Figure 26-1. Experimental setup for one-dimensional steady-state measurements of thermal conductivity in rock samples (modified from Birch, Francis, 1950, Flow of heat in the Front Range, Colorado, Geological Society of America Bulletin, v. 61, Figure 14, p. 601). Modified with permission of the publisher, the Geological Society of America, Boulder, Colorado USA. Copyright ©1950 Geological Society of America.
INTRODUCTION

Terrestrial heat flow is the quantity of heat transferred from a variety of heat sources in the interior of the earth to the earth’s surface. This heat is predominantly radiogenic in nature. Within the crust, heat is transferred largely by crystalline-lattice conduction, although considerable convective heat transfer may occur where local geological conditions involve substantial movement of fluids. Heat transfer by radiation, at temperatures prevalent within the upper crust, is negligible.

In a geologic context, terrestrial heat-flow density \( q_z \) is expressed as the vertical component of heat conducted through the drillable portion of the outer crust:

\[
q_z = k_z \frac{dT}{dz}
\]

where \( \frac{dT}{dz} \) is the rate of increase of temperature \( T \) with depth \( z \) within a rock formation of thermal conductivity \( k_z \), measured in the vertical \( z \) direction. This is a simplification based on the assumption that heat is transferred through the crust solely by conduction, along a path normal to the surface of the earth. Therefore, anomalies in measured heat-flow density may be attributed either to anomalous thermal properties of the rock formation, an anomalous heat source, and/or convective heat transfer. Measurements of heat-flow density are generally reported in SI units as mW/m² (milliwatts per square meter; 1 mW = 0.000239 calorie per second).

THERMAL CONDUCTIVITY OF ROCK FORMATIONS

Thermal conductivity of a rock formation is usually estimated on the basis of laboratory measurements conducted on selected rock samples. Various transient methods of in situ measurements have been developed (Beck, 1957), but their applicability is commonly limited to fine-grained unconsolidated sediments, particularly in marine or lacustrine heat-flow surveys. Conductivity measurement techniques may be divided into two broad categories: (a) one-dimensional steady state, and (b) two-dimensional transient or steady state.
In the most commonly used one-dimensional steady-state technique, a disc machined out of the rock sample is sandwiched between two reference discs of known conductivity. This entire “sandwich” is placed between a heat source and a heat sink (Figure 26-1). Temperature gradients measured across the two reference discs are used to determine the quantity of heat flowing per unit time across a unit area down the stack. This quantity is used in conjunction with the temperature gradient measured across the central disc to determine the thermal conductivity of the sample. A detailed description of the technique and a review of the inherent experimental errors are given by Beck (1965). Descriptions of alternative steady-state techniques are presented by Benfield (1939), Schroder (1963), and Creutzburg (1964).

Two-dimensional thermal conductivity measurements employ a long, cylindrical heat source inserted into the investigated medium (usually unconsolidated sediment), which is heated at a carefully controlled rate per unit length. The resulting rise of temperature at a given radial distance from the linear heat source is a function of the thermal conductivity of the surrounding material. The theoretical background for this methodology is described in Carslaw and Jaeger (1959).

Thermal conductivity of rock materials is dependent on both temperature and pressure. Conductivity is directly proportional to pressure (Hurtig and Brugger, 1970) and, in general, is inversely proportional to temperature, within the range of crustal temperatures (Kappelmeyer and Haenel, 1974).

Thermal conductivity is commonly expressed in either of the two following units:

1 Watt m\(^{-1}\)K\(^{-1}\) = 2.391 mcal cm\(^{-1}\)s\(^{-1}\)oC\(^{-1}\)

Typical values of most common rocks, at 25°C and 1 atmosphere, range from an average of 0.35 W/m\(^{1}\)K\(^{-1}\) for obsidian to about 6.71 W/m\(^{1}\)K\(^{-1}\) for quartzite (Kappelmeyer and Haenel, 1974).

A detailed treatment of the methodology for measurements of thermal conductivity in sediments and rock formations can be found in Beck (1988) or Jessop (1990).

GEOTHERMAL GRADIENT

The geothermal gradient is commonly defined as the rate of increase of temperature (T) with depth (z) within a rock formation (\(\frac{dT}{dz}\) in equation 1). The gradient is usually measured in boreholes at discrete intervals, utilizing thermistors, or as a continuous temperature log.

The effects of numerous cyclic and noncyclic environmental temperature perturbations are superimposed upon the geothermal gradient. Short-term cyclic effects, such as those associated with the diurnal or annual oscillations of surface temperature, are generally attenuated within 20 to 50 m (66 to 164 ft) of the surface. The effects of long-term cyclic effects, such as Pleistocene glaciation, may be observed to depths of hundreds of meters, and the raw data must be corrected (Birch, 1948; Cermak, 1976). One of the noncyclic temperature perturbations most commonly observed is caused by variations in topography and may be corrected in a manner similar to that used in gravity surveys. An excellent summary of the various methods used to correct for topographic effects is given by Blackwell and others (1980). Other perturbations resulting from uplift, erosion, subsidence, sedimentation, surface drainage, and groundwater movement may require individual corrections (Kappelmeyer and Haenel, 1974).

An extensive description of the techniques of measuring the geothermal gradient and the associated problems can be found in Beck and Balling (1988).

TERRESTRIAL HEAT-FLOW DENSITY IN PENNSYLVANIA

Nine measurements of heat-flow density have been made in Pennsylvania (Figure 26-2). The initial measurement, 54 mW/m\(^{2}\), was taken in the north-central part of Pennsylvania by Joyner (1960). Six additional measurements have been made by Urban (1971). The two most recent measurements were made by the authors in 1979 (Maurath, 1980). Measurements range from a low of 38 mW/m\(^{2}\) in the southeastern part of the state to a high of 84 mW/m\(^{2}\) in Venango County. The mean heat-flow-density value is 57 mW/m\(^{2}\), which closely approximates the global mean heat flow of 59 mW/m\(^{2}\) (Chapman and Pollack, 1975).

A generalized picture of the regional thermal regime in Pennsylvania may be derived from the map of the geothermal gradient approximated from the groundwater temperatures at about 30-m (100-ft) depth and the bottom-hole temperatures for 439 oil and gas wells in Pennsylvania and adjacent states (Figure 26-2). Whenever possible, temperatures were obtained from continuous-temperature logs of deep wells. Where these logs were not available, recorded bottom-hole temperatures or maximum recorded temperatures from various other geophysical logs were used.

DISCUSSION

Because only nine measurements of heat-flow density have been made within Pennsylvania, a detailed analysis of regional heat flow is not possible. The mean observed heat-flow density of the six measurements in the Appalachian Plateaus province prob-
ably represents the regional conductive heat flow. The
two measurements of heat-flow density made by the
authors (84 and 77 mW/m²) are anomalously high as
a result of local geologic conditions. Thickening of
highly radiogenic Pennsylvanian and Mississippian
shales could result in local increases in the concen-
tration of heat-producing elements within the upper
crust. The three low values observed in eastern Penn-
sylvania are part of a southwest-northeast-trending
regional heat-flow low associated with the folded Ap-
palachians (Urban, 1971) and are not associated with
the adjacent Reading Prong. Assuming an upper man-
tle heat-flow density of 16.7 mW/m² (Roy and others,
1968) and a two-layer crustal model, the observed
heat flow in the Appalachian Plateaus province of
Pennsylvania indicates an average crustal thickness
of 45 km (28 mi). Although this estimate is based
upon many simplifying assumptions, it is in general
agreement with a crustal thickness of 36 to 45 km
(22 to 28 mi) determined by other geophysical meth-
ods (Pakiser and Steinhart, 1964).

PROBLEMS AND FUTURE RESEARCH

There is an extensive gap in heat-flow-density data for central Pennsylvania. Considerable work needs
to be done within the eastern part of the Appalachian
Plateaus province and the Ridge and Valley province
that could possibly provide information useful in con-
straining models of the Appalachian orogeny, crustal
thickness, and maturation and distribution of the Ap-
palachian hydrocarbons. Unresolved problems include
an unexplained temperature-gradient high in the Pitts-
burgh area, which may be an artifact of thermal con-
vection in Paleozoic aquifers in the area. In addition,
the subnormal heat flow observed in eastern Pennsyl-
vania needs to be examined in much greater detail. High
heat-flow-density values should be expected in the re-
region of the Reading Prong, where Eckstein and others
(1982) reported high radiogenic heat production.

RECOMMENDED FOR FURTHER READING

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