# Heat Flow

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## GLOSSARY

**Geothermal gradient** The rate of increase in temperature with depth in the Earth. By convention, this quantity is positive when increasing downward.

**Paleo-heat flow** The magnitude of heat flow deduced to have been at an earlier time in Earth history.

**Radiogenic heat production** The rate of heat generated in a unit volume of rock by spontaneous radioactive decay of natural unstable isotopes in the rock. The primary heat-producing isotopes in the Earth are $^{232}$Th, $^{238}$U, $^{40}$K, and $^{235}$U (listed in modern order of decreasing importance).

**Tectonothermal age** The age of the last major tectonic or magmatic event to affect a region.

**Terrestrial heat flow** Heat flow from the hot interior of the Earth to its relatively cool exterior and specifically its heat loss through its surface.

**Thermal conductivity** The steady-state rate of thermal energy transfer through a rock in the direction of a unit thermal gradient per unit area and per unit thickness.

**HEAT FLOW** is the energy flow in planets, including Earth, that drives the internal geological processes that build the surface of the planet. These processes include tectonism, mountain building, and volcanism. Although atmospheric processes driven by solar energy have strongly modified the surface of Earth, the primary differences among the surfaces of the terrestrial planets, including Earth's Moon, are associated with their internal thermal budgets and their different mechanisms of heat loss. The smaller bodies, the Moon, Mercury, and Mars, have surfaces dominated by impact craters and volcanism, suggesting that they lost heat early in their evolution through crustal formation, then by volcanism and thermal conduction. Venus, slightly smaller than Earth, is enigmatic. None of its crust is very old, but it lacks the clear distinction of continental crust and oceanic crust seen on Earth, and nothing betrays signs of active volcanism or tectonism. It seems to record one massive event of heat loss by global volcanism and resurfacing sometime in the past half billion years or so, followed by relative quiescence. Perhaps this behavior is periodic in nature, but we have a record of but one event. Earth, the largest of the terrestrial planets, is the most active, and heat flow provides the energy flow driving the rich assortment of geological processes that are associated with plate tectonics.

## I. HEAT FLOW, SOURCE AND THEORY

Terrestrial heat flow is the flow of heat from the hot interior of the Earth to its relative cool exterior. Temperatures at the surface of the Earth are controlled by the balance of energy received from the Sun and reradiated back into space, modified and redistributed by Earth's atmosphere and oceans. Below the surface, however, temperatures increase in response to the outward flow of heat. This thermal energy flow is responsible for the geologically dynamic nature of our planet, and similarly the dynamic natures and histories of other planets and planetary bodies in our solar system. Evidence of the hot interior of the Earth is most readily seen in thermal springs and volcanoes, and can be directly measured in deep mines and drill holes. Global heat loss is about $4 \times 10^{13}$ W, or an average of about 80 mW m$^{-2}$. This energy flux is more than two orders of magnitude less than the energy received by the Earth from the Sun, but as the solar energy is mainly reradiated into space, it has little effect on temperatures deeper than a few meters in the Earth. Effectively, transfer of solar energy controls the temperature of the Earth's surface, and internal heat controls the Earth's internal temperatures.

Earth's internal heat has two basic origins, heat associated with its formation, or primordial heat, and heat derived from the decay of long-lived radioactive isotopes in the Earth. The generally accepted hypothesis for the formation of the solar system, the nebular hypothesis, postulates that Earth was formed by the cold accretion of a rotating, collapsing gas cloud most of which nucleated to form our Sun. The early Earth could have been heated by a number of processes including kinetic energy from the accreting masses, solar heating, short-lived radioactive isotopes, compressional heating, and potential energy released during core formation. There is evidence from other planetary bodies, such as the Moon, that these heat sources would have been sufficient to have raised the temperature of the early Earth close to its solidus temperature with major melting episodes in its rocky mantle probable, and a fully molten iron core. Heat from the events associated with the formation and initial differentiation of the Earth were all released early in Earth history, and the Earth has been slowly losing this heat since.

The long-lived radioactive isotopes, primarily $^{232}$Th, $^{238}$U, $^{40}$K, and $^{235}$U (listed in modern order of decreasing importance), have half-lives of the order of $10^9$ years and would not have had an immediate significant thermal effect after formation of the Earth. Using the Moon as a guide, we can estimate that heat from these radioactive isotopes would probably have started to have a major heating effect about $0.7 \times 10^9$ years after the Earth was formed (about 3.8 billion years ago). This event was recorded on the Moon by a resurgence of volcanic activity with the initiation of eruption of the lunar marie basalts. The approximate timing of this event also roughly coincides with the end of late heavy bombardment on the Moon, the end of a distinct phase of large crater formation about 3.9 billion years ago, and an event that almost certainly also occurred on Earth. Most of the lunar crust (the lunar highlands) predates late heavy bombardment, but all rock units on Earth examined so far have ages approximately synchronous or younger than this event, and show no evidence of shock structures. After heat from the long-lived radioactive isotopes built up in the Earth and reached a state of dynamic equilibrium with heat loss through the surface of the Earth, heat from this source has gradually declined as the absolute quantities of these isotopes have decayed. The modern Earth is estimated to have a balance of about 80% radiogenic heat and 20% primordial heat of Earth formation and differentiation. Most of this heat is lost through the Earth's surface: a small fraction is converted to other forms of energy that drive tectonics, magmatism, and other internal dynamic processes.

Three heat transfer mechanisms are thought to be effective in the Earth: (1) lattice (or phonon) conduction, (2) convection or advection, and (3) radiation (or photon conduction). Over most of the Earth's surface heat is lost by conduction and heat transfer is governed by Fourier's law of conduction. To a first approximation, heat flows vertically from the Earth, and can be represented by the relationship

$$q = -K(\partial T/\partial z), \qquad (1)$$

where heat flow $q$ is given by the product of rock thermal conductivity $K$ and $\partial T/\partial z$, the rate of increase of temperature $T$ with depth $z$ (by convention, $z$ is defined as positive downward). Conductive heat flow $q$ is the parameter commonly quoted with reference to terrestrial heat flow and is the usual quantity measured by geophysicists. In stable regions, this conductive thermal regime is thought to extend to the base of the lithosphere at depths of about 100–200 km. However, heat flow commonly decreases with depth in the lithosphere either due to transient effects or due to radiogenic heat production in the lithosphere, as discussed below.

In localized regions at the Earth's surface, during tectonic deformation of the lithosphere, and below the base of the lithosphere, heat flow is dominated by convection in which heat is transferred by physical movement of Earth materials across a temperature difference (i.e., rock, or ascending and descending fluids, such as water or magma). Heat flow by convection is given by

$$q = \Delta c v (T_1 - T_2), \qquad (2)$$

where $q$ is the heat flow in the direction of flow, $\Delta$, $c$, and $v$ are the density, specific heat, and velocity of the flowing medium, respectively, and $T_1 - T_2$ is the temperature difference in the flow. (*Note*: This expression must be modified for flow in a porous medium, such as water flow in an aquifer, but the basic parameters and principles are the same.) The threshold velocity at which convection becomes more efficient than conduction depends upon material thermal properties ($K$, $\Delta$, and $c$), but for most terrestrial heat flow problems this velocity is of the order of a few centimeters per year, or about $10^{-9}$ m sec$^{-1}$. Convection can be driven by thermally induced buoyancy forces (free convection), as in the circulation of seawater through young oceanic crust (see below) or in mantle and core convection. Convection may also be externally driven (forced convection), as in regional gravity-driven groundwater flow and tectonic deformation of the lithosphere (see below). A significant portion of the Earth's heat loss is convected to the surface at midocean ridges, and convection is thought to be the dominant heat transfer process at depths greater than 100–200 km (below the base of the lithosphere). The term advection is commonly used in tectonics specifically to indicate heat transfer by convection where there is no return flow as in magma intrusion, rather than generally to indicate forced convection, as it is more widely used in physics.

Radiative heat transfer becomes increasingly effective at high temperatures and can be defined by the same relationship as conductive heat transfer [Eq. (1)] by replacing the term for thermal conductivity $K$ by an equivalent and strongly temperature-dependent radiative conductivity. The role of radiative heat transfer in the Earth is poorly defined, but in olivine, which is common in the upper mantle, radiative heat transfer becomes efficient relative to phonon conduction at temperatures in excess of about 400°C. Experiments on the thermal conductivity of xenolith samples of the upper mantle indicate that this property of olivine is transferred to rocks typical of the mantle lithosphere. A typical conductivity versus temperature profile suggested for the mantle lithosphere is shown in Fig. 1.

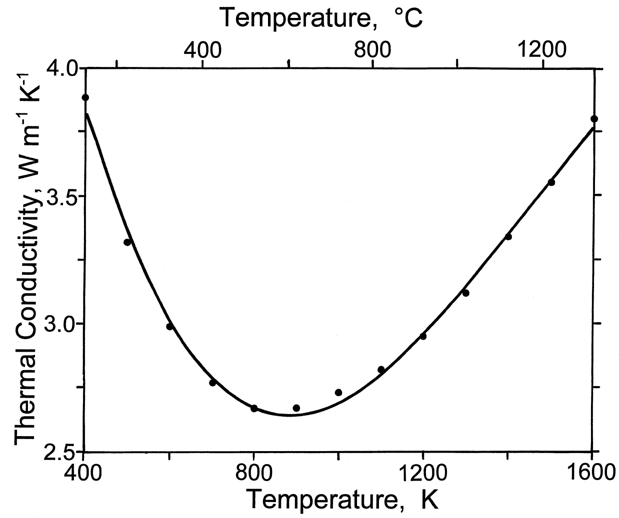


**FIGURE 1** Thermal conductivity of Olivine versus temperature in degrees Celsius (upper scale) and absolute temperature in Kelvins (lower scale). Data from Schatz and Simmons (1972) are shown by dots. Third order polynomial fit is shown by the solid curve.

## II. MEASUREMENT OF TERRESTRIAL HEAT FLOW

The only quantity that can be directly measured related to heat flow in the Earth is the near-surface heat flow. Conductive heat flow is usually determined by making separate measurements of the thermal gradient $\partial T/\partial z$ and the thermal conductivity $K$ of the rocks in which the gradient is measured, and calculating the heat flow from Eq. (1).

On land, gradients are usually measured using an electrical thermometer in boreholes, 100 m or more in depth, below the transient effects of surface temperature variations and below the groundwater infiltration zone. Thermal conductivities are generally measured by steady-state thermal-comparator techniques or transient heating techniques on core samples or drill-cutting samples from the boreholes. At sea and in some deep lakes, gradients are measured in the upper few meters of sediment using a string of electrical temperature sensors on a sediment-penetrating probe. The same probe may include a heating element for *in situ* thermal conductivity measurements, or conductivities may be measured on sediment cores. Stable thermal conditions at depth in the oceans and in deep lakes allow heat flow determinations to be made over a much shallower depth interval at these sites than for land measurements. A number of corrections may be applied to the heat flow measurements to account for local thermal effects.

Occasionally estimates of convective surface heat transfer can be made by measuring the discharge rate and

**TABLE I Basic Parameters in Terrestrial Heat Flow**

| Parameter | Working unit | Typical range |
|---|---|---|
| Vertical temperature gradient | °C/km (mK $m^{-1}$) | 5–50 |
| Thermal conductivity | W $m^{-1}$ $K^{-1}$ | 1–4 |
| Heat production | $\mu$W $m^{-3}$ | 0–8 |
| Heat flow | mW $m^{-2}$ | 5–125 |

temperatures of flowing hot springs, through airborne infrared measurements or from chemical analyses of discharging groundwater. This convective flux is an important part of the global heat flow budget, as discussed below: however, because of the difficulties of accurately measuring its magnitude and its extremely localized nature, unfortunately it is not generally included in large-scale heat flow maps.

An additional parameter essential for understanding heat flow in the Earth is the intrinsic radiogenic heat production of Earth materials. This parameter is usually measured in the laboratory on approximately 1-kg samples by measuring the radioactive isotope abundances through their natural gamma decay and using the isotope half-lives and decay energy to convert abundances to heat production. The heat-producing isotopes are isotopes of Th, U, and K, which are incompatible elements (not readily incorporated into mineral lattices). They tend to be concentrated in lower melting-point silicic igneous rocks, although their abundances do not closely follow major element abundances.

The main measured parameters in heat flow are given in Table I together with typical ranges of these parameters.

## III. HEAT FLOW AND INTERNAL TEMPERATURES

Regionally, near-surface heat flow varies from zero or near zero to more than 100 mW $m^{-2}$. The low limit of regional heat flow is imposed by the requirement that temperature must increase with depth (the Earth is losing heat), but the geothermal gradient can be reduced to almost zero over the descending portions of convection systems. The upper limit of regional heat flow (typically about 125 mW $m^{-2}$) is probably controlled by melting in the crust: higher heat flow requires super-solidus temperatures in the crust or uppermost mantle that typically result in magmatic convection and local redistribution of heat. Locally, a much wider range of heat flow is measured, either associated with redistribution of heat by groundwater on land, seawater convection through oceanic crust, and/or magmatism.

The temperature profile within the Earth, or geotherm, is constrained by the surface heat flow, but, as the effectiveness of different heat transfer mechanisms at depth and the distribution of heat production in the Earth are somewhat uncertain, it must be constrained at depth by indirect evidence and theoretical considerations. A global average geotherm is shown schematically in Fig. 2. Temperature increases rapidly in the upper 100 km or so as heat is conducted through the lithosphere. There is some radiant heat transport in the lower lithosphere. In the mantle below the lithosphere, an absence of evidence for widespread melting, seismic, and other geophysical evidence indicates that heat is transferred by solid-state convection and radiant heat transport at temperatures just below the solidus, and the geotherm is less steep. Variations in the geotherm in this zone are related to the details of mantle convection. Seismic data reveal topography on the core–mantle boundary, which is interpreted to indicate a dynamic thermal interchange between the core and mantle and the presence of a thermal boundary layer with a relatively rapid temperature increase near the base of the lithosphere.

Seismic data indicate that the base of the mantle is solid but that the outer core is liquid. The temperature of the core–mantle boundary is therefore constrained to be between the solidus of the lower mantle and the liquidus of the outer core. Seismic data also indicate that the inner core is solid, and thus the geotherm within the core is constrained to cross the solidus/liquidus within the core. Models of the geomagnetic dynamo responsible for the Earth's main magnetic field suggest that it is associated with convection currents in the outer core and that the energy for this field is either the radioisotope $^{40}$K in the core or, perhaps more likely, segregation and growth of a pure iron and nickel inner core from a lighter component of FeO or FeS in the outer core. The main heat generation and heat transfer mechanisms in the Earth are shown in Fig. 3.

Apart from convective instabilities in the mantle (thermal plumes), major departures from the global average geotherms are primarily in the uppermost thermomechanical layer of the Earth, the lithosphere, and the thermal structure of this zone is the subject of the remainder of this discussion.

The magnitudes and major modes of heat loss from the Earth are summarized in Table II. Almost 75% of the global heat loss is through the oceans, approximately 85% of which is associated with the creation of new oceanic lithosphere. In contrast, more than 60% of the 25–30% of the global heat loss from the continents is generated by radioactive decay in the crust. Just as continents and oceans are fundamentally different with respect to crustal thickness, bulk composition, and mean age, there are fundamental differences in their thermal regimes.

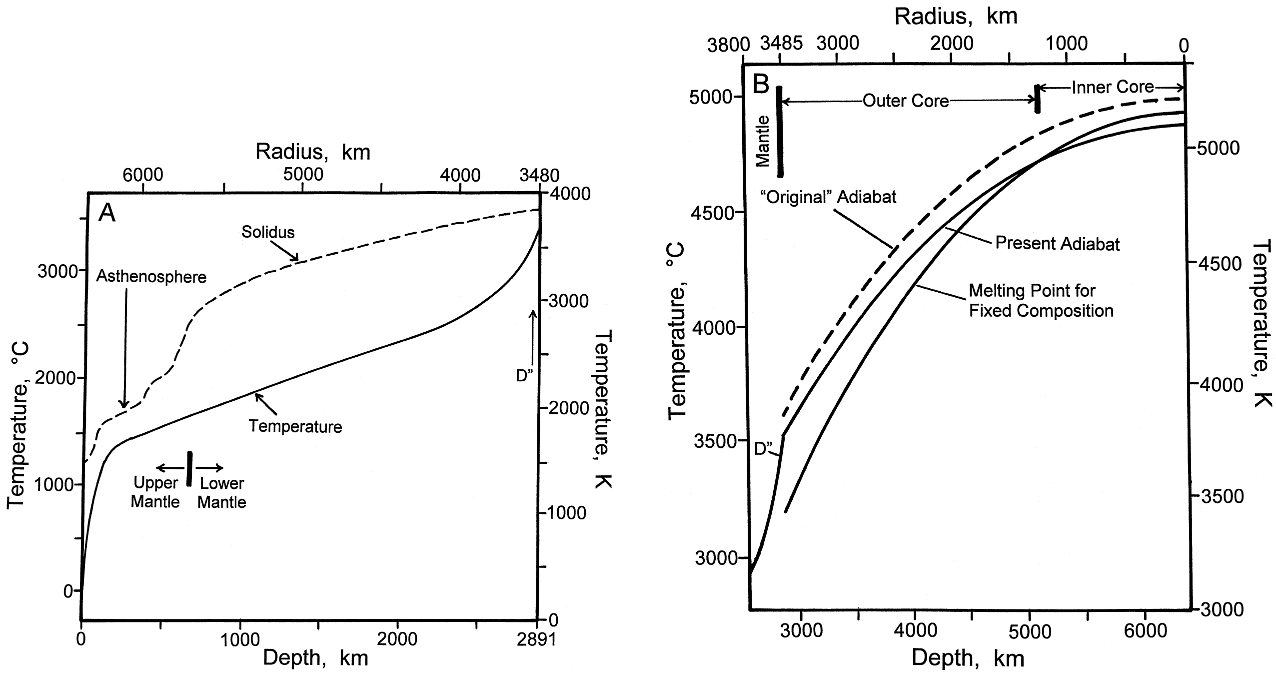


**FIGURE 2** Model temperatures in the earth as a function of earth radius (upper scale) and depth (lower scale). (Modified from Stacey, 1993.) Temperatures are given in degrees Celsius on the left and absolute temperatures in Kelvins on the right of each plot. Plot A shows temperatures as a function of depth in the mantle. Steps in the solidus curve primarily represent phase transitions in mantle mineralogy. The rapid temperature increase with depth in the uppermost mantle roughly corresponds with the lithosphere, and beneath this zone mantle temperatures are closest to the solidus, corresponding to the rheologically weak asthenosphere. Temperatures increase rapidly with depth again at the base of the lithosphere, roughly corresponding with the "D" layer. Plot B shows temperatures as a function of depth in the core. The difference in temperature between the "original" adiabat and present adiabat curves represents the estimated secular cooling of the core since its formation. Crossover of the present adiabat and melting point curves explains the solid inner core but molten outer core. The core probably includes a lighter element such as sulfur or oxygen, in which case some chemical differentiation will occur during solidification and a more complex melting curve relation should be applied. However, the basic results will be unchanged.

## IV. OCEANIC HEAT FLOW

When heat flow was first measured in the oceans in the early 1950s, expectations were that oceanic heat flow would be much lower than continental heat flow because thin, basaltic oceanic crust is much less rich in the heat-producing isotopes than thick, granitic continental crust. Early data indicated that oceanic heat flow was approximately equal to continental heat flow, however, and current data indicate that mean oceanic heat flow is approximately 70% higher than mean continental heat flow. Clearly, therefore, most of the heat from the Earth's interior is lost through oceanic crust, and this heat loss is part of the process of seafloor spreading through the creation of new oceanic crust at midocean ridges.

Heat is convected upward at the midocean ridges as the mantle rises and partially melts due to pressure release. These melts continue to rise and eventually cool to form new oceanic crust with a very high near-surface geothermal gradient and heat flow. Newly formed oceanic lithosphere moves approximately symmetrically away from the midocean ridge and cools by conduction to the surface. As it cools, the geothermal gradient in the lithosphere and surface heat flow decrease. General cooling in the lithosphere–asthenosphere system is accompanied by thermal contraction and an increase in the mean density of the system, which is manifested by isostatic sinking of the ocean floor. These features of the thermal development of new oceanic lithosphere are shown in Fig. 4.

Cooling of oceanic lithosphere is essentially one-dimensional and the thermal characteristics of all oceanic lithosphere are primarily dependent upon its age (i.e., they are independent of spreading rate and distance from the midocean ridge). A variety of models of lithospheric development associated with cooling have been proposed, all of which explain the basic observable thermal and elevation characteristics of oceanic lithosphere, and contrasting examples of which are shown in Fig. 4B.

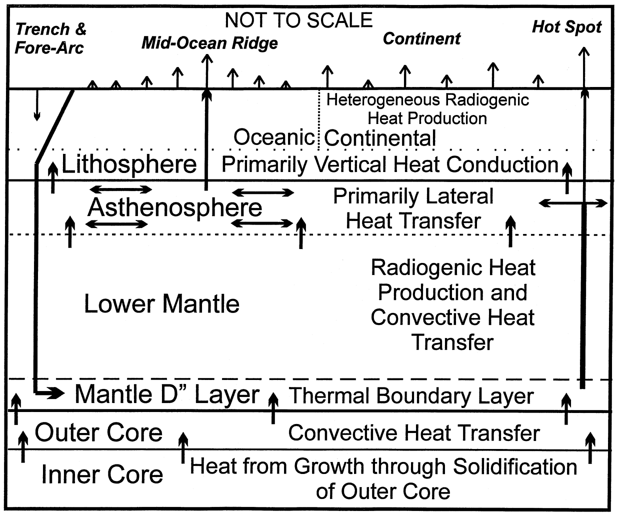


**FIGURE 3** Schematic representation of heat sources and transport in the earth and heat loss at the surface. The two primary heat sources are radiogenic heat production (~80%) and secular cooling of the earth (~20%). Lithospheric extension, magmatism, and upward water flow increase surface heat flow by advection. Subduction, compression, and downward water flow decrease surface heat flow unless accompanied by magmatism. Changes in surface temperature associated with climate change, erosion, and sedimentation cause transient changes in surface heat flow.

None of the models, however, explains all known features of oceanic lithosphere development.

There are two major departures in the observed parameters from the simple model predictions. The first departure is in the measured heat flow. Observed heat flow is generally much lower and more erratic than predicted, as shown in Fig. 5A. Heat flow data from all oceans are in reasonable agreement with the theoretical cooling curves for older oceanic lithosphere (greater than 10–70 Ma), and sonar and seismic observations of the sea floor indicate that this agreement is achieved when a coherent layer of sediment first covers the igneous oceanic crust. Where sedimentation rates are high, observations and predictions agree in relatively young crust; where sedimentation rates are low, agreement is not achieved until much older crust. With the exception of a few very high heat flow measurements, observed heat flow is always much less than predicted by the models in very young (less than 5 Ma) crust.

Measured heat flow is much less than predicted by the models, especially near midocean ridges, because much of the heat is removed from young oceanic crust by the convection of seawater through the crust. Measured heat flow is low because only the conductive component of the total heat loss is measured, and, by redistribution of heat flow, the convecting seawater results in large areas of recharge with low heat flow and very small areas of discharge with high heat flow. This convection system is best developed at the midocean ridges, where the discharging submarine hot springs produce spectacular black-smoker chimneys rich with metalliferous effluent and remarkable local biological communities. Access for seawater to the igneous oceanic crust is eventually cut off by the buildup of sediments on the ocean floor, and as soon as an impermeable layer of sediments is developed, all heat is lost by conduction.

The second departure from predictions of the simple thermal models is the flattening of the ocean floor with increasing age, in contrast to continuously increasing depth with increasing age predicted by the unconstrained cooling models (Fig. 5B). Unfortunately, the predicted heat flow variations for old ocean floor where the observed depths deviate from predictions cannot be resolved with available data, but it is likely that heat flow also does not continue to decrease indefinitely with increasing ocean-floor age. The thickness of the oceanic lithosphere does not thicken indefinitely with age and several models have been proposed to explain this behavior.

The simplest model used to explain the thermal behavior of older oceanic lithosphere is the model of a cooling constant-thickness plate with a fixed temperature at its base. This model successfully predicts the observable parameters associated with aging of the ocean floor, but is difficult to justify through geological arguments. Other models are based on a maximum thickness of the lithosphere controlled by heat input to the base of the lithosphere by shear heating, small-scale convection, radiogenic heat production in the upper mantle, or mantle plumes impinging on the base of the lithosphere.

A dramatic example of the probable effect of a mantle plume interacting with the lithosphere is the Hawaiian Swell and the string of volcanic islands that run along its

**TABLE II Major Modes of Heat Loss from the Earth**[a]

| Component | Value |
|---|---|
| Heat loss though the continents | $1.2 \times 10^{13}$ W |
| Heat loss through the oceans | $3.1 \times 10^{13}$ W |
| Total | $4.2 \times 10^{13}$ W |
| Heat loss by hydrothermal circulation | $1.0 \times 10^{13}$ W |
| Heat lost in plate creation | $2.6 \times 10^{13}$ W |
| Mean heat flow | |
| Continents | 50 mW m$^{-2}$ |
| Oceans | 100 mW m$^{-2}$ |
| Global | 84 mW m$^{-2}$ |
| Convective heat transport by surface plates[b] | ~65% heat loss |
| Radioactive decay in crust | ~17% heat loss |

[a] After Sclater *et al.* (1981).
[b] Includes lithospheric creation in oceans and magmatic activity in continents.

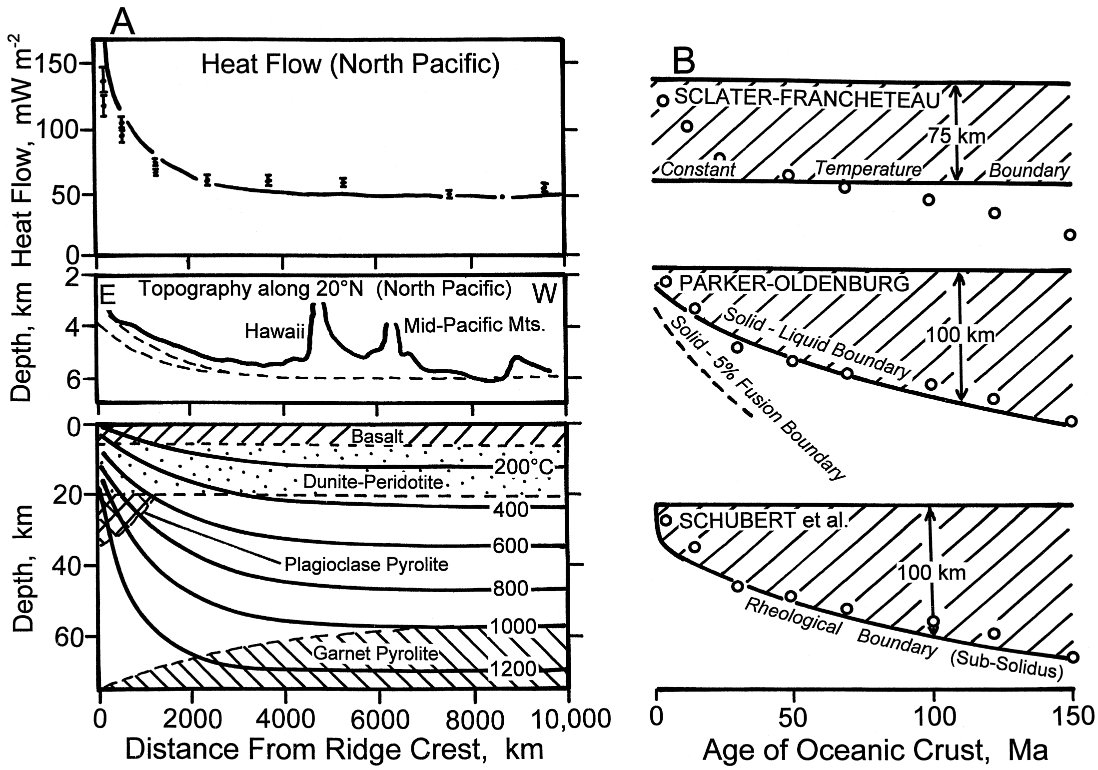


**FIGURE 4** Thermal development of oceanic lithosphere. (A) Lower: isotherms and mineralogy in oceanic lithosphere as it moves away from a mid-ocean ridge with a horizontal scale appropriate to the northern Pacific. Middle: topography along 20°N (solid curve) compared with theoretically predicted topography (dashed curve) for models including only thermal expansion (upper) and bother thermal expansion and mineral phase transitions (lower). Upper: Observed heat flow averages with error bars (symbols) and calculated heat flow (solid curve). (B) Models of the cooling oceanic lithosphere suggested by different workers. Upper: constant thickness (plate) model in which temperatures are perturbed relative to a constant-temperature lower boundary. Middle: increasing thickness model in which the lower boundary of the lithosphere is defined by the solidus temperature. Lower: increasing thickness model in which the base of the lithosphere is defined by a subsolidus rheological constraint. Circles show the estimated thickness of the lithosphere based on surface wave seismic data. (Modified from Bott, 1982.)

axis. This swell, some 1000 km wide and 5000 km long, slopes gently west from the Hawaiian Islands (Fig. 4A) and has a small heat flow anomaly (5–10 mW m$^{-2}$) relative to other ocean floor of the same age. The simplest explanation of this behavior is that the age of the Pacific plate is reset to a younger age as it passes over the Hawaiian plume or hot spot: heat input from the mantle plume thins the lithosphere and the cooling process starts again as the oceanic lithosphere moves laterally westward away from the plume. Another well-defined example of a plume or hot spot is Iceland, where the effects of the plume are superimposed on a midocean ridge. Numerous other smaller mantle plumes probably cause similar but less pronounced deviations from the simple cooling behavior of oceanic lithosphere.

To predict the thermal properties of oceanic lithosphere, Stein and Stein (1992) derived a plate model of oceanic lithosphere (GDH1) in which the lithosphere was found to have a best fit to the data with a plate thickness of $95 \pm 15$ km, a basal temperature of $1450 \pm 250$°C, and a volume coefficient of thermal expansion of $3.1 \pm 0.8 \times 10^{-5}$ K$^{-1}$ (all uncertainties ± one standard deviation). From this model oceanic depth $d(t)$ is related to age of the ocean floor $t$ by

$$\begin{aligned} d(t) &= 2600 + 365t^2 \text{ m} && t < 20 \text{ Ma}, \\ d(t) &= 5651 - 2473 \exp(-0.0278t) \text{ m} && t \geq 20 \text{ Ma}, \end{aligned} \tag{3}$$

where $t$ is age in Ma, and heat flow $q(t)$ is give by

$$\begin{aligned} q(t) &= 510t^{-1/2} \text{ mW m}^{-2} && t \leq 55 \text{ Ma} \\ q(t) &= 48 + 96 \exp(-0.0278t) \text{ mW m}^{-2} && t > 55 \text{ Ma} \end{aligned} \tag{4}$$

Oceanic lithosphere is returned to the mantle at subduction zones. Cool, sinking lithosphere in these zones convects heat downward, and surface heat flow is significantly

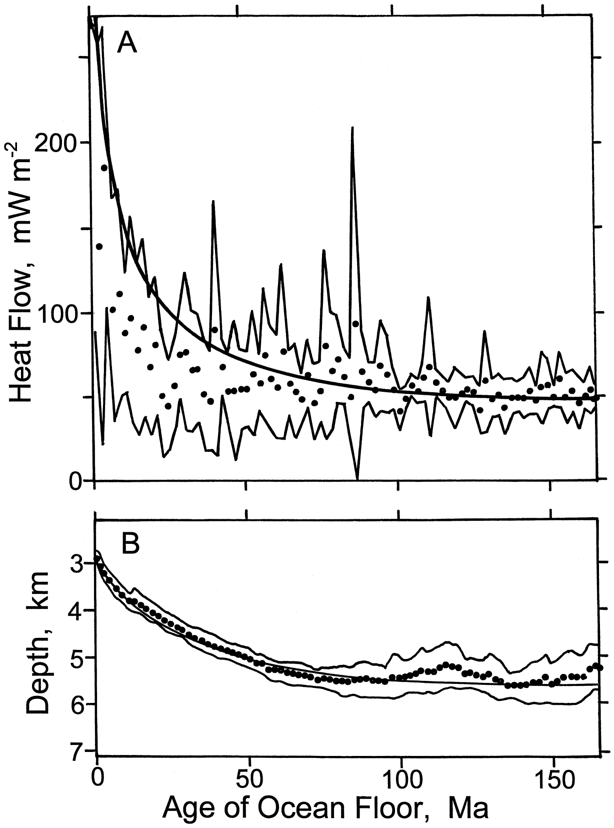


**FIGURE 5** Data and models for heat flow (A) and ocean depth (B) as a function of age of ocean floor. Depths are an average of values in the North Pacific (north of the equator) and the Northwest Atlantic (15°N to 45°N and 40°W to 80°W); heat-flow values are averages of values in these regions. The data, shown by solid circles, are averaged in two million year (Ma) bins, and the envelope shows one standard deviation about the mean. The solid curve shows the plate model of Stein and Stein (1992) described in the text. (Modified from Stein and Stein, 1992.)

reduced, sometimes to effectively zero. Melting and magmatism associated with the descending plate reverses this trend, however, and heat flow is high in zones of active volcanism behind subduction zones (volcanic arcs). The depression of isotherms associated with subduction and the dual low/high heat flow signature of a typical subduction zone are illustrated schematically in Fig. 6. Active or very recent volcanism must be present for the high heat flow zone to exist, and the high heat flow is clearly associated with upward advection of heat by magmas. In zones where arc volcanism is no longer active (e.g., northern central Andes) or has moved back from the trench (e.g., U.S. Cascades), heat flow is low. If the stress field is favorable, subduction zones can also be associated with extension and the creation of new oceanic lithosphere in zones of back-arc spreading. Seafloor spreading in these zones tends to be less well organized than at midocean ridges,

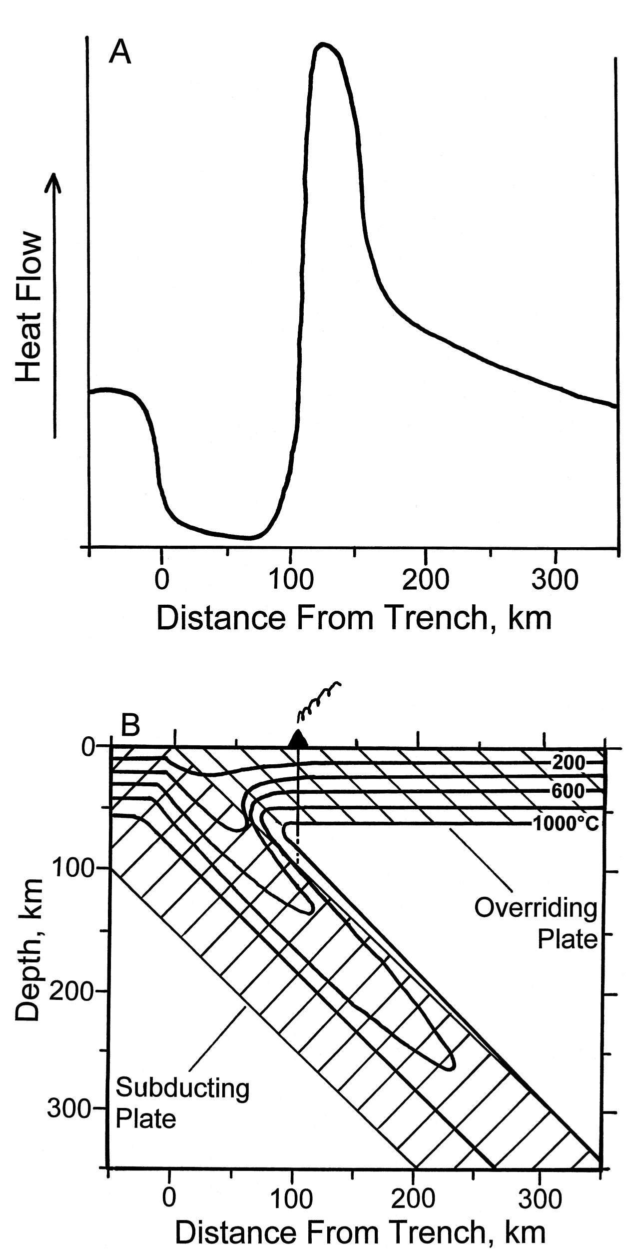


**FIGURE 6** Surface heat flow (A) and the general thermal structure of the lithosphere associated with subduction (B). Details of local thermal modification of the lithosphere associated with subduction-related magmatism are not shown. (See also Anderson, 1986.)

but their thermal structure is essentially identical to that of their midocean counterparts.

## V. CONTINENTAL HEAT FLOW

Compared with the relatively short life (<200 Ma) and rapid turnover of oceanic lithosphere, continental lithosphere has a long life span and plays a relatively passive role in plate tectonics. It has a rate of creation and destruction at least an order of magnitude slower than the creation and destruction of oceanic lithosphere, and is chemically much more heterogeneous than oceanic lithosphere, particularly with respect to the heat-producing isotopes. Oceanic lithosphere is created by an apparently uniform thermal process and does not experience major tectonics, apart from the effects of mantle plumes, before destruction. Continental lithosphere typically experiences a variety of tectonic and thermal events of differing magnitudes during its lifetime. The transient thermal processes operating in continental lithosphere are conceptually similar to the processes operating in oceanic lithosphere, but their magnitudes are much less predictable and their signals are mixed with thermal variations associated with the chemical heterogeneity of the continents.

A compilation of measurements of continental heat flow is plotted in Fig. 7 as a function of the tectonothermal age, the age of the last major tectonic or magmatic event in the region of each measurement site. Comparison of the continental data with oceanic data (Figs. 4 and 5) suggests that the continental data follow a similar trend to the oceanic data as a function of "age," but with more scatter in the continental data, especially as "age" increases. Detailed inspection of individual datasets indicates

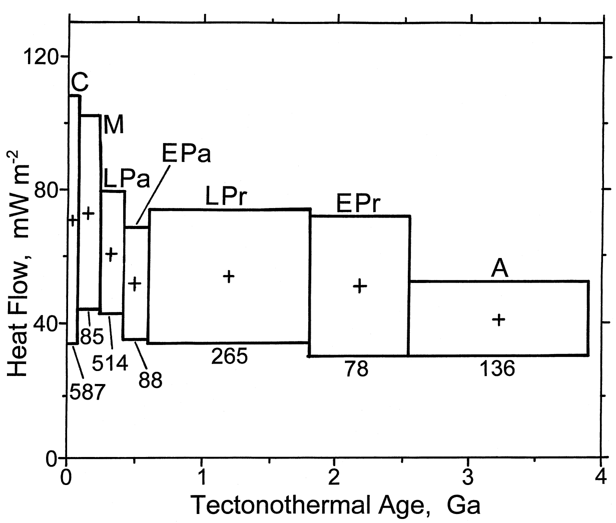


**FIGURE 7** Average continental heat flow versus tectonothermal age in billions of years (Ga), the age of the last major tectonic of magmatic event at the heat flow site. Data are grouped in geological age ranges: C–Cenozoic, M–Mesozoic, LPa–Late Paleozoic, EPa–Early Paleozoic, LPr–Late Proterozoic, EPr–Early Proterozoic, A–Archean. Crosses are plotted at the mean heat flow and the midpoint of the age range. Box widths indicate age range and box heights indicate ± one standard deviation of the heat flow data about the mean. Numbers below the boxes indicate the number of data in each group. (From Morgan, 1984.)

more fundamental differences between the two settings, however.

Continental tectonothermal settings of Cenozoic age include zones of extension, compression, strike–slip motion, and volcanism. High heat flow is generally measured in zones of extension as heat is advected upward by magmas and by the general upward movement associated with thinning of the lithosphere in response to extension. The advective processes in these zones are similar to advection in midocean ridges, and there is good evidence that in some continental rifts they may be the precursors of midocean ridges.

Similar zones of high heat flow are associated in general with magmatism in the continents in all tectonic settings. The dominant process operating in these zones is upward advection of heat by ascending magmas. Basaltic magmas tend to either pond in the upper mantle or lower continental crust or to erupt, and their surface thermal anomalies are of relatively short duration. In contrast, intermediate and felsic magmas (often derived in association with basaltic magmatism) commonly form middle crustal and shallow magma chambers which have more significant and durable thermal anomalies, often associated with surface thermal manifestations such as fumaroles and geysers.

High heat flow has been predicted, but not unambiguously observed, in association with shear heating on continental transform (strike–slip) faults. The rate of heating $q_f$ is given simply as the mechanical rate of work on the fault,

$$q_f = \sigma A \upsilon, \tag{5}$$

where $\sigma$ is the stress on the fault, $A$ is the area of the fault, and $\upsilon$ is the rate of movement on the fault. For major continental transform faults, such as the San Andreas fault in California, the rate of movement is well constrained and the area may be reasonably estimated, but there has been great debate as to the order of magnitude of the stress on the fault. Heat flow data from the San Andreas fault suggest that the stress on the transform fault is relatively low, resulting in no significant thermal anomaly, but earthquake data indicate much higher stresses associated with thrust faults that branch from the main strike–slip fault.

Locally, and even regionally, heat in the upper continental crust is redistributed by convecting groundwater, somewhat analogous to the convection of seawater through young oceanic crust. On continents the fluid motions are commonly driven by hydraulic gradients associated with variations in the water table elevation in addition to flow driven by thermal buoyancy.

Most geothermal gradient data include small, depth-dependent temperature perturbations caused by climatically induced changes in mean surface temperature, particularly warming during the latter part of the 20th century.

These perturbations may be used to infer the timing and magnitude of climatic changes, particularly during the past 200 years, and, if unrecognized, may result in a small error in heat flow determinations, typically an underestimate of a few mW m$^{-2}$. However, separation of effects of the local microclimate, local topography, local geology and hydrology, and relevant thermal history of surface around the measurement site often makes meaningful extraction of more regional climate signals from individual geothermal gradient data problematic.

On a regional scale there seems to be an approximate upper limit to the average heat flow of about 125 mW m$^{-2}$. At this level of heat flow, near-solidus temperatures are indicated near the Moho. Higher regional heat flow averages may be prevented by buffering of maximum temperatures in the crust by melting.

When sources of heating in continental lithosphere cease to operate, it cools and subsides in a similar manner to oceanic lithosphere. However, it is less easy to demonstrate this behavior in the continents. Yellowstone National Park (the Yellowstone caldera), Wyoming, in the western United States, is perhaps the clearest example of the effects of a mantle plume in continental lithosphere, and this hot spot has a trace that extends west–southwest along the eastern Snake River Plain from the Yellowstone hot spot. Like midocean ridges, elevations decrease along the Snake River Plain away from the heat source (at the Yellowstone caldera) at a rate consistent with a cooling lithosphere model in which the age of cooling increases approximately linearly with distance along the plain. Cooling and subsidence appear to cease in this example at a cooling age of approximately 20 Ma, when the heat flow has decreased to the high background heat flow of the western United States. In place of the large constructional basaltic volcanoes associated with the Hawaiian plume, basaltic volcanism associated with the Yellowstone/Snake River Plain system heats the crust causing massive silicic eruptions, and locally makes the crust more basic and more dense, resulting in a topographically depressed plain relative to the surrounding terrain.

A second good example of cooling and subsidence in a continental setting is found in some sedimentary basins of extensional origin, in particular, basins on passive continental margins. Sediments in these basins generally record a history of decreasing subsidence rate with time, consistent with cooling and subsidence models of the lithosphere. Subsidence is amplified in these basins by the loading effects of the sediments, and high heat flow in the early stages of basin formation can be recorded in temperature-dependent processes, such as radiometric argon loss, fission-track annealing, or hydrocarbon maturation in the older sedimentary rocks deposited in the basins.

Subduction zones on continental margins have similar thermal structure to subduction zones in completely oceanic settings, as shown schematically in Fig. 6. The contrasting adjacent hot/cold thermal regimes in these zones are preserved in paired metamorphic belts in older terranes, with blueschist-grade metamorphism marking the high-pressure, low-temperature conditions over the subducting slab between the trench and subduction-related volcanism, the fore-arc zone, and amphibolite-grade metamorphism recording the moderate-pressure, high-temperature crustal conditions associated with the volcanism.

Similar, but more complex thermal effects to subduction zones are found in zones of continent–continent collision such as the Alpine–Himalayan belt. Where simple or imbricate thrusts result in the underthrusting of one continental mass beneath another, the net movement of material in the lithosphere is downward, resulting in downward advection of heat and low surface heat flow. Heating of the resulting thickened crust in these zones commonly results in magmatism, however, which results in elevated surface heat flow. Thus, zones of continent–continent collision may be expected to be associated with complex interspersed high- and low-heat-flow regions. Unfortunately, however, because of the logistical problems of measuring heat flow in these zones and redistribution of heat by flowing groundwater in the resulting rugged topography, data defining this pattern are sparse.

A final complication to continental heat flow in regions of recent tectonothermal activity results from the effects of advection of heat by sedimentation and erosion. Sedimentation results in a net downward movement of material, advecting heat downward and depressing surface heat flow; erosion has the opposite effect, advecting heat upward and increasing surface heat flow. These effects can extend the time period for thermal relaxation of continental lithosphere (the time for thermal equilibrium to be attained after tectonothermal activity), as they result in advection of heat in the lithosphere as the lithosphere responds isostatically to its changing geotherm. Examples of possible surface heat flow recovery paths for a variety of initial tectonothermal disturbances and the effects of sedimentation and erosion are shown in Fig. 8.

Large scatter in heat flow in continental regions of young tectonothermal age (Cenozoic and Mesozoic in Fig. 7) is a result of a mixture of different complex thermal processes with different magnitudes and signs operating in these zones. Relative to oceanic data, however, even taking into account the complexities of the thermal processes perturbing continental geotherms and the temporal extensions of these anomalies through erosion and sedimentation, continental heat flow values in stable regions

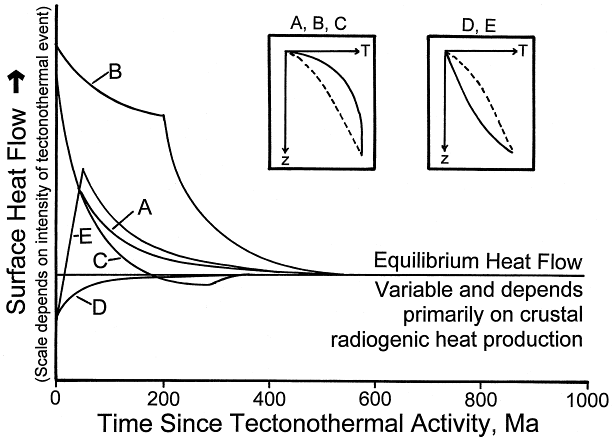


**FIGURE 8** Schematic examples of surface heat flow relaxation following continental tectonothermal activity. Insets on upper right show initial geotherms (temperature, T, versus depth, z) by solid curves relative to the dashed curve of the equilibrium geotherm. The left inset shows an initial hot geotherm (rifting, mod-plate magmatism, hot spot, subduction- or collision-related magmatism). The right inset shows an initially cold geotherm (fore-arc subduction or compression without magmatism). Different post-tectonothermal scenarios are represented: A–cooling without sedimentation or erosion, B–cooling with erosion for 200 Ma, C–cooling with sedimentation for 300 Ma, D–reheating without sedimentation or erosion, E–reheating with rapid erosion for 50 Ma. (From Morgan and Sass, 1984.)

do not converge to a relatively small range of values similar to heat flow in old oceanic crust. There is still a large scatter about the mean in the continental data from regions of older tectonothermal age (greater than about 250 Ma), and this scatter reflects the chemical heterogeneity of continental crust.

In general, continental heat flow has three significant components: (1) heat loss from the deep interior of the Earth, (2) heat generated within the continental lithosphere, and (3) advection within the lithosphere associated with tectonothermal activity. The third component should be insignificant in regions with tectonothermal ages of Paleozoic or older. Studies by F. Birch, R. F. Roy, E. R. Decker, and D. D. Blackwell in the late 1960s showed that for measurements in major silicic plutons the first two components can be separated. These workers discovered linear relationships between surface heat flow and surface heat production for silicic igneous sites in different thermal settings. The heat flow intercept of these relationships, or reduced heat flow, gives the surface heat flow for zero surface heat production, and is interpreted to be the heat flow from below the zone of upper crustal enrichment in the heat-producing isotopes. The slope of the linear relationship has the units of length, and is interpreted to be a scaling parameter for the depth distribution of heat production in the crust. A. H. Lachenbruch has shown that for the relationships to survive the effects of differential erosion, heat production should decrease exponentially from the surface with an exponential decrement defined by this scaling parameter.

The linear heat flow–heat production relationship has been extended to high-grade metamorphic terranes in addition to silicic igneous terranes, although the quality of the relationship is poorer in metamorphic terranes. Relationships have been established for 17 regions, or heat flow provinces, covering all continents except Antarctica. In provinces with young tectonothermal ages, the reduced heat flow values show a wide range, correlating with the style of tectonothermal disturbance (e.g., see Fig. 8). In provinces with tectonothermal ages Paleozoic or older, however, the reduced heat flow is remarkably constant, with a mean and standard deviation of $27 \pm 4$ mW m$^{-2}$ ($n = 10$). Therefore the deep component of continental heat flow in stable regions appears to be approximately constant, and the scatter in surface heat flow is primarily a function of lateral variations in crustal heat production. As shown in Fig. 7, the means and standard deviations of heat flow data from Paleozoic and Proterozoic tectonothermal age provinces are insignificantly different, consistent with their similar heat flow at depth and similar ranges in crustal heat production. Mean heat flow for Archean tectonothermal age sites is lower than for younger sites, however, and the scatter in these data is significantly less. Examination of heat flow–heat production relationships for Archean terranes indicates that reduced heat flow in these provinces is insignificantly different from that in younger, stable terranes, but that high heat production crust is much rarer in Archean terranes than in younger terranes. This conclusion is consistent with geochemical evidence that Archean crust is significantly statistically lower in incompatible elements, including Th, U, and K, than younger crust.

Models of variations in crustal heat production in stable continental terranes, with an approximately uniform deep component of heat flow, indicate that stable continental lithosphere has a range in thickness, higher heat production crust resulting in thinner lithosphere than low heat production crust (Fig. 9). Geotherms are constrained in the lower lithosphere using xenolith data, but regardless of these parameters, variations in upper crustal heat production cause a spread in geotherms at depth which is probably equivalent to a difference between minimum and maximum thicknesses of the stable continental lithosphere of approximately a factor of two.

## VI. PALEO-HEAT FLOW

Variations in surface heat flow for the modern Earth are summarized schematically in Fig. 10 relative to the global

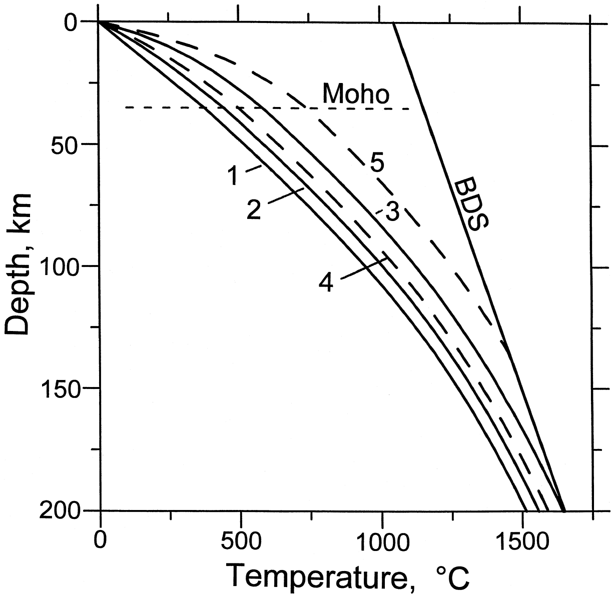


**FIGURE 9** Model geotherms for stable continental lithosphere with the same reduced heat flow and a range in crustal heat production. The geotherms correspond to a reduced heat flow of 27 mW $m^{-2}$, a crustal thermal conductivity of 2.5 W $m^{-1}$ $K^{-1}$, and a temperature dependent mantle conductivity shown by the solid curve in Fig. 1. A uniform heat production of 0.05 $\mu$W $m^{-3}$ was assumed for the mantle, with a crustal thickness of 35 km, and the crust/mantle boundary is shown on the plot by the horizontal dashed line labeled Moho. The solid line labeled BDS is an approximation of the basalt dry solidus and gives one interpretation of the maximum possible thickness of the lithosphere. Solid curves labeled 1, 2, and 3 correspond to geotherms calculated for a crustal heat production depth scaling parameter of 10 km with an exponential decrease from the surface starting at 0, 2, and 6 $\mu$W $m^{-3}$, respectively. Dashed curves labeled 4 and 5 correspond to geotherms calculated for a crustal heat production depth scaling parameter of 15 km with an exponential decrease from the surface starting at 2 and 6 $\mu$W $m^{-3}$, respectively.

mean heat flow. Earlier in Earth history this global mean must have been significantly higher, as the abundances of the heat-producing isotopes within the Earth were higher and the component of heat flow associated with secular cooling of the Earth was larger (Fig. 11). As heat loss from the Earth is closely related to tectonic and magmatic activity, the mode by which this extra heat was lost from the Earth is closely related to the tectonic style of the early Earth.

The largest fraction of the global heat budget in the modern Earth is lost in association with the creation of new oceanic lithosphere. The record of seafloor spreading preserved beneath the current oceans indicates that the rate of creation of new oceanic lithosphere has not been constant over the last 150 Ma. A peak in seafloor spreading occurred in the late Cretaceous, approximately 75 Ma ago, associated with the opening of the Atlantic. This spreading peak was probably associated with a peak in mean oceanic heat flow of about 125 mW $m^{-2}$ and a peak in mean global heat flow of about 100 mW $m^{-2}$, relative to the present mean global value of about 80 mW $m^{-2}$. Therefore, much of the additional heat available in the early Earth could have been lost by an accelerated rate of oceanic lithosphere creation, either through seafloor spreading at a faster mean rate than at present or through smaller plates and greater ridge length than at present. Hot, young oceanic lithosphere subducts more slowly than cold, old lithosphere in present-day plate tectonics, suggesting that the early Earth was covered by many small plates moving slowly: If Archean heat flow was three times that of the present, 27 times as much ridge would have been required to lose this heat, assuming the same balance between continental and oceanic heat loss mechanisms in the Archean and modern Earth.

Other heat loss mechanisms may have also been significant in the early Earth. The lack of preservation of crust prior to 4.0 Ga, roughly coincident with the end of the last heavy meteorite bombardment of the Earth, suggests that this bombardment disrupted the stabilization of early continents and may have caused a constantly changing pattern of heat loss through plate convection and magmatism. More heat may have been lost in the early Earth by conduction into the base of continental lithosphere, and calculations indicate that the present stable reduced heat flow can be approximately doubled to about 55 mW $m^{-2}$ before an unacceptable amount of melting is predicted in the lower crust. As discussed earlier, however, it is unlikely that peak heat flow values were higher in the Archean than at present, as peak heat flow is buffered even today by crustal melting, although very high heat flow may have been more common and more widespread in the early Earth.

Tectonic processes may have been slightly different in the early Earth from modern processes, and it has been suggested that ancient oceanic crust was thicker than modern oceanic crust in association with more efficient advection of heat at ancient midocean ridges. The occurrence of komatiites, very high temperature magnesium-rich lavas in Archean greenstone belts, suggests a more dynamic mantle thermal regime in the Archean, although the dating of Archean age diamonds, which require the conditions of a relatively cool geotherm for formation, indicates at least some local “cool” spots in the Archean upper mantle. The problem in understanding the thermal regime of the early Earth lies in finding data to constrain the processes in operation rather than in finding mechanisms capable of explaining the additional heat loss.

The statistical chemical difference (low heat production) between surviving Archean continental crust and younger crust may also be related to higher heat loss during the Archean. As shown in Fig. 9, low-heat-production

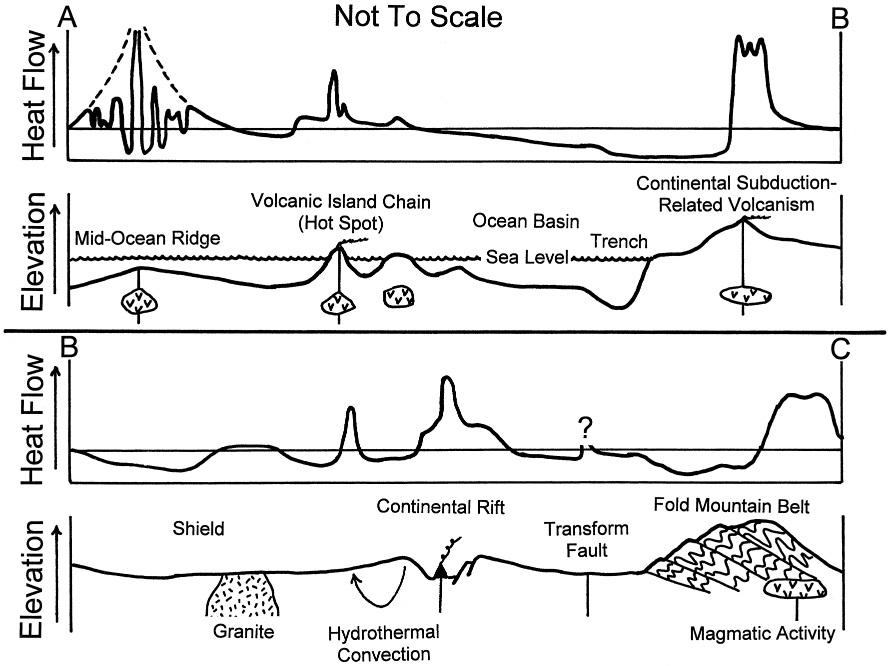


**FIGURE 10** Schematic heat flow associated with different oceanic and continental thermal and tectonic regimes. The thin line on the heat flow plots indicates mean global conducted heat flow. *Note*: Heat flow and topographic variations are shown qualitatively only. Features shown from A to B: high irregular heat flow associated with hydrothermal convection at mid-ocean ridge, reheating of oceanic lithosphere associated with volcanic island chain (hot spot), stable low heat flow in ocean basin, very low heat flow over descending lithosphere in trench and fore-arc region of subduction zone, and high and variable heat flow over the subduction-related volcanism with associated hydrothermal convection, returning to normal values in the stable continent behind the subduction zone. Features shown from B to C: generally low, but variable heat flow in the continental shield with higher heat flow over the granite rich in heat-producing radiogenic isotopes; local heat-flow high associated with discharge of deeply circulating groundwater; variable high heat flow associated with general tectonic and local magmatic advection of heat in extensional continental rift zone; possible frictional heating associated with high-stress transform fault (as yet undetected), and offset in heat flow across fault caused by juxtaposition of different thermal regimes across transform fault; low heat flow associated with underthrusting in compressional fold belt adjacent to high heat flow associated with magmatic activity caused by lower crustal melting in the belt. (Modified from Morgan, 1989.)

crust has a cooler geotherm than high-heat-production crust and is mechanically stronger by virtue of its lower temperature. It requires more heating before it can be melted. If reduced heat flow was generally significantly higher in the Archean than at present, possibly high-heat-production crust was selectively reworked during Archean orogenesis and low-heat-production crust was selectively preserved. Samples of the lower lithosphere are provided by xenoliths, wall rocks that are entrained in magmas that are erupted from depth. These samples yield estimates of the mantle lithosphere geotherm from specific pressure- and temperature-sensitive mineral compositions. Xenolith data indicate that the mantle lithosphere has evolved through time and that the crust and underlying mantle remain together for very long time periods. These observations reinforce the observation that remaining Archean lithosphere is chemically and physically different from younger lithosphere.

## VII. CONCLUDING REMARKS

Much progress has been made in understanding the thermal regime of the Earth during, and in association with, the development of the concepts of plate tectonics. In particular our understanding of the thermal development of oceanic lithosphere as well as of the components of continental heat flow has greatly improved. More study is required to constrain the details of heat loss associated with continental tectonism and magmatism and the thermal process(es) associated with flattening of the ocean floor with age. Through study of these processes and through study of the tectonothermal processes recorded in ancient continental rocks, we will gain more insight into the thermal regime of the early Earth.

The thermal structure of the lower lithosphere is steadily being revealed through studies of xenoliths. Where these xenoliths have been brought to the surface recently, the

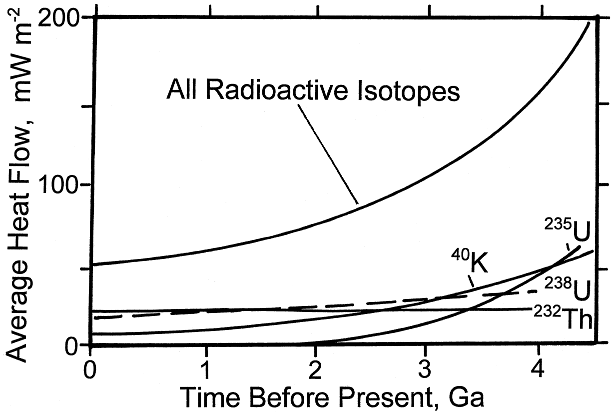


**FIGURE 11** Average surface heat flow in mW m$^{-2}$ from the earth generated from each of the major radiogenic heat producing isotopes, and all of the isotopes, as a function of time before present in billions of years (Ga). An additional 20% of heat flow from the modern earth is estimated to come from secular cooling of the earth. (Modified from Turcotte and Schubert, 1982.)

relations between surface heat flow measurements and this deeper thermal structure can be investigated. Xenoliths from older eruptions provide a window back into earlier thermal regimes, but have yet to indicate anything significantly different from modern thermal regimes.

## ACKNOWLEDGMENT

The work of many colleagues, mentors, and heroes over the past 65 years is summarized in this article. I hope that I have been fair to them. This is contribution number 243 of the ARC National Key Centre for Geochemical Evolution and Metallogeny of Continents.

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