of the oceanic lithosphere

We need to work out the equation which describes the cooling of the lithosphere.

To do this we shall neglect the surface topography and consider the cooling of a half-space. This can be justified since the cooling extends down to depths $\gg$ the variation in topography.



1m x 1m Consider a small block of material 1 cm $\times$ 1 cm laterally between z and $z + dz$ in depth.

We'll assume there is no heat flow through the sides of the block.

This can be justified since the isotherms are nearly flat except very near the rice crest.

Let $\mathcal{U}$ be the total heat energy content in the box (the internal energy, to be more precise).

The rate of change of $\mathcal{U}$ with time is given by

$$\frac{d\mathcal{U}}{dt} = \underset{\text{bottom}}{q_{in}} - \underset{\text{top}}{q_{out}}$$

$[q]$ = ergs / cm^2 - sec or J/m^2-sec or W/m^2
$[\mathcal{U}]$ = ergs or calories or J

But from Fourier's law

$$q_{out} = k\,(dT/dz)_{z}$$

$$q_{in} = k\,(dT/dz)_{z+dz}$$

Thus

$$dQ/dt = k\left[(dT/dz)_{z+dz} - (dT/dz)_z \right]$$

Now we must relate dQ/dt to the rate of change of temperature.

The quantity which does this is called the specific heat (at const. pressure) c_p.

By defn, c_p = the amount of heat energy required to raise the temp. of one gm. of material by 1°C.

The specific heat of water is

$$c_p(H_2O) = 1 \text{ cal / gm °C; at STP}$$

This is the definition of the calorie. In SI units

$$c_p(H_2O) = 4184 \text{ J / kg °C.}$$

The specific heat of a typical igneous rock is about one-quarter this. Sclater and Parsons (1977) adopted value:

$c_p = 1170 \frac{J}{kg°C}$ $c_p = 0.28$ cal / g °C.

The specific heat of all inorganic <u>solids</u> at room temp or above is about the same if expressed in terms of cal / mole °C.

The "law" of Dulong and Petit:

$$c_p \text{ (solids)} \sim 6 \text{ cal / mole °C.}$$

i.e. mol. wt. per atom

The average molecular weight of olivine is $M \sim \cancel{50}\,20$, which gives

$$c_p \sim 0.3 \text{ cal / g °C.}$$

Mg_2SiO_4

The "law" of Dulong and Petit is very well understood in terms of the energy of molecular excitation.

$48+28+64$
$= 140 \div 7$
$= 20$

We thus have

$$dU/dt = \rho \cdot c_p \cdot 1^2 \cdot dz \cdot (dT/dt)$$

$$\text{cal/sec} = \frac{\text{gm}}{\text{cm}^3} \cdot \frac{\text{cal}}{\text{gm °C}} \cdot \text{cm}^3 \cdot \frac{\text{°C}}{\text{sec}}$$

units check.

$$\rho c_p \frac{dT}{dt} = \kappa \underbrace{\left[\frac{(dT/dz)_{z+dz} - (dT/dz)_z}{dz} \right]}_{\text{looks like derivative}}$$

Now taking limit as $dz \to 0$ we get

$$\rho c_p \frac{dT}{dt} = K \frac{d^2 T}{dz^2}$$

Actually T is a function of both z and t and the usual notation for the above derivatives is the <u>partial</u> derivative

$$\rho c_p \frac{\partial T}{\partial t} = K \frac{\partial^2 T}{\partial z^2}$$

In the above argument we have neglected the possibility of <u>radioactive heating</u>. The decay of K, U and Th in the rocks can also <u>increase</u> the heat content of our block.

Let $\cancel{h}$ A = <u>heat productivity</u> by radioactive heating, units cal / cm^3-sec (amount of energy or heat added per second). Better to call this A for later agreement.

This can be measured by measuring <u>radioactivity</u> <u>content</u> of rocks.

Clearly ~~h~~ A must be added to our eqn as follows:

$$\boxed{\underbrace{\rho c_p \frac{\partial T}{\partial t}}_{\substack{\text{total rate}\\ \text{of change of}\\ \text{heat energy}\\ (\text{cal}/\text{cm}^3-\text{sec})}} = \underbrace{k \frac{\partial^2 T}{\partial z^2}}_{\substack{\text{change due}\\ \text{to heat}\\ \text{flow thru}\\ \text{top + bottom}}} + \underbrace{\cancel{h}\ A}_{\substack{\text{change due}\\ \text{to radio-}\\ \text{active}\\ \text{heating}}}}$$

call this A instead

This is partial differential equation: called the 1-d heat flow (or diffusivity) eqn.

In oceanic crust and lithosphere A ~~h~~ is quite small and can be neglected: Table 7.2 of Stacey, page 186.

material	~~h~~ A instead (10^{-6} W/m^3)
due to U, Th, K large radius elements granites	2.8
tholeitic basalts	0.079 (35 times smaller)
dunites, peridotites	0.005 (600 times smaller)

Contribution from radioactive decay more important in granitic cont. crust, negligible in mantle and in oceanic crust.

Rewrite equation as

$$\boxed{\partial T/\partial t = k\, \partial^2 T/\partial z^2} \quad *$$

where $\boxed{k = \kappa/\rho c_p, \text{ the thermal diffusivity}}$

Units of k: $\dfrac{\text{cal}}{°\text{C. cm sec}} \times \dfrac{\text{cm}^3}{\text{gm}} \times \dfrac{\text{gm }°\text{C}}{\text{cal}}$

$$= \frac{\text{cm}^2}{\text{sec}}$$

This agrees with units in equation:

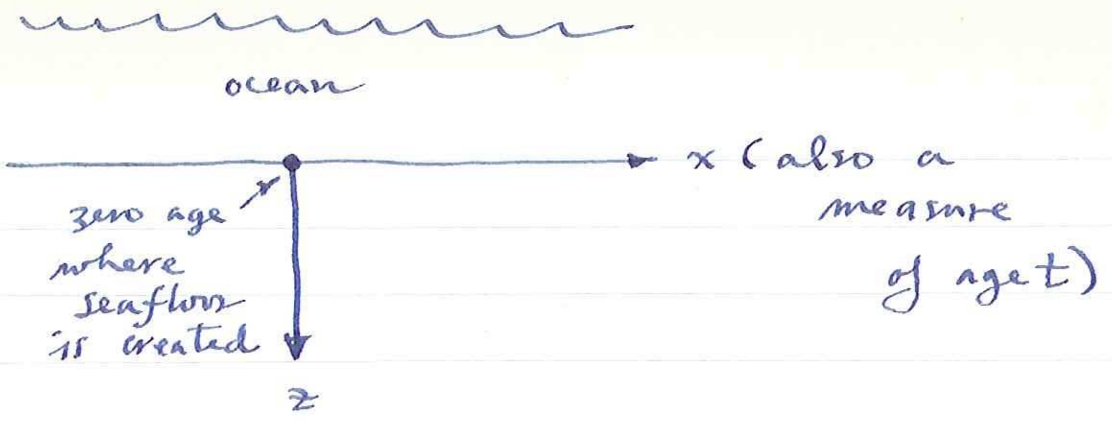
$$\frac{°\text{C.}}{\text{sec}} = \frac{\text{cm}^2}{\text{sec}} \times \frac{°\text{C.}}{\text{cm}^2}$$

i.e. 0.8 mm^2/sec

Typical value $k \sim 0.008$ cm^2/sec

The p.d.e. * is just about the world's simplest p.d.e. Whole books have been written on how to solve it in more complicated geometries.

We know we need boundary and initial conditions to be specified.



We specify $T(0, t) = 0$ temp on sea bottom just 1° or 2° above 0°C.

Also $T(\infty, t) = T_m$, ambient temperature in mantle, much hotter of order 1300°C.

Also $T(z, 0) = T_m$ temp at zero age is T_m at all depths.

This is known to be just the right ~~number~~ kind of b.c. and i.c. for the heat flow eqn — a parabolic p.d.e.

Instead of stopping to learn how to solve p.d.e.'s let us just verify the solution.

Consider, then, the solution

$$\boxed{T(z,t) = \frac{2}{\sqrt{\pi}}\, T_m \int_0^{z/\sqrt{4kt}} e^{-u^2}\, du \qquad **}$$

Let us verify this is the solution:

1. $T(0,t) = \int_0^0 du$ vanishes, satisfies b.c. on top.

2. $T(\infty,t) = T(z,0)$

$$= \frac{2}{\sqrt{\pi}}\, T_m \underbrace{\int_0^\infty e^{-u^2}\, du}_{\text{definite integral has value } \sqrt{\pi}/2.}$$

$= T_m$. Thus temp at 0 age and at ∞ depth is that of ambient mantle.

3. $$\partial T/\partial t = \frac{2}{\sqrt{\pi}}\, T_m\, e^{-z^2/4kt} \left(-\tfrac{1}{2} t^{-3/2}\right) \frac{z}{\sqrt{4k}}$$

$$= -\frac{T_m}{2\sqrt{\pi}}\, e^{-z^2/4kt} \left(\frac{z}{\sqrt{k}\, t^{3/2}}\right)$$

$$\partial^2 T / \partial z^2 = \frac{2}{\sqrt{\pi}} T_m \, e^{-z^2/4kt} \frac{1}{\sqrt{4kt}} \left(-\frac{2z}{4kt}\right)$$

$$= -\frac{T_m}{2\sqrt{\pi}} \, e^{-z^2/4kt} \left(\frac{z}{\sqrt{k}\, t^{3/2}}\right) \frac{1}{k}$$

Thus $\partial T / \partial t = k \, \partial^2 T / \partial z^2$, so ** is our solution.

Conventional notation: define the so-called <u>error function</u> (arises in theory of Gaussian errors — <u>not</u> named after the ~~...~~ famous mathematician Error) Wolfgang Windgassen Error

$$\boxed{\operatorname{erf} q = \frac{2}{\sqrt{\pi}} \int_0^q e^{-u^2} \, du}$$

Then the solution can be written

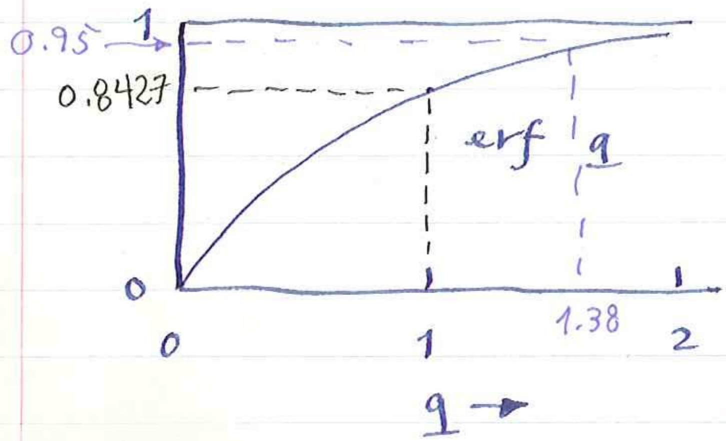
$$\boxed{T(z, t) = T_m \operatorname{erf}\left(z / \sqrt{4kt}\right)}$$

This gives temp. at time t and depth z in a cooling half-space.

Properties of erf:

$$\mathrm{erf}\ 0 = 0$$
$$\mathrm{erf}\ \infty = 1$$

Plot of erf looks like

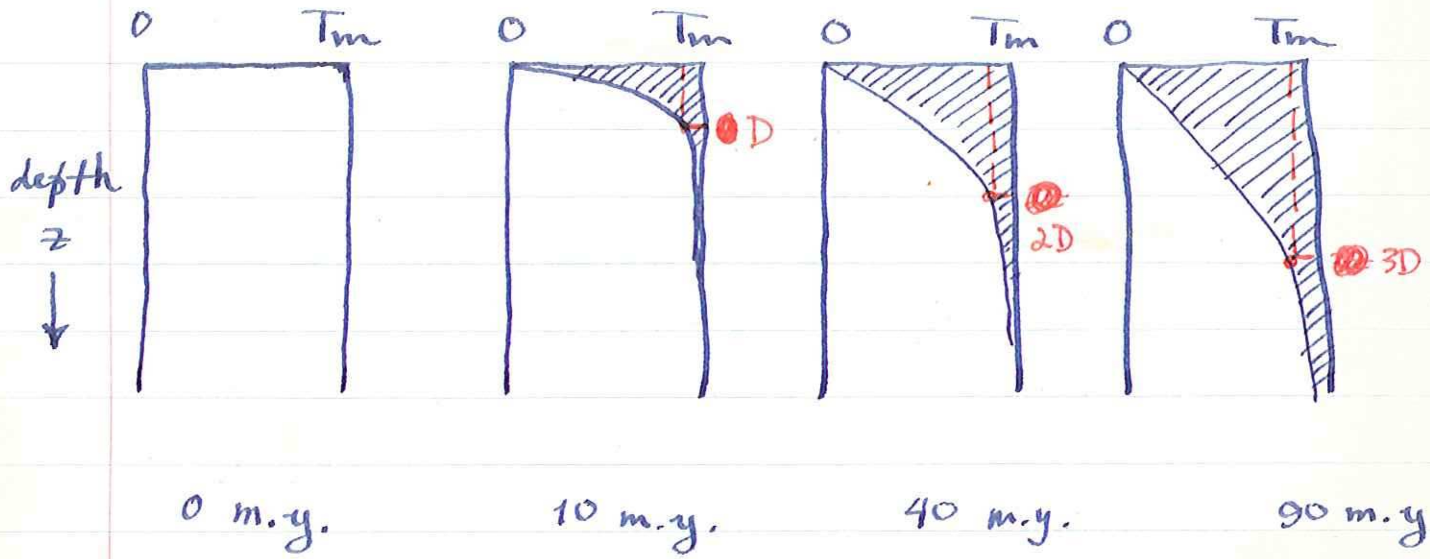


For example

$$\boxed{\begin{aligned} \mathrm{erf}\ 1 &= 0.8427 \\ \mathrm{erf}\ 1.38 &= 0.95 \end{aligned}}$$

Note that T depends only on the combination $z/\sqrt{4kt}$.
This means that at 40 m.y. the cooling has gone down 2 times as far as at 10 m.y. and at 90 m.y. it will have gone down 3 times as far.

Plots of T vs. z at various times thus look like red line denotes $T = 0.95 T_m$



The cooling proceeds like $\sqrt{t}$.

At what depth $\cancel{z}\ D$ is $T = 95\%$ of T_m?

Answer:

$$\cancel{z}\,D / \sqrt{4kt} = 1.38$$

$$\cancel{z}\,D = 2.76\,(kt)^{1/2}$$

We just decide to take this as depth to which appreciable cooling has penetrated.

If $k = 0.008\ \mathrm{cm}^2/\mathrm{sec}$

$$\boxed{\cancel{z}\,D = 13.9\ t^{1/2} \quad \text{where } t \text{ in m.y. and } \cancel{z}\,D \text{ in km}}$$

Heat — a form of energy — measured in joules

$$1\ J = 1\ \frac{kg\ m^2}{s^2}$$

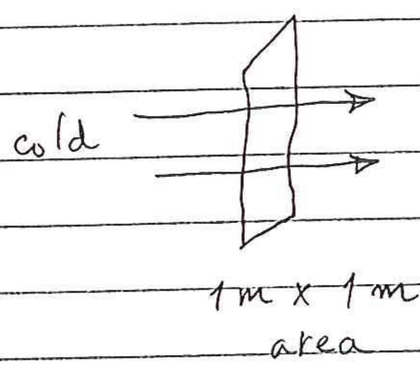
Power — rate at which energy is produced or consumed, etc. — measured in watts

$$1\ W = 1\ \frac{J}{s}$$

Calorie: amount of heat required to raise temperature of 1 gm of H_2O by 1°C.

$$1\ \text{calorie} = 4.184\ J$$

Conductive heat flow within the Earth



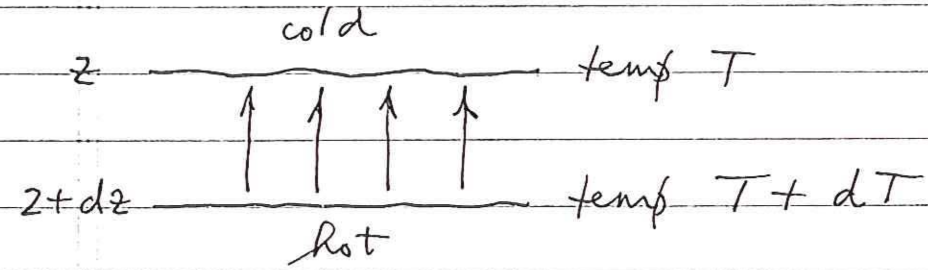
$$\frac{J}{m^2\ s} = \frac{W}{m^2}$$

$$1\ \frac{mW}{m^2} = \frac{1}{1000}\ \frac{W}{m^2}$$

Mean surface heat flow: $\bar{q} = 60\ mW/m^2$

$\bar{q} \times (\text{surface area of } \oplus) = 3 \cdot 10^{13}\ W$

Heat conduction is governed by Fourier's law:



$$\underline{q} = \kappa \left[\frac{T + dT - T}{z + dz - z} \right] = \kappa \frac{dT}{dz}$$

heat flow = thermal conductivity × temperature gradient

Thermal conductivity κ : $\dfrac{W/m^2}{°C/m} = \dfrac{W}{m\,°C}$

material	κ (W/m °C)	
Cu	400	
Al	240	conductors
Fe	80	
dry rock	3	
water	0.6	
soil	~ 0.6	insulators
styrofoam	0.03	

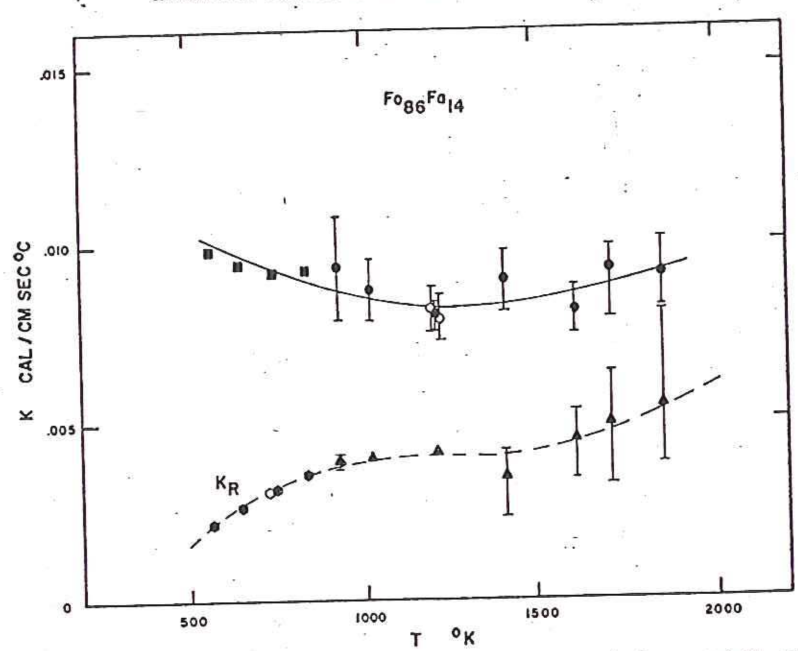


Fig. 5b. Total and radiative thermal conductivities in olivine single crystal $Fo_{86}Fa_{14}$. Open circles are points measured at descending temperature. Solution below 900°K are obtained by using assumed values of κ_L.

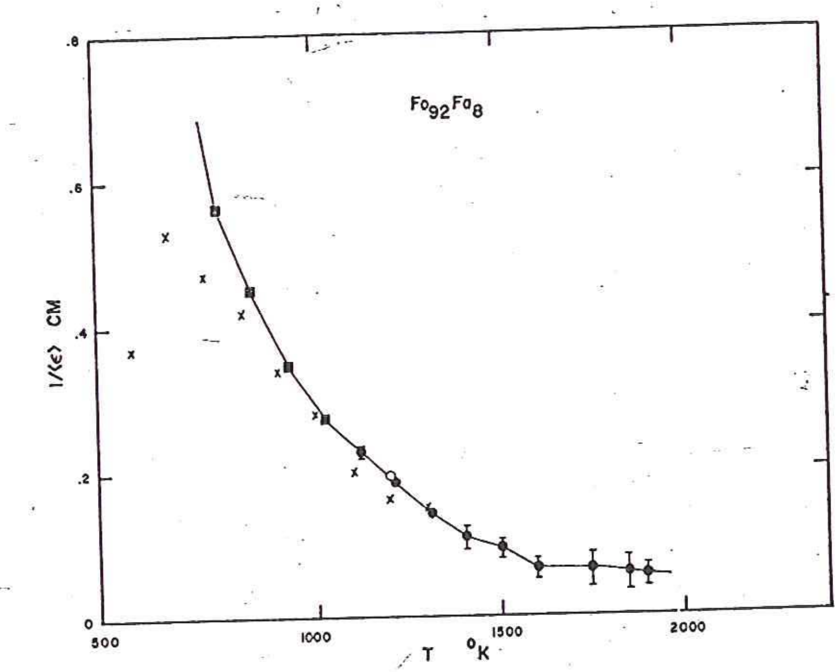
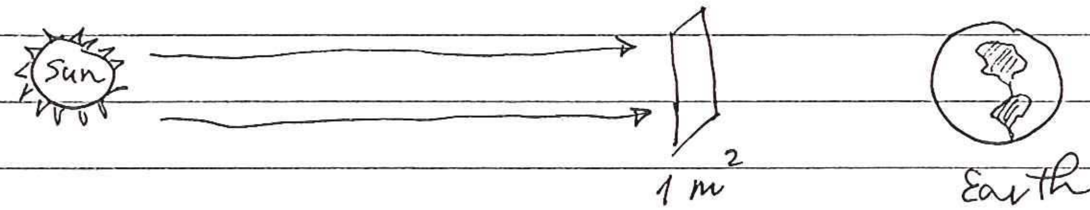


Fig. 6a. Photon mean free path in olivine single crystal $Fo_{92}Fa_8$. The open circle is a point measured at descending temperature. Solutions below 1100°K are obtained by using assumed values of κ_L. The crosses are data of *Fukao et al.* [1968] for a crystal of olivine composition $Fo_{88}Fa_{12}$.

Solar constant: 1370 W/m^2



Averaged over $\oplus$ surface: $\frac{1370}{4} = 340\ \frac{W}{m^2}$

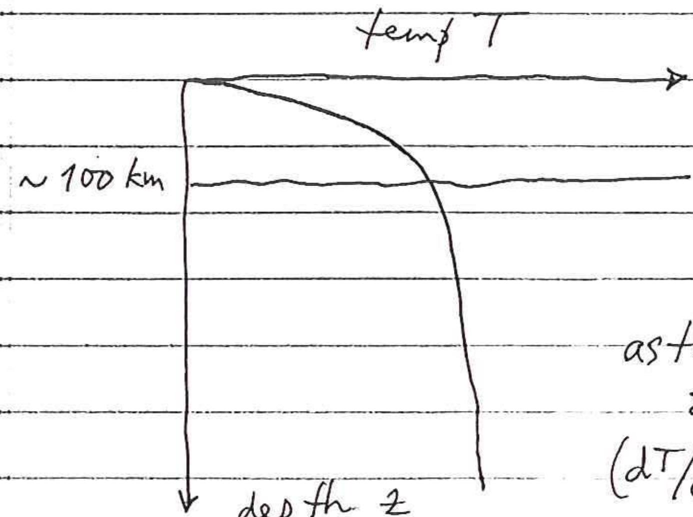
Total solar energy striking the $\oplus$:

$$340 \times 4\pi\,(\text{radius})^2 = 1.7 \cdot 10^{17}\ W$$

Albedo: 30% reflected back into space

Solar energy incident at $\oplus$ surface
= 4000 × (total conductive heat flow from interior)

Typical geotherm at $\oplus$ surface: 20 °C/km



lithosphere: dominated by conduction
$dT/dz \sim 20\,^\circ C/km$

asthenosphere: dominated by subsolidus convection
$(dT/dz)_{adiabatic} \sim 0.6\,^\circ C/km$

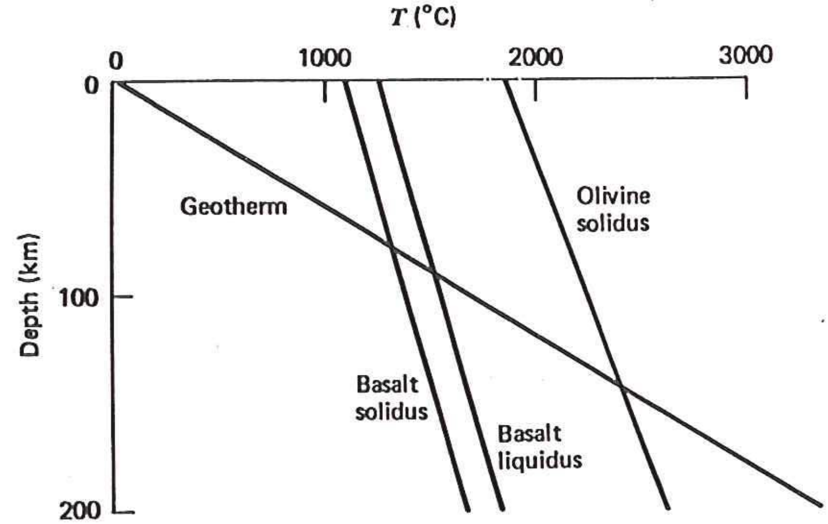
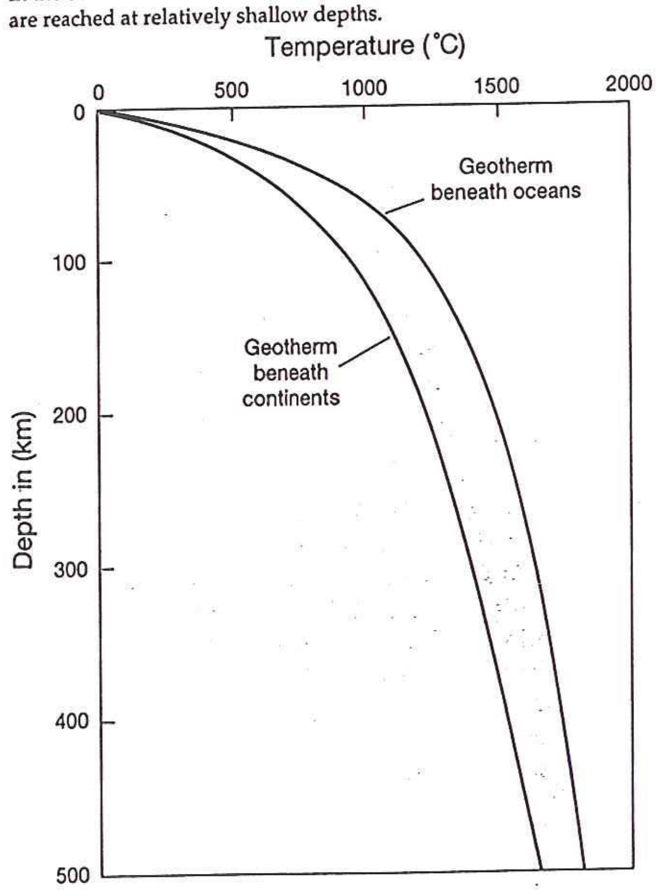
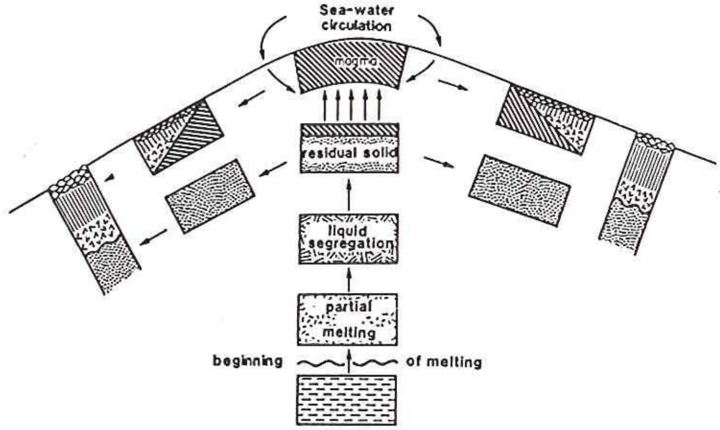


Figure 4-8 Temperature as a function of depth within the earth assuming heat transport is by conduction (conduction geotherm). Also included are the solidus and liquidus of basalt and the solidus of peridotite.

FIGURE 8.12 Estimated average geotherms in continental and oceanic lithosphere. The convecting mantle is nearer to the surface in the ocean basins than under the continents, so high temperatures are reached at relatively shallow depths.





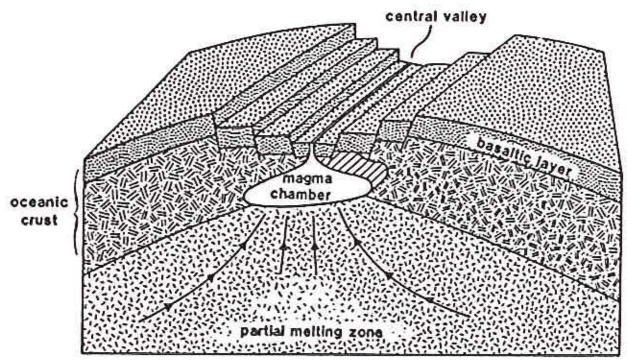


FIGURE 55 The creation of oceanic lithosphere. Material from the mantle rises beneath a ridge axis. During its ascent melting occurs and the liquid thus formed rises faster than surrounding unmelted material. Near the surface the liquid (which forms the crust) cools and interacts with seawater. Then the liquid and residual material spread horizontally away from the spreading axis. On the bottom is a block diagram of a slow-spreading ridge (spreading at, say, a rate on the order of 2 centimeters per year) that shows how the processes of lithosphere creation may occur in a more realistic scenario.

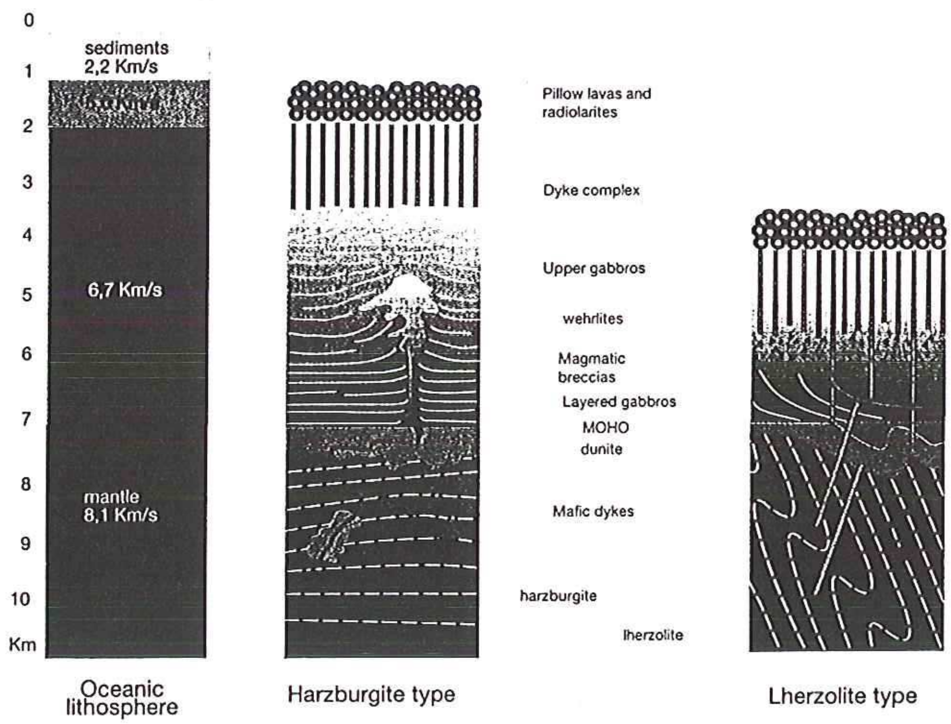


Figure 5.8

Columns comparing the structure of the oceanic crust as defined seismically with the two main types of ophiolites: the harzburgite type as illustrated by the Oman ophiolites and the lherzolite type, by the Trinity ophiolites of California. (After F. Boudier and A. Nicolas 1985, Earth Planet. Sci. Lett., 76, 84–92)

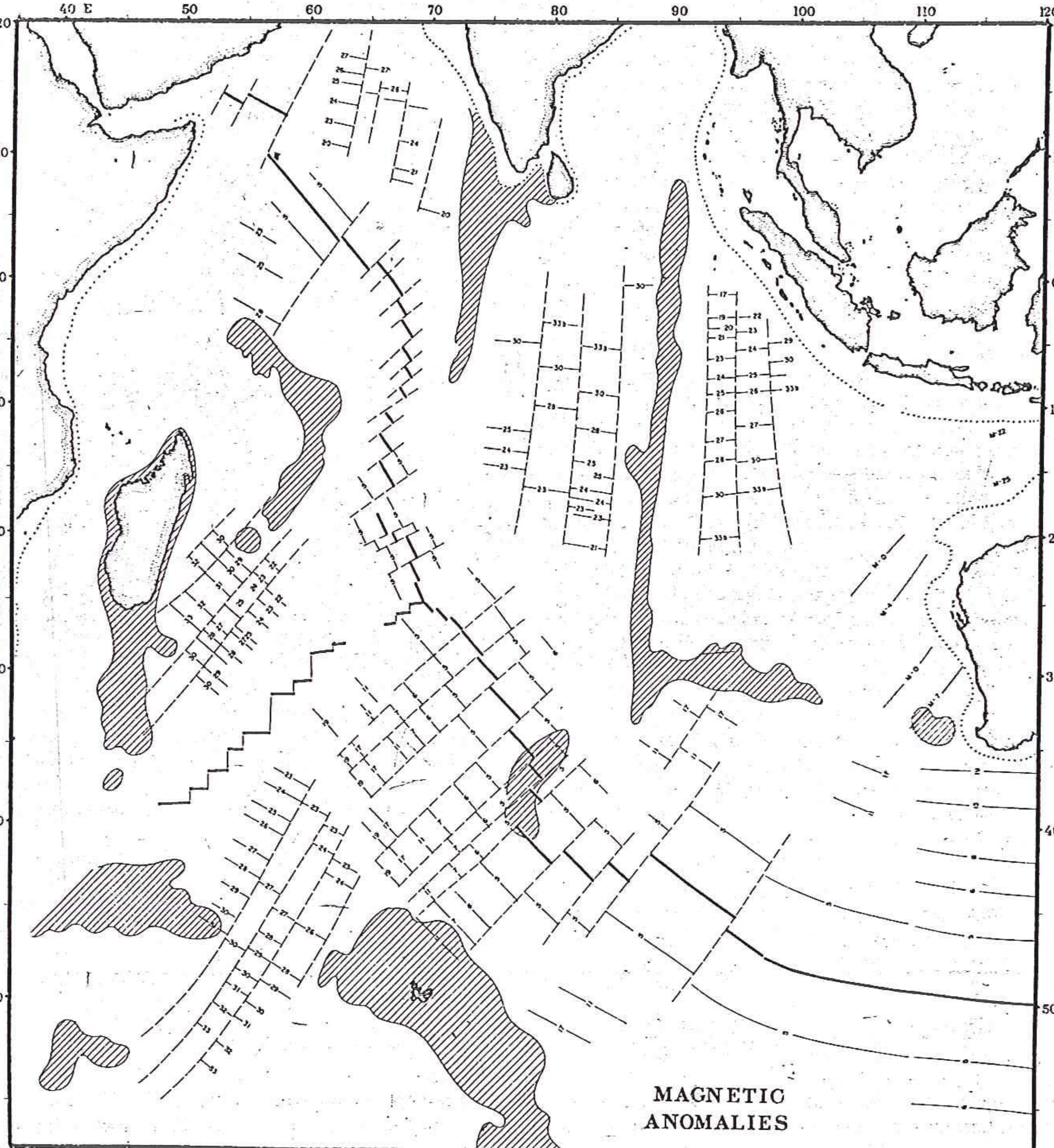


Fig. 6. Identifiable magnetic anomalies in Indian Ocean. For sources, see text. Hachures represent shoal regions of Figure 1.

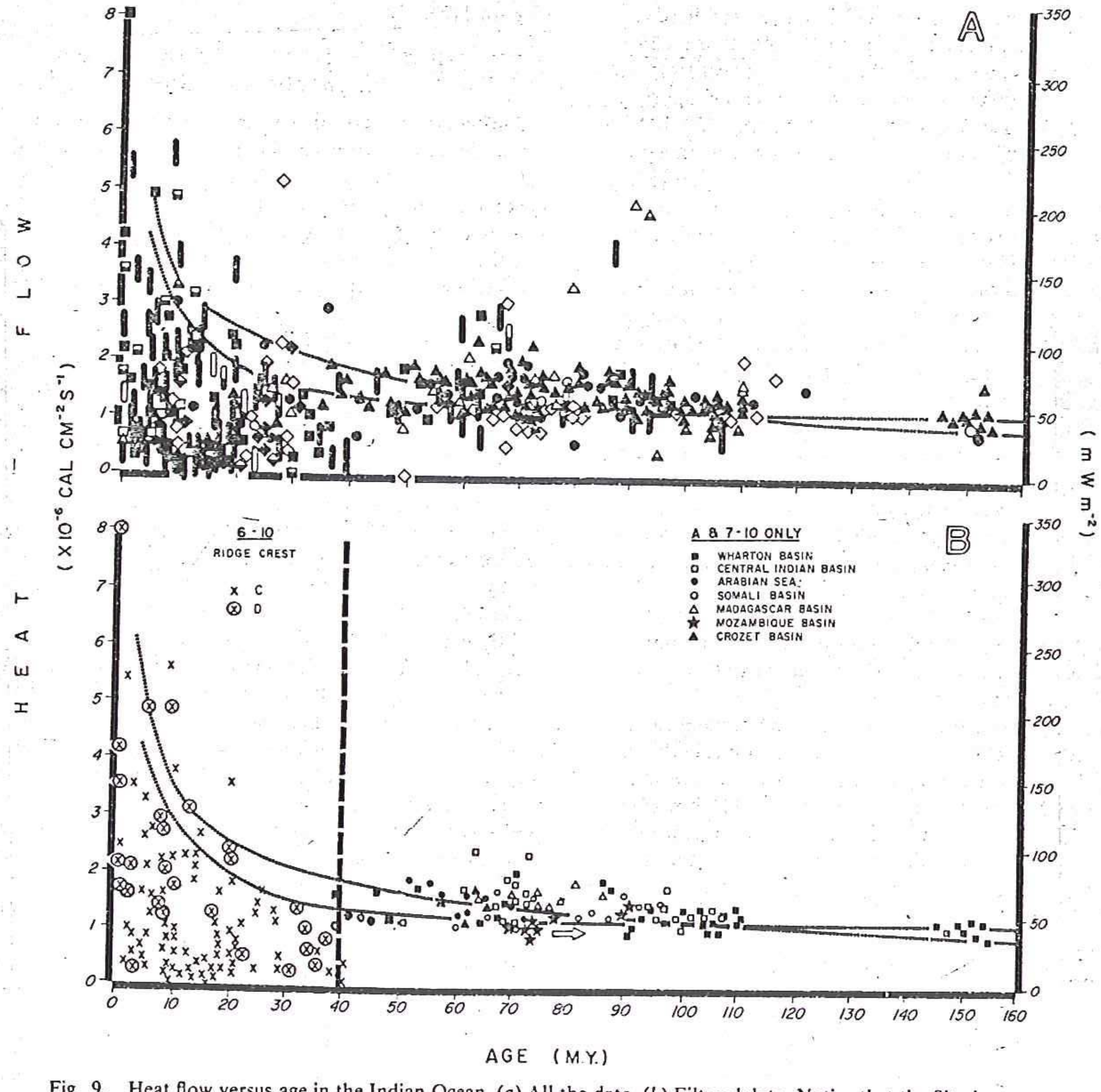


Fig. 9. Heat flow versus age in the Indian Ocean. (*a*) All the data. (*b*) Filtered data. Notice that the filtering removes much scatter but does not change the mean of values older than 50 m.y. B.P. The arrow refers to the Mozambique Basin values that appear to be from older sea floor than that indicated by DSDP hole ages in the basin. Solid curves are from *Sclater and Francheteau* [1970] and *Parker and Oldenburg* [1973].

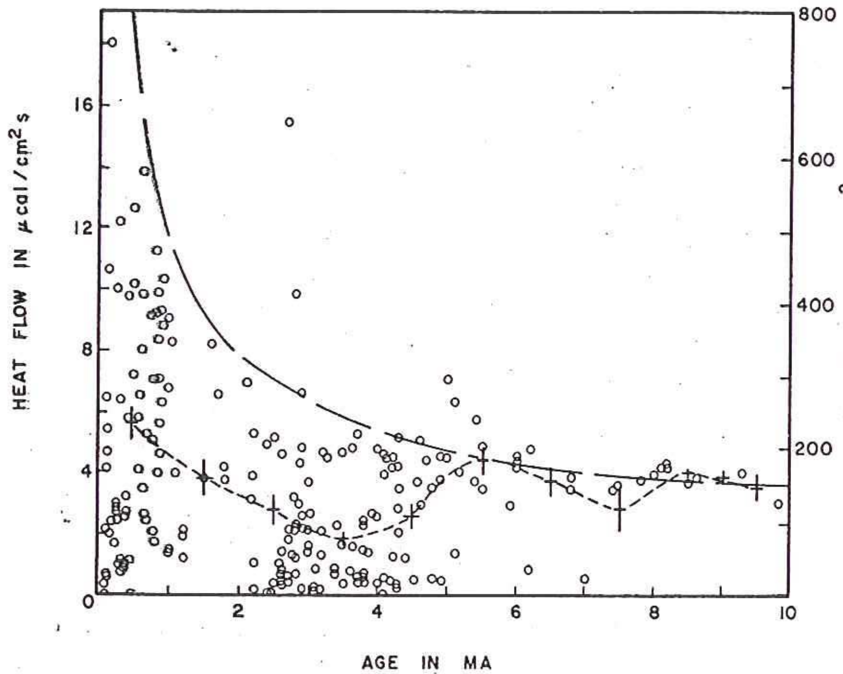


Fig. 5. Heat flow values plotted as a function of age on the Galapagos spreading center. Only those values which are on oceanic crust of well-defined age were used for this plot. Circles represent heat flow values. Pluses are 1-m.y. means. The long-dashed curve is the heat flow expected from the thermal model of *Parsons and Sclater* [1977], and the short-dashed curve connects the mean of the observed data [after *Anderson and Hobart,* 1976].