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Exploration for Hot Dry Rock Geothermal Resources in the Midcontinent USA

*Hot Dry Rock Conceptual Models for Exploration,
HDR Test Site Investigations, and the
Illinois Deep Drill Hole Project*

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Hot Dry Rock Conceptual Models for Exploration, HDR Test Site Investigations, and the Illinois Deep Drill Hole Project

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PREFACE

This report summarizes the integrated effort of several investigators at different academic institutions to use geological, geophysical, and geochemical data to develop a strategy for the location of potential hot dry rock (HDR) geothermal resources in the Midcontinent of the United States. HDR sites in the Midcontinent are of considerable interest because of the high population density and the strong demand upon conventional energy sources. However, the lack of surface manifestations of potential HDR sites necessitates a strategy based largely upon subsurface investigations.

To conduct this study, a group of investigators gathered to combine their interests and expertise. L.W. Braile, W.J. Hinze, and R.R.B. von Frese from Purdue University coordinated the project and conducted studies related to the exploration strategy, the southeast Michigan test site, and the use of magnetic methods in HDR investigations. E.G. Lidiak and R.E. Denison, University of Pittsburgh, were responsible for basement geology studies. G.R. Keller and R.F. Roy of the University of Texas at El Paso were concerned with gravity anomaly and heat flow data respectively and together worked on the Mississippi Embayment test site. C.A. Swanberg then of New Mexico State University was involved with geochemical data and, with C.L.V. Aiken of the University of Texas at Dallas, was concerned with the western Nebraska test site. Major contributions to the western Nebraska test site study were also made under separate funding by W.D. Gosnold of the University of North Dakota. Paul Morgan, then at the Lunar and Planetary Institute, Houston, was primarily concerned with thermal gradient data and HDR models.

The research program as originally designed and implemented was intended for a multiyear lifetime with increased funding in successive years. Unfortunately, funding limitations and uncertainties made it impossible to meet the ultimate goal of the research. Nonetheless, significant progress has been made which will serve as an excellent basis from which future research can logically progress when funding becomes available.

The emphasis of the program was diverted at the request of the Los Alamos National Laboratory when a unique opportunity became available to investigate the Illinois deep drill holes, which penetrated nearly a vertical kilometer of granitic basement rock in two locations in northern Illinois. Hinze, Keller, Lidiak, and Roy all became involved in scientific investigations at the

Illinois site, and Hinze served as a member of the Steering Committee for the holes.

This report is divided into two volumes. Volume I consists of an introduction, a geologic overview of the Midcontinent, and data acquisition and evaluation. Volume II contains a section on conceptual models for hot dry rock geothermal exploration; local test site evaluations of geothermal prospects in western Nebraska, the Mississippi Embayment area, and southeastern Michigan; and studies related to the deep drill holes in northern Illinois. Full integration of this report has not been possible due to funding limitations; for example, geological and geochemical discussions of the local test sites are given in the geologic overview and data reduction and evaluation sections, respectively. Thus, the two volumes of this report should be used as a single unit rather than as two separate volumes.

We thank Barbara Carlos of the Los Alamos National Laboratory for numerous helpful and constructive comments on a draft of this final report. S. Aldis, D. Chady, M. Dickey, R. Dotson, K. Hrametz, C. Kohring, and P. Thompson assisted P. Morgan at the Lunar and Planetary Institute in the final preparation and production of this report. The Lunar and Planetary Institute (LPI) is operated by the Universities Space Research Association under contract no. NASW-3389 from the National Aeronautics and Space Administration. This is LPI contribution No. 527.

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EXPLORATION FOR HOT DRY ROCK GEOTHERMAL RESOURCES IN THE MIDCONTINENT USA

Hot Dry Rock Conceptual Models for Exploration, HDR Test Site Investigations,
and the Illinois Deep Drill Hole Project

by

W. J. Hinze, L. W. Braile, R. R. B. von Frese, E. G. Lidiak, R. E. Denison,
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ABSTRACT

The Midcontinent of North America is commonly characterized as a stable cratonic area which has undergone only slow, broad vertical movements over the past several hundreds of millions of years. This tectonically stable crust is an unfertile area for hot dry rock (HDR) exploration. However, recent geophysical and geological studies provide evidence for modest contemporary tectonic activity in limited areas within the continent and, therefore, the possibility of localized thermal anomalies which may serve as sites for HDR exploration. HDR, as an energy resource in the Midcontinent, is particularly appealing because of the high population density and the demand upon conventional energy sources. Five generalized models of exploration targets for possible Midcontinent HDR sites are identified: (1) radiogenic heat sources, (2) conductivity-enhanced normal geothermal gradients, (3) residual magmatic heat, (4) sub-upper crustal sources, and (5) hydrothermal generated thermal gradients. Detection of possible HDR targets is a formidable problem in the Midcontinent where there are no direct surface manifestations (e.g., recent volcanism) of the thermal anomalies. Exploration must be based on subsurface information derived from a variety of geological, geochemical, and geophysical evidence. Thus, exploration strategy in the stable craton is different from that used for high-temperature geothermal sites which occur in tectonically active regions.

Three potential sources of HDR, each covering approximately a 2° x 2° area, were identified and subjected to preliminary evaluation with ad hoc exploration strategies. In the Mississippi Embayment test site, lateral thermal conductivity variations and subcrustal heat sources may be involved in producing abnormally high subsurface temperatures. Studies indicate that enhanced temperatures are associated primarily with basement rift features where vertical displacement of aquifers and faults cause the upward migration of hot waters

leading to anomalously high, local, upper crustal temperatures. The Western Nebraska test site is a potential low temperature HDR source also related, at least in part, to groundwater movement. There appear to be much more widespread possibilities for similar HDR sites in the Great Plains area. The Southeast Michigan test site was selected for study because of the possible presence of radiogenic plutons overlain by a thickened sedimentary blanket. There is no direct information on the presence of abnormally high temperatures in this area, but the study does show that a combination of gravity and magnetic anomaly mapping with regional geological information derived from sparse drill holes in the Phanerozoic rocks is useful on a widespread basis for focusing on local areas for detailed evaluation.

A variety of studies related to the Illinois Deep Drill Hole Project illustrate the utility of regional and local geophysical studies in delineating radiogenic plutons in the basement rocks. In addition, evidence indicates that despite the high uranium and thorium content of the granitic pluton penetrated by the Illinois Deep Drill Hole, the heat flow is only slightly above normal, probably as a result of the limited depth extent of the pluton.

IV. HOT DRY ROCK CONCEPTUAL MODELS FOR EXPLORATION STRATEGY

In order to develop an exploration strategy for Hot Dry Rock (HDR) resources it is necessary to develop conceptual models of the sources of potential HDR resources. As HDR technology is at present best suited to low permeability crystalline rocks, the exploration strategy focuses upon areas of the crystalline basement where the mean geothermal gradient down to the top of the basement is high enough for economic extraction of heat from the basement. In this section models are developed that emulate geological situations which cause anomalously high basement temperatures and therefore potential HDR resources. Five basic models have been examined which may have relevance to the Midcontinent: (1) radiogenic heat sources; (2) conductivity-enhanced gradients with normal heat flow; (3) residual magmatic heat; (4) sub-upper crustal heat sources; and (5) hydrothermal heat transfer. These five conceptual models are illustrated in Figure IV.1. Each model is discussed below, and with simplified geometries, the parameters which determine the effectiveness of each model are evaluated. These parameters are then related to measurable geological and geophysical parameters as relevant to an exploration strategy. Finally, the exploration significance of each model is discussed.

LATERAL CONDUCTIVITY VARIATIONS

HYDROTHERMAL

RADIOGENIC HEAT SOURCE

SUB-UPPER CRUSTAL SOURCE

RESIDUAL MAGMATIC HEAT

Figure IV.1.
Conceptual models for Midcontinent hot dry rock geothermal resources.

A. Target 1 - Radiogenic Heat Sources

Most of the earth's heat at the present stage in its evolution is thought to be generated by the decay of the isotopes ^{235}U , ^{238}U , ^{232}Th , and ^{40}K . The bulk of these isotopes probably reside in the mantle, but through differentiation of the mantle to produce as an end product the continental crust, the greatest concentrations of the isotopes are found in the continental crust. These concentrations are variable, and available data indicate that about 50% of the surface heat flow from stable continental regions is derived from crustal radiogenic heat production, although this contribution to surface heat flow varies from a few percent to greater than 65% (e.g., Morgan, 1984). Where concentrations are higher than average, the surface heat flow is correspondingly higher, and it is these areas that we seek as an HDR exploration target.

Uranium and thorium are incompatible elements, and thus their concentrations cannot be accurately predicted from bulk chemical descriptions of rocks. The only generalization that can be made is that they tend to be concentrated in low-melting-temperature ("acidic," "felsic," or "granitic") rocks. In addition, concentrations of these elements can be changed by a variety of geological transport processes, including primary differentiation of the crust, magmatic and metamorphic fluid circulation, and meteoric water circulation. Studies of uranium in metamorphic rocks indicate that it is essentially immobile up to amphibolite grade (although it can be transported by groundwater) but it is transported through migrating fluid phases during granulite metamorphism (Dostal and Capedri, 1978). Presumably thorium is mobilized at a similar metamorphic grade (McLennan and Taylor, 1980). However, despite the wide range of concentrations of these heat producing elements in typical crustal rocks, their contribution to the surface heat flow from subsurface medium to high-grade metamorphic and plutonic rocks can be predicted if the concentrations of these elements at the top of the basement are known.

Two components to heat flow measured in stable continental regions are generally acknowledged: heat flow from the upper mantle and heat generated by heat production in the crust (Birch et al., 1968; Roy et al., 1968a; Lachenbruch, 1968). The key to separating these components from heat flow determinations in plutonic rocks was the discovery by Birch et al. (1968) of a linear relationship between heat flow and surface heat production of the form:

$$q_s = q^* + A_s b$$

where q_s is the measured heat flow, and A_s is the heat production of the plutonic rocks in which the measurement was made. q^* and b are the constants of the linear relationship, interpreted to be the heat flow from below the crustal heat producing layer, commonly known as the reduced heat flow, and the depth scaling parameter for the heat producing layer, respectively. This discovery led to the definition by Roy et al. (1968a) of a heat flow province as an area in which single values of q^* and b are valid. Roy et al. (1972) later redefined the term heat flow province to be a region with the same heat flow from the mantle (q^*), essentially removing the variability in crustal heat production from the definition. Subsequently, the linear heat flow/heat production relationship has also been established for sites in metamorphic rocks (e.g., Richardson and Oxburgh, 1978), and heat flow provinces have now been defined for seventeen continental areas, as listed in Table IV.1. These linear relationships are only valid for crystalline terranes, but they can be applied to sediment covered basement with the modification that surface refers to basement surface. Radiogenic heat production in sediments is not generally significant for geothermal considerations and is ignored in the present discussion.

The depth parameter, b , defined in the equation above may have some significance in defining the relative depth distribution of heat production in an area for which a single linear relationship of this type is valid, but does not uniquely define the form of this distribution. Under the assumption that heat production does not increase with depth at any level, two distributions of heat producing elements have been proposed which satisfy the heat flow/heat production relationship. In the first, the step model, the heat production is assumed to remain constant from the surface to a depth given by b (Birch et al., 1968). In the second, the exponential model, heat production is assumed to decrease exponentially from the surface (Lachenbruch, 1968), such that heat production A_z at depth z is given by

$$A_z = A_s \exp(-z/b).$$

This latter relationship has the advantage that the linear relationship between heat flow and heat production is preserved during differential erosion

TABLE IV.1. Heat Flow, Heat Production, and Age Parameters for Several Heat Flow Provinces^a

Province	Age		q_s	q^*	q_A	b	n	Reference
	1)	2)						
<u>North America</u>								
Basin and Range	C	C	-	59±14	-	9.4	12	1
ditto	C	C	92±33	69±34	23	(10.0)	86	2
Southeastern Appalachians	P1	P1	49±11	28±1	21	7.2	3	3
Sierra Nevada	M	C	37±13	18±3	19	10.1	10	2
Eastern United States	Pr	Pr	57±17	33±4	24	7.5	15	4
Superior	A	A	34±8	21±1	13	14.4	11	5
ditto, corrected	A	A	-	28±1	-	13.6	11	5
<u>South America</u>								
Brazilian coastal shield	LPr	LPr	56±15	28±7	28	13.1	4	6
<u>Europe</u>								
Baltic Shield	APr	APr	36±8	22±6	14	8.5	3	7
Bohemian Massif	CM	C	73±18	44±1	29	3.8	2	8
Ukraine	A	A	37±8	25±2	12	7.1	12	9
England and Wales	P1	P1	59±23	23±2	36	16.0	10	10
<u>Africa</u>								
Niger	LPr	LPr	20±8	11±8	9	(7.5)	-	11
Zambia	LPr	LPr	67±7	40±6	27	(7.5)	-	12
ditto	LPr	CM	67±7	40±6	27	(7.5)	-	12
<u>Australia</u>								
Western	A	A	39±8	26±8	13	4.5	9	13
Central	EPr	EPr	83±21	27±6	56	11.1	10	14
Eastern	P1	C	72±27	57±22	15	(11.1)	-	14
<u>Asia (Indian Shield)</u>								
Late Proterozoic	LPr	LPr	71±11	38±2	33	14.8	4	15
ditto	LPr	CM	71±11	38±2	33	14.8	4	15
Archean Shield	A	A	49±8	33±1	16	7.5	2	15

^a Ages assigned to each province are (1) crustal age and (2) age of last tectonothermal event (ages estimated from Condie, 1976), Figure 5-10, page 110, Sclater et al., 1980, and as discussed in text); Age key - C Cenozoic, M Mesozoic, CM Cenozoic-Mesozoic, Pl Paleozoic, Pr Proterozoic, LPr Late Pr, EPr Early Pr, A Archean, APr Archean-Proterozoic.

q_s is mean surface heat flow (mW/m^2) $\pm$ standard deviation. q^* is reduced heat flow (mW/m^2) $\pm$ uncertainty. q_A is the estimated crustal heat production (mW/m^2), calculated by $q_A = q_s - q^*$. b is the depth parameter (km) for upper crustal heat production - see text. Assumed b values are given in parentheses. n is the number of heat flow/heat production data used to define the heat flow province.

References: 1 Roy et al. (1968a,b) see also Blackwell (1978); 2 Lachenbruch and Sass (1978); 3 Costain (1977); 4 Roy et al. (1968a); 5 Jessop and Lewis (1978); 6 Vitorello (1978); 7 Rao and Jessop (1975); 8 Cermak (1975); 9 Kutas (1977); 10 Richardson and Oxburgh (1978); 11 Chapman and Pollack (1974); 12 Chapman and Pollack (1977); 13 Jaeger (1970), Sass et al. (1976b); 14 Sass et al. (1976b); 15 Rao et al. (1976). Table modified from Vitorello and Pollack (1980).

(Lachenbruch, 1968), and is consistent with heat generation data from batholiths at different depths within the crust (Swanberg, 1972). Costain and Glover (1979), however, have recently published data from the southern Appalachians of the USA which they interpret as supporting the step model. As discussed by Blackwell (1971), the actual distribution is likely to be much more complex than either model.

Recent attempts to understand the distribution of heat producing elements with depth have focused on the processes of transport of uranium, thorium, and potassium in the crust and indicate that there may be different depth parameters controlling the distributions of these three elements in the crust, and/or different distributions at different depths (e.g., Costain, 1978; Gosnold and Swanberg, 1980; Jaupart et al., 1981). Heat production data have recently been published by Nicolaysen et al. (1981) for a 16-km, originally vertical section of Archean granitic crust, exposed in the core of an updomed and overturned sequence of strata in the Vredefort structure in South Africa. Individual heat production values for uranium, thorium, and potassium are plotted as a function of depth in Figure IV.2. It is clear that the data do not follow simple distributions nor the same distributions with depth,

Figure IV.2.

Heat production as a function of depth in the Vredefort structure, South Africa (data from Nicolaysen et al., 1981). Actual data are shown by open circles for uranium, dots for thorium, and x's for potassium. Smooth curves show the results of exponential regression analyses of the uranium and thorium versus depth data and the mean of the potassium data. (From Morgan, 1984)

but the uranium and thorium data can be approximated by exponential curves, as shown in the figure. Potassium does not vary significantly with depth. Other distributions, such as a more complex exponential distribution proposed by Gosnold and Swanberg (1980), could also be fitted to the data. The distribution of heat production in the crust is clearly complex, and additional data are required before generalizations about the distribution can be made.

Allis (1979) and Rybach and Buntebarth (1982) have proposed relationships among density, seismic velocity, heat generation, and mineralogy from which it may be possible to predict heat production from other physical properties. From these relationships, Allis (1979) has proposed that an exponential decrease with depth in heat producing elements is only applicable to the upper crust, and that the lower crust is uniformly depleted with a mean heat production of $0.2 \pm 0.1 \mu\text{W/m}^3$. With the data available at present, the only possible conclusion is that the distribution of heat producing elements in the upper crust appears to be bracketed by the step and exponential models. If a different distribution is applicable to the lower crust, it would be undetectable in the available heat flow/heat production data, and it is insignificant for HDR exploration purposes because its contribution to surface heat flow would be relatively low.

Vitarello and Pollack (1980) have suggested that crustal heat production tends to decrease with crustal age due to increasing erosion of the heat producing elements with time. However, a plot of the contribution of crustal heat production to surface heat flow as a function of crustal age, shown in Figure IV.3, shows no decrease in heat production with crustal age. In fact, the highest crustal heat production contribution comes from an Early Proterozoic province. The only generalization that is suggested by the limited data available is that Archean provinces generally have low crustal heat production and without later magmatic activity would not be obvious sites for radiogenic heat source HDR exploration.

From heat flow/heat production studies in the regions surrounding the Midcontinent region (see Table IV.1), it is likely that the effective thickness of the heat producing layer in the Midcontinent is between 7 and 14 km. The highest concentrations of heat producing isotopes are likely to be found in "granitic" plutons that have not been subjected to high-grade metamorphism, but it is impossible to predict these concentrations from the bulk

Figure IV.3.

Crustal heat production (defined by mean surface heat flow minus reduced heat flow) as a function of crustal age for heat production provinces listed in Table IV.1. See Table IV.1 for province codes and age assignments. Heat flow error bars are plotted to represent standard deviations of the mean surface heat flow data, and data are displaced as necessary on the age axis to clarify the plot. Data derived from assumed heat production depth parameters are indicated by codes in parentheses. (From Morgan, 1984)

chemistry of the plutons or from their age. We thus focus upon buried granitic plutons as a model for a radiogenic heat source (see Figure IV.1), and we seek a model to predict the thermal anomaly due to a radiogenic pluton and to relate this model to other geological and geophysical parameters.

The important parameters for potential radiogenic HDR resources therefore are those which identify and delineate plutons with a high concentration of

radiogenic isotopes. The concentration of heat producing isotopes in a pluton is not a function of the bulk properties of the pluton, but to a first approximation, the spatial distribution of these isotopes is controlled by the shape of the pluton. Simmons (1967a) suggested that this distribution is the same as the mass distribution for the pluton. This approximation is invalid for calculating geotherms within and below a pluton if the isotope distribution is nonuniform, but the approximation is reasonable for computing heat flow above a pluton for most likely isotope distributions. Assuming a conductive heat flow regime, and recognizing the equivalence of the potential fields generated by heat sources and mass anomalies, Simmons (1967a) presented the following relation between gravimetric and thermal anomalies:

$$q_a = F(A/2\pi G\rho)$$

where q_a is the anomalous heat flow, F is the anomalous gravitational attraction, A is the rate of heat production per unit volume, G is the gravitational constant, and ρ is the density contrast. This relation can be used in two ways: (1) to predict surface heat flow anomalies due to radiogenic plutons if the geometry and heat production of the plutons are known, using standard gravity modeling techniques (e.g., Talwani et al., 1959; Talwani and Ewing, 1960; Simmons, 1967a); or (2) to predict surface heat flow anomalies from gravity anomalies from plutons if the heat production and density contrasts of the plutons are known or if there are some surface heat flow data over the pluton. An example of this latter use of the relation between gravity and heat flow anomalies is given by Bott et al. (1972) for the Weardale granite in northeast England.

For HDR exploration, basement lithology must be supplemented by a catalog of heat production measurements on samples from wells that penetrate basement. From this catalog, potentially "hot" plutons could be identified. Geophysical techniques will provide information to locate and delineate these plutons, and the form and magnitude of the thermal anomaly can be predicted from the pluton gravity anomaly and some thermal data over the pluton anomaly, or the heat production/density contrast ratio for the pluton. From this analysis, a preliminary evaluation of the pluton as a heat source can be made, and the most favorable site for exploratory drilling identified.

B. Target 2 - Conductivity-Enhanced Gradients with Normal Heat Flow

Under steady-state conditions, and in the absence of heat sources, the conducted component of heat $\underline{q}$ is given by

$$\underline{q} = -K\nabla T$$

where K is thermal conductivity and T is temperature. In general in the earth, especially in the upper layers of the crust, K is neither spatially constant nor isotropic. Thus, even if the heat flow is constant, ∇T will be laterally variable. As a target for HDR exploration we seek conditions which will significantly enhance the vertical component of ∇T . This enhancement can have two forms: (1) enhancement by a reduction of K with respect to vertical heat flow; and (2) enhancement by an increase in the vertical component of $\underline{q}$ through thermal refraction.

Thermal conductivity depends on a number of factors including temperature, mineralogy, and porosity. Data for selected rocks have recently been compiled by Roy et al. (1981). Crustal materials have a wide range in conductivity at surface temperatures, ranging from less than 1 W/m•K for unconsolidated sediments to greater than 6 W/m•K for halite and quartzite. Most of this variation comes from the effects of low conductivity fluids filling pore space and high conductivity monomineralic rocks, both typically found at relatively shallow crustal depths. Anisotropy is controlled by factors such as bedding, schistosity, and fracturing; and conductivity in a rock can vary by more than a factor of two as a function of the orientation of heat flow (e.g., Clark, 1966). Conductivity is generally lowest perpendicular to the long axis or axes of the mineral or fabric orientation.

If conductivity varies only in the vertical direction, and if the regional heat flow q is known, the mean vertical temperature gradient is given by

$$dT/dz = q/K_m$$

where K_m is the weighted (proportional to thickness of each conductivity layer) harmonic mean conductivity of the section given by

$$\frac{D}{K_m} = \int_0^D \frac{1}{K(z)} \cdot dz$$

where $K(z)$ is the conductivity at depth z in the section, and D is the total thickness of the section. As $K(z)$ is not generally a smooth function of depth, the harmonic mean conductivity is usually computed as a summation of discrete layers defined by the layers in the section (e.g., Bullard, 1939). The effects on the geothermal gradient of conductivity variations with depth are clearly illustrated in Figure IV.4 with data from a drill hole in the southeastern corner of Kansas (from Blackwell and Steele, 1981). High gradients (averaging $53.4 \pm 1.5^\circ\text{C}/\text{km}$) were measured between 95 and 205 m depth, below which the gradients average less than $20^\circ\text{C}/\text{km}$. The gradient transition marks the contact between Pennsylvanian shales (low conductivity) and a predominantly limestone-dolomite (high conductivity) section of Mississippian and older age.

Thus, for high geothermal gradients due to vertical variations in thermal conductivity, of interest for HDR exploration, a thick sedimentary section with a low mean thermal conductivity is an exploration target. In the Midcontinent region, the most obvious sedimentary lithology with this desired property is flat-lying shale.

Increases in the vertical component of heat flow by thermal refraction are caused by lateral variations in thermal conductivity, including conductivity anisotropy. Unfortunately it is possible to write analytical solutions for thermal refraction for only a very limited set of geometries of conductivity contrast. These geometries include ellipsoids, spheroids, cylinders, spheres, and a planar interface (e.g., Jaeger, 1965; Lee and Henyey, 1974; Lee, 1975). For more complex geometries it is necessary to use numerical solutions, such as the finite difference or finite element techniques to compute thermal refraction (e.g., Simmons, 1965; Lee and Henyey, 1974) or electrical analogue models (e.g., Bullard, 1939). Although the details of thermal refraction are complex, the following generalizations can be made: (1) the effects of refraction on the vertical component of heat flow are largest close to rapid lateral changes in conductivity; (2) except in cases of very complex geometry, higher surface heat flow will occur on the side of the conductivity contrast in which the mean conductivity is highest and the high conductivities come closest to the surface; (3) the magnitude of

Figure IV.4.

Temperature-depth and gradient-depth curves for Kansas hole 31S/20E-22cac. Five-meter gradient intervals are plotted. (From Blackwell and Steele, 1981)

the thermal anomaly depends on the absolute conductivity contrast causing the thermal refraction and on the geometry, particularly the vertical extent of the conductivity contrast; and (4) the surface thermal anomaly decreases as the depth of the conductivity contrast increases.

The main shallow (less than 5-km) conductivity contrasts likely to occur in the Midcontinent study region are between the relatively high conductivity crystalline basement and low conductivity sediments, and between high conductivity evaporites and low conductivity sediments. Lateral conductivity contrasts are thus expected where faulting juxtaposes sediment against basement, or diapirism juxtaposes evaporites against other sediments. Gentle basement topography or folding and faulting in sediments without evaporites

are not expected to cause thermal anomalies significant to HDR exploration studies.

The effect of thermal refraction associated with basement faulting is illustrated by the example in Figure IV.5 (from Blackwell and Chapman, 1977). This model shows the refraction effects at the edge of a 1.5-km-deep sediment-filled graben with basement coming within 100 m of the surface at the graben margin. The surface heat flow is increased by approximately 30% close to the fault on the upthrown side. Similar magnitude anomalies were predicted by Simmons (1965). From Figure IV.5 it can be seen that the geothermal gradient is actually lower in the high conductivity basement, although the heat flow is higher close to the fault. This situation could therefore only produce an HDR resource if an additional thickness of sediments occurs over the basement uplift in which the high heat flow could generate a high gradient. However, the refracted high heat flow predicted by this model would be somewhat dissipated toward the surface by additional sedimentary layers, as is shown by the effect of depth of burial in a spherical conductivity contrast, shown in Figure IV.6 (from Lee, 1975). Thus, the maximum gradient enhancement over a basement fault with 1.5 km vertical throw, with the upthrown basement covered by 2 km of sediment, and with a 1:2 ratio between

Figure IV.5.

Configuration of isotherms and heat flow pattern across a range-bounding fault using a typical Basin and Range background heat flow. (From Blackwell and Chapman, 1977)

Figure IV.6.

Theoretical heat flow profiles running radially from the epicenter of a buried spherical conductivity contrast. Upper: depth to center of sphere, b , divided by radius of sphere, c , = 3.0; Middle: $b/c = 2.0$; Lower: $b/c = 1.1$. Distance is normalized to c . Hatched half-circle represents the buried sphere. Curves for different sphere to country rock conductivity ratios, K_1/K_2 , are shown by different lines with key in lower right. (From Lee, 1975)

sediment and basement conductivities is estimated to be on the order of 10-15%. Because of the difficulties of accurately predicting thermal refraction effects, however, numerical models should be computed for specific geometries and conductivity contrasts relevant to specific Midcontinent structures.

It may be possible to estimate the thermal refraction effects of a salt diapir by the analytical solution for a spherical conductivity contrast, as suggested by Lee (1975). As halite can have a very high conductivity (greater than 6 W/m $\cdot$ K), diapirs can result in very high (approximately 5:1) lateral conductivity contrasts. In addition, the thermal anomaly over a diapir can be

further enhanced by the focusing effect of the upward convex shape of the top of a diapir. The modeled results of the thermal refraction effects of the Vejrum salt diapir, Denmark, are shown in Figure IV.7 (from Jensen, 1983). This model was computed using the finite element technique with complex conductivity layering and some radiogenic heat production in the sediments. The ratio of halite to sediment conductivities ranges from approximately 2:1 to 3:1. A comparison of the results of the analytical and numerical models shown in Figures IV.6 and IV.7, respectively, indicates that the analytical solution gives a very good approximation of the shape and magnitude of the anomaly. The complex form of this solution and the nature of thermal refraction, however, prevent a direct computation of the thermal anomaly from other geophysical anomalies.

Thermal anomalies on the order of a few hundred percent can easily be generated with the high conductivity contrasts associated with salt diapirs if the top of the diapir is close to the surface. However, this is not an ideal situation for an HDR resource as high gradients associated with the anomaly will exist only in the thin section of sediments above the diapir. For a potential HDR resource, the top of the diapir must be covered by at least 2 km of sediments, and as shown in Figure IV.6, the effects of thermal refraction will be somewhat dissipated by this overlying sedimentary pile. However, significant (30% or higher) gradient anomalies could still exist over buried diapirs, making the diapirs targets of HDR exploration.

The significant measurable parameters associated with conductivity-enhanced gradients with normal heat flow relate to lithology and structure. A catalog of lithologies with thermal conductivities should be compiled for the Midcontinent. Other geological and geophysical techniques should concentrate on the location of thick sections of low conductivity strata (primarily shales) and the locations of major faults in the basement/sediment interface and salt diapirs.

C. Target 3 - Residual Magmatic Heat

Residual magmatic heat is perhaps the most attractive HDR exploration target as magmas can bring high temperatures close to the surface. Due to interaction with groundwater and rapid cooling by the atmosphere, high temperatures associated with volcanic rocks last only a very short time (maximum a few years unless continually replenished). As there are no active volcanic systems in Midcontinent, volcanic heat sources are not considered in

The absolute value of temperature gradient around the Vejrum salt diapir.

Surface heat flux values around the Vejrum salt diapir.

Figure IV.7.

Calculated thermal conditions around the Vejrum salt diapir. Upper: contours of absolute temperature gradient; Lower: surface heat flux around the diapir. (From Jensen, 1983)

the HDR models. Thermal anomalies associated with plutonic activity can last up to a few million years after the last plutonic activity, and as it is impossible to rule out the remote possibility of plutonic activity in the Midcontinent during the last few million years (see, for example, discussion on the Mississippi Embayment, Section V.B, below), the conditions favorable for residual magmatic heat will be examined.

The most optimistic estimates of pluton cooling for HDR exploration purposes (those which give the maximum thermal anomaly for the longest time) are based on the assumption that the pluton cools conductively. A common modeling technique assumes that the pluton intrudes instantaneously, that the pluton and host rock have the same isotropic, homogeneous, temperature-independent thermal properties, and that there is no groundwater or magma convection. The surface boundary condition (temperature equals arbitrary zero) is maintained by placing a negative image source equal in strength to the pluton above the surface, and for calculation of anomalous temperatures the intrusion temperature is given by the magma temperature minus the host rock temperature at the mean depth of the intrusion. The intrusion temperature or the effective magma-specific heat may be modified to include the effects of magma latent heat (e.g., Jaeger, 1968). If the pluton can be approximated by a simple geometric shape, the anomalous temperatures associated with the pluton can be calculated using analytical solutions at any point in or around the pluton, at any time after intrusion (e.g., Lovering, 1955; Carslaw and Jaeger, 1959; Jaeger, 1964, 1968; Simmons, 1967b). Absolute temperatures can be calculated by adding the anomalous temperature to the regional undisturbed geothermal gradient, and gradient anomalies can be calculated by differentiating the temperature solutions with respect to depth.

Conductive cooling solutions for instantaneous intrusions can be superposed, allowing complex pluton shapes to be approximated by adding simple geometric shapes, one of the most useful of which is the rectangular parallelepiped (Carslaw and Jaeger, 1959, p. 56; Blackwell et al., 1974). Alternatively, anomalies can be calculated for complex pluton shapes and heterogeneous thermal properties using numerical approximations (finite difference and finite element solutions, e.g., Lee et al., 1980). Generally, however, computation of anomalies due to pluton models is a smaller problem than constraining the geometry and time of intrusion of the pluton.

The magnitude and duration of anomalies associated with plutons depend upon (1) size and geometry of the pluton, (2) depth below surface, (3) intrusion temperature relative to country rock, and (4) intrusion and host rock thermal properties. As the rock thermal properties are not likely to vary by more than a factor of 2, the intrusion temperature is likely to be within the range 650 - 1200°C, and the country rock temperature depends on (2), the depth of the pluton below the surface, the main parameters

controlling the magnitude and duration of the thermal anomaly associated with a pluton are the size and geometry of the pluton and its depth of emplacement. Both magnitude and duration of the anomaly increase with increasing pluton size, but the magnitude decreases while the duration increases with increasing depth of emplacement. The effect of pluton geometry is complex, but generally both the magnitude and duration of the thermal anomaly will decrease as the ratio of surface area to volume for the pluton increases.

Some estimate of the size, depth of emplacement, and age of a pluton for a significant thermal anomaly to exist can be obtained by examining the solution for an instantaneous point source of heat in a semi-infinite medium with image source to maintain surface zero temperature. If the heat source is located at (x', y', z') (z axis directed vertically down) and releases a total quantity of heat $S\rho c$ (ρ is density and c is specific heat), the anomalous temperature T due to the source and its image is given by

$$T(x, y, z, t) = S[\exp(-r_1^2/4kt) - \exp(-r_2^2/4kt)]/[8(\pi kt)^{3/2}]$$

where $r_1^2 = (x - x')^2 + (y - y')^2 + (z - z')^2$ and $r_2^2 = (x - x')^2 + (y - y')^2 + (z + z')^2$, and where k is diffusivity and t is time after source emplacement (Carslaw and Jaeger, 1959; Simmons, 1967b). If the source is on the axis $x = y = 0$, i.e., at $(0, 0, z')$, immediately above the source at the surface $(0, 0, 0)$ the heat flow is given by

$$q = -SKz'\exp(-z'^2/4kt)/(8\pi^{3/2}k^{5/2}t^{5/2})$$

where K is thermal conductivity given by $K = k\rho c$. The time- and depth-dependent portion of this equation, $(z'/t^{5/2})\exp(-z'^2/4kt)$, is plotted in Figure IV.8 for the time intervals for which an intrusion would have a significant effect on the surface heat flow for intrusion depths z' of 1, 5, 10, and 15 km, assuming a thermal diffusivity of $32 \text{ km}^2/\text{m.y.}$ ($1.0 \text{ mm}^2/\text{s}$). As shown in Figure IV.8, the maximum effect of a point source intrusion at a depth of 1 km occurs approximately 0.003 m.y. (3,000 years) after intrusion. The corresponding times for intrusions at 5, 10, and 15 km are 0.075, 0.33, and 0.7 m.y. (75,000, 330,000, and 700,000 years), respectively. The life-times of the anomalies, measured as the time that the anomaly is greater than a tenth of its peak value, are 0.016, 0.44, 1.45, and 3.75 m.y. (16,000,

Figure IV.8.

Plot of time and depth portion of equation for instantaneous point source intrusion (see text) as a function of time after intrusion for different point source depths. Depths of point source, z , given in km. Uniform diffusivity of $32 \text{ km}^2/\text{m.y.}$ (approximately $1 \text{ mm}^2/\text{s}$) used to compute curves.

440,000, 1,450,000, and 3,750,000 years) for source depths of 1, 5, 10, and 15 km, respectively. Thus, it is clear that thermal anomalies associated with intrusions, even taking the slowest estimate of conductive cooling, will last no more than a few millions of years.

The rapid decrease in peak surface anomaly with source depth is also shown in Figure IV.8. The peak anomaly decreases by almost five orders of magnitude as the source depth increases from 1 to 15 km. This effect is partially offset by the possibility of increasing source size with depth through larger intrusions at depth, but as the temperature contrast between intrusion and host rock decreases with depth, the general trend of decreasing

anomaly with increasing source depth cannot be reversed. Below a depth of 15 km or so, a single intrusion is unlikely to have a major effect (greater than 10%) on the near-surface geothermal gradient unless the intrusion is very large (greater than $5,000 \text{ km}^3$). Multiple intrusions may produce some surface gradient anomaly, but these intrusions are likely to be associated with a significant mantle thermal anomaly and elevated regional heat flow and will be considered in the next section. Examples of the results of surface thermal anomaly calculations for upper crustal intrusions are shown in Figures IV.9 and IV.10.

Groundwater convection associated with an igneous intrusion causes more rapid cooling of the intrusion but increases the near-surface geothermal

Figure IV.9.

Maximum surface gradient anomaly (gradient over center of body) as a function of time for cubic intrusions defined on figure, assuming an intrusion temperature of 1200°C (including latent heat), a diffusivity of $1 \text{ mm}^2/\text{s}$, and purely conductive cooling.

Figure IV.10.

Vertical and horizontal gradients at the edge of a 20 x 20 x 20-km pluton as a function of time after intrusion. Depth to top of pluton is given by d . Dashed lines give measured gradients in Los Alamos Fenton Hill well EE-2.

gradient over the area where the heated groundwater rises toward the surface (e.g., Norton, 1977; Norton and Knight, 1977; Norton and Knapp, 1977). The conduction cooling calculations outlined above give the slowest possible cooling rates for an upper crustal intrusion as it is inevitable that some groundwater convection will be associated with these intrusions. Convection calculations require numerical modeling techniques and are dependent upon parameters which are only poorly known, such as host rock and magma permeabilities. Thus, for an exploration strategy, conduction calculations are probably the most appropriate tool for estimating the effect and duration of an intrusion, but it must be remembered that convection will locally enhance near-surface anomalies and speed cooling of the intrusion.

Exploration for HDR resources associated with residual magmatic heat should focus upon techniques to locate very young intrusions (less than a few million years old). Geological exploration should focus upon very young surface manifestations and hydrothermal activity and deposits. Geophysical

techniques should attempt to locate the young intrusions by surface deformation studies, local heat flow, gravity and magnetic surveys, seismic reflection and refraction studies, and passive seismic (microearthquake) surveys to locate local areas of microseismic activity caused by thermal fracturing associated with intrusions (e.g., Knapp and Knight, 1977).

D. Target 4 - Sub-Upper Crustal Heat Sources

Studies of heat flow from below the upper crustal layer enriched in radiogenic heat producing elements (see Section IV.A.) indicate that this quantity is relatively constant, $27 \pm 4 \text{ mW/m}^2$ ($n = 10$), for stable continental areas (e.g., Morgan, 1984). Any significant increase in this lower crustal-mantle contribution to the surface heat flow is classified as a sub-upper crustal heat source. As there is no evidence for significant lateral variations in radiogenic heat production in the lower crust or upper mantle, increased sub-upper crustal heat flow must come from dynamic crustal and/or mantle processes. Specifically, high mantle heat flow is usually associated with lithospheric thinning or the thermally equivalent concept of the penetration of an asthenolith into the lithosphere. Heat flow is increased as the geotherm is raised by bringing hot asthenosphere closer to the surface than it would be in a stable "equilibrium" setting.

Thin lithosphere and high apparent mantle heat flow occur in areas of active tectonics and magmatism, areas in which shallow thermal processes often mask and make it difficult to measure the mantle-reduced heat flow component (e.g., Lachenbruch and Sass, 1978). Seismic and other data do not indicate large volumes of partial melt in the mantle in these regions; therefore, some indication of maximum mantle temperatures can be obtained from mantle solidus temperatures. Experimental studies on the types of magmas erupted in continental rift settings constrain the mantle solidus to around 1200°C (e.g., Wendlandt and Egglar, 1980), and thus this temperature can be taken as the temperature of the base of the lithosphere (e.g., Wendlandt and Morgan, 1982; Morgan, 1983). Alternatively, as basalt is a common mantle magmatic product, the dry basalt solidus can be used to define the temperature at the base of the lithosphere (temperature = $1050 + 3z^\circ\text{C}$, where z is the depth in km; Lachenbruch and Sass, 1978), which ranges from 1170 to 1650°C as the depth increases from 40 to 200 km. As maximum reduced heat flow corresponds to minimum lithospheric thickness, at which both solidi are similar, the choice

of mantle solidi for HDR models is unimportant. The maximum temperature likely at the base of a 35-km-thick crust is around 1200°C.

Moho temperature in stable continental lithosphere depends primarily upon the quantity and distribution of upper crustal radiogenic heat production, and to a lesser extent upon crustal thickness. For a 35-km-thick crust, Morgan (1984) estimates a possible range of Moho temperatures from 400 to 775°C. Thus, an increase in Moho temperature to 1200°C represents an increase in mantle heat flow of 57 to 30 mW/m², assuming a mean crustal thermal conductivity of 2.5 W/m·K. Thus an increase in mantle heat flow of a factor between 2 and 3 may be expected if the lithosphere is thinned to the Moho.

Additional lithospheric thinning can occur by penetration of the asthenosphere into the crust. Here maximum temperatures will be buffered by crustal solidi which probably range from 800 to 1000°C in the lower crust (e.g., Wyllie, 1977). Such buffering will be associated with significant magmatic activity, and the deep thermal regime will be masked by the effects of shallow magmatic activity, the effects of which have been discussed above (Section IV.C.). High mantle heat flow can cause repeated crustal magmatism, prolonging the thermal effect of individual intrusions, but the thermal effects of greatest duration, and those probably most appropriate to Midcontinent HDR resources, will be associated with the high mantle heat flow and lithospheric thinning. In the Midcontinent these phenomena are most likely to be associated with thermal thinning and rifting of the lithosphere (Morgan, 1983; Keller et al., 1983). These effects can potentially double the shallow geothermal gradient and thus halve the depth required for HDR drilling.

The duration of high mantle heat flow following a tectonic and/or magmatic event may be examined by plotting reduced heat flow data as a function of age of last tectonothermal activity of the heat flow sites, as shown in Figure IV.11. This plot indicates that mantle thermal anomalies decay in a period of a few hundred million years or less. Morgan (1984) and Morgan and Sass (1984) estimated that the stable continental lithosphere probably ranges in thickness from around 80 to 220 km, depending upon the amount and distribution of crustal heat production. If we assume a diffusivity of 1.2 mm²/s, this range suggests lithospheric thermal relaxation time constants from 60 m.y. for the thinnest to around 500 m.y. for the thickest lithosphere (Figure 9 of Lachenbruch and Sass, 1977), in good

Figure IV.11.

Reduced heat flow as a function of age since last tectonothermal activity for heat flow provinces listed in Table IV.1. (From Morgan, 1984)

agreement with the reduced heat flow data. The form of the decay of a triangular temperature disturbance in a 150-km-thick lithosphere is shown in Figure IV.12 and indicates that much of the surface heat flow anomaly decays in less than 100 m.y. Thus, although sub-upper crustal heat sources can cause significant shallow thermal anomalies, they must have been active during the last 100 m.y., and preferably during the last 50 m.y., to have a significant effect on shallow temperatures and potential HDR resources.

Exploration for sub-upper crustal heat sources in the Midcontinent should focus upon the detection of phenomena associated with lithospheric thinning which has occurred during the last 100 m.y. Thermal thinning of the continental lithosphere without significant crustal extension causes uplift due to thermal buoyancy, as indicated in Figure IV.13. Thus, Cenozoic volcanism associated with regional domal uplift would indicate significant

Figure IV.12.

Conductive relaxation of a triangular-form thermal perturbation in a fixed thickness lithosphere, assuming a steady-state geotherm equivalent to a 150-km-thick lithosphere and a maximum temperature perturbation at a depth of 30 km. The three sets of curves are for three different lower boundary conditions. Upper left: constant temperature; upper right: lithosphere and asthenosphere, a continuous conductive medium; lower: no heat flux. (From Morgan, 1984; adapted from Vitorello and Pollack, 1980)

Figure IV.13.

Uplift as a function of lithospheric thickness for slow thinning and rapid thinning (see text). Upper: initial lithosphere thickness 100 km; lower: initial lithosphere thickness 200 km. (From Morgan, 1983)

lithospheric thinning and a potential HDR resource. With crustal extension either uplift or downwarp can occur, but the extension and lithospheric thinning can be identified by rifting and possibly volcanism. Again, significant activity should have occurred during the Cenozoic for reasonable HDR potential.

Residual effects of lithospheric thinning may be manifest in geophysical data by high heat flow, anomalously low mantle seismic velocities, long wavelength negative Bouguer gravity anomalies, and anomalously shallow depths to the Curie isotherm (e.g., Morgan, 1983; Fairhead and Reeves, 1977; Brown and Girdler, 1980). Where the surface manifestations of rifting are concealed by sedimentation, modification of the crust by rifting may be detected geophysically. Gravity, magnetic, and seismic techniques should be used to detect major crustal intrusive bodies, predominantly basic in composition, and to detect anomalous deep crustal and Moho structure which may be associated with rifting (e.g., Keller et al., 1983; Mooney et al., 1983). If magmatic

activity is suspected to be very young (less than 5-10 m.y.), seismic methods, supported by other geophysical techniques, should be used to attempt to locate lower crustal upper mantle magma chambers or pockets of partial melt (low seismic velocity, high attenuation, large rigidity contrast, low density). The combination of these factors with Cenozoic magmatism and/or rifting with uplift or downwarp would indicate a potential HDR resource.

E. Target 5 - Hydrothermal Systems

Although hydrothermal systems are not generally considered as an HDR exploration target, there are conditions where it may be desirable to exploit a hydrothermal anomaly by HDR technology. These conditions include hydrothermal systems in which natural rock permeability is not sufficient for adequate production of the intrinsic hydrothermal fluid, and those in which hydrothermal temperatures are insufficient for a resource but high, useful temperatures exist in the sub-flow basement.

There are basically two types of hydrothermal systems: free convection systems in which fluid flow is driven by thermal buoyancy forces and forced convection systems where the driving force is externally derived, either from hydrostatic or lithostatic forces. Elevated geothermal gradients (greater than 50-200°C/km) and low permeabilities are required for free convection to occur (e.g., Donaldson, 1962; Lowell, 1979), and this secondary mechanism for elevating shallow temperatures is commonly associated with young intrusions (e.g., Morgan et al., 1977). In the Midcontinent, forced convection is likely to be the primary type of hydrothermal system. Forced convection driven by lithostatic forces has recently been discussed by Cathles and Smith (1983) with reference to the genesis of Mississippi Valley-type lead-zinc deposits. They conclude that this may be an efficient mechanism for bringing temperatures of the order of 150°C within 1 km of the surface, but that it is most likely to occur during basin subsidence and sediment dewatering. This process may be occurring in the Gulf Coast at the present, but it is probably finished in the older Midcontinent basins. Hydrostatically driven forced-convection hydrothermal systems are a consequence of variations in water table elevation which are related to modern topography and hydrologic conditions, and they are thus the most probable hydrothermal systems to be found at present in the Midcontinent region.

Thermal perturbations due to groundwater flow depend on three basic factors: the piezometric gradient driving the flow, the permeability

structure of the rock through which the groundwater is flowing (these two factors define the flow field), and the heat flow into the base of the flowing system and thermal properties of the fluid and rock matrix. Detailed modeling of the thermal effects of groundwater flow systems requires a detailed knowledge of these factors (e.g., Smith and Chapman, 1983), but often this information, especially the detailed permeability structure, is not available. An alternative approach is to use simple analytical models to predict conditions for the onset of forced convection to assess areas for further exploration. A useful model in this respect was given by Domenico and Palciauskas (1973). In this model thermal effects of basin flow are modeled assuming homogeneous permeability in the basin, two-dimensional basin geometry with a rectangular cross section parallel to flow, and a simple circular function form for the water table (Figure IV.14). An approximation used in this analytical solution limits its use to flows close to those where forced convection starts to become effective, but this limitation can be used to determine the minimum permeability of a basin for forced convection to occur (e.g., Morgan et al., 1981). Examples of temperature perturbations within a basin and surface geothermal gradient perturbations along a basin, as predicted by this model, are shown in Figure IV.15.

The models shown in Figure IV.15 were calculated for conditions appropriate to groundwater flow in Rio Grande rift basins; a piezometric gradient based on the Rio Grande gradient indicates that forced convection must be occurring in these basins even with effective permeabilities down in the semipermeable range (Morgan et al., 1981). Significant temperature perturbations occur when vertical flow velocities exceed $3 - 10 \times 10^{-9}$ m/s (100-300 mm/year). The shallow temperature field associated with the discharge portion of one of these basins is shown in Figure IV.16. The models predict temperatures under conditions of steady fluid flow and steady-state heat flow and indicate that not only the shallow temperatures but also the basement temperatures at the discharge end of the basin are elevated. Flow systems need not discharge to the surface, but high subsurface temperatures will occur wherever an impermeable subsurface barrier causes upward groundwater movement. It is these conditions that make hydrothermal systems an attractive target for HDR exploration.

Flow conditions similar to those modeled for the Rio Grande rift may exist in some of the Midcontinent basins. An obvious candidate for such a

Figure IV.14.

Groundwater flow model for forced convection calculations. Basin is z_0 deep and L long. Water table has a cosine form along the basin length, as shown, with amplitude $2B$. Flow lines (with arrows) and lines of equal pressure in basin are shown. (From Domenico and Palciauskas, 1973)

Figure IV.15.

Examples of forced convection calculations showing isotherms in basins (lower) and perturbation of surface geothermal gradient (upper). (From Morgan et al., 1981)

system is the Denver Basin, a section through which is shown in Figure IV.17. This basin has a strong west to east gradient (approximately twice that of the Rio Grande), and the Dakota sandstone at a depth of up to approximately 2.5 km below the land surface provides a permeable conduit for deep basin flow. Flow through this system may be related to high heat flow in western Nebraska (Gosnold, 1980a), and thus it appears that hydrothermal systems are effective in producing thermal anomalies in the Midcontinent.

Figure IV.16.

Temperatures in $^{\circ}\text{C}$ at 90 m in the Las Alturas geothermal anomaly, New Mexico, an example of a discharge area for a basin groundwater flow system. (Adapted from Icerman and Lohse, 1983)

Geological exploration for hydrothermal systems in the Midcontinent should focus upon basin and aquifer geometry, geohydrology, and water table configuration, with the aim of locating the ascending limbs of the flow systems. Gravity, magnetic, and seismic studies should be used to aid the study of basin geometry, and reflection seismic studies can be used to provide details of aquifer geometry. Electrical studies can provide information on the water table configuration and locate subsurface impermeable structural high areas where upward flow is possible.

F. Combined Model Mechanisms for Hot Dry Rock Resources

Conceptual models of five types of HDR resources are presented above as isolated mechanisms for bringing high temperatures close to the surface. In practice it may be hoped that factors from two or more of the models may combine to give favorable conditions for an HDR resource. For example, fluid

Figure IV.17.

Configuration of land surface and top of the Dakota Group for the Denver Basin in a profile from southeast Wyoming to western Nebraska. (From Gosnold, 1980a)

flow temperatures in a basin will be higher if the upper sediments in the basin have a low thermal conductivity and/or the heat flow into the basin is high due to high crustal radiogenic heat production and/or high mantle heat flow. Thus, the models should not be considered in isolation of each other in planning an HDR exploration strategy, and the most favorable sites for HDR exploration will be those where more than one mechanism for producing shallow high temperatures exists.

V. HDR TEST SITE INVESTIGATIONS

A. Nebraska Geothermal Resources

1. Introduction

Nebraska is part of a large cratonic region in the north-central United States where Precambrian basement is covered by Phanerozoic strata. The relatively flat-lying sedimentary veneer in Nebraska ranges in thickness from about 300 m in the northeast to greater than 3000 m in the west (Figure V.1). The Precambrian surface is characterized by arches and basins that are considered to be dominantly epeirogenic in origin, and irregularities in this surface are interpreted to be paleotopographic features of various ages. The Precambrian rocks of Nebraska and their relation to exposed Precambrian rocks have been discussed by Lidiak (1972) and are also discussed in Section II.C. The Precambrian of Nebraska is composed mainly of a complex of igneous and metamorphic rocks, overlain in the southeast by basalts and associated sedimentary rocks related to the Midcontinent gravity high.

The geothermal regime of Nebraska was not known from any direct evidence prior to the commencement of a U.S. Department of Energy sponsored study in 1979. Although some interpretations of this geologic province indicated a typical Midcontinent heat flow in this region (Roy et al., 1972; Combs and Simmons, 1973; Lachenbruch and Sass, 1977), other studies suggested that anomalously high subsurface temperatures might exist in parts of the state (see Figure V.2 and Swanberg and Morgan, 1981; Schoon and McGregor, 1974). Direct thermal measurements, primarily by W.D. Gosnold and his colleagues, have confirmed the presence of significant thermal anomalies in Nebraska.

2. Subsurface Temperature and Heat Flow Data

Bottom Hole Temperature Data. A preliminary geothermal gradient map of Nebraska based on exploration well bottom hole temperature data is shown in Figure V.3 (this map is an update of the AAPG/USGS study, Section III.A). High gradients are indicated in western Nebraska, and other thermal data confirm this anomaly (see below). Most features of this map are not consistent with other data, however, and Gosnold and German (1982) and Gosnold (1982) suggest that the bottom hole temperature data are of very limited use for geothermal exploration in this state.

Heat Flow Data. Gosnold and German (1982) have reported heat flow values from 29 sites in Nebraska (Figure V.4). Heat flow for most of Nebraska ranges

Figure V.1.

Configuration of the Precambrian surface in Nebraska (from published state survey reports).

Figure V.2.

Geothermal gradients ($^{\circ}\text{C}/\text{km}$) for Nebraska based on AAPG/USGS (1976) temperature data and the limit of the western Nebraska groundwater silica geothermal anomaly. Gradient contour interval is $3.7^{\circ}\text{C}/\text{km}$. (Redrawn from Gosnold, 1980b)

from 40 to $60 \text{ mW}/\text{m}^2$, but values of 10 sites exceed $80 \text{ mW}/\text{m}^2$, and heat flow values as high as $120 \text{ mW}/\text{m}^2$ have been determined (Gosnold et al., 1981). All but one of the high heat flow sites fall into one of two areas: a broad band striking approximately NNW through the Nebraska panhandle and an area adjacent to the Nebraska/South Dakota boundary in north-central Nebraska (see Figure V.4). These two areas are thought to be of particular geothermal significance. The remaining isolated high heat flow site in southeastern Nebraska is a local thermal anomaly surrounded by lower values and is not considered to be of significance for geothermal exploration.

Figure V.3.
Temperature contours ($^{\circ}\text{C}$) at a depth of 1 km in Nebraska derived primarily from bottom hole temperature data (from Gosnold, 1980a).

Figure V.4.
Nebraska heat flow data. Approximate zones of high heat flow are shown by line pattern (from Gosnold and German, 1982).

Equilibrium Temperature Log Data. Equilibrium temperature measurements have been obtained from more than 100 wells in Nebraska in addition to those used for the heat flow determinations (Gosnold and German, 1982). Gradients derived from a portion of these data from western and north-central Nebraska are contoured in Figure V.5. High gradients are indicated in the two areas of high heat flow delineated in Figure V.4, and thus these gradient data are consistent with the presence of significant geothermal anomalies in western and north-central Nebraska.

Other Geophysical Data. Areas of anomalously high subsurface temperatures indicated by groundwater chemistry data overlap the high heat flow and high geothermal gradient areas (see Section III.B). High gradients are also indicated in western Nebraska by the AAPG/USGS (1976) bottom hole temperature map and the gradient map of Kron and Stix (1982) (see Section III.A), but the anomalous areas are poorly delineated by these data. Gravity and magnetic data were compiled as part of the present study for use in the investigation of the possible cause(s) of the Nebraska geothermal anomalies.

Three sets of gravity data were obtained for Nebraska: gravity data which cover the entire United States from the Defense Mapping Agency and from the U.S. Geological Survey, and data from the Nebraska State Survey which covers only the eastern portion of the state but in greater detail than the national coverage. These three data sets were merged and edited into a Nebraska data file containing 31,377 stations. Magnetic anomaly data were compiled for Nebraska from data collected during the Department of Energy's National Uranium Resource Evaluation-Aerial Radiometric Reconnaissance Survey (NURE-ARRS) program. New gravity anomaly maps were prepared for Nebraska, contoured at 2, 5, and 10 mgal intervals, but the detail shown by these maps is too great to be reproduced in this report. Similarly, a composite air-magnetic map was prepared, but cannot be reproduced here. Copies of these maps may be obtained from Professor C.L.V. Aiken (University of Texas at Dallas). Unfortunately the present project was curtailed before detailed analyses of these maps with reference to the Nebraska geothermal anomalies could be performed. Preliminary analyses (C.L.V. Aiken, unpublished study) indicate that a magnetic high and gravity low may be associated with the western Nebraska geothermal anomaly. Further analysis of these data are required, however, to define their relationship with the geothermal anomalies.

Figure V.5.
Geothermal gradients ($^{\circ}\text{C}/\text{km}$) in western and north-central Nebraska contoured from equilibrium temperature log data (from Gosnold and German, 1982).

3. Discussion of the Nebraska Geothermal Anomalies

High heat flow in western Nebraska appears to be related to the slow, updip flow of groundwater in deep aquifers from the eastern Denver-Julesburg Basin over the Chadron and Cambridge Arches (see Figure V.6). Gosnold and Eversoll (1981) and Gosnold and German (1982) have suggested that the lower sand units of the Cretaceous Dakota group are the most important deep aquifer for thermal convection in this

system. Structural contours and projected temperatures on the top of the Dakota group are shown in Figures V.7 and V.8, respectively (see also Figure IV.17). Locations of control points for the subsurface temperature

Figure V.6.
Principal structural features of Nebraska (from Gosnold and Eversoll, 1982).

Figure V.7.

Generalized structural contours of the top of the Cretaceous Dakota Group in Nebraska (based on Carlson and Reed, 1961; from Gosnold and Eversoll, 1982).

Figure V.8.

Projected temperature contours at the top of the Dakota Group in Nebraska (from Gosnold and Eversoll, 1982).

Figure V.9.

Location of control points for subsurface temperature projections contoured in Figure V.8 (from Gosnold and Eversoll, 1982).

projections are shown in Figure V.9, and simplified geologic cross sections showing the approximate configuration of the Dakota Group are shown in Figure V.10. Gosnold and German (1982) have calculated that updip water flow in the lower sand units of the Dakota Group with a flow rate of about 0.4 to 0.6 m/yr could produce the subsurface temperatures observed in western Nebraska. Thus, all evidence and modeling suggest that the primary cause of the western Nebraska geothermal anomaly is a regional hydrothermal system. Temperatures in this flow system are increased by a thick sequence of low conductivity Cretaceous shale which overlies the Dakota Group.

The high heat flow in north-central Nebraska is not easy to explain. Heat flow values from four sites in the area range from 100 to 140 mW/m², and temperature gradients logged in 14 deep water wells range from 50 to 90°C/km. Updip water flow is an unlikely explanation for the high heat flow because the sediments dip at very shallow angles in this area. A more likely explanation is that upward flow of groundwater along fractures enters aquifers such as the Dakota Group and flows laterally causing a region of high heat flow. A third possibility is that the Precambrian basement rocks contain high concentrations of potassium, uranium, and thorium and have high heat generation values.

Figure V.10.
Geologic cross sections showing approximate configuration of the Dakota Group in Nebraska (from Gosnold and Eversoll, 1982).

Equilibrium temperature data from a deep heat flow hole in this area that bottomed in Precambrian granite indicate that high heat flow continues to basement in this area (information supplied by D.D. Blackwell and W.D. Gosnold). There is no evidence of significant Cenozoic tectonic activity in this area, and likely conductivity contrasts cannot explain the geothermal

anomaly. Thus, unless groundwater is convecting heat in the basement, the north-central Nebraska geothermal anomaly appears to be caused at least in part by high radiogenic heat production in the basement rocks. Again, temperatures in this system are increased by low conductivity Cretaceous shales in the sedimentary section.

The results of the Nebraska geothermal studies are of major importance for Midcontinent geothermal exploration. Two major geothermal anomalies have been discovered in an area in which there are no surface geothermal manifestations or a priori reasons to suspect geothermal anomalies. The two anomalies appear to have different primary sources, but temperatures in both systems are increased by low conductivity overburden. These results are very encouraging for Midcontinent geothermal prospects.

4. Hot Dry Rock Geothermal Prospects

Although there appears to be abundant water in the sediments in the Nebraska geothermal anomalies, these hydrothermal resources are only at low to intermediate geothermal temperatures (less than 120°C). Temperatures are higher in the Precambrian basement underlying the sediments, but permeabilities are probably low, and hot dry rock technology will be required to extract heat from these units. The best areas for hot dry rock prospects are the areas of the geothermal anomalies as the geothermal gradient to the top of basement is greatest in these areas. It would still be advantageous to select basement areas in these anomalies with high radiogenic heat production, however, as basement geothermal gradients are likely to be highest in these rocks. Under favorable conditions, temperatures of around 200°C may be found at depths around 5 km in both Nebraska geothermal anomalies.

B. Mississippi Embayment Test Area

The Mississippi Embayment extends well into the craton of North America and intersects the continental margin at a high angle (Figure V.11). As suggested by Burke and Dewey (1973), it is a likely candidate for classification as a failed-arm rift. Ervin and McGinnis (1975) proposed a failed-arm model to explain the Mississippi Embayment. In their model, the Reelfoot rift formed in late Precambrian-early Paleozoic time as Proto-North America broke up to begin the Wilson cycle that formed the Appalachian-Ouachita system (Figure V.12). Their evidence consists of similarities in timing and geometry with the well-known southern Oklahoma aulacogen, orientation with respect to the continental margin, the presence of anomalously high seismic velocities in

Figure V.11.
Tectonic setting of the northern Mississippi Embayment.

the lower crust as mapped by McCamy and Meyer (1966), and a prominent regional gravity maxima which extends from the Gulf Coast area into southern Illinois. Ervin and McGinnis (1975) produced a gravity model which represents their synthesis of the data then available. They also suggested that this old aulacogen was reactivated in Mesozoic time to form the present-day embayment and cited the presence of Mesozoic intrusives as evidence.

This rift model has motivated many subsequent studies. Cordell (1977) removed the gravitational effects of the sedimentary rocks and better defined the regional anomaly. Seismic studies of deep structure (Mitchell and Hashim, 1977; Austin and Keller, 1982; Mooney et al., 1983) have corroborated the Ervin and McGinnis (1975) model (Figure V.13). More detailed gravity and magnetic studies (Hildenbrand et al., 1977) have mapped what appears to be a deep-seated graben in the upper Mississippi Embayment region that correlates

Figure V.12.

Compilation of Eocambrian rifts in the Midcontinent. RFR = Reelfoot rift; SOA = Southern Oklahoma aulacogen; DA = Delaware aulacogen; RCG = Rough Creek graben; RT = Rome trough; PA = Pennsylvanian aulacogen; SGH = Scranton gravity high (modified from Keller et al., 1983).

with the New Madrid seismic zone. This feature correlated with a low velocity layer detected by Mooney et al. (1983) and would appear to represent the buried central graben of the rift. Knowledge of this structure is important to efforts to assess earthquake hazards in the area because of the association of the graben with linear alignments of epicenters (Hildenbrand et al., 1977).

The model of Ervin and McGinnis (1975) seems consistent with all available data and has become widely accepted because it presents a unified explanation of a variety of geological and geophysical data. The work of Hildenbrand et al. (1977) is also very convincing, and our research group has devoted

Figure V.13.
Generalized crustal cross section of the Mississippi Embayment.

considerable effort to further mapping of the central graben. The major features identified by Hildenbrand et al. (1977) are two parallel, linear, northeast trends of gravity and magnetic anomalies which are interpreted as being the signatures of plutons which are associated with the boundaries of the central graben (Figure V.14). Braile et al. (1982a) noted that, with some disruption in southern Illinois, these same trends can be traced through the Wabash valley area into the vicinity of Indianapolis, Indiana (see Figure V.12). Soderberg and Keller (1981) have outlined an apparent deep-seated graben in western Kentucky. This east-west trending feature extends into southern Illinois and intersects the feature proposed by Braile et al. (1982b). The timing of the origination of this feature suggests it is probably related to the Reelfoot rift.

As pointed out by Braile et al. (1982b), a prominent trend of parallel gravity and magnetic highs extends along the Mississippi River from southern Illinois to the St. Louis, Missouri area. The combination of these trends produces a northern three-arm terminus for the Reelfoot rift which is very similar to that of the Benue trough in West Africa, a classic well-defined failed-arm rift associated with Mesozoic opening of the South Atlantic (Braile et al., 1982b).

Thus, the Mississippi Embayment provides an interesting place to test hot dry rock conceptual models in a Midcontinent area which is tectonically

Figure V.14.
Generalized fault map of the New Madrid region showing plutons, approximate rift boundaries, and epicenters of earthquakes detected from July 1974 to June 1977.

anomalous. An extensive data base of gravity and magnetic measurements was compiled, and portions of contour maps generated from these data are shown in Figures V.15 and V.16. These maps are very useful because they delineate the large plutonic masses in the area, the areas in which the sedimentary rocks attain their greatest thickness, and the major fault systems in the area and can be used in future studies to generate interpretive maps for geothermal exploration.

The most prominent local gravity and magnetic highs in the study area lie along a NE-SW trend and are thought to represent plutons emplaced along the major faults of the Eocambrian rift during a Mesozoic reactivation. Several of these plutons are exposed in the Little Rock, Arkansas area, and they are composed of nepheline syenite. Roy et al. (1982) reported a heat flow measurement in one of these bodies (ARK 5, Table V.1, Figure V.17) and determined its radioactive heat production. These results were not encouraging from a geothermal resource point of view. If all the intrusive bodies are similar in radiogenic heat producing isotopes to the one at Little Rock, they have little geothermal potential. However, there may be undetected felsic bodies, more enriched in heat producing isotopes, which were emplaced during the initial Eocambrian rifting. These bodies would probably produce small negative gravity anomalies, and although they are too deeply buried to be detected without much more study, they could cause increased gradients in the sedimentary cover and high basement temperatures. More analysis of the gravity data is required to search for this type of pluton before their HDR potential can be evaluated.

The thick sediments in the central rift graben and the major fault zones bounding this feature provide another potential HDR target. Hildenbrand et al. (1982) suggest that the total sedimentary thickness in the graben is approximately 6 km (20,000 ft). Also, the active seismicity in the area (e.g., Stauder, 1982) suggests the old faults in the area are not healed and could act as conduits for deeply circulating waters. Studies of the hot springs near Little Rock suggest circulation of groundwater as a mechanism for the elevation of the water temperature. The relatively low geothermal gradient in the area and the shallow circulation produce temperatures of about 60°C at these springs. Heat flow studies in the Mississippi Embayment have been limited, and some of the determinations have large uncertainties (Roy et al., 1982; Swanberg et al., 1982). However, the available data do suggest values

Figure V.15.
 Bouguer gravity map of the northern Mississippi Embayment. Contour interval 5 mgal. Reduction density 2.67 g/cm^3 . Sea level datum.

Figure V.16.

Aeromagnetic map (total intensity) of the northern Mississippi Embayment. Contour interval 100 gammas (nT).

TABLE V.1. Summary of Results for Heat-Flow Survey of Arkansas.

Hole No.	Location	Long./Lat.	Elevation (meters)	Depth Used (meters)	Gradient (°C/km)	Mean Conductivity (mcal/cm/s/°C)	Heat Flow ($\times 10^{-6}$ cal/cm ² /s)	Acknowledgments
ARK 1	Near Glenwood Montgomery Co.	93°50' 34°23'	317	40-310	21.1	5.17 (8)	1.09 ^a $\pm$ 0.01	F.C. Gale, Milchem, Inc.
				320-530	18.46	5.92 (20)	1.09 ^a $\pm$ 0.02	" "
ARK 2	Zinc, Boone Co. T19N R18W S20	92°55' 36°17'	274	480-610	20.3	13.91 (5)	2.82 $\pm$ 0.11	J.R. Reeves, Placer Amex, Inc.
ARK 3	Pindall, Searcy Co. T16N R18W S2	92°52' 36°03'	360	450-590	13.0	12.42 (5)	1.61 ^a $\pm$ 0.02	W. Moore Placer Amex, Inc.
ARK 4	Marianna, Lee Co. T1N R4E S9	90°42' 34°42'	61	30-320	23.8	13.17 (8)	3.13 $\pm$ 0.06	J. Edds, U.S.G.S.
ARK 5	Little Rock, Pulaski Co. T1N R12W S25	92°15' 34°42'	87	65-175	25.5	4.16 (14)	1.06 ^a $\pm$ 0.01	Granite Mountain Quarries
ARK 6	Horatio, Sevier Co. T10S R31W S1	94°15' 33°56'	117	40-90	19.6	10.20 (1)	2.0 $\pm$ 0.01	R. Gage, J. Gray

^aTerrain correction applied; numbers in brackets refer to the number of samples used.
(From Roy et al., 1982).

somewhat above the eastern United States average (Figures V.17 and V.18, Table V.1). Also, there is an interesting local anomaly in the New Madrid seismic zone area (Swanberg et al., 1982). The thick sediments, major fault zones which can act as conduits, and the possibility of slightly enhanced geothermal gradients suggest that a hydrothermal system in which groundwater is heated by deep circulation and then trapped by an impermeable barrier could form a potential HDR resource as well as a hydrothermal source. Available data do not give strong indications of an economic HDR resource in this area, but more data are required to complete this evaluation.

A final possibility is that there may be an active source of magmatic heat beneath the Mississippi Embayment, presumably another reactivation of the original Eocambrian rift. The seismic activity could suggest such a feature, but focal mechanisms (Herrmann and Canas, 1978) indicate strike-slip movements which can easily be interpreted as reactivation of old faults with maximum compressive stress oriented ENE. There is a regional magnetic low in the area which has been detected by the MAGSAT satellite. However, there are many

Physiographic provinces of Arkansas and location of heat-flow holes.

Pseudogeothermal gradients in the coastal plain of Arkansas.

Figure V.17.
Geothermal data from Arkansas (from Roy et al., 1982).

possible explanations for this anomaly which do not require increased heat and decreased Curie depths.

Location of heat-flow sites. Heat-flow values are given in parentheses in units of $\text{cal}/\text{cm}^2\text{-sec}$. The small solid circles denote earthquakes located by the Saint Louis University seismograph network between June 29, 1974, and December 31, 1977.

Temperature profiles for the heat-flow sites located in figure 1: 1, Fort Pillow, Tenn.; 2) Ridgely, Tenn.; 3, Wilson, Ark.; and 4) Caruthersville, Mo.

Locations of wells having abnormally high temperature gradients as calculated from bottom-hole temperatures. Solid circles, triangles, and open circles denote gradients that are $>30^\circ$, $24^\circ\text{--}30^\circ$, and $<24^\circ/\text{km}$, respectively. The dark shading delineates the region of maximum seismicity (Stauder and others, 1976), and the light shading indicates the approximate extent of the Mississippi Embayment.

Locations of the hottest wells at depths of 90, 120, 150, 180, 210, 240, 375 and 525 ± 16 m. Solid circles, triangles, and open circles denote the hottest wells, marginally hot wells, and normal wells, respectively. The dark shading delineates the region of maximum seismicity (Stauder and others, 1976), and the light shading indicates the approximate extent of the Mississippi Embayment.

Figure V.18.
Geothermal data from the northern Mississippi Embayment (from Swanberg et al., 1982).

In summary, deep circulation along old rift-related faults seems to provide the best potential for a hydrothermal resource and possibly a related HDR resource in the Mississippi Embayment area.

C. Geophysical Investigation of a Potential Hot Dry Rock Site in Southeastern Michigan

1. Introduction

Southeastern Michigan was chosen as a test area for hot dry rock (HDR) geothermal exploration because, although few thermal data were available for the area, many other geophysical studies were available. In this area intrusives with high concentrations of heat producing radioisotopes are potentially viable HDR sites, particularly where they are covered by a thermally insulating sedimentary rock blanket. Specific potential radiogenic heat sources include both felsic (e.g., Wolf River Batholith) and alkalic (e.g., Coldwell Complex) intrusives reported from this area. These felsic intrusives are commonly characterized by gravity minima of the order of a few tens of mgals and negative magnetic anomalies, although exceptions exist, particularly in the magnetic signature. In our evaluation of the HDR potential of southeastern Michigan, we have concentrated on the potential field signatures of intrusives with potentially high radiogenic heat production.

The Oakland-Macomb Counties, Michigan gravity minimum (Figure V.19), situated north of Detroit, may indicate the presence of a pluton of interest for HDR exploration. Preliminary geophysical and geological data suggest the gravity anomaly source is in some of the youngest basement rocks of the Midcontinent, within a basin that is covered by a thick blanket of low thermal conductivity sedimentary rocks and is situated favorably for utilization purposes near a major metropolitan area. The gravity low is bracketed within a regional positive anomaly to the north caused by metavolcanics and mafic gneisses (Figure V.20) and a positive regional gravity anomaly to the south associated with a late Precambrian rift. The source of the gravity low is probably a basement granitic intrusive associated with the Grenville Front. Further evidence indicating that the gravity anomaly may be caused by variations in intrabasement lithology is given by the magnetic data (Figure V.21) which characterize the study region as a negative magnetic anomaly discontinuously ringed by magnetic maxima. Basement lithologies for this region are shown in Figure V.20 and are generalized from geophysical and

Figure V.19.

Location of southeast Michigan HDR test area and associated Bouguer gravity anomaly data from Hinze et al. (1975) and heat flow values (in HFU) from Judge and Beck (1973). Study area is indicated by box.

Figure V.20.
Basement province map for the southeast Michigan test area (from Hinze et al., 1975). Study area is indicated by box.

Figure V.21.
Total magnetic intensity anomaly data for the southeast Michigan test area
(from Hinze et al., 1975). Study area is indicated by box.

limited drilling information. Canadian airborne radiometric surveys (Darnley et al., 1971) indicate that, to the north where these units are exposed, the regional contact of the Grenville Group with underlying granitic basement is the locus of a zone of radioelement enrichment over 500 km in length. By extrapolation, the same zone of radioelement enrichment may be expected along the same contact zone in the study area (Figure V.20). In the following sections, geophysical and geological evidence is synthesized and modeled to evaluate the hot dry rock potential of the Oakland-Macomb Counties site.

2. Description of Available Geophysical/Geological Data Set

A variety of data sets have been assembled to provide fundamental information regarding the HDR potential for the Oakland-Macomb Counties site. Sparse heat flow data (Figure V.19) are available for the study region (Combs and Simons, 1973; Judge and Beck, 1973). Geochemical data from the U.S. Geological Survey water quality file (WATSTORE) have also been assembled. In situ groundwater temperatures, temperatures derived from Si, Na-K-Ca, and Na-K-Ca-Mg geothermometry, and groundwater fluoride concentrations, all based upon 1° by 1° quadrilateral means, are shown in Figures V.22, V.23, V.24, V.25, and V.26, respectively. In general, these data do not indicate elevated temperatures for the Oakland-Macomb Counties site.

The gravity and magnetic anomaly data for the study region are derived from the geophysical study of Michigan's basement geology by Hinze et al. (1975). The gravity anomaly data shown in Figure V.19 are available at a 10-km grid interval, whereas the magnetic anomaly data of Figure V.21 are available on a 5-km flight line interval. The thickness of Phanerozoic shale units is evaluated in Figure V.27 from well logs from wells distributed across the State of Michigan and indicates that the total thickness of low thermal conductivity shale overlying the basement in the southeast Michigan region is around 400 m.

3. Data Analysis and Results

To assess the HDR potential of the Oakland-Macomb Counties site, it is desirable to evaluate the vertical and lateral geometric characteristics of the source of the gravity and magnetic anomaly minima in this area. To achieve this, the gravity anomaly data contoured in Figure V.19 for the region (42.25-43.25)°N, (83.75-82.50)°W were gridded at a 0.2° interval and contoured as shown in Figure V.28. Smooth-contouring the gravity anomaly data of Figure V.19 through the test site region resulted in the NW/SE trending regional

Figure V.22.

Temperature of groundwaters ($^{\circ}\text{C}$) averaged over $1^{\circ} \times 1^{\circ}$ areas for the southeast Michigan test area. Approximate location of study area is indicated by box.

Figure V.23.

Silica geotemperatures ($^{\circ}\text{C}$) (see text) averaged over $1^{\circ} \times 1^{\circ}$ areas for the southeast Michigan test area. Approximate location of study area is indicated by box.

gravity minimum gridded in Figure V.29. Subtracting this regional anomaly from the observed data of Figure V.28 yields the -30-mgal residual anomaly shown in Figure V.30. Sources that may be postulated for this residual gravity minimum include a thickening of Silurian or Devonian evaporites, a thickening of Phanerozoic or late Precambrian clastics, or an intrabasement lithological variation.

Nearly 4 km of borehole gravimetry in the Michigan Basin (Hinze et al., 1978) provide densities for estimating the gravity effects of sources, such as evaporites, within the Phanerozoic sedimentary rocks. Accordingly, the average density of the Phanerozoic section including evaporites is approximately 2.61 gm/cm^3 , whereas the average density for the Phanerozoic section without the evaporites is roughly 2.72 gm/cm^3 . The average density of the evaporites is approximately 2.16 gm/cm^3 so that according to the Bouguer slab effect formula the thickness of evaporites needed to produce the -30-mgal residual anomaly is in the range of 1.5 to 1.6 km. Such great thicknesses of

Figure V.24.
Na-K-Ca geotemperatures (°C) (see text) averaged over 1° x 1° areas for the southeast Michigan test area. Approximate location of study area is indicated by box.

Figure V.25.
Na-K-Ca-Mg geotemperatures (°C) (see text) averaged over 1° x 1° areas for the southeast Michigan test area. Approximate location of study area is indicated by box.

evaporites are not evident in the drill holes across the site, as shown in Figures V.31, V.32, and V.33.

Negative gravity anomalies may also be related to clastic wedges associated with the Precambrian rift as observed elsewhere along the Midcontinent Rift System (Craddock, 1972), but the occurrence of well-defined magnetic anomalies coincident with the gravity anomaly in the study area strongly suggests that the source is an intrabasement feature. To develop the residual magnetic anomaly, the observed anomaly data for the test site region in Figure V.21 were gridded and contoured as before as shown in Figure V.34 and smooth-contoured to develop the regional anomaly given in Figure V.35. This regional anomaly, in turn, was subtracted from the observed gridded data of Figure V.34 to obtain the residual magnetic anomaly for this region shown in Figure V.36. The residual magnetic anomaly for the test site region shows

Figure V.26.

Fluoride concentrations of groundwater (ppm) averaged over $1^\circ \times 1^\circ$ areas for the southeast Michigan test area. Approximate location of study area is indicated by box.

a well-defined magnetic minimum discontinuously ringed by a series of magnetic maxima which have been connected by the dashed contour.

These results suggest that the source of the test site anomalies may be due to an intrabasement lithologic variation. Further details concerning the vertical and lateral dimensions of the anomaly source are obtained from gravity modeling by Gauss-Legendre quadrature integration. The average density of the mafic basement rocks is assumed to be approximately 2.85 gm/cm^3 , whereas the mean density of felsic intrusives can be approximated by the value 2.62 gm/cm^3 . Hence, the density contrast used for the modeling is the preliminary value of -0.2 gm/cm^3 . This density contrast represents an upper limit on the possible density contrast for the the source which, in turn, puts a lower limit on the source volume.

The results of the gravity modeling are indicated in Figure V.37. The outline of the geometry of the modeled source is shown on the figure, and as indicated, the source decreases in area from a maximum at the basement surface, 2.4 km, to a minimum at a depth of 6 km. Magnetic bodies were modeled to simulate the ringed magnetic residual anomalies of Figure V.30. Calculated anomalies and the bodies from which they were calculated are shown in Figure V.38. Susceptibility contrasts of $+0.004$ and -0.003 emu/cm^3 were used at different depths in the bodies as indicated on the figure.

4. Conclusions

Several aspects of the Oakland-Macomb Counties study area in southeastern Michigan suggest that it may be a potential HDR target. Gravity and magnetic

Figure V.27.
Total thickness of all shale units in Michigan. Contour interval 50 m.

Figure V.28.

Observed Bouguer gravity anomaly data gridded and contoured for the southeast Michigan HDR test area. Contour interval is 5 mgals.

studies indicate a potential radiogenic heat source is in some of the youngest basement rocks of the Midcontinent in this area, within a basin which is covered by a thick blanket of low thermal conductivity sedimentary rocks and is situated favorably near a major metropolitan area for utilization purposes. A typical example of a felsic intrusive, such as that which may be responsible for the study area anomalies, is the 1500 m.y. granite pluton drilled over a vertical range of nearly 1 km in northern Illinois. Analyses of the core indicate an abnormally high U and Th content and a mean heat generation of roughly $40 \times 10^{-13} \text{ cal/cm}^3 \text{ sec}$. If we assume only one-half this mean value of the heat generation for the Oakland-Macomb Counties source and the thermal insulating characteristics of a 2.4-km-thick sedimentary section, temperatures ranging from 100° to 150°C are predicted at the base of the Phanerozoic sedimentary rocks in this area (Figure V.39). If future studies confirm that the source of the Oakland-Macomb Counties gravity and magnetic anomalies is a basement pluton with high radiogenic heat production, this site would be a promising HDR target.

Figure V.29.

Smooth-contoured regional gravity anomaly data gridded and contoured for the southeast Michigan HDR test area. Contour interval is 5 mgals.

Figure V.30.

Residual Bouguer gravity anomaly data gridded and contoured for the southeast Michigan HDR test area. Contour interval is 5 mgals.

Figure V.31.
Phanerozoic sequence encountered by drill holes across the southeast Michigan HDR test area.

Figure V.32.
Phanerozoic sequence encountered by drill holes across the southeast Michigan HDR test area.

Figure V.33.
Phanerozoic sequence encountered by drill holes across the southeast Michigan HDR test area.

Figure V.34.

Observed total magnetic intensity anomaly data gridded and contoured for the southeast Michigan HDR test area. Contour interval 100 nT.

Figure V.35.

Smooth-contoured regional total magnetic intensity anomaly data gridded and contoured for the southeast Michigan HDR test area. Contour interval 50 nT.

Figure V.36.

Residual total magnetic intensity anomaly data gridded and contoured for the southeast Michigan HDR test area. Contour interval 50 nT. The dashed line is drawn to outline the magnetic minimum surrounded by magnetic maxima.

Figure V.37.

Modeled Bouguer gravity residual anomaly data for the southeast Michigan HDR test area. Lateral geometries for the crustal prisms which constitute the model are superimposed, where shading corresponds to prism thickness as indicated.

Figure V.38.
 Modeled total magnetic intensity residual anomaly data for the southeast Michigan HDR test area. Lateral geometries for the crustal prisms which constitute the model are superimposed, where shading corresponds to prism thickness as indicated.

Figure V.39.
Effects of thermal blanketing and variations in crustal radiogenic heat production on the geotherm (from Costain and Glover, 1982).

VI. THE ILLINOIS DEEP DRILL HOLE PROJECT

A. Overview

In 1980, Commonwealth Edison completed the drilling of three engineering test holes in northwestern Illinois. The three holes (UPH-1, -2, and -3) were 0.6, 1.6, and 1.6 km deep, respectively, and the latter two penetrated nearly a vertical kilometer of Precambrian granitic basement. Continuous core from all three holes and access to UPH-3 were made available for scientific investigations through the Illinois Deep Hole Project. The drill hole and cores provided a unique scientific opportunity because (1) the basement geology of the region was poorly known, and it represents the basement of a vast region of the Midcontinent; (2) a large vertical depth of continuous core was available; and (3) one of the holes was available for testing, and corresponding core was available. The principal study of these holes was by integrated surface and in-hole experiments, plus detailed petrologic, geochemical, and physical property studies of the core samples. The combined results of these studies contribute significantly to our understanding of the regional geology of the poorly known Precambrian basement, the regional geophysics and tectonics, and the correlation between geophysical well logging and the true character of the rocks penetrated as evidenced in the drill core. All of these are important to Midcontinent HDR investigations, and in addition, significant thermal data were obtained from a previously little-known area.

Scientific studies related to the Illinois Deep Hole Project were published in 1983 in volume 88, issue B9, of the Journal of Geophysical Research. Three of the papers in this publication which were authored or co-authored by members of the Midcontinent HDR investigation team are presented as Appendices to this report. A discussion of the Illinois Deep Hole Project and the results to date of the related scientific investigations are presented by Coates et al. (1983) in Appendix VI.1.

Regionally, the drill holes lie within a northeast trending zone of positive magnetic anomalies that corresponds to a broad positive gravity trend (see Appendix VI.1). In detail, the holes correlate with a negative anomaly which is interpreted as originating from a granitic pluton (see Appendix VI.3) and a correlative magnetic minimum. Two main rock types have been identified in the core samples, a porphyritic biotite granite and a fine-grained granite

gneiss which occurs in the upper part of UPH-3 and is apparently older than the granite (Appendix VI.2). Geochronologic studies on UPH core samples show that these units are about 1460 m.y. old, part of a widespread 1450-1500 m.y. felsic, anorogenic terrane (Appendix VI.1).

The availability of the deep continuous core has permitted studies of various geochemical processes and in particular phenomena near the basement surface involving U, Th-Pb systematics, and associated U-series disequilibria (Stuckless et al., 1981; Rosholt, 1983; Doe et al., 1983). These studies have provided significant information on the leaching of uranium near the top of the Precambrian basement, the depth to which such processes have operated, and the degree to which present crustal water circulation is actively leaching uranium or its decay products from the rock. Fission track studies (Zimmerman, 1981; 1983) suggest that considerable sedimentary cover (0.5-3.0 km) has been eroded in the past 130 m.y., that the paleogeothermal gradient was greater, or some combination of both.

In-hole measurements consisted of hydrofracturing stress, rock-mass permeability, and in-hole geophysical logging. Stress measurements indicate conditions favorable for strike-slip faulting striking northeast below 1100 m (Haimson and Doe, 1983), and extremely low permeabilities are found in the basement rocks decreasing with depth from 10^{-4} darcy at 700 m to 10^{-8} darcy at 1600 m. An extensive suite of geophysical logs was run in the hole and compared with core measurements (Daniels et al., 1983). This comparison will be useful in evaluating geophysical logs in holes where core is unavailable. Furthermore, the in-hole measurements were useful in localizing fractured zones and in identifying an approximately 300-m zone of nonfractured granite at the bottom of UPH-3.

Of particular interest to the Midcontinent HDR project are investigations related to the thermal characterization of the rocks encountered in the Illinois deep holes. Thermal conductivity measurements on the granitic rocks of UPH-3 and temperature gradient observations, which have a mean value of $23.0^{\circ}\text{C}/\text{km}$ (Rahman, 1982), result in an average heat flow of $75 \text{ mW}/\text{m}^2$. This is higher than the normal heat flow in the eastern U.S., probably as a result of the high total radioactivity (K, U, Th) of the granitic rocks (Rahman and Roy, 1981; Rosholt, 1983). Rahman (1982) determined a mean heat generation of $16 \mu\text{W}/\text{m}^3$ in the basement rocks of UPH-2 and -3. In UPH-3, the heat generation decreases with depth, but in UPH-2 the values are relatively low to the 1150-m

level, where high values occur to the bottom of the hole. The measured heat flow is considerably less than that predicted for the eastern U.S. considering the relatively high heat generating capacity of the granitic rocks (Birch et al., 1968; Roy et al., 1972). This discrepancy is attributed by Rahman (1982) to the limited vertical extent of the granite which is confirmed by interpretation of the gravity data (Aiken et al., 1983; Appendix VI.3).

B. Geothermal Analysis of the Illinois Deep Drill Holes

1. Introduction

The purpose of this study was to measure the thermal conductivity and radioactivity of core samples from the Illinois deep drill holes near Winslow, Illinois and apply the results to current geothermal and crustal models. Additional information used in the study included temperature depth and electrical logs provided by the U.S.G.S. in Denver [Olhoeft et al., 1981 and by J.H. Scott in 1981 and D.D. Blackwell in 1982 and gravity interpretations by Aiken et al. (1983)].

The Illinois deep drill project began as a result of a plan by the Consolidated Edison Company of Illinois to maximize energy efficiency in their nuclear power plants. The proposal called for the excavation of a large volume of rock ($2.6 \text{ km}^2 \times 0.03 \text{ km}$) within the granitic basement along the Illinois-Wisconsin border. This chamber would be used for temporary water storage. Excess energy from the nuclear power facilities would be used to pump water from this chamber to the surface. During peak demand times, additional energy would be derived from turbines driven by water returning to the chamber.

As a part of the study, three test holes were drilled. The first hole, UPH-1, which penetrated less than 30 m of the Precambrian basement, was drilled to a depth of 636 m. Three kilometers to the north, drill holes UPH-2 and UPH-3 ($42^{\circ}30'N \times 89^{\circ}46'W$) were completed. UPH-2 was drilled to a total depth of 1659 m and encountered the basement at 665 m. UPH-3 reached a final depth of 1600 m and encountered the basement at 664 m. Continuous solid core was obtained from the basement in all three drill holes. UPH-1 and UPH-2 were cemented and sealed shortly after the completion of the drilling operations. The core samples from all holes and the drill hole UPH-3 were then made available to the scientific community through a steering group established in February 1980. Geological, geophysical, and geochemical studies were

performed on the core samples. All well logs were made in UPH-3, the location of which is shown in Figure VI.1.

2. Measurement Procedures

Thermal Conductivity. Thermal conductivity measurements on core samples were made with a standard steady-state divided-bar apparatus; details of the construction, calibration, and use of the apparatus are given by Johnson (1974) and Taylor (1981). The thermal conductivity values of natural quartz (heat flow perpendicular to the optic axis) and silica glass as given by

Figure VI.1.
Location of drill hole UPH-3.

Ratcliffe (1959) were used for calibration. Solid core samples approximately 3.81 cm thick and 4.77 cm in diameter were used for thermal conductivity measurements. First, samples were measured dry, then measured wet after being saturated with boiled, distilled water by evacuation of the samples for 48 hours, followed by pressurization under water at 2,500 psi (172 bars) for an hour or more. Sample ends were coated with KY Jelly (manufactured by Johnson and Johnson) to reduce contact resistance, and conductivities were measured under 14 bars axial pressure after a wait of 20 to 30 minutes for thermal equilibrium to be established.

Uranium, Thorium, and Potassium Concentrations and Heat Generation.

Uranium, thorium, and potassium concentrations in the samples were measured using gamma-ray spectrometry techniques using two 15-cm-diameter by 10-cm-thick NaI (TI) crystal detectors as described by Rye and Roy (1978) and Smith (1972). Heat generation values were calculated for each sample from uranium, thorium, and potassium contents using the heat generation constants given by Birch (1954) and assuming a bulk density of 2.67 g/cm^3 for the samples.

Temperature Log. Temperature-depth data for UPH-3 were provided for this study by D.D. Blackwell of Southern Methodist University. These data were recorded to a temperature precision of 0.001°C at 1-m intervals in the drill hole and were collected 18 months after drilling, so thermal transients associated with drilling should be minimal in these data. Temperatures at 50-m intervals in UPH-3 are shown in Figure VI.2.

3. Results and Interpretation

A mean gradient of 23.0°C/km was calculated in the granite section of UPH-3 from the temperature-depth data. Significant changes in gradient become obvious when the gradient is plotted as function of depth (Figure VI.3). For the first 400 m of granite, the gradient is between 22 and 23°C/km . The gradient then increases over the next 100 m to 23.5°C/km and remains between 23 and 24°C/km for 200 m. The gradient within the granite then decreases slightly, then increases toward the bottom of the hole. These changes are too abrupt to be attributable to thermal transients related to surface climatic effects and are probably due to small differences in thermal conductivity related to differential crystallization within the granitic basement (Daniels et al., 1983).

Figure IV.2.
Temperature versus depth plot for UPH-3.

Thirty-nine solid core samples from various depths of UPH-1, UPH-2, and UPH-3 were measured in the divided bar apparatus with an average sample temperature of 25°C. Virtually all measured conductivity values were very close to the mean of 3.31 ± 0.04 W/m·K. A small temperature correction (Birch and Clark, 1940) was applied to all values to match the conductivities to their in situ conditions, and the mean of the revised conductivities is 3.27 ± 0.05 W/m·K. Individual sample conductivities are listed in Table VI.1. The calculated mean conductivity at 25°C of 3.31 W/m·K compares well with Eckstein's result of 3.33 W/m·K reported at the A.G.U. convention of May 1981, in Baltimore (although an erroneous value is given in their abstract, Eckstein et al., 1981).

Heat flow values were calculated as the product of geothermal gradient and thermal conductivity; a heat flow of 75 mW/m^2 was calculated for UPH-3 using temperature corrected conductivities. This value is significantly higher than the average of 46 mW/m^2 for the eastern U.S. (Roy et al., 1968a, b).

Figure VI.3.
Temperature gradient versus depth for UPH-3.

Forty-four crushed rock samples from various depths from each drill hole were analyzed for uranium, thorium, and potassium concentrations, and heat generation values were then calculated for each sample. These data are listed in Table VI.2. The overall concentration averages were as follows: 34 ± 3 ppm uranium, 92 ± 6 ppm thorium, and $5.03 \pm 0.09\%$ potassium. Concentration versus depth plots were made for each element, and averages were also computed for portions of each drill hole. These plots and averages for UPH-2 and UPH-3 are shown in Figures VI.4, VI.5, and VI.6.

In UPH-2, both uranium and thorium values are fairly low until about 1,150 m where there is a sharp increase in values. Concentrations below 1,150 m generally are high and show little, if any, trend with depth. In UPH-3, uranium and thorium concentrations generally decrease with increasing depth to a depth of 1,250 m; one sample at 1,070 m has concentrations anomalously higher than the general trend. This sample was reanalyzed at the University of California at Berkeley, Lawrence Radiation Laboratory (information supplied by A.R. Smith, 1981) with almost identical results, and the sample appears to be anomalous. From 1,250 m, uranium and thorium concentrations generally

TABLE VI.1. Conductivity Values (W/m·K)

Drill Hole	Sample Depth (meters)	Conductivity (uncorrected)	Conductivity (temperature corrected)
UPH-1	611	3.27	3.30
UPH-1	625	3.43	3.45
UPH-1	630	3.06	3.08
UPH-2	873	3.28	3.28
UPH-2	905	3.24	3.22
UPH-2	973	3.17	3.14
UPH-2	1033	3.84	3.80
UPH-2	1121	3.61	3.57
UPH-2	1152	2.62	2.58
UPH-2	1219	3.59	3.54
UPH-2	1281	3.68	3.62
UPH-2	1325	3.56	3.50
UPH-2	1343	3.25	3.18
UPH-2	1401	3.58	3.51
UPH-2	1455	2.96	2.88
UPH-2	1576	3.53	3.44
UPH-2	1649	3.29	3.19
UPH-3	673	3.53	3.55
UPH-3	692	3.37	3.38
UPH-3	698	3.35	3.37
UPH-3	731	3.43	3.43
UPH-3	738	3.46	3.46
UPH-3	800	3.35	3.35
UPH-3	851	3.11	3.11
UPH-3	905	3.16	3.15
UPH-3	960	3.53	3.51
UPH-3	1016	3.03	3.00
UPH-3	1071	2.89	2.86
UPH-3	1128	2.96	2.94
UPH-3	1183	2.99	2.94
UPH-3	1237	3.02	3.00
UPH-3	1293	3.33	3.27
UPH-3	1351	3.23	3.16
UPH-3	1405	3.20	3.13
UPH-3	1429	3.43	3.36
UPH-3	1460	3.68	3.60
UPH-3	1487	3.37	3.29
UPH-3	1515	3.18	3.09
UPH-3	1541	3.07	2.99
Averages:		3.31 ± 0.04	3.27 ± 0.05

TABLE VI.2. Gamma-Ray Spectrometry Results

Drill Hole		Sample Depth (meters)	Uranium (ppm)	Thorium (ppm)	Potassium (%)	Heat Generation ($\mu\text{W}/\text{m}^3$)
UPH-1		611	12.8	40.3	5.7	7.1
UPH-1		636	21.8	67.0	4.3	10.1
UPH-2	[1]	665	13.4	45.5	5.6	7.3
UPH-2	[2]		21.9	38.1	5.7	9.0
UPH-2		726	14.0	38.3	5.1	6.9
UPH-2		873	10.0	35.6	6.4	5.6
UPH-2		905	13.5	36.8	4.5	6.6
UPH-2		973	13.4	55.8	3.9	7.9
UPH-2		1033	12.0	44.9	3.1	6.6
UPH-2		1121	17.2	40.5	5.6	7.9
UPH-2		1151	20.0	28.9	4.5	7.7
UPH-2		1219	64.4	163.8	5.1	29.0
UPH-2		1281	55.5	149.6	5.5	25.7
UPH-2		1325	67.3	119.2	4.4	26.5
UPH-2		1343	48.6	132.5	5.5	22.6
UPH-2		1401	38.9	105.7	5.0	20.3
UPH-2	[1]	1455	14.6	28.4	4.9	6.3
UPH-2	[2]		14.8	46.9	5.1	7.7
UPH-2		1519	36.9	94.1	5.5	17.4
UPH-2		1576	45.6	125.0	4.7	21.3
UPH-2		1649	27.8	88.4	5.4	13.8
UPH-3	[1]	673	50.0	129.6	5.2	22.8
UPH-3	[2]		60.7	144.3	6.3	26.7
UPH-3		692	49.4	125.0	4.7	22.2
UPH-3		698	53.8	134.7	5.1	24.1
UPH-3		717	53.0	147.3	5.0	24.8
UPH-3		731	52.5	149.3	5.1	24.8
UPH-3		738	47.8	112.6	4.6	21.0
UPH-3		800	46.6	131.4	5.2	22.0
UPH-3		851	51.9	129.4	5.1	23.2
UPH-3		905	47.4	127.5	5.0	21.9
UPH-3		960	38.2	112.2	5.7	18.5
UPH-3		1016	33.5	113.2	5.0	17.3
UPH-3		1071	66.7	145.8	4.3	28.2
UPH-3		1128	28.9	82.8	4.8	13.9
UPH-3		1183	22.5	60.7	4.5	10.6
UPH-3		1237	17.6	54.5	5.1	9.0
UPH-3		1293	19.3	55.1	5.4	9.5
UPH-3		1351	26.5	81.1	5.1	13.2
UPH-3		1405	36.3	100.6	5.4	17.1
UPH-3		1429	31.7	88.0	5.0	15.0
UPH-3		1487	39.8	108.3	4.7	18.5
UPH-3		1515	27.1	85.5	4.8	13.4
UPH-3		1541	26.2	84.0	4.9	13.3
Averages:			34 $\pm$ 3	92 $\pm$ 6	5.0 $\pm$ 0.1	16 $\pm$ 2

Figure VI.4.
Uranium concentration versus depth for UPH-2 and UPH-3.

increase in value almost down to the bottom of the hole. Potassium concentrations are relatively uniform in both UPH-2 and UPH-3.

Heat generation versus depth plots for both UPH-2 and UPH-3 are shown in Figure VI.7. Predictably, both curves follow the same patterns as the uranium and thorium curves. In UPH-2, the heat generation values are relatively uniform and low down to 1,150 m, below which higher values predominate to the bottom of the hole. In UPH-3, values start high but generally decrease with depth. A U.S. Geological Survey gamma-ray log of UPH-3 (Figure VI.8) has the same basic characteristics as the heat generation versus depth plot for UPH-3 (Figure VI.7).

Examination of the solid cores from UPH-2 and UPH-3 revealed that grain size varied slightly with the amount of heat production for a given range of depths. In UPH-2, the high heat production zone coincides with samples of porphyritic granite containing large feldspar crystals in a fine-grained groundmass. Finer-grained cores predominated in the top 300 m of the UPH-3 basement where the highest heat production values are found. A general decrease in heat generation with depth in UPH-3 is probably due to a decrease

Figure VI.5.
Thorium concentration versus depth for UPH-2 and UPH-3.

Figure VI.6.
Potassium concentration versus depth for UPH-2 and UPH-3.

Figure VI.7.
Heat generation versus depth for UPH-2 and UPH-3.

Figure VI.8.
Gamma-ray log of UPH-3.

in the felsic mineral component of the granite with depth due to fractionation during crystallization (Daniels et al., 1983). Excluding the anomalous sample from 1,070 m, generally low values between 1100 and 1350 m coincide with a major fracture zone in the granite; this fracture zone is clearly identified by the physical property logs from UPH-3 and interpreted to occur along a zone of weakness resulting from an interruption of the cooling history of the intrusion (Daniels et al., 1983).

The overall heat generation average in the Illinois drill holes is $16 \mu\text{W}/\text{m}^3$ which is extraordinarily high. Normally, granites average only 3-4 ppm uranium, 10-15 ppm thorium, and 3-4% potassium, equivalent to a heat generation range of 2 to $4 \mu\text{W}/\text{m}^3$ (Adams, 1964).

High heat generation in the basement granite implies that heat flow should decrease significantly with depth in the granite. We have calculated that, for a heat generation value of $16 \mu\text{W}/\text{m}^3$, the heat flow should decrease by about $15 \text{ mW}/\text{m}^2$ between 670 and 1600 m in UPH-3, assuming vertical heat flow at this location. From this decrease in heat flow, a decrease in gradient of about $4.6^\circ\text{C}/\text{km}$ over this depth range in UPH-3 is predicted, in contrast to the increase of about $1^\circ\text{C}/\text{km}$ that was measured in the hole. Most of the observed gradient increase in UPH-3 can be explained as conductivity related, assuming a temperature coefficient of conductivity of $-0.2\%/^\circ\text{C}$. It is still necessary to explain why the decrease in gradient with depth predicted from heat generation considerations is not observed, however.

We suggest that the most probable explanation of why no gradient decrease was observed is that the temperature log includes a transient component of heat flow caused by changes in ground surface temperature during Quaternary climatic fluctuations. Using the climatic model of Emiliani (1978) and the computation technique described by Jessop (1971) and assuming a uniform thermal diffusivity of $1.5 \text{ mm}^2/\text{s}$, we have calculated a variation in gradient of about $4.5^\circ\text{C}/\text{km}$ between 670 and 1600 m depth due to recent climatic fluctuations. This calculation is subject to many uncertainties as many of the parameters used were poorly defined, but it suggests that the decrease in gradient predicted from heat generation considerations may be approximately balanced by an increase in gradient resulting from climatic fluctuations.

4. Discussion and Conclusions

The heat flow of $75 \text{ mW}/\text{m}^2$ and heat generation of $16 \mu\text{W}/\text{m}^3$ determined for the Illinois drill holes are not typical of values determined for the eastern

United States. Birch et al. (1968) and Roy et al. (1968a, b) have shown that most heat flow and heat generation data from the eastern U.S. follow a simple linear relationship, but the Illinois drill hole data do not plot on this line (Figure VI.9; see also Section IV.A).

Certain factors should be noted before interpreting this discrepancy. First, the eastern U.S. heat flow heat production relationship was established using shallower holes and lower heat generation values than those in the Illinois drill holes. Next, the average heat generation value of the upper 550 m of the basement granite penetrated by UPH-2, assuming a heat flow of 75 mW/m^2 , plots near the eastern U.S. line. The abrupt change in values seen throughout the lower part of UPH-2 basement could be the result of uranium and thorium enrichment in a late phase of the pluton cooling (Daniels et al., 1983) or a fault repetition of the upper part of UPH-3, although there is no direct evidence for faulting in the core samples. The differences in grain size seen in the UPH-2 basement granite samples support either of these possibilities. The heat generation values of UPH-3 generally decrease with depth, but the mean heat generation value for this hole is much higher than that predicted by the eastern U.S. heat flow/heat generation relationship (Figure VI.9).

Figure VI.9.

Heat flow heat generation relationship for the eastern U.S. with mean heat flow heat generation values for UPH-2 and UPH-3.

Heat flow lower than that expected for an abnormally high heat generation value can be explained by assuming a smaller depth extent for the pluton than the 7.5-km value implied by the eastern U.S. relationship. A pluton only 2.6 km thick with an average heat generation of $16 \mu\text{W}/\text{m}^3$ would yield a heat flow near $75 \text{ mW}/\text{m}^2$ if we assume a reduced heat flow of $33.5 \text{ mW}/\text{m}^2$, which is the value deduced from other studies (Birch et al., 1968). Data from the deep holes suggest, therefore, that the holes penetrate a thinner than expected pluton, the flanking edge of a thicker pluton, or a nonsystematic distribution of radioelements in the pluton.

Gravity data (Aiken et al., 1983) show a significant gravity low to the north of the deep hole sites (see Appendix VI.3). The computer modeling of these authors indicates that a granite pluton about 5 km thick underlies the deep holes and thickens to the north. If the heat generation of the pluton is about $5 \mu\text{W}/\text{m}^3$ for the 4 km below UPH-3, without a corresponding change in density, the heat flow/heat production data would be consistent with this model. A rapid decrease in heat generation with depth must occur beneath the Illinois deep drill holes to be consistent with the measured surface heat flow. Without additional data, however, all models of this decrease are speculative.

The upper basement of UPH-2 yields relatively uniform heat generation values which are consistent with the eastern U.S. heat flow heat generation relationship and the Birch et al. (1968) simple step model for the depth distribution of the heat producing isotopes (see Section IV.A). In contrast, heat generation values in UPH-3 generally decrease with depth and are more consistent with the Lachenbruch (1968) exponential radioelement distribution model. Available data indicate that while the granitic basement is not uniform with respect to uranium and thorium distribution, some important factors are apparent. UPH-2 appears to penetrate a highly radioactive, fractured pluton which possibly experienced interruptions in cooling and crystallization. The decrease of UPH-3 heat generation values with respect to depth occurs at a greater rate than has been observed in other deep holes (Lachenbruch and Bunker, 1971). A fractured, highly radioactive pluton overlying either another granitic pluton or the same pluton at a more depleted level at a fault contact or contacts could account for this overall decrease.

Based on these data, the following conclusions can be made. The deep hole penetrates a petrologically complex granitic pluton which is fragmented

and intruded. The temperature gradient is expected to decrease with depth in the pluton. If this pluton is equivalent to the single structure identified in the gravity models, it may be expected to thicken to the north. If heat generation values remain constant, higher heat flow and higher temperature gradients may be expected over the thicker pluton to the north. Further exploration is required to evaluate the HDR prospects of sites to the north.

APPENDIX VI.1.

INTRODUCTION TO THE ILLINOIS DEEP HOLE PROJECT

Reprinted from Coates et al. (1983)

INTRODUCTION TO THE ILLINOIS DEEP HOLE PROJECT

M. S. Coates¹, B. C. Haimson², W. J. Hinze³, and W. R. Van Schmus⁴

Abstract. Commonwealth Edison recently drilled three engineering test holes in northwestern Illinois. The three holes (UPH 1, 2, and 3) were 0.6, 1.6, and 1.6 km deep, respectively, and the latter two penetrated nearly 1 km of Precambrian granitic basement. In 1980, continuous core from all three holes and access to UPH 3 were made available for scientific investigations through the Illinois Deep Hole Project. The drill hole and cores provided an excellent opportunity for "piggyback" scientific investigations because (1) the basement geology of the region was poorly known, (2) a large vertical depth of continuous core was available, and (3) one of the holes was available for testing where corresponding core is available. The principal study of these holes was an integrated investigation of surface and in-hole experiments, plus detailed petrologic, geochemical, and physical property studies of the core samples. The combined results of these studies contributed significantly to our understanding of the regional geology of the Precambrian basement, the regional geophysics and tectonics, and the correlation between geophysical well logging and the true character of the rocks penetrated. Furthermore, this project has shown that substantial scientific benefit can be obtained from continental drilling "holes of opportunity" if members of the scientific community are able to respond in a timely and coordinated fashion.

Introduction

Commonwealth Edison, a Chicago-based electric utility, recently drilled three deep drill holes near Winslow, Stephenson County, northern Illinois, as part of a feasibility study for an underground pumped hydroelectric (UPH) storage plant [Coates, 1980, 1982]. All three holes penetrated Precambrian granite, with UPH 2 and UPH 3 each extending about 1 km into the basement (Table 1). The first two holes were back filled as soon as drilling and basic logging were completed. However, in January 1980, the third hole came to the attention of one of us (B. C. Haimson), who determined that Commonwealth Edison was receptive in principle to the idea of add-on (piggyback) experiments utilizing the hole, core, and logs. Consequently, this was an excellent "hole of opportunity," provided that organization and funding could be mobilized to take advantage of the drilling rig, while it was still on site,

and that arrangements could be made to keep the hole open after drilling was completed. In view of this urgency, the Continental Scientific Drilling Committee (CSDC) and the U.S. Geodynamics Committee promptly convened a task group meeting in February 1980 to develop recommendations regarding significant scientific experiments that could be carried out in conjunction with this drill hole.

As a result, the Illinois Deep Hole Project Steering Group was established to develop and coordinate plans for utilizing UPH 3 for in-hole experiments and for studying cores and logs of all three holes. The steering group (the active members of which are the authors of this paper) immediately prepared a document describing the project and a request for letters of intent from potential participants. These were widely disseminated throughout the scientific community, and more than two dozen scientists expressed serious intent to take part in the study. Based on these responses, the steering group prepared a scientific plan for the Illinois Deep Hole Project, which was endorsed by CSDC at its meeting of April 10-12, 1980, and by Commonwealth Edison. The scientific results are briefly summarized in this paper and in more detail in the articles published in this special volume and elsewhere.

During the 6-week period between the endorsement of the scientific plan and completion of UPH 3, three different institutions came to our assistance to ensure that the most important of the in-hole tests would be conducted on time. The National Science Foundation expediently funded the stress and permeability testing, Commonwealth Edison contributed by agreeing to pay for the drilling rig costs, and the U.S. Geological Survey (USGS) used its in-house research budget to undertake a full range of downhole logging. In addition, the Madison, Wisconsin, office of the USGS Water Resources Division provided funding to buy the casing in UPH 3 so the hole could remain open, and the USGS also provided funding to W. R. Van Schmus to defray the costs of collecting, curating, and distributing core samples.

Initial sampling of the available crystalline rock core was carried out in May 1980. These samples were taken at approximately 70-m intervals from portions of core that represented at least 1 m of homogeneous lithology without fractures, veins, xenoliths, dikes, etc. The samples, which generally were about 30 cm long, were referenced to the adjacent pieces of core. These samples were inventoried, cataloged, processed, and distributed to the scientific community from the University of Kansas under the direction of W. R. Van Schmus (limited material is still available, and interested parties should contact him). For distribution, samples were labeled according to drill hole identity, approximate depth (in feet), and sub-sample (e.g., UPH-1/4610G). Several specialized studies required specific types of samples that could not

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TABLE 1. Commonwealth Edison Drill Holes, Stephenson County, Illinois

Hole No.	Location	Total Depth, m	Depth to Precambrian, m	Hole Condition
UPH 1	42°30'17"N, 89°51'07"W	639	608	back filled
UPH 2	42°26'34"N, 89°52'00"W	1659	664	back filled
UPH 3	42°26'13"N, 89°51'27"W	1607	664	open

be removed until detailed logging of the core was completed. Thus cores from the Paleozoic sedimentary rock section and special features in the Precambrian crystalline rocks were not sampled until September 1980. One lesson learned from this is that effective cataloging, preparation, and distribution of core material is time consuming and costly. This is an aspect of similar, future projects that will require careful planning, coordination, and funding.

Summary of Results

Regional Geology.

The Commonwealth Edison deep drill holes are located near the axis of the Wisconsin Arch, which extends south from central Wisconsin into northern Illinois as a basement structural high (Figures 1 and 2). The Precambrian geologic history of the region is only approximately known because of the paucity of basement samples and the difficulty of extrapolating geologic information from the nearest basement outcrops, a few hundred kilometers away to the north in central Wisconsin and to the south in the St. Francois Mountains of Missouri. Petrologic and geochronologic studies [Lidiak et al., 1966; Bickford et al., 1981a; Van Schmus and Bickford, 1981; Hoppe et al., this issue; Denison et al., 1983] indicate that the basement rocks of northern Illinois are part of a vast terrane of largely felsic igneous rocks which extends over the central midcontinent from southern Wisconsin to Kentucky to New Mexico. The ages of the rocks range from 1760 m.y. in southern Wisconsin, to 1450-1500 m.y. in Missouri, Illinois, and Indiana, to 1350-1400 m.y. in southern Kansas and Oklahoma. The predominant rock types are granite and rhyolite. Chemical and petrographic data suggest that the terrane may be largely anorogenic [Bickford et al., 1981b; Cullers et al., 1981; Bickford and Van Schmus, 1981; Schulz, 1981; Anderson, 1982; Lidiak and Denison, this issue] and may overlie an older Proterozoic basement [Nelson and DePaolo, 1982]. However, there are few data to confirm the presence of older basement, its age, or its character. Geochronologic studies on UPH core samples show that the basement at these drill sites is about 1460 m.y. old and part of the 1450-1500 m.y. terrane [Hoppe et al., this issue; Peterman et al., 1983]. Wide spread degradation of Rb-Sr and K-Ar age systems is shown by young apparent ages throughout the region. This is generally thought to be due to ancient exchange with meteoric groundwater, as shown for the St. Francois Mountains terrane by Wenner and Taylor [1976].

Oxygen isotope data for the UPH samples also show evidence of exchange with meteoric groundwater [Sheih, this issue; Kombrink and Clayton, this issue] and further support this contention.

Despite the relatively similar felsic basement terrane suggested by subsurface samples, the area contains considerable magnetic and gravity character (Figures 3a and 3b). The anomalies present are probably derived not only from magnetic polarization and mineralogic density variations in the shallow basement rocks but also from lithologic and structural variations within basement rocks that presumably underlie them. The UPH drill holes lie in an area not previously covered by published aeromagnetic data (Figure 3a), but aeromagnetic coverage of northwestern Illinois (including the sites of the UPH drill holes) has recently been released by the U.S. Geological Survey [1982]. The UPH drill holes lie on the flank of a minimum whose axis strikes roughly north-south and located approximately 10 km to the east. This minimum correlates with a negative gravity anomaly which is interpreted as originating from a granitic pluton [Aiken et al., this issue]. Regionally, the drill holes are located within a northeast trending zone of positive magnetic anomalies (Figure 3a) that do not show marked

Fig. 1. Approximate topography of top of Precambrian basement for region around Commonwealth Edison drill holes. Numbers give elevation in kilometers relative to sea level [after Bayley and Muehlberger, 1968]. Circle in NW Illinois gives approximate location of drill holes (Figure 2).

Fig. 2. Basement topography in vicinity of Commonwealth Edison drill holes based on vibroseis survey [Coates, 1981]. Primary contouring was in feet; numbers in parentheses give approximate elevation in meters. See Figure 1 for general location.

internal linear trends [Chandler et al., 1982]. In this regional context, the Illinois drill holes lie on the southwestern margin of a particularly intense, positive magnetic anomaly that covers a broad area of southeast Wisconsin. Yaghubpur [1979] interpreted the extension of this zone of anomalies in southeast Iowa as a series of granitic plutons. A recently obtained basement sample from one of them in Jackson County, Iowa, has not only confirmed that interpretation, but also has shown that it has the same age and general lithology as the UPH samples [Hoppe et al., this issue].

Comparison of the gravity anomaly map (Figure 3b) with the magnetic anomaly map (Figure 3a) shows a broad positive gravity trend that corresponds approximately in location to the belt of magnetic anomalies mentioned above. In addition, the northwestern boundary of the magnetic anomaly belt roughly coincides with a marked change in character of the gravity anomaly patterns, as shown by the heavy dashed lines in Figures 3a and 3b and in more detailed maps presented by Aiken et al. [this issue]. Southeast of this line the gravity anomalies are broader and have lower horizontal gradients, suggesting that the sources of the anomalies are considerably deeper. Thus, based on geophysical data, it appears that the UPH drill holes have penetrated one of several plutons that lie along the northwestern boundary of the 1450-1500 m.y. old rhyolite-granite terrane of the central United States. In this context, then, extensive

petrographic and chemical data obtained from core samples from these sites [Lidiak and Denison, this issue; Peterman et al., 1983; Schulz, 1981] will prove invaluable in helping to define the origin of this terrane.

The Phanerozoic sedimentary rocks that unconformably overlie the Precambrian basement reflect the slow regional epeirogenic movements which were initiated in late Cambrian or early Ordovician time and dominated the tectonism of the area in the Paleozoic Era. Late Paleozoic or subsequent deformation have caused minor faulting in northern Illinois and southern Wisconsin. Current tectonism is evident only in the occasional earthquake (Figure 4). A crudely northeast striking band of epicenters transects northern Illinois and southern Wisconsin and correlates with a northeast striking long-wavelength magnetic anomaly. Regionally, observed heat flow provides no evidence of recent thermal events or related tectonism. Thermal conductivity measurements on the granitic rocks of UPH 3 and temperature gradient observations, which have a mean value of $23.0^{\circ}\text{C km}^{-1}$ [Rahman, 1982], result in an average heat flow of 75 mW m^{-2} . This value is higher than normal heat flow in the eastern United States, probably due to the high total radioactivity (K, U, Th) of the granite [Rahman and Roy, 1981; Rosholt, this issue]. Rahman [1982] determined that the heat generation in the granitic rocks of both UPH 2 and UPH 3 is abnormally high, with a mean value of roughly 16 mW m^{-2} . In UPH 3 the heat

Fig. 3. (a) Total magnetic intensity anomaly map [after O'Hara, 1982; Hinze et al., 1982]. Contour interval 200 gammas (nT). Rectangle shows approximate locations of drill holes (Figure 2). Heavy dashed line marks approximate northwestern boundary of NE-SW trending belt of positive anomalies. (b) Bouguer gravity anomaly map [after O'Hara, 1982; Hinze et al., 1982] for region around Commonwealth Edison drill holes (heavy rectangle; Figure 2). Contour interval 5 mGal. Heavy dashed line marks approximate NW boundary of region of broad, low-amplitude anomalies.

generation decreases with depth, but in UPH 2 the values are relatively low to a depth of 1150 m, where high values occur to the bottom of the hole. The measured average heat flow is considerably less than that predicted for the eastern United States considering the relatively high heat-generating capacity of the UPH granites [Birch et al., 1968; Roy et al., 1972]. This discrepancy is attributed by Rahman [1982] to a relatively thin pluton source, which is confirmed by interpretation of the gravity data [Aiken et al., this issue].

Geochemical Processes.

Because holes UPH 2 and UPH 3 each provide more than 900-m continuous sample of the crystalline basement, this is an excellent opportunity to examine a number of geochemical processes as a function of depth. These include both near-surface processes, principally weathering, and deeper-seated processes involving crustal water [Courtne et al., this issue] and the geothermal gradient. In addition, the granitic body encountered appears to have a pronounced stratification or compositional gradient, so that detailed petrologic and chemical studies may help in understanding the processes of magmatic crystallization.

One of the principal studies involving near-surface phenomena was the U, Th-Pb systematics and associated U-series disequilibrium [Stuckless et al., 1981; Rosholt et al., this issue; Doe et al., this issue]. This yielded important information on the degree of leaching of uranium in the rock near the top of the Precambrian basement, the depth to which such processes have operated, and the degree to which present crustal water circulation is actively leaching uranium or its decay products from the rocks. Such data will not only help in understanding the mobility of uranium and lead in the crust but will also contribute to better understanding of rock-water interactions in general. Significant data relating to these problems also came from stable isotope studies [Shieh, this issue; Kombrink and Clayton, this issue].

Fission track studies are helping to define the uplift history of the region. Progressively younger fission track ages were found along the 900-m depth of the hole that can be translated into approximate uplift rates [Zimmerman, 1981, 1983]. The results suggest that considerable sedimentary cover (0.5-3 km) has been eroded since 130 m.y. ago, that the past geothermal gradient was greater, or some combination of both.

Physical Property Studies.

Core samples that are representative of both the major and minor rock units have been studied to provide fundamental physical properties of the crystalline rock units. Laboratory physical property measurements on core samples include density, specific gravity, magnetic susceptibility, induced polarization, sonic velocity, electrical resistivity, thermal conductivity, and radioactivity [Daniels et al., this issue; Kowallis and Wang, this issue; Eckstein et al., 1981, this issue; Hamison and

Fig. 4. Earthquake epicenters, 1534-1971, for eastern United States [after York and Oliver, 1976].

Doe, 1981; Lidiak and Denison, this issue; Olhoeft et al., 1981; Rahman and Roy, 1981]. Results of these studies provide basic data needed for better interpretation of regional geophysics, crustal hydrology, crustal stress measurements, etc. In addition, these measurements make it possible to correlate petrophysical quantities with specific lithologies and with results of geophysical logs. Several different measurements confirm the homogeneity and lack of alteration and fracturing over considerable vertical intervals within the crystalline rocks. In addition, Carter [1981] studied the structural deformation in the crystalline rocks, which provides a very useful supplement to the stress measurements by Hamison [1981] and Hamison and Doe [this issue]. Detailed fracture logging of the core [Hamison and Doe, 1981; Kowallis and Wang, this issue] is also of assistance in investigating the stress history of the granite and is applicable for correlation with various borehole experiments and the study of crustal fractures as a function of depth.

Regional Surface Geophysics.

The regional significance of many results from measurements within the Illinois deep drill UPH 3 hole and its core are dependent upon defining the basement geologic framework. Two scales of definition are important. One involves characterizing the three-dimensional geometry of the specific feature which has been drilled, and the second relates to the regional basement geology of the northern midcontinent. Current

direct knowledge of the basement is restricted by the lack of deep drill holes and basement outcrops and surface geophysical measurements are essential in providing indirect information.

Gravity measurements have been particularly useful in outlining granitic bodies. Previous gravity measurements on approximately a 2-km grid are available over Illinois [McGinnis et al., 1976]. Aiken et al. [1981, this issue] expanded this network into Wisconsin, where only reconnaissance gravity observations were previously available [Ervin and Hammer, 1974]. They have isolated a Bouguer gravity anomaly minimum over the drill holes and found that it extends to the north into Wisconsin. They interpret the minimum as caused by a granite body that covers an area of roughly 1500 km² and has a thickness of 4-5 km. Recently released aeromagnetic coverage in northwestern Illinois [U.S. Geological Survey, 1982] shows a magnetic minimum correlative with the gravity negative anomaly. These data, as discussed in the section on regional geology, establish the regional basement geology framework. A seismic reflection survey [Coates, 1981] over the drill hole sites shows the modest relief of the basement surface (Figure 2).

In-Hole Experiments

Two of the three drill holes had been plugged before the Illinois Drill Hole Project began. Thus the only hole available for testing was UPH 3. Casing has been left in this hole throughout the sedimentary section in order to maintain wall integrity; Figure 5 summarizes some of the logging information on UPH 3 for the Paleozoic section. Thus in-hole experiments were restricted to the Precambrian granite section. The 76-mm diameter of this section further limited the variety of tests that could be performed. Three major series of measurements were conducted, hydrofracturing stress, rock-mass permeability, and in-hole logging.

The stress measurements [Haimson and Doe, this issue] were the first at such depths (650-1500 m) in the Precambrian granite of North America, and their results are significant to the understanding of the stress regime in the midcontinent. The stress profile suggests that conditions favorable to strike slip faulting may exist below 1100 m, in accord with a focal mechanism solution of a 1972 earthquake some 100 km south of the UPH 3 site. The single packer permeability tests [Doe and Haimson, 1981] show overall extremely low permeabilities, decreasing with depth (10⁻⁴ to 10⁻⁶ darcy) and strongly affected by two fracture zones in the hole.

Accessibility of drill hole UPH 3 to geophysical logging, and thus measurement of in-place physical properties for their intrinsic value as well as for comparison with measurements on the core samples, is a significant aspect of the scientific importance of the Illinois Deep Drill Hole Project. All three holes were logged geophysically for Commonwealth Edison to determine basic geologic, engineering, and hydrologic characteristics. Subsequently, temperature logs of UPH 3 have been recorded by Olhoeft et al [1981] and have been used by Rahman and Roy [1981] to measure the heat flow. In

addition, Daniels et al. [this issue] have recorded resistivity, gamma ray, neutron, induced polarization, self potential, magnetic susceptibility, gamma-gamma density, and nonlinear complex resistivity logs over the nearly 1-km depth of granite drilled in UPH 3. They have correlated these observations with physical properties obtained from laboratory measurements on core samples and the geologic log. Thus these measurements provide an insight into the usefulness of geophysical logs for characterizing the geologic section where core samples are not available for direct study. Furthermore, the in-hole observations are useful in localizing fractured zones and in identifying an approximately 300 m zone of nonfractured granite at the bottom of the hole.

Conclusions

The Illinois deep drill holes and core have provided a unique scientific opportunity to a wide variety of geoscientists. The uniqueness is derived from (1) the geographic locations within an area in which the basement and lower Paleozoic geology is poorly known because of the paucity of deep drill holes, (2) the great vertical depth of continuous core of both the Paleozoic sedimentary rocks and the Precambrian crystalline basement, (3) the geological locations on the Wisconsin Arch, and (4) the availability of hole UPH 3 to in-hole testing and experiments within crystalline rocks where core is available for geological and petrophysical studies. The wide spectrum of in-hole and surface experiments and core studies summarized above and detailed in the following and other papers testify to the importance put upon the scientific opportunities by a broad range of disciplines. The results have provided a new insight into the age, origin, and subsequent geological processes acting upon the Precambrian basement rocks and have provided important information on geophysical parameters and in situ characteristics which are necessary for defining past and present tectonic regimes. Furthermore, the comparison of surface, in-hole, and core-derived parameters will be useful in interpreting future investigations where all three types of measurements are not available.

The results of the Illinois deep drill hole studies presented here should have a much broader range of significance than indicated by the point location of the drill holes. This is true geographically, but just as important, it is true in terms of understanding geologic processes and the character of deep basement rocks. These results should have far-reaching implications to pressing national societal concerns. Furthermore, the scientific results as well as the lessons learned regarding the development and coordination of the scientific program will be useful in planning and conduction of future add-on studies for "holes of opportunity." We believe that important ingredients to the success of the Illinois deep drill program were (1) early recognition of the scientific importance of add-on experiments to the drill holes, (2) cooperation from the owner and operator of the drilling program, (3) organization of a steering committee composed of interested individuals with diverse discipline backgrounds that could enlist

Fig. 5. Summary of well log data for Paleozoic section of Commonwealth Edison drill UPH 3 [after Harza Engineering Company, 1980].

the cooperation of a broad range of the scientific community, (4) limited "seed" money to provide for program development and coordination, and (5) scientists and their supporting agencies that were willing to modify existing programs to take advantage of the unique opportunities provided by the "holes of opportunity."

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APPENDIX VI.2.

PETROLOGY AND PHYSICAL PROPERTIES OF GRANITES FROM THE ILLINOIS
DEEP HOLE IN STEPHENSON COUNTY

Reprinted from Lidiak and Denison (1983)

PETROLOGY AND PHYSICAL PROPERTIES OF GRANITES FROM THE ILLINOIS DEEP HOLE
IN STEPHENSON COUNTY

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Abstract. Two main basement granitoid types have been identified in core samples from the Illinois deep hole project. The main variety is a medium- to coarse-grained porphyritic biotite granite with phenocrysts of microcline perthite and less commonly quartz and sodic plagioclase in a matrix of these minerals and biotite, muscovite, fluorite, magnetite, ilmenite, zircon, hornblende, apatite, sphene, monazite, rutile, and clinopyroxene (relict). The texture is typically hypidiomorphic gradational to recrystallized xenomorphic. The second main granitoid, which occurs in the upper part of hole UPH 3, is a fine-grained granoblastic to lepidoblastic gneissic granite that is distinct from and possibly older than the nonfoliated granite. The textures of both rocks have been modified by a mild cataclastic shearing that has partially recrystallized the more susceptible mineral phases. Thin fracture planes that crosscut the earlier foliations are common. Microprobe analyses indicate that biotites in the gneissic granite are chemically distinct from those in the granite. Biotites in the gneissic granite have higher Fe/Fe + Mg ratio, FeO, and Al_2O_3 and lower MgO and SiO_2 . Reflection microscopy and microprobe analyses indicate that the oxide phases in the two rocks are also different. The oxides in the granite are magnetite and ilmenite, whereas hematite and pseudobrookite occur in the gneissic granite. The biotite and Fe-Ti oxide data represent additional evidence in support of the fact that the granite and gneissic granite are distinct rocks and probably not part of a continuous comagmatic sequence. Major element chemical analyses indicate that the granites have affinities to anorogenic rapakivi granites. The granites in the deep holes are high in SiO_2 , alkalis ($Na_2O + K_2O$), F, FeO/MgO; low in Al_2O_3 , FeO, Fe_2O_3 , TiO_2 , MnO, and P_2O_5 ; and slightly low in MgO and CaO. Magnetic susceptibility and density measurements correlate generally well with magnetic susceptibility and density logs. Magnetic susceptibilities range from 20 to 17,000 micro SI units; higher values correlate mainly with greater amounts of Fe-Ti oxides and biotite; lower values with gneissic rocks and muscovite content. Density measurements range from 2.595 to 2.685 Mg/m^3 ; denser rocks correlate positively with higher amounts of biotite and Fe-Ti oxides and inversely with microcline perthite content.

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Introduction

Thirty nine samples from the Commonwealth Edison deep drill holes UPH 2 (sec. 12-28N-5E) and UPH 3 (sec. 7-28N-6E) in Stephenson County, Illinois, provided an excellent opportunity to undertake a detailed study of the petrology and physical properties of the core samples. Study of these samples is important because they lead to a better understanding of the basement in this portion of northern Illinois, an area of sparse basement well control and thus poorly known basement geology. The nearest basement well control occurs about 80 km to the west and south-east, in Dubuque County, Iowa, and in Winnebago and Boone counties, Illinois, respectively, where biotite granites have been encountered in both areas. The nearest Precambrian outcrops occur about 110 km to the north where the Baraboo Quartzite is exposed; Precambrian granites and rhyolites crop out about 160 km farther north in central Wisconsin.

The area of northern Illinois is part of a large granite-rhyolite terrane [Lidiak et al., 1966; Van Schmus and Bickford, 1981; Denison et al., 1983] that extends over much of the central Midcontinent from central Wisconsin to northern New Mexico and Arizona. This vast terrane consists mainly of unmetamorphosed rhyolitic and trachytic volcanic rock and epizonal granitic rock with minor basalt, sedimentary rock, and metamorphic rock. Cores from UPH 2 and UPH 3 are examples of porphyritic granite from this terrane.

The main purpose of this study is to report on the petrographic, chemical, and mineralogical characteristics of the granitic rocks. A second objective is to correlate the magnetic susceptibility and density measurements on the core samples with the main mineralogical variations and with the magnetic susceptibility and density logs.

Petrography

Two main rock types are present in the Illinois deep holes. The main variety, which is present in both UPH 2 and UPH 3, is a medium- to coarse-grained, moderately porphyritic biotite granite. The second main type is a fine-grained granoblastic to lepidoblastic gneissic granite that occurs in the upper part of UPH 3 in the depth interval 673-731.

Modal analyses indicate that microcline perthite generally predominates over quartz and plagioclase as the essential minerals. The granites (Figure 1) plot in both the granite A and granite B fields of Streckeisen's [1976]

Fig. 1. Modal analyses of granites from Illinois deep holes UPH 2 and UPH 3.

classification. Each mode was determined by counting 1000-1500 points on one thin section from each sample. The scatter shown on Figure 1 reflects in part variation in phenocryst content and large grain size as well as variation in primary mineralogy.

The change in mineralogy with depth in each hole is shown on Figure 2. There is no pronounced systematic change with depth in either hole except that biotite and Fe-Ti oxides show some slight tendency to increase with depth in both UPH 2 and UPH 3 and that muscovite tends to decrease with depth in UPH 3. The latter observation is mainly an expression of the fact that muscovite predominates over biotite in the gneissic granite but is subordinate to it in the granite.

Gneissic Granite

A fine-grained gneissic granite (Figure 3) is found in the upper portion of the UPH 3 well. The silicate mineralogy and modal composition of the gneissic granite are essentially identical to the massive granites and are particularly close to the medium-grained granite. It may represent a finer-grained, foliated phase of the massive granites. However, the texture shows significant differences, and on this basis the rock may be a separate and distinct body. On the logs it shows a higher gamma ray radioactivity and a lower magnetic susceptibility than the adjacent massive granite.

The gneissic granite is composed chiefly of perthitic microcline, quartz, and plagioclase. Muscovite (secondary) and biotite are the most common accessories with lesser to trace amounts of Fe-Ti oxides (hematite and pseudobrookite), hornblende, sphene, zircon, fluorite, rutile, and a carbonate mineral. Chlorite and feldspar alterations are found in each interval examined.

The microcline is a mesoperthite, containing locally developed string lamellae. It is generally very fresh and well twinned but does contain a minor amount of local hematite dust and vacuoles.

Quartz is generally moderately stained. It contains inclusions of rutile as fine hairlike needles. There are minor trains of inclusions that appear associated with fractures and pass beyond crystal boundaries into adjacent feldspars.

Plagioclase is generally fresh and well twinned. The twinning is commonly bent and fractured. The composition is sodic oligoclase-calcic albite.

Muscovite is conspicuous as a crystalloblastic mineral. It has a rude preferred orientation parallel to shearing. The mica is bent in larger crystals. There is a slight pale olive-greenish cast to the muscovite that is locally developed near grain margins or near tiny inclusions of undetermined composition.

Biotite occurs as small very strongly pleochroic crystals. It has a distinctly dark olive brown color and contains abundant indeterminate specks. It shares a slight preferred orientation with muscovite.

Fluorite is a widely distributed and relatively abundant accessory mineral. It has a strong purple color developed locally but is generally colorless. Some larger crystals have rutile needles developed along cleavage.

Zircons are found as tiny, rather sparse crystals. Small rare hornblende crystals occur mainly as inclusions in other minerals. Fe-Ti oxides form sparse equidimensional crystals. A high relief carbonate mineral, probably siderite, is an apparent metamorphic mineral.

Chlorite and feldspar alterations are minor but widely distributed secondary minerals and

Fig. 2. Variation in mineralogy with depth, Illinois deep holes UPH 2 and UPH 3. Q, quartz; KF, K feldspar; PL, plagioclase; stippled area, biotite; solid area, muscovite; open area to right of micas, accessory minerals.

probably represent a low-grade metamorphic event.

The rock is generally fine grained with a curious sheared granoblastic to lepidoblastic texture (Figure 3). The average grain size is generally less than 1 mm with larger crystals as large as 5 mm locally developed. The gneissic fabric shows little evidence of mineral layering, and the gneissic character is defined by the dimensional orientation of the mineral phases and a parallel shearing plane. As noted previously, the muscovite best shows the slight preferred orientation. The biotite is less abundant and is in stubbier crystals and shows its orientation to a lesser extent. The principal mineral phases all show a well-defined elongation in the plane of shear foliation. Large microclines tend to contain numerous inclusions. Some of these inclusions are plagioclase and microcline in random orientations, but most are quartz. The quartz inclusions occur mainly as unoriented blebs and less commonly as crystallographically oriented crystals. These crystals are interpreted as poikiloblasts and are apparently of metamorphic origin.

The gneissic fabric is not well defined in each of our thin sections. We ascribe this to the sections not being cut normal to foliation. The foliation is not conspicuous megascopically. Nonetheless, the metamorphic gneissic fabric is clearly evident in every section in the interval 673-731 and is in clear contrast to the texture of the other granites. We interpret the gneissic fabric as having developed prior to the intrusion of the coarse granites by virtue of the extensive recrystallization and penetrative deformation not seen in the massive granites and thus suggest that the gneissic granite may be older than the coarser-grained granite. An alternative interpretation is that

the gneissic granite is a textural variant of the pervasively sheared, coarser granite and that the gneissic fabric reflects more extensive development of the shearing in a finer-grained rock.

In addition to the metamorphic gneissic fabric, thin, cataclastic shear surfaces are present in both granitoid types. Partial recrystallization and modification of earlier textures are common along these irregular surfaces. These surfaces are interpreted as postdating both the gneissic granite and the granite. Additional thin fracture planes crosscut these earlier foliations.

Granites

Both a medium-grained (Figure 3) and a coarser, strongly porphyritic granite are found in the drill holes. The mineralogy of the two is exceptionally close but not identical. Information about the contacts between the granites is based upon limited observations during core examination, inspection of gamma ray and magnetic intensity logs, and generalized lithologic logs provided by M. S. Coates of Harza Engineering Company. The difference is better marked megascopically. The contacts in the core are generally abrupt but in detail are gradational over an interval of 50-100 mm. The coarser, porphyritic granite is less porphyritic at contacts and contains rather conspicuous coarse bluish quartz. We interpret the medium-grained granite as a later crystallization stage of the same body. The medium-grained granite has a higher gamma ray intensity and, to a lesser extent, a lower magnetic susceptibility than the porphyritic granite. An example of these changes at a contact can be seen in UPH 2 near 776 m. Some contacts do not appear to be marked by strong gamma ray

Fig. 3a. Photomicrograph of sheared granoblastic to lepidoblastic texture in gneissic granite, core 673, Illinois deep hole, UPH 3. Photo width = 30 mm. Crossed polarized light.

contrasts. None of these contacts was observed during core examination.

The medium-grained and porphyritic granites are very clearly closely related. There is considerable overlap of all mineralogic and petrographic characteristics. The rather abrupt contacts marked by changes in gamma ray and magnetic intensity indicate that they are not simply textural variations of an otherwise homogeneous granite body but probably represent related intrusive stages.

The granites are composed mostly of microcline perthite, quartz, and plagioclase with biotite as the most conspicuous accessory

mineral. In addition minor to trace amounts of muscovite, Fe-Ti oxides (magnetite and ilmenite), sphene, rutile, zircon, hornblende, monazite, and fluorite are widely distributed. Minor secondary chlorite after biotite and feldspar alterations are ubiquitous. Rare crystals of relict clinopyroxene, which occur as resorbed cores in hornblende megacrysts, are also present.

Quartz forms large single crystals showing extensive straining and some reorganization of crystal margins with other quartz. There are locally abundant, randomly oriented hairlike inclusions in quartz believed to be rutile. These inclusions probably cause the bluish cast

Fig. 3b. Photomicrograph of granitic texture in granite, core 1325, Illinois deep hole, UPH 2. Photo width = 30 mm. Crossed polarized light.

Fig. 3c. Photomicrograph of thin cataclastic shear surfaces and xenomorphic texture in granite, core 1016, Illinois deep hole, UPH 3. Photo width = 3 mm. Crossed polarized light.

seen in hand specimen. The quartz also contains trains of equidimensional inclusions of uncertain composition, some of which pass undisturbed between crystals.

The plagioclase shows mild zoning, although zoning is not characteristic of all plagioclase. Most of the feldspar alterations are formed at the expense of plagioclase. Much of the alteration is the development of clay minerals and fine sericite with little apparent pattern but more abundant near the cores of crystals. There is also a systematic development of white micas showing very distinctive preferred orientation. These white micas occur along

several planes that are almost certainly crystallographically controlled. They tend to have sharp boundaries and are generally in slender, well-formed books. Locally but not characteristically the plagioclase is bent and broken.

The microcline perthite is generally exceptionally fresh and well twinned. The perthite is both a patch and string type. The patch type is characteristic of the larger crystals, and the string type more common in the smaller. The larger crystals contain poikilitic inclusions of small quartz and plagioclase in an apparently random orientation.

Fig. 3d. Photomicrograph of crystoblastic quartz-biotite-FeTi oxide restite in granite, core 1455, Illinois deep hole, UPH 2. Photo width = 3 mm. Plane polarized light.

TABLE 1. Chemical Analyses of Biotites

Rock type No. of grains	Sample			
	666 granite 4	905 granite 5	1455 granite 7	731 gneiss 7
SiO ₂	35.86	37.33	35.90	32.26
TiO ₂	2.46	2.43	2.05	2.02
Al ₂ O ₃	15.48	15.45	17.89	18.92
FeO†	25.93	23.48	23.25	33.25
MnO	0.41	0.57	0.23	0.28
MgO	6.69	7.88	6.87	1.41
CaO	0.15	0.15	0.19	0.21
Na ₂ O	0.09	0.07	0.11	0.10
K ₂ O	8.64	8.75	9.65	8.61
Total**	95.71	96.11	96.14	97.06
Cations Per 22 Oxygen				
Si	5.613	5.731	5.528	5.177
Aliv	2.387	2.269	2.472	2.823
Alvi	0.468	0.562	0.776	0.756
Ti	0.290	0.280	0.238	0.244
Fe	3.394	3.015	2.994	4.463
Mn	0.055	0.074	0.030	0.038
Mg	1.561	1.804	1.577	0.337
Ca	0.025	0.025	0.031	0.036
Na	0.028	0.020	0.033	0.031
K	1.725	1.714	1.895	1.763
Total	15.546	15.458	15.574	15.605
Fe/Fe+Mg	0.685	0.626	0.655	0.930

*Sample numbers in Tables 1-3 refer to depth interval (in meters) of cores in Figure 1.

**Fluorine was detected in all samples but was not quantitatively determined.

The biotite is strongly pleochroic. It is characteristically but not uniquely olive brown in color in the medium-grained granites and a reddish brown in the coarser. The biotite is almost always associated with other accessory minerals, mainly zircon and Fe-Ti oxides. Zircon inclusions cause strong pleochroic halos in the biotite. Apple green chlorite is formed locally at the expense of biotite, but the biotite is generally fresh. The coarser porphyritic granite has larger biotite books than the medium grained. Although most biotite is in well-formed books, some is in delicate sievelike crystals.

Zircons are numerous and widely distributed. They characteristically occur as euhedral crystals in close association with biotite. They are commonly zoned and rather turbid near the cores. Monazite (?) is invariably found near zircon. It forms larger crystals that are partly altered to reddish brownish indeterminate material. Apatite is characteristic only of the coarse porphyritic granite where it occurs as small but numerous prisms associated with other accessory minerals.

Fe-Ti oxides (magnetite and ilmenite) are found as discrete equidimensional crystals and as larger, rodlike crystals in association with

biotite. They are much more common in the coarse granite. Sphene is a sparse mineral found near the margins of some biotite crystals.

Muscovite is more characteristic of the medium-grained granite, much of it in intimate association with biotite. It may either be in crystallographic orientation with the dark mica or in random orientation and is of probable subsolidus origin. The muscovite is also found as rather late intergranular crystals, some of which occupy a rather large area as incipient sievelike crystals.

Fluorite is unusually abundant and widely distributed as an accessory mineral. It is generally colorless, but the cores of some crystals have a distinctive purple color. There are tiny, crystallographically oriented needles of rutile(?) in many of the larger fluorites.

The characteristic texture is porphyritic and hypidiomorphic, later modified by a mild to moderate cataclastic shearing and partial recrystallization that produces local xenomorphic texture (Figure 3). The average grain size, as measured in thin section, is in the 4- to 7-mm range with microcline phenocrysts commonly larger than 10 mm. Where the grain boundaries are the least modified they tend to be straight and

TABLE 2. Chemical Analyses of Fe-Ti Oxides

	Sample				
	905	1455	1455	731	731
Mineral Rock type No. of grains	magnetite granite 7	magnetite granite 4	ilmenite granite 8	hematite gneissic granite 4	pseudobrookite gneissic granite 5
SiO ₂	0.02	0.04	0.05	0.21	0.26
Al ₂ O ₃	0.17	0.11	0.27	0.41	0.57
Fe ₂ O ₃	67.03	67.45	10.51	96.26	20.41
FeO	30.54	30.39	37.23	0.29	18.18
MgO	0.01		0.01		0.01
CaO	0.03	0.01	0.09	0.04	0.09
TiO ₂	0.28	0.16	45.44	0.28	51.94
MnO	0.20	0.18	3.78	0.19	0.44
Total	98.28	98.34	97.38	97.68	91.90
Oxygen	4	4	3	3	5
Si	0.001	0.002	0.001	0.006	0.011
Al	0.008	0.005	0.008	0.013	0.028
Fe+3	1.975	1.986	0.206	1.964	0.649
Ti	0.008	0.005	0.890	0.006	1.650
Fe+2	1.000	0.994	0.811	0.007	0.641
Mg					0.001
Ca	0.001		0.003	0.001	0.004
Mn	0.007	0.006	0.083	0.004	0.016
Total	3.000	2.998	2.002	2.001	3.000

smooth, typical of anorogenic granites. Modification of the grain boundaries and mild recrystallization are seen in all samples.

The quartz-quartz contacts are the most sensitive to recrystallization. These boundaries are delicately sutured and rarely have a modest mortar structure. The quartz-feldspar contacts are reflected in the feldspars as bent and broken twinning planes. The recrystallization in the micas is also pervasive but not as dramatically so as in the quartz.

The micas tend to have locally frittered edges that have clearly grown into the adjoining quartz-feldspar. The muscovite appears texturally late and is commonly developed along shear dislocations in feldspar. In some intervals are complex symplectic intergrowths of quartz-biotite-Fe-Ti oxides that suggest a crystalloblastic restite origin (Figure 3).

Mineralogy

Co-existing biotites and Fe-Ti oxides from both the granite and the gneissic granite were analyzed with a fully automated ARL microprobe and corrected with the Bence and Albee [1968] correction procedures. The reported analyses are averages of three to five points on each of four to eight grains per thin section. Total Fe was analyzed as FeO but was recalculated on the basis of mineral stoichiometry according to the method of Finger [1972] in order to estimate

the ferrous and ferric proportions. On the basis of these calculations, only the Fe-Ti oxides contain ferric cations.

Biotite. Microprobe analyses of four biotites, three from the granite and one from the gneissic granite, are listed in Table 1. Biotites from the granites have Fe/Fe+Mg ratios between 0.6 and 0.7 and are typical of most granitic biotites in which Fe predominates over Mg [Foster, 1960; Dodge et al., 1969; Neilson and Haynes, 1973; Barriere and Cotten, 1979]. In contrast, biotite from the gneissic granite has an Fe/Fe+Mg ratio of 0.93. The high Fe and Al and low Mg contents indicate that it is siderophyllite. Such Fe-rich biotites characteristically occur in greisen, pegmatites, and rapakivi granites [Foster, 1960; Simonen and Vorma, 1969]. The very high Fe content is further borne out by the fact that biotite in the gneissic granite has distinctly darker olive brown pleochroism than does biotite in the granite.

On the basis of the available analyses, two populations are apparently present. Compared to the biotite in the granite, the biotite in the gneissic rock is higher in FeO and Al₂O₃ and lower in MgO and SiO₂. These compositional differences suggest that the granite and the gneissic granite are distinct rock types and, although possibly comagmatic, not part of a continuous crystallization sequence.

Fe-Ti oxides. Contrasting Fe-Ti oxide

TABLE 3. Chemical Analyses of Granites

	Sample						
	717	905	1071	1429	1281	1325	1455
	gneissic granite UPH 3	granite UPH 3	granite UPH 3	granite UPH 3	granite UPH 2	granite UPH 2	granite UPH 2
SiO ₂	74.85	71.95	75.53	75.46	75.57	75.01	70.62
TiO ₂	0.16	0.49	0.18	0.32	0.09	0.11	0.55
Al ₂ O ₃	12.92	13.36	12.66	11.77	12.31	13.19	13.78
Fe ₂ O ₃	0.58	0.86	0.47	0.84	0.81	0.39	1.47
FeO	1.13	1.82	0.97	1.90	0.62	0.95	2.47
MnO	0.03	0.04	0.03	0.05	0.01	0.03	0.07
MgO	0.05	0.38	0.05	0.11	0.04	0.03	0.58
CaO	1.16	1.48	1.11	1.42	0.98	1.38	1.72
Na ₂ O	3.28	3.13	3.28	3.28	3.28	3.96	3.74
K ₂ O	5.84	5.55	5.68	4.40	6.38	4.32	4.58
H ₂ O ⁺	0.65	0.70	0.45	0.54	0.29	0.58	0.53
H ₂ O ⁻	0.02	0.08	0.03	0.03	0.04	0.02	0.03
P ₂ O ₅	0.01	0.07	0.01	0.04	0.01	0.02	0.19
F	0.21	0.19	0.28	0.58	0.36	0.42	0.36
Y ₂ O ₃	0.01	0.01	0.01	0.01	0.01	0.01	0.01
ZrO ₂	0.02	0.02	0.02	0.03	0.01	0.02	0.02
Total	100.92	100.13	100.76	100.78	100.81	100.44	100.72
CIPW Norms							
Q	30.75	28.28	32.28	35.84	30.83	32.14	26.16
Or	34.51	32.79	33.57	26.00	37.69	25.55	27.05
Ab	27.74	26.49	27.74	27.74	27.74	33.51	31.62
An	3.28	6.01	3.06	4.40	0.03	5.45	7.32
Di	2.18	0.73	2.08	2.32	0.84	1.21	0.02
Wo	—	—	—	—	1.58	—	—
Hy	0.36	2.47	0.17	1.41	—	0.73	3.98
Mt	0.83	1.25	0.69	1.23	1.18	0.53	2.13
Il	0.30	0.93	0.35	0.61	0.17	0.21	1.05
Ap	0.03	0.17	0.03	0.10	0.04	0.03	0.44
Z	0.04	0.04	0.04	0.04	0.02	0.04	0.04

Analyzed by H. Dehn and E. G. Lidiak.

minerals are present in the granite and in the gneissic granite. Magnetite and subordinate ilmenite occur in the granite, whereas hematite and pseudobrookite characterize the gneissic granite. These minerals are distinguishable by reflection microscopy and have the following properties: magnetite: brownish gray, moderate reflectivity, no birefractance, and isotropism; ilmenite: brownish gray, moderately low reflectivity, moderate birefractance, and distinct anisotropism; hematite: white, high reflectivity, weak birefractance, and distinct anisotropism; pseudobrookite: rectangular to irregular shape, gray with reddish brown margins, internal reflections, low reflectivity, no birefractance, and weak anisotropism.

Representative microprobe analyses are listed in Table 2. The magnetite and ilmenite are typical of Fe-Ti oxides from granites in that both are low in TiO₂, MgO, and FeO/Fe₂O₃ ratio. Application of the Buddington and Lindsley [1964] magnetite-ilmenite geothermometer/oxygen barometer using Lindsley and Spencer's [1982]

calculation procedures on the coexisting magnetite-ilmenite pair from granite sample 1455 yields a temperature of less than 600°C and an oxygen fugacity of about 10⁻¹⁵ Pa. These values probably represent subsolidus equilibration conditions.

Hematite and pseudobrookite are important phases in the gneiss. Whether they are primary minerals or represent oxidation products of preexisting magnetite and ilmenite was not determined. At least some of the hematite is late, as it mantles some of the pseudobrookite and occurs as veinlets in the gneissic granite. Of particular interest to the origin of the rock is the pseudobrookite. Pseudobrookite is a high-temperature mineral which is unstable below about 600°C. It occurs in a variety of high-temperature igneous associations, mainly as a high-temperature alteration product of Ti-rich mafic rocks [Frenzel, 1954], as a primary intratelluric oxidation phase in basalt [Conquere and Girod, 1968], and as a primary quench phase in basaltic glass [Anderson and Wright, 1972]. In metamorphic rocks, it is

Fig. 4. Plot of aluminum saturation of granites from UPH 2 and UPH 3, Illinois deep holes.

formed by high-temperature contact metamorphism and partial melting of pelites [Smith, 1965] and by high-temperature contact metamorphism of weathered basalt [Agrell and Langley, 1958]. Its presence is regarded as strong evidence that the oxidation in the gneissic granite is not a low-temperature deuteric phenomenon associated with cooling of the granite. In addition, as the granite does not show high-grade metamorphic effects, a later high-temperature thermal event is apparently ruled out. The gneissic granite appears to be distinct from the granite. It may be a genetically related, earlier intrusive body. Alternatively, the gneissic granite could be a relict fragment or xenolith, which possibly is part of an older quartz-feldspathic basement.

Microcline. K-feldspars from both the granite and the gneissic granite are slightly perthitic microclines. Triclinicity values [Goldsmith and Laves, 1954] on 15 samples from the granite have a mean of 0.925 ± 0.038 and 7 samples from the gneissic granite have a mean of 0.945 ± 0.034 . Both sample populations approach maximum microcline, reflecting equilibration down to about 350°C [Ribbe, 1975]. The values are not distinguishable statistically.

Rock Chemistry

Major element analyses have been obtained by standard wet chemical techniques. Analyses of six granites and one gneissic granite from UPH 2 and UPH 3 are summarized in Table 3. Analyses of the granites and the gneissic granite

are closely similar and not distinguishable on the basis of major element composition.

The granites are relatively high in SiO_2 , alkalis ($\text{Na}_2\text{O} + \text{K}_2\text{O}$), F, and FeO/MgO ; low in Al_2O_3 , FeO, Fe_2O_3 , TiO_2 , MnO, and P_2O_5 ; and slightly low in MgO and CaO. These chemical characteristics are closely similar to rapakivi granites from Finland [Sahama, 1945; Savolahti, 1962] and from the Wolf River batholith, Wisconsin [Van Schmus et al., 1975; Anderson and Cullers, 1978]. On this basis, the granites from UPH 2 and UPH 3 are regarded as being rapakivi granites. They have anorogenic affinities and can be regarded as being A-type granites as defined by Loiselle and Wones [1979].

The rapakivi granites are distinguishable from the more aluminous and calcic calc-alkaline granites on the basis of alumina saturation [Shand, 1943]. Figure 4 shows that rapakivi granites are mainly metaluminous and generally distinguishable from calc-alkaline granitic rock associations such as the Sierra Nevada batholith [Bateman et al., 1963]. Other plots such as $\text{Al}_2\text{O}_3/\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$ versus $\text{FeO} + \text{Fe}_2\text{O}_3/\text{FeO} + \text{Fe}_2\text{O}_3 + \text{MgO}$ (not shown) yield similar results.

The chemical variation among the seven analyses is shown on Figure 5. There is a curvilinear relationship between MgO and SiO_2 , TiO_2 , MnO, P_2O_5 , Y_2O_3 , and to a lesser extent Al_2O_3 , Fe_2O_3 , FeO and CaO. The variation is probably the result of modal differences of the main essential and accessory minerals rather than of magmatic fractionation. Biotite, although clearly not the only control mineral, is an impor-

Fig. 5. MgO versus other main oxides of granites from UPH 2 and UPH 3, Illinois deep holes.

tant phase in accounting for the variation. The positive slopes for TiO_2 , MnO , Al_2O_3 , and FeO and the negative slope for SiO_2 shown on Figure 5 are all consistent with biotite being a major mineral in controlling the chemical variation. The differences in MgO and FeO contents between the average granitic biotite of Table 1 and the seven granites show this dependence particularly well. The FeO and MgO contents of the biotite lie along a linear extension of the trend shown on Figure 5. Application of the lever rule indicates that removal (or addition) of about 8 wt % biotite can account for variation in FeO and MgO in the granites. This amount is consistent with the variation in percent of modal biotite (recalculated to modal weight percent) that occurs in the analyzed granites (Figure 2). Other chemical variations shown on Figure 5 require removal (or addition) of additional coexisting phases.

Correlation With Physical Properties

Magnetic susceptibility and density measurements were carried out on samples from both UPH 2 and UPH 3. Susceptibilities were measured on cylindrical cores having a diameter and length of 25.4 mm using a Bison model 3101 Susceptibility Meter. Densities were determined

on the same cores using a standard dry bulk density method in which the volume and weight were measured in air.

Magnetic susceptibility. The magnetic susceptibility log shown on Figure 6 is taken from Daniels et al. [this issue]. Also plotted on Figure 6 are the magnetic susceptibility measurements on the core samples. In general, there is good correlation between the log and the measured susceptibility values, with both showing a general increase with depth. The susceptibilities are clearly related to magnetite contents, and the modal percentage of Fe-Ti oxides generally correlates well with the susceptibility log and the measured values (Figure 6). Comparison of Figures 2 and 6 also shows a crude inverse relation between muscovite percent and susceptibility and a generally positive relation between biotite percent and susceptibility. The latter effect is attributed to the fact that biotite and Fe-Ti oxides are commonly associated as adjacent grains.

The contrasting oxide mineralogy in the gneissic granite and granite is clearly reflected in the susceptibility log and in the magnetic susceptibility of core samples. The presence of hematite and pseudobrookite as the apparently sole oxide phases in the gneiss is the obvious explanation of the low susceptibility response

Fig. 6. Relation between magnetic susceptibility and density logs [Daniels et al., this issue] and laboratory measured values (solid dots), Illinois deep hole, UPH 3.

in the depth interval 673-731 m. Similarly, the higher response at greater depths in UPH 3 is attributed to the presence of magnetite as the main oxide phase in the granite. High concentration of magnetite is particularly evident in the depth interval 1127-1348 m. and at 1541 m. Both of these zones are regions of high magnetic susceptibility response.

Density. Bulk density measurements on core samples range from 2.595 to 2.685 Mg/m^3 . The denser rocks tend to have greater amounts of biotite and Fe-Ti oxides and lesser amounts of microcline perthite.

A plot of the density of core samples against the drill hole density log is shown on Figure 6. There is generally good correlation between the densities obtained from measurements on the core and in situ densities of the log to a depth of about 1070 m and at depths greater than roughly 1370 m. The lower density measurements in the intervening 300 m of the core generally correspond to the section of the hole which Daniels et al. [this issue] identify as a fractured zone on the basis of low resistivity values and low neutron well log responses. It is probable that these core densities which

were measured at atmospheric pressure are lower because of the effect of the opening of fractures within this segment of the hole.

Conclusions

Petrographic data indicate that two main rock types are present in the Illinois deep core holes. The main variety is a massive medium- to coarse-grained porphyritic granite. The other main type, which occurs in the upper part of the basement in hole UPH 3, is a fine-grained granoblastic to lepidoblastic gneissic granite that displays extensive recrystallization. Superimposed on both rock types are subtle to moderately developed cataclastic shear surfaces that are imprinted on all rocks found in the core holes. The shearing is accompanied by subtle partial recrystallization of the more susceptible mineral phases. Later, thin, crosscutting fracture planes occur throughout the core interval.

Mineralogical data in support of the presence of two different rock types in UPH 3 come from the contrasting Fe-Ti oxides in both rocks. Hematite and pseudobrookite are the oxide phases in the gneissic granite, whereas magnetite and ilmenite occur in the granite. In addition, the composition of the biotites is different in the granite and in the gneissic granite. The biotites in the gneissic rock have higher Fe/Fe + Mg ratio, FeO, and Al₂O₃ and lower MgO and SiO₂.

The gneissic texture, penetrative deformation, and extensive recrystallization in the gneissic granite suggest that it is the older of the two rock types. It possibly is an older intrusion or, less likely, part of the basement into which the massive granite intruded. An alternative interpretation of the two rocks is that the gneissic granite is a fine-grained variant of the coarse granite and that the gneissic fabric developed preferentially in the finer-grained rock. We favor the interpretation that the gneissic granite is the older rock. This view does not rule out the possibility that the two rocks are related, and their similar modal and major element compositions suggest that they may be cogenetic.

The granites have chemical affinities to rapakivi granites. They are relatively rich in silica, K₂O, Na₂O, and F and poor in Al₂O₃, FeO, Fe₂O₃, MgO, CaO, TiO₂, MnO, and P₂O₅ and have a high FeO/MgO ratio. They are anorogenic or A-type granites.

Magnetic susceptibility and density measurements correlate reasonably well with the magnetic susceptibility and density logs. Variations in these parameters are related at least in part to differences in mineral content.

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APPENDIX VI.3.

GEOLOGICAL SIGNIFICANCE OF SURFACE GRAVITY MEASUREMENTS IN THE
VICINITY OF THE ILLINOIS DEEP DRILL HOLES

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GEOLOGICAL SIGNIFICANCE OF SURFACE GRAVITY MEASUREMENTS
IN THE VICINITY OF THE ILLINOIS DEEP DRILL HOLES

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Abstract. A gravity study of northeastern Illinois and southwestern Wisconsin was undertaken to determine the three-dimensional geometry of the Precambrian granitic body encountered in the Illinois deep drill holes. Gravity data acquired during this study and existing observations show a marked negative gravity anomaly in the vicinity of these drill holes which is interpreted as originating from a negative density contrast between the granitic body and the intruded crystalline rocks. A fifteenth-order polynomial surface fitted to the Bouguer anomaly data (which includes only wavelengths greater than ~ 35 km) was chosen as best displaying the effects of the granitic body, and profiles of this surface were modeled employing a variety of techniques and assumptions. The preferred model of the causative body has a depth to the top of 2 km, a density contrast of -100 kg m^{-3} , and a maximum vertical extent of 5 km. An isopach map of the body constructed from the modeling places the Illinois deep drill holes on the edge of an oval-shaped mass.

Introduction

In 1980, the Commonwealth Edison Corporation completed the drilling of three closely spaced deep drill holes in northern Illinois in Stephenson County (Figure 1). The cores from these holes were made available to the scientific community and one hole (UPH 3) was made available for a variety of geological and geophysical studies. Two of the holes penetrated approximately 1 km of granitic rock which underlies roughly 700 m of Paleozoic sedimentary rocks. The three-dimensional geometry of the granitic body sampled by the drill holes is significant to the interpretation and results of several other investigations. Therefore a gravity study was conducted of the region surrounding the drill holes.

Granitic plutons are commonly less dense than the rock which they intrude. Thus the gravity method is widely used to determine their geometric configuration by analyzing the negative gravity anomaly they produce. A negative gravity anomaly associated with the granitic rocks encountered was isolated and modeled using the geological constraints derived from previous studies and studies of cores from the drill holes. Basement depths and rock types were taken from the published drill hole data of Thawites [1923, 1957], Lidiak et al. [1966], and Goldich et al. [1966]. The density values used in modeling were acquired from Hinze

[1959], Bott and Smithson [1967], and Lidiak and Denison [this issue]. The major end product of this study is an interpretative isopach map illustrating the three-dimensional geometry of the granitic body encountered in the deep drilling. This map was derived from inversion of a series of gravity profiles using a 2 1/2-dimensional formulation of the forward gravity interpretation problem.

Data Collection and Reduction

Existing gravity coverage in northern Illinois was adequate for the study [McGinnis et al., 1976], but additional gravity data were required in southern Wisconsin. Thus 1450 gravity stations were observed in southwestern Wisconsin in 1980 and were merged with the approximately 7500 previously available gravity stations. The resulting data base provided a station spacing of roughly 1.6 km in the area within a radius of 30 km of the drill holes and more regional coverage in adjacent areas of Wisconsin and Iowa. Therefore approximately 9000 gravity stations (Figure 2) were used to produce a simple Bouguer gravity anomaly map (Figure 3) which encompasses an area of approximately 55,000 km².

Gravity observations were made with two LaCoste-Romberg gravimeters using normal field procedures [Nettleton, 1940]. Observations were made at locations with established elevations as shown on U.S. Geological Survey topographic maps. The total error due to imprecision in elevation, horizontal position, and gravimeter drift is estimated to be less than $0.3 \times 10^{-5} \text{ m s}^{-2}$ (mGal). All observations were reduced to Bouguer anomaly values by standard procedures using a density of 2670 kg m^{-3} . No terrain corrections were needed because of the low surface relief of the area.

Data Processing

The simple Bouguer gravity anomaly map in Figure 3 was contoured by gridding observed values at a 4-km interval employing the method of minimum curvature [Briggs, 1974; Swain, 1976]. A variety of techniques were tested in order to isolate the portion of the gravity field associated with the pluton penetrated by the deep drill holes. In particular, extensive processing was done using least squares polynomial surfaces fitted to the data by employing an orthogonal polynomial routine developed by Lance [1982]. When employing the polynomial technique, regional trends can be represented by a low-order polynomial (i.e., low-pass filter) surface, while residuals with respect to such a surface represent a high-pass filtered map

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Fig. 2. Index map of gravity station locations (crosses).

Regional Gravity Field

The regional gravity field of Wisconsin has previously been discussed by Ervin and Hammer [1974] and the Illinois gravity anomaly map has been interpreted by McGinnis et al. [1976]. Generally, the Wisconsin Bouguer gravity anomaly field is dominated by a broad anomaly minimum which reaches values lower than -90×10^{-5}

m s^{-2} (mGal). In contrast, the Bouguer gravity anomaly field of Illinois increases to the south, reaching low positive values in the southernmost portion of the state.

A crude inverse relationship between gravity anomaly values and depths to the basement surface as well as a mosaic pattern of anomalies have often been noted in the midcontinent [Woollard, 1962, 1966; Hinze, 1963; McGinnis, 1966, 1970].

Fig. 3. Simple Bouguer gravity anomaly map of the study area. Datum, sea level; reduction density, 2670 kg m^{-3} , contour interval, $2 \times 10^{-5} \text{ m s}^{-2}$ (mGal).

The drill hole control available (Figure 1) indicates that the depth to the basement surface generally increases from north to south across the study area. However, the slope is gentle and is not a major source of regional anomalies in this study. A mosaic pattern of anomalies is observed (Figure 3), and the steep gradients present suggest an upper crustal source. However, there is no correlation between Bouguer anomalies and basement lithologies in the area, and there is little vari-

ation in the densities of the predominant basement lithologies (granite and quartzite [Hinze, 1959]). Thus major anomalies such as the low associated with the Illinois deep drill holes must be due to structural and/or petrologic variations within the deeper basement rocks which underlie the predominantly felsic upper portion of the basement.

The Illinois deep drill holes lie on a shallowing gravity anomaly trough extending from the southern margin of a broad gravity minimum which

Fig. 4. Fifteenth-order polynomial surface which was fitted to the simple Bouguer anomaly values. Contour interval, $2 \times 10^{-5} \text{ m s}^{-2}$ (mGal). Lines with bars on the end show the locations of the profiles that were modeled. The location of the Illinois deep drill holes is shown by a circle containing a cross.

covers central Wisconsin (Figures 3 and 4). A parallel but more subdued gravity minimum lies approximately 50 km to the east. This anomaly, which is terminated immediately south of the latitude of the Illinois deep drill holes, probably has an origin similar to the anomaly associated with the drill holes. The southern margin of the central Wisconsin minimum is marked by a steep

east-west trending gradient which has been interpreted by Hinze [1959] as abrupt northward thickening of the felsic uppermost basement rocks. This occurs in Wisconsin at roughly a latitude of $43^{\circ}20'N$ (Figure 3).

The Illinois deep drill holes lie on the southern margin of a band of relatively intense local positive magnetic anomalies which trends

northeast across Illinois from Iowa into northern Wisconsin [Coates et al., this issue]. The northwestern boundary of this zone of magnetic anomalies corresponds with a major change in the gravity anomaly pattern (Figure 3), extending from $43^{\circ}10'N$, $88^{\circ}30'W$ southwesterly into Iowa. Southeast of this line, the source of the anomalies appears deeper because the anomalies are more subdued and the horizontal gradients decrease. This boundary marks a fundamental break in the basement rocks which is not manifested in a recognizable way within the Phanerozoic sedimentary rocks. It is significant to note that the gravity minimum which is identified with the granitic rocks of the Illinois deep drill holes crosses this boundary and shows a similar change in character, suggesting that the boundary postdates the intrusion.

Source of Gravity Minimum

The principal objective of this study was to determine the three-dimensional configuration of the source of the gravity anomaly minimum which is related to the granitic rocks encountered by the Illinois deep drill holes. However, with the limited geological control available, the results of modeling are subject to considerable ambiguity. In an effort to minimize this ambiguity, an east-west gravity profile through the drill holes was selected for initial modeling using varied inversion techniques while using differing geological assumptions. In order to remove the effects of local density variations, the smooth profiles obtained from the fifteenth-order polynomial surface which approximates the observed gravity anomaly values were used in all the inversions.

A modified version of the two-dimensional gravity modeling technique employing a finite strike length (2 1/2 dimensional) source [Pederson, 1977] was used in a generalized linear inverse scheme [Cribb, 1976; Lai, 1980] to study the effect of varying depth to the top of the anomalous source and the homogeneous density contrast. An east-west profile passing through the Illinois drill holes was modeled. The results are shown in Figure 6 and indicate that varying the depth to the source from 1 to 4 km by 1-km increments has little effect upon the shape of the two-dimensional source which best fits the observed gravity anomaly data. Within the context of the known geology, the depth of the anomalous source is the total thickness of the sedimentary rocks present plus the uppermost, felsic basement rocks which

Fig. 5. Comparisons of observed gravity with profiles across polynomial surfaces of various orders. Location of this E-W profile is $42^{\circ}30'N$ (Figure 4).

Fig. 6. Gravity models depicting the results of varying the inversion formulation as well as the density contrast and depth to the top of the body. All models are for the profile at $42^{\circ}30'N$. (a) Results of the Backus-Gilbert inversion assuming a density contrast of -100 kg m^{-3} . (b) Results of the generalized linear inverse technique employing varying depths to the top of the body. MSQERR denotes the mean squared error between the observed and calculated values. (c) Results of the generalized linear inverse technique employing density contrasts of -100 kg m^{-3} and -50 kg m^{-3} . In Figures 6b and 6c, the calculated values are essentially identical for the various models, so only one curve is shown.

Figure 7. Panel diagram showing the results of the generalized linear inversion technique when applied to the profiles shown in Figure 4. A depth to the top of the body of 2 km and a density contrast of -100 kg m^{-3} were assumed.

have little or no density contrast with the intruding granites. A density contrast of -100 kg m^{-3} was assumed in this calculation based upon a mean density of approximately 2650 kg m^{-3} for the granitic rocks of drill hole UPH 3 [Daniels et al., this issue; Lidiak and Denison, this issue] and an assumed density of roughly 2750 kg m^{-3} for the metamorphic rocks underlying the upper basement, felsic layer. This latter density value is poorly constrained. Thus the effect of varying the density was also investigated by using homogeneous density contrasts of -50 and -100 kg m^{-3} . The results show the profound effect that the density contrast has upon the extent of the source. As a cross-check, a final inversion was performed employing a Backus-Gilbert inversion, using a com-

puter program written by Hong [1981]. The results shown in Figure 6, based on a horizontal upper surface and a density contrast of -100 kg m^{-3} , are in excellent agreement with the edges and bottom of the source computed from the generalized linear inverse scheme using the same assumptions.

A total of seven east-west profiles and one north-south profile along $89^{\circ}45'W$ longitude were modeled (Figure 4), using the generalized linear inverse scheme employing strike lengths appropriate for the positions of the profiles relative to the gravity minimum. The upper surface of the source was assumed to be horizontal and at a depth of 2 km, and the source was assigned a density contrast of -100 kg m^{-3} with the adjacent rocks. The model studies and available geological data suggest that these assumptions are at least reasonable. Their geological consequence is that the source of the anomaly minimum is interpreted to be due to thickening (with respect to its regional thickness of approximately 1 km) of the felsic layer comprising the uppermost portion of the basement and/or intrusion of granitic rocks into the underlying layer of metamorphic rock which has a mean density of roughly 2750 kg m^{-3} .

Figure 7 is a panel diagram showing the models derived by inversion of the eight profiles. An isopach map of the source body (granitic pluton) was constructed from the modeling results and indicates that the Illinois deep drill holes are located near an oval-shaped area where the source body is approximately 5 km thick (Figure 8). This area is interpreted as the approximate outline of the granitic mass encountered in the drill holes. In a similar manner, the other oval-shaped areas where the source body is thickened probably represent granitic intrusives although structural control cannot be dismissed. In particular, faulting is suspected where sharp horizontal gradients are present. The results of this analysis are necessarily ambiguous because of the limited geological constraints, particularly in regard to the base-

Fig. 8. Isopach map of the body delineated by this study. This map is based on extrapolations of the results shown in Figure 7. Contour interval is 0.5 km.

ment country rocks which surround the granitic rocks of the drill holes. However, the gravity studies indicate that the thickness of the granitic rocks probably does not exceed 5 km. This is in agreement with heat flow studies and measurements of heat generation in the granitic rocks which suggest that a thickness of approximately 3 km for these rocks in the vicinity of the Illinois deep drill holes would account for the excess heat flow observed (Rahman and Roy, 1981)

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