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Temperatures, heat flow, and water chemistry from drill holes in the Raft River geothermal system, Cassia County, Idaho

by

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ABSTRACT

The Raft River area of Idaho contains a geothermal system of intermediate temperatures (≈150°C) at depths of about 1.5 km. Outside of the geothermal area, temperature measurements in three intermediate-depth drill holes (200-400 m) and one deep well (1500 m) indicate that the regional conductive heat flow is about 2.5 μcal/cm² sec or slightly higher and that temperature gradients range from 50° to 60°C/km in the sediments, tuffs, and volcanic debris that fill the valley. Within and close to the geothermal system, temperature gradients in intermediate-depth drill holes (100-350 m) range from 120° to more than 600°C/km, the latter value found close to an artesian hot well that was once a hot spring. Temperatures measured in three deep wells (1-2 km) within the geothermal area indicate that two wells are in or near an active upflow zone, whereas one well shows a temperature reversal. Assuming that the upflow is fault controlled, the flow is estimated to be 6 liter/sec per kilometer of fault length. From shut-in pressure data and the estimated flow, the permeability times thickness of the fault is calculated to be 2.4 darcy m.

Chemical analyses of water samples from old flowing wells, recently completed intermediate-depth drill holes, and deep wells show a confused pattern. Geothermometer temperatures of shallow samples suggest significant re-equilibration at temperatures below those found in the deep wells. Silica geothermometer temperatures of water samples from the deep wells are in reasonable agreement with measured temperatures, whereas Na-K-Ca temperatures are significantly higher than measured temperatures. The chemical characteristics of the water, as indicated by chloride concentration, are extremely variable in shallow and deep samples. Chloride concentrations of the deep samples range from 580 to 2200 mg/kg.

INTRODUCTION

The Raft River geothermal area in Idaho is under investigation by the Idaho National Engineering Laboratory as a site for demonstrating the generation of electricity from an intermediate-temperature reservoir. The purpose of this paper is to report temperature measurements and chemistry of waters from shallow and intermediate-depth drill holes and deep wells in order to describe the characteristics of the geothermal system and its setting in the regional heat-flow regime. The locations of drill holes for which temperature and chemical data have been obtained are shown in Figure 1, and a more detailed map for the area around the deep geothermal wells is shown in Figure 5 (boundary shown on Figure 1). The regional heat-flow setting is discussed first using data from drill holes that appear to be beyond the influence of the geothermal system. Data for drill holes near and within the geothermal system are presented next, organized in order of presentation by hole depth. The water-chemistry data are discussed after the temperature data for the intermediate-depth drillholes.

The drill hole and well locations are shown on the generalized topographic map (Figure 1). The area is in the northern part of the Basin and Range physiographic province. Basalt associated with the Snake River Plain province outcrops about 15 km to the north. The Jim Sage Mountains (center, Figure 1) are made up of Tertiary rhyolites and tuffaceous sediments that define a broken antiform structure (Williams et al, 1976). The Albion Mountains to the west and the Raft River Mountains to the south expose Precambrian adamellite (quartz monzonite) mantled by Precambrian and lower Paleozoic metasedimentary rocks and by allochthonous upper Paleozoic sedimentary rocks. The narrow fault-defined valley separating the Albion and Jim Sage Mountains (Williams et al, 1976) rises about 600 feet (180 m) from its low point in the south (Upper Raft River Valley) to its saddle near Elba. To the east, the Black Pine Mountains consist mainly of faulted Pennsylvanian and Permian sedimentary rocks. The valley fill is composed of sand, gravel, silt, and clay of Pleistocene age and tuffaceous sandstone, siltstone, and conglomerate of Tertiary age (Williams et al, 1976).

Figure 2 shows the same area as Figure 1, along with the faults (Williams et al, 1976) and gravity contours (Mabey and Wilson, 1973). Both the Raft River Valley and Upper Raft River Valley have associated gravity lows. A gravity high (-160-mGal contour, Figure 2) centered on drill-hole I.D. 5 does not conform to the topography (Figure 1). This gravity high trends northwest across the Jim Sage Mountains and may be an extension of the Big Bertha gneiss dome of the Albion Mountains (Williams et al, 1976). The Narrows structure (shown on Figure 2 as a set of double broken lines) trends from the southwest to the northeast (Williams et al, 1976; Mabey et al, 1978). It is easily visible in the gravity contours in the area south of the Jim Sage Mountains, but its northeast and southwest extensions are not as apparent. Additional geophysical information is given in Mabey et al (1978), Ackermann (1979), and Keys et al (1979).

THERMAL DATA

The temperature measurements reported here were made using four-conductor cables with thermistors as sensors and digital multimeters as detectors. The relative accuracy is better than 0.002°C, and the absolute accuracy about 0.02°C. Observations were made at discrete depths on some occasions and continuously (300 m/h) on others. Some details concerning instrumentation are given by Sass et al (1971). Wells with positive pressure were logged by lowering the cable through a packing gland mounted on top of a standpipe. Thermal conductivities have been measured using the needle-probe technique by drilling a hole into core samples that were waxed in the field (Sass et al, 1971).

REGIONAL HEAT FLOW

Although only one drill hole (I.D. 5) in the Raft River area was designed for regional heat-flow determination, several wells and drill holes spread over an area of several hundred square kilometers provide corroborating data. Figure 3 shows a temperature profile and thermal conductivities for heat-flow hole I.D. 5, cased and cemented to total depth. Conductivities measured using the needle-probe method in cores average 4.7 mcal/cm °C sec from 76 to 128 m and 5.7 mcal/cm °C sec from 140 to 216 m. With measured gradients of 63°C/km in the upper zone from 76 to 128 m and 45°C/km from 140 m to the bottom, the heat flows are 3.0 (76-128 m) and 2.6 $\mu cal/cm^2$ sec (140-216 m). The average heat flow is 2.8, and the value corrected for three-dimensional terrain is 2.7 $\mu cal/cm^2$ sec. Similar values of 2.2 and 3.1 have been obtained at Mahogany and Murphy south of the Snake River Plain but in western Idaho (Urban and Diment, 1975). The value of 2.7 is greater than the value of 2.5 used by Lachenbruch and Sass (1977) to define the Battle Mountain high (Sass et al, 1971), but it is within one standard deviation of their average for the Basin and Range province (2.1 $\pm$ 0.71) excluding the Battle Mountain high and Eureka low.

The regional significance of the heat-flow measurement is confirmed by temperature measurements in the Strevell well and the Almo 2 drill hole shown in Figure 3 and in the Griffith-Wight well shown in Figure 4. Lithologic data for the Strevell well are given by Oriel et al (1978). Temperature gradients in the Strevell well are rather consistent with depth and average 56°C/km. Almo 2 shows a disturbed zone from 145 to 183 m, but a log obtained before the well was cleaned out and pumped shows no such disturbance. Almo 2 has a gradient of 52°C/km.

Figure 4 shows three temperature logs of the Griffith-Wight well. Lithologic and geophysical logs are available in Oriel et al (1978). The wellhead pressure is 3.2 bars gauge as measured 1.2 m above ground level, and this well is frequently flowed during the winter to prevent the valve from freezing. The log of 20 October 1975 was obtained with the well shut-in, although there could be a disturbance caused by previous episodes of flow. The log of 9 August 1976 was obtained after a flow of 10 mL/sec was measured. Birch (1947) and Boldizar (1958) analyzed the distribution of temperature in a flowing well. From measurements of the flow and the distribution of temperature T during flow, the original earth temperature T_g as a function of depth can be obtained from the formulas

$$T_g = T - A \frac{dT}{dz} \tag{1a}$$

$$A = M\ c\ f(t_d) / \ 2 \pi \ k_m \tag{1b}$$

where M is the mass flow of water, c is the specific heat of water, $f(t_d)$ is a function of dimensionless time for a cylindrical source of heat or temperature, and k_m is the thermal conductivity of the rock/water mixture. If we use flow rate of 10 mL/sec obtained before the 9 August 1976 temperature log in equations (1) together with a thermal conductivity of 5 mcal/cm $^{\circ}C$ sec, we obtain a difference between measured temperature and original earth temperature of about 1°C below 200 m. Clearly this is much too small. The 1975 temperature log is almost 5°C cooler than the 9 August 1976 log. The discrepancy is most likely caused by a disturbance in the temperatures that remains from previous episodes of flow. Both shallow temperature logs have a conspicuous break in slope near the depth of the casing at 193 m. This change in gradient can be explained if we assume that both shallow temperature logs reflect decaying temperature disturbances caused by previous episodes of flow and that the value of A in equation (1a) depends on whether flow is in a cased or uncased part of the hole. At low flows, the theory of equations (1) predicts a constant offset in temperature between that measured in the well and the true ground temperature (at some distance above the point of water entry). Below 200 m the gradient in the 9 August 1976 temperature log is 53°C/km, and in the 20 October 1975 log it is 55°C/km. The near constancy of the gradient below 200 m seems to indicate that the temperatures are simply offset by a constant amount from the true ground temperatures. The log of 18 December 1976 (Figure 4) was obtained while intermittently flowing the well to get past zones where the temperature probe stuck, and this is the reason for the various breaks. Below 1320 m the gradient is quite linear at 52°C/km, a value that agrees closely with that obtained from the shallow logs. A projection of the deep temperatures to the surface gives an intercept temperature of 18°C, significantly higher than the value of about 11°C one would expect on the basis of data from other drill holes. This result may indicate that some small waterflow occurs even at the greatest depth logged. (The well was originally drilled to 2068 m, much deeper than the logged depth of 1489 m.) Another explanation is that the gradient is not so uniform as the comparison between the shallow and deep data would indicate. No conductivity data are available for the Strevell, Almo 2, and Griffith-Wight drill holes, however a value of 5 mcal/cm ^{o}C sec would give heat flows compatible with that measured in I.D. 5.

SHALLOW AUGER HOLES

Several dozen shallow auger holes were drilled to depths of as much as 30 m at the Raft River area in 1974 for hydrologic investigations (Crosthwaite, 1974). Plastic pipe was placed in the holes, and the annulus back-filled with cuttings. Most of these holes were sited along a linear trend from the Schmitt hot well to The Narrows; the locations of several of these holes are shown in Figure 5 along with elevation contours and the county dirt road. Because the completion technique of the auger holes may allow water to flow between different horizons and because of the shallow depth of the holes, some of the temperature measurements are of limited usefulness. Comparison of two sets of data obtained in winter and summer of 1976 shows that the form of the temperature-depth profile in the upper 7 m is determined by decay of the annual wave of surface temperature whereas this perturbation is small at a depth of 10 m. For a thermal conductivity of 2.5 mcal/^{o}C cm sec and a volumetric specific heat of 0.6 cal/cm^3 ^{o}C, the annual temperature wave attenuates to 3 percent of its surface value at a depth of 7 m and to 1 percent at a depth of 9 m (Carslaw and Jaeger, 1959, p. 66). These values agree well the with measured differences between winter and summer temperatures. Because of the large effect of the annual wave on the form of the temperature profile above 7 m, we present data only for holes that are deeper than 7 m.

Figure 6 shows the temperature profiles for holes marked "I" in Figure 5 and of A.H. 13A and I.D. 4 further to the southwest. The temperature reversal in A.H. 13-N indicates a horizontal flow of hot water above a flow of colder water. The colder temperatures in A.H. 13A, compared to A.H. 13-N to the north and I.D. 4 to the south, indicate that there are separate flows of hot water near the surface. I.D. 4 is near The Narrows spring, in which a temperature of 38^{o}C has been measured.

The three auger holes A.H. 6A, 11A, and 13-N all seem to be situated within the thermal infrared anomaly of Watson (1974). For a thermal conductivity of 2.5 mcal/cm sec ^{o}C and near-surface gradients ranging from 3.5^{o} to 7.0^{o}C/m in these holes, the conductive heat flow is from 90 to 180 μcal/cm^{2}sec. Although the data are insufficient to establish a boundary of the thermal infrared anomaly in terms of a value of measured heat flow, the appropriate value would seem to be less than 100 μcal/cm^{2}sec.

Temperature profiles of the auger holes marked "II" on Figure 5 are shown on Figure 7, along with profiles of A.H. 1-S and A.H. 6 further to the northeast. Maximum temperatures measured in the group II auger holes are cooler than those in group I. Only A.H. 7-S shows a clear reversal, although the other auger holes have gradients that decrease with depth. The temperature profiles of these holes indicate pervasive movement of hot water at shallow depths, but the only known occurrence of hot water at the surface is near I.D. 4. The three auger holes to the north of the road (A.H. 5A, 3A, and 9A) are all much cooler than the nearby auger holes to the south of the road (A.H. 11A, 7A, and 8A). This would seem to rule out flow of hot water from under the mountains at the locations with pairs of hot and cold auger holes. The pattern of temperatures decreasing toward the northeast from A.H. 7-S to 7A to 8A in group II could be interpreted to indicate flow in the direction of decreasing temperatures. The locations of the auger holes essentially along a single line and their shallow depths preclude any definitive statement as to the direction of flow. The data from the auger holes show that the quantity of hot water flowing in the near surface is significantly greater than that indicated by flow from the hot spring near drill hole I.D. 4.

SHALLOW- AND INTERMEDIATE-DEPTH TEMPERATURES

Additional data on shallow- and intermediate-depth thermal regime within the geothermal system come from two sources: two artesian hot wells, and four coreholes that have been drilled in the area of these hot wells for geothermal information.

The Crank well (Figure 5) is 165 m deep and produces 93°C water at the surface; the Schmitt well is 126 m deep and produces 90°C water at the surface. Stearns et al (1938, p. 170) state "Before the [Schmitt] well was drilled there was a warm moist spot of ground at this place stained with spring deposits." High temperatures at shallow depths in the two flowing artesian wells indicate significant vertical flow of hot water. Another drill hole SMHW was recently placed about 20 m from the Schmitt well; measured temperatures are plotted in Figure 8.

Other drill holes in the area are I.D. 1, 2, and 3 (Figure 5); basic lithologic and geophysical data are given in Crosthwaite (1976). Temperature logs and thermal conductivities are shown in Figure 8, along with the temperature log of I.D. 5A, drilled 30 m away from I.D. 5 (Figure 1) but to a greater depth. These drill holes (except for I.D. 5A) have only been partly cased and cemented, and so some variations in gradient undoubtedly reflect water movement within the holes rather than original ground temperatures. A representative gradient for I.D. 1 is 120°C/km, and measured thermal conductivities average 2.6 mcal/cm °C sec. The heat flow is thus about 3 $\mu cal/cm^2 sec$, only slightly above the value measured in I.D. 5 outside the geothermal area; thus I.D. 1 may reflect one edge of the geothermal anomaly. Drill hole I.D. 2 has a gradient of 210°C/km and thermal conductivities average 3.1 mcal/cm °C sec, so the heat flow is 6.5 μcal/cm^2 sec. Drill hole I.D. 3 has an average gradient of over 200°C/km and an average thermal conductivity of 4, so the conductive heat flow in I.D. 3 is higher than in I.D. 2. The high gradients in I.D. 2 and 3 reflect shallow movement of hot water and indicate that the movement of hot water is more pervasive than is evidenced by the surface discharge.

The model that emerges from the temperature data can be summarized as follows: Hot water from a geothermal reservoir is leaking to the surface at three known places: near The Narrows, at the Schmitt well, and at the Crank well. Two of these flows are most likely structurally controlled; a fault lies near the Schmitt well, and the hot spring near I.D. 4 is in The Narrows structure. In addition to these flows of hot water that reach the surface, the drill holes indicate that near-surface aquifers are being charged by hot water. These near-surface flows of hot water have caused significant hydrothermal alteration (Keys and Sullivan, 1979, Ackermann, 1979). This model is both clarified and confused by the water-chemistry data.

GEOCHEMISTRY

Two objectives in looking at the chemistry of waters at Raft River are: 1) to determine if the geothermometer temperatures obtained from near-surface samples agree with the temperatures measured in the deep wells, and 2) to see how the water in the near-surface flows relates to the deep water; i.e., is the water in the near-surface cooled by conductive heat loss or by mixing with cold water? To a large extent, the water-chemistry data are not illuminating for these two objectives. The geothermometer temperatures obtained from shallow samples are not a good predictor of deep temperatures. Furthermore, geothermometer temperatures of waters from the deep wells also are not in very good agreement with measured temperatures. These data are not useful for relating shallow and deep waters, because the chemistry of water samples from the deep wells show a large variation from well to well.

Table 1 lists chemical analyses and geothermometer temperatures of waters from the different wells and drill holes. These data are summarized in Figure 9, which shows silica concentrations versus chloride concentrations for the various waters. Temperatures obtained from the Na-K-Ca geothermometer of Fournier and Truesdell (1973), using the magnesium correction of Fournier and Potter (1979) where necessary, are shown in parentheses at each data point. Silica concentrations are dependant on temperature (Fournier and Rowe, 1966) in a nonlinear fashion. Horizontal lines are drawn at silica concentrations corresponding to quartz geothermometer temperatures of 140°, 150°, and 160°C. In those cases where the quartz and Na-K-Ca geothermometer temperatures agree within a few degrees Celsius, the Na-K-Ca temperature is marked by an asterisk following it in Figure 9. The reason for plotting silica versus chloride in Figure 9 is that the mixing of cold and hot waters and the loss of silica caused by conductive cooling are easily depicted. Chloride is normally assumed to be a conservative constituent in geothermal waters. During mixing of cold with hot waters, the cold water normally contains low chloride and silica. If no re-equilibration occurs, silica is conserved during mixing, and chloride and silica concentrations should be linearly related if only a single source of geothermal water exists. The geothermometer based on Na-K-Ca involves the ratios of these components, so it is less affected by dilution. Silica may be lost from a geothermal water if the water flows so slowly to the surface that it is able to cool conductively, and thus the chloride concentration is unaffected.

Geothermometers generally are most accurate for flowing springs, and so the data for The Narrows spring and the two wells that have been flowing for a long time are discussed as a group. The Schmitt well at one time was a hot spring, and the Crank well has such high temperatures at shallow depth that it too may have been a hot spring. The Schmitt well has a Na-K-Ca temperature that agrees closely with the measured reservoir temperature of 140° to 150°C. The silica temperature of 126°C could be explained either by mixing or silica loss (Figure 9) during upflow. For The Narrows spring and the Crank well, the silica temperature agrees with the Na-K-Ca temperature, but both temperatures are significantly below the measured reservoir temperatures. These waters may have re-equilibrated. The low chloride and high magnesium contents in The Narrows spring compared to the Schmitt well indicate that waters in The Narrows spring probably mixed before re-equilibrating. The data for the Schmitt well and The Narrows spring can be combined to give a common parent water if we assume that the Schmitt well lost silica while cooling during flow to the surface and that The Narrows spring is a mixed water (Figure 9). The data from the Crank well are not compatible with this picture because the chloride content of its water is much too high for it to have the same parent hot-water with a single value of chloride and enthalpy as the Schmitt well and The Narrows spring.

The water sample from I.D. 3 indicates equilibrium because its silica temperature of 108°C is essentially the same as the Na-K-Ca temperature of 103°C. The maximum measured temperature in the drill hole is 89°C, in reasonable agreement with the geothermometer temperatures. Silica and Na-K-Ca (Mg corrected) temperatures for drill-hole I.D. 2 are 131° and 133°C, respectively, but these geothermometer temperatures are significantly higher than the maximum measured temperature of 54°C. The chloride concentration for I.D. 2 is much lower than that found in its near neighbors Crank and I.D. 3. The sample from I.D. 1 is somewhat strange; the maximum measured temperature is 39°C, and yet the Na-K-Ca temperature is 220°C. The silica and chloride contents are high, although the drill hole seems to be at the edge of the geothermal system.

That the deep system is also confusing is reflected in the water samples from deep wells RRGE- 1, 2, and 3 (Figure 9). Chloride contents vary considerably, whereas Na-K-Ca temperatures are similar for the three wells but significantly above any of the measured temperatures. Silica concentrations in RRGE- 3 are appropriate to the measured temperature, but silica temperatures in RRGE- 1 and 2 are a bit high. The waters in the deep wells are quite variable in composition and do not indicate a single source of water of unique chloride composition and temperature. Because the chloride concentration of the deep water varies so much, the near-surface waters can be related to the deep waters by more than one process. For example, the composition of water in the Schmitt well (Figure 9) can be obtained from that of RRGE- 1 water by silica loss, or by mixing from a parent water of a composition between those of RRGE- 1 and RRGE- 3 waters. However, the temperature data for RRGE- 1, discussed in the next section, indicate that conductive cooling is the most likely explanation.

Three alternative hypotheses might explain the chemistry of the water in the deep wells. The first is that a deep hot water of low chloride concentration picks up chemicals as it passes through a zone of easily dissolved material. The second is that the Raft River system was much hotter in the past; as temperatures have decreased, the ability of the circulating water to dissolve constituents from rocks at great depth has likewise decreased. This circumstance would explain the high geothermometer temperatures in the deep wells as a relic of the past, reflecting the inability of these waters to re-equilibrate. The third is that there are two waters of similar enthalpy but with different amounts of dissolved chemicals (Kunze et al, 1977, Allen et al, 1979). The various waters could thus be produced by mixing. Allen et al (1979) have used lithium, strontium, and fluoride concentrations to argue that there must be mixing of two deep waters at Raft River. Because lithium and strontium increase with chloride concentration, their contents could be explained either by the mixing of two deep waters or by the addition of salt. Fluoride decreases with increasing chloride concentration, and so its behavior would support only the mixing of two waters. However, calculations of fluoride and calcium activities from measured concentrations, using the method of Truesdell and Jones (1974), show that the fluoride concentrations are determined by the solubility of fluorite and not by a mixing relation. Thus, the available data can be explained either by the mixing of two deep waters or by the addition of salt.

THE PROBLEM OF ALMO 1

The Almo 1 drill hole provides a set of data that may indicate an extension of the geothermal system to the Upper Raft River Valley or a totally distinct system. Almo 1 is situated in the Upper Raft River Valley on the west side of the Jim Sage Mountains (Figure 1). A temperature log is given in Figure 10 and a chemical analysis of the water in Table 1 with a data point plotted on Figure 9. At least some of the temperature pattern shown in Figure 10 reflects vertical water movement in the drill hole. Before logging, about 3 mL/sec was flowing from the valve at the surface. Using the theory from equations (1), we can estimate the magnitude of the temperature error. For a thermal conductivity of 3 mcal/cm sec ^{o}C and an $f(t_d)$ value of 6, the value of A in equation (1a) is 9.5 m for a flow of 3 mL/sec. The temperature gradient in the upper 30 m is about 1.3^{o}C/m, and so the measured temperatures should be high by about 12^{o}C. The projection of measured temperatures to the surface gives an intercept of 28^{o}C, about 18^{o}C above the actual mean annual ground temperature; thus the error from flow up the well estimated by equations (1) is the correct order of magnitude. Because the flow history of the well is unknown, the failure of equations (1) to predict the correct surface temperature exactly is not surprising. Even though the temperature pattern of Almo 1 is not indicative of original ground temperatures, the temperature in the well is still over 70^{o}C at 100 m. Although, the geothermometry of the well indicates temperatures of 140^{o} and 143^{o}C, the chloride content is only 76 mg/kg. The quartz geothermometer may be significantly in error because of the high pH of the water (Fournier, 1973). The failure of the anions and cations to balance on analyses in different laboratories of separately collected waters may indicate that the Na-K-Ca geothermometer is also in error. Because the temperatures are still anomalous, the geothermometers may actually be accurate. Whether the Almo 1 well reflects a continuation of the Raft River geothermal system or a separate system involving deeply circulating groundwater will remain undetermined until further data have been obtained.

DEEP WELLS

Temperatures measured in three of the deep wells drilled at Raft River for production of geothermal fluids (RRGE- 1, 2, and 3) are shown on Figure 11, with the well locations on Figures 1 and 5. Physical data concerning the wells are listed in Table 2. These wells have had times ranging from 3 to 6 months for temperature recovery from the drilling disturbance but have also been disturbed by production since drilling was completed. One measure of the degree of nonequilibrium is that the mean annual ground temperature in the area of the deep wells is 10^{o} to 11^{o}C, while the near surface temperatures measured in the wells range from 20^{o} to 27^{o}C. During drilling, lost-circulation zones were encountered in each of the three deep wells. In RRGE- 1 an especially large zone of lost circulation was encountered at approximately 460 m. Adding up the volumes noted in the driller's log, about 5 million liters were lost. The temperature reversal at that depth in RRGE- 1 is the remaining disturbance after the zone of lost circulation was cased off 8 months before logging. In addition to disturbances from drilling and production, well RRGE- 2 shows the effects of injection. The three temperature reversals below the cased depth are interpreted to be perturbations remaining from injection (Stoker et al, 1977). After RRGE- 3 was drilled to 1784 m, two additional holes were drilled by sidetracking below the casing. Since the three legs have rather different flow properties (Covington, 1977c), water may be flowing up one leg and down another, and this may explain the sharp break in gradient at around 1310 m.

The general shapes of the temperature profiles of RRGE- 1 and 2 show curvature with the gradient uniformly decreasing with depth. Several interpretations of these data are possible. A horizontal flow of hot water throughout the entire thickness of the wells could cause the curvature shown. However, discharge at the surface from the Schmitt well suggests that vertical flow is more important. The vertical flow can be interpreted as an upwelling of hot water over a broad area as in the model of Bredehoeft and Papadopulos (1965); however, the two wells are sited along a fault. RRGE- 1 is interpreted to have intersected this fault at depth (Williams et al, 1976) while RRGE- 2 did not (Covington, 1977b). Consistent with the geologic interpretation that the two wells are sited along a fault, we can analyze their temperature profiles by assuming that flow is restricted to a thin zone. Nathenson et al (1979) presented an approximate solution for the temperature pattern caused by the flow of hot water up a fault zone and applied this analysis to the temperatures measured in RRGE- 1 to estimate an upward flow of 6 L/sec per kilometer of fault length. The smaller curvature in RRGE- 2 could be interpreted to reflect either a smaller vertical flow or that RRGE- 2 is located a little farther from the fault. For simplicity, we assume that RRGE- 2 lies farther from the fault. The flow rate can be used to estimate a permeability-thickness product k h for the fault by rewriting Darcy's law in the form

$$Q = -\frac{k\ h}{\mu}\left(\frac{dp}{dz} - \rho g\right) \qquad (2)$$

where Q is the flow per unit fault length, and the term in parenthesis is the pressure gradient excess above hydrostatic. RRGE- 1 has an overpressure of 9 bars at a reservoir depth of 1200 m, and so an average value for the term in parenthesis is 0.008 bar/m. Substituting the flow value and an average viscosity of 0.3 cp over the temperature range 25° to 150°C into equation (2), we obtain 2.4 darcy m. For comparison, Witherspoon et al (1978) found from an interference test between RRGE- 1 and 2 a value for k h of 69 darcy m for horizontal fluid flow. These values are not directly comparable as a measure of permeability, because the thickness of the aquifer is likely to be differ significantly from that of the fault. The difference between the vertical and horizontal permeability-thickness products indicates that the propensity towards horizontal flow is much greater than towards vertical flow. However, the high overpressure at depth causes vertical flow to dominate in the natural system.

The temperature profile of well RRGE- 3 differs significantly from those of RRGE- 1 and 2. Temperatures in RRGE- 3 show a reversal from around 570 to 1150 m, and the thickness of the zone defined by this reversal is large enough that it cannot easily be explained by lost circulation, although the absence of repeated logs makes any conclusion tentative. This reversal can be explained by horizontal flows of hot water above cold water or by a transient caused by a flow of hot water starting in the recent past. No other wells are available to give a clue to the possible direction of a horizontal flow of hot water, and so further interpretation must await more deep-drilling data.

CONCLUSIONS

Temperatures measured in wells and drill holes, and chemical analyses of water samples show that the geothermal system at Raft River is quite complex. Wide variations in the composition of the waters indicate that no unique value of chloride is associated with the deep geothermal water. The appearance of flowing hot water at the surface at The Narrows, in the Schmitt hot well, and in the Crank hot well indicates active upflow. Temperature profiles of the deep wells RRGE-1 and 2 indicate active upflow from depths of more than a kilometer. Shallow drill holes at The Narrows and temperatures measured in RRGE- 3 show reversals indicating that vertical flows of hot water also charge near-surface aquifers with subsequent horizontal flow. The data are insufficient to calculate the total anomalous heat flow from the system accurately. From the calculated vertical flow of water in RRGE- 1 of 6 L/sec per km of fault length and the distance to RRGE- 2 of about 1 km, we obtain a minimum estimate of flow of 6 L/sec. Because the other heat-flow anomalies at The Narrows and at the Crank well are likely to be of the same order of magnitude, the total convective flow is likely to be about 20 L/sec. This flow rate corresponds to a convective heat flow of 2×10^6 cal/sec, a value that falls toward the low end of heat flows measured for other systems in the Basin and Range province (Olmsted et al, 1975). If this flow is to be maintained by a steady-state gathering of 1 $\mu cal/cm^2 sec$ from the regional heat flow, 200 km^2 of area would be required.

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Table 1. Water chemistry of selected wells and springs in southern Raft River Valley. All analyses in mg/kg.

Well or drill hole	Collection date	Lab No.	Depth interval (m)	Temperature (°C)	pH	SiO_2	Ca	Mg	Na	K	Li	HCO_3	SO_4	Cl	F	T_{quartz}[4] conductive (°C)	$T_{Na-K-Ca}$[5] 4/3 (°C)	$T_{Na-K-Ca}$[5] 1/3 (°C)	$T_{Na-K-Ca}$[6] Mg cor. (°C)	
I.D. 1	12/05/74[2]	RR-69	86-336	26	7.8	88	300	1.4	2000	270	1.3	58	45	3900	3.9	131	221	220	--	--
RRGE-2	11/06/76[1]	T-76-1	1288-1994	62	8.0	149	32	0.12	378	35	1.0	77	61	578	9.3	161	158	187	--	--
Schmitt	10/06/76[1]	T-76-4	?-126	90	7.6	80	47.7	0.16	545	28	1.38	79	63	833	7.0	126	141	159	--	--
RRGE-1	10/06/76[1]	T-76-2	1104-1521	--	7.9	137	44.5	0.08	451	40	1.57	69	66	748	7.3	156	157	185	--	--
I.D. 3	12/06/74[2]	RR-72	60-434	82	8.1	56	56	0.5	1300	14	1.8	63	52	2000	5.0	108	119	103	--	--
Crank	7/13/74[1]	T-74-4	45-165	93	6.7	87.4	130	0.37	1180	33	2.4	122	60	1850	5.7	130	130	135	--	--
RRGE-3	10/06/76[1]	T-76-3	1297-1803	--	7.6	123	194	0.28	1260	115	2.73	95	61	2200	4.7	149	178	192	--	--
I.D. 2	1/14/75[2]	RR-16	197-198	30	7.7	88	35	3.9	370	34	0.64	176	32	570	2.8	131	154	185	122	133
Spring at The Narrows	9/14/74[2]	RR-60	surface	27	---	68	56	5.8	260	15	---	123	41	430	4.6	117	101	151	101	112
I.D. 4	3/28/75[2]	RR-14	19-77	40	6.8	37	58	9.0	240	13	0.68	138	44	380	4.4	89	.94	147	91	89
Almo 1	10/07/76[1,3]	T-76-5	79-150	60	10.0	104	5.5	0.04	115	5	0.20	160	57	76	7.3	140	108	143	--	--

1. Collector: A. H. Truesdell; analyst: J. M. Thompson.

2. Analyst: U.S. Geological Survey laboratory, Salt Lake City, Utah (Crosthwaite, 1976).

3. Analyses of separately collected samples from Almo 1 have differences in anions and cations of about 10% in milliequivalents.

4. Silica geothermometer of Fournier and Rowe (1966), as given by Truesdell (1976).

5. Na-K-Ca geothermometer of Fournier and Truesdell (1973), as given by Truesdell (1976). Best temperature underlined.

6. Magnesium-corrected Na-K-Ca geothermometer of Fournier and Potter (1979). Best temperature underlined.

Table 2. Physical data for Raft River geothermal exploration wells.

Well	RRGE-1	RRGE-2	RRGE-3
Drilling started	4 January 1975	27 April 1975	28 March 1976
Drilling completed	31 March 1975	21 March 1976	25 May 1976
Cased depth (m)	1104	1288	1291
Casing size (cm),((in))	34(13-3/8)	34(13-3/8)	34(13-3/8) to 422 m, 24.5(9-5/8) to 1291 m
Total depth (m)	1521	1994	1803 (leg C)
Fluid production before logging (10^6L)	36	3.8 produced*, 0.08 injected	6.4
Date of production before logging	April 1975	20 July 1976	mid-June 1976
Shut-in pressure (bar)	10.0	8.7	8.2

*RRGE-2 was drilled to 1825 m in June 1975. Between June 1975 and March 1976, approximately 120 million liters were produced and 40 million liters injected; injection followed most of the production. In March 1976, the well was deepened to 1994 m. Production and injection amounts shown in table took place after this deepening.

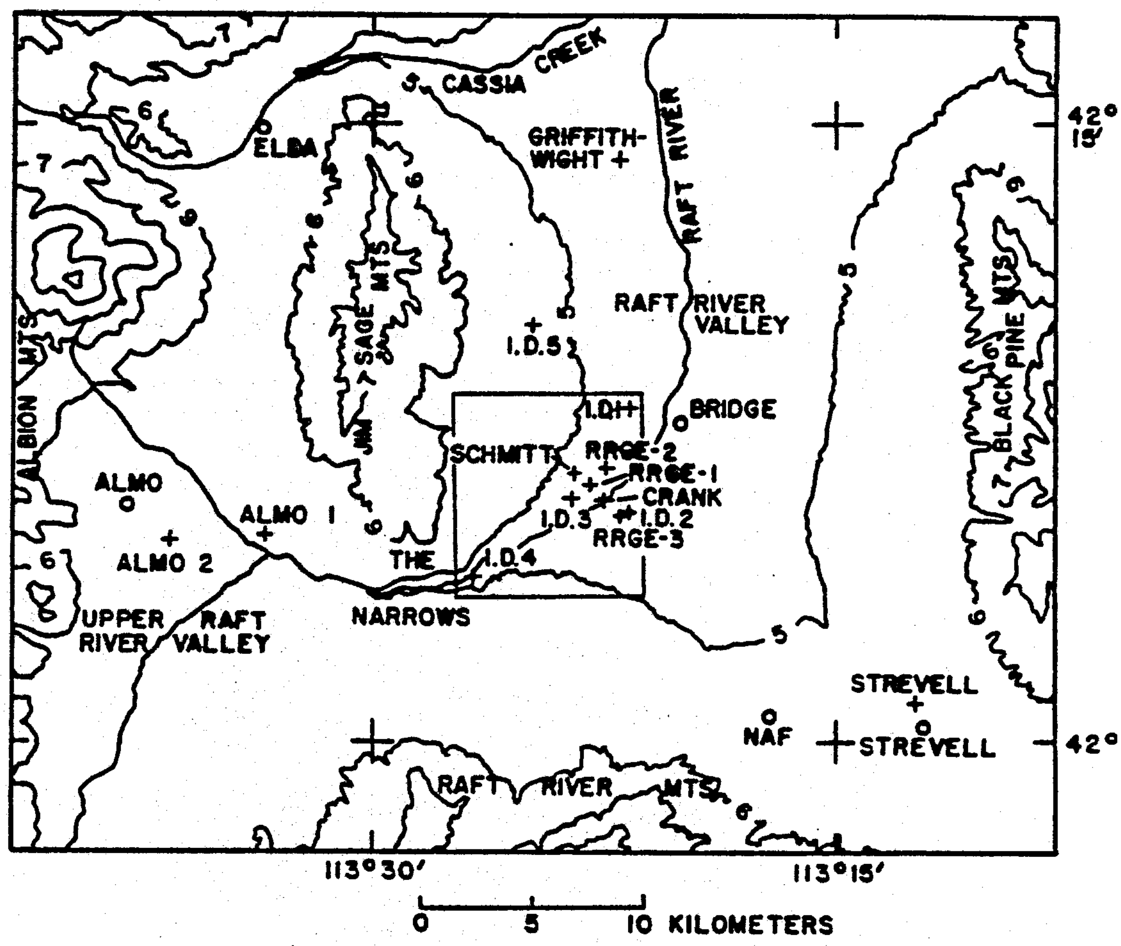


FIG. 1. Topographic map (1000-foot contours) of Raft River geothermal area and environs. Crosses, well and drill-hole locations; circles, towns. Area of Figure 5 is outlined.

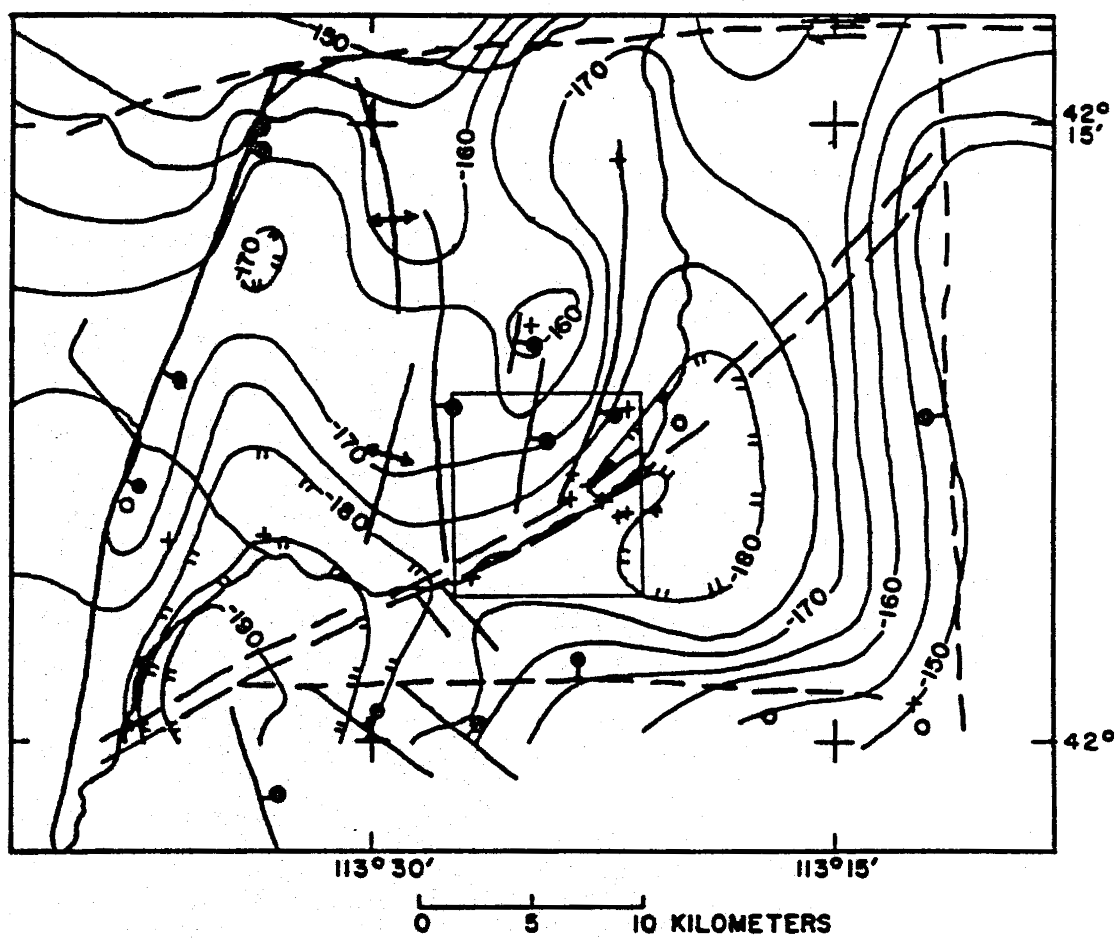


FIG. 2. Map of Bouguer anomaly values calculated with a 2.67-g/cm^3-density factor (Mabey and Wilson, 1973); contour interval, 5 mGal (1 Gal = 1 cm/sec^2). Areas of low gravity are shown with hachured contours. Major faults (bar and ball on downthrown side—dashed where inferred) and anticline (Jim Sage Mountains only) from Williams et al (1976). Crosses, wells; circles, towns. Area of Figure 5 is outlined.

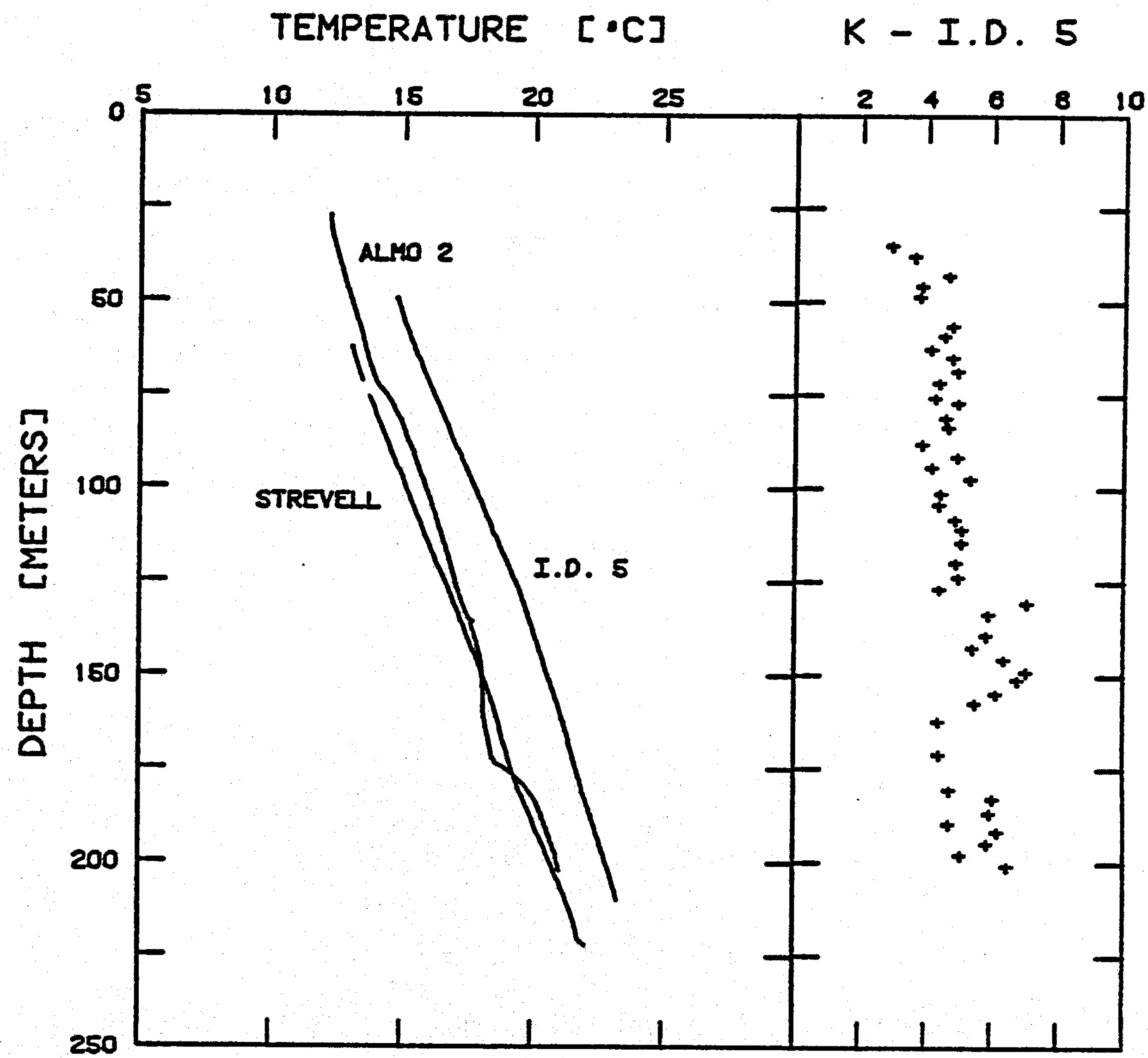


FIG. 3. Temperature logs in drill holes Strevell (17 October 1975), I.D. 5 (6 August 1976), and Almo 2 (8 August 1976). Strevell cased to logged depth, I.D. 5 cased and cemented to total depth, and Almo 2 cased to 141 m but cemented only to 70 m. Thermal conductivities for I.D. 5 are in mcal/cm sec ^{o}C.

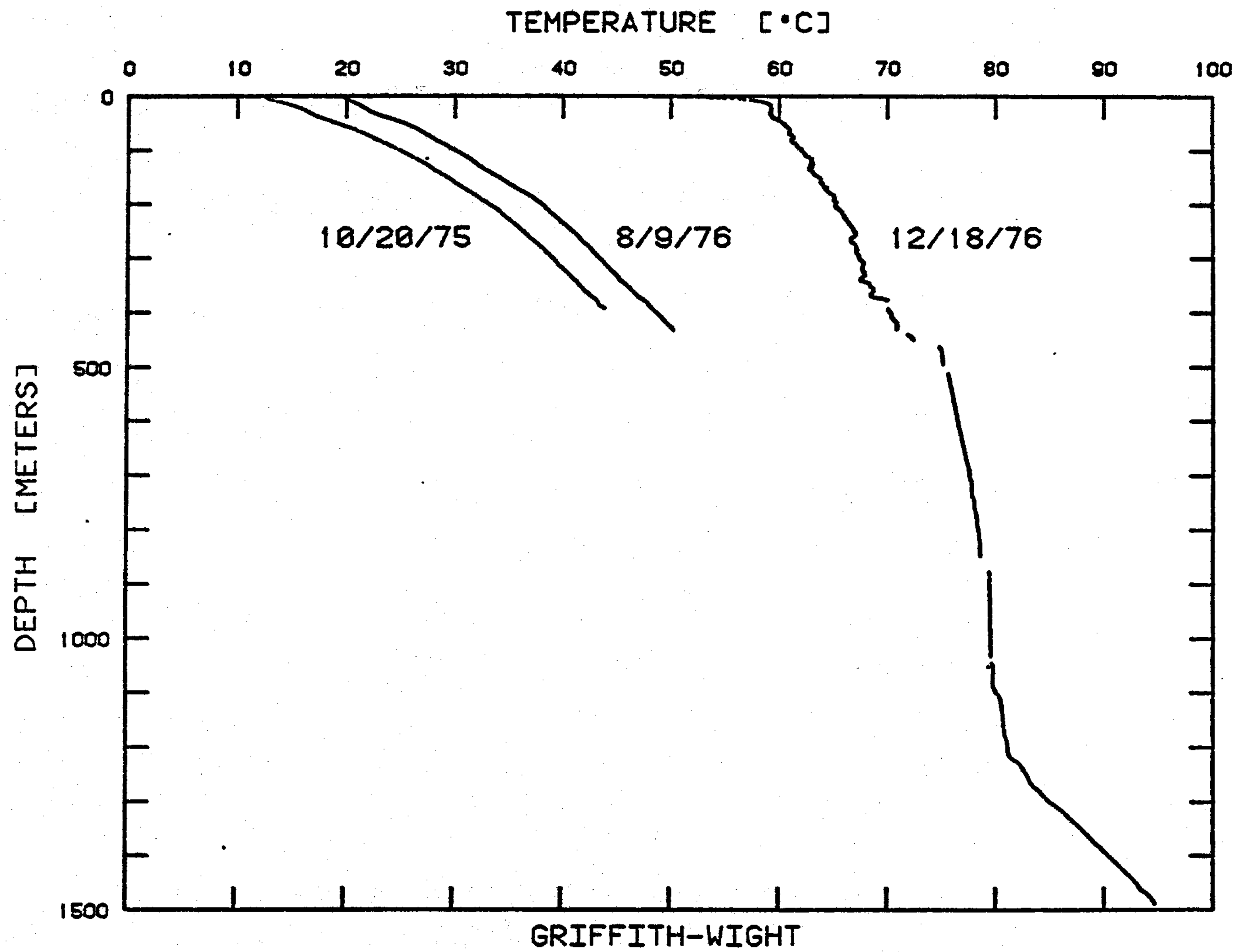


FIG. 4. Temperature logs in Griffith-Wight well. Log of 20 October 1975 was taken with no flow. Log of 9 August 1976 taken after well had been flowing 10 mL/sec for some time. Log of 18 December 1976 taken while intermittently flowing the well to get past zones where the temperature probe stuck.

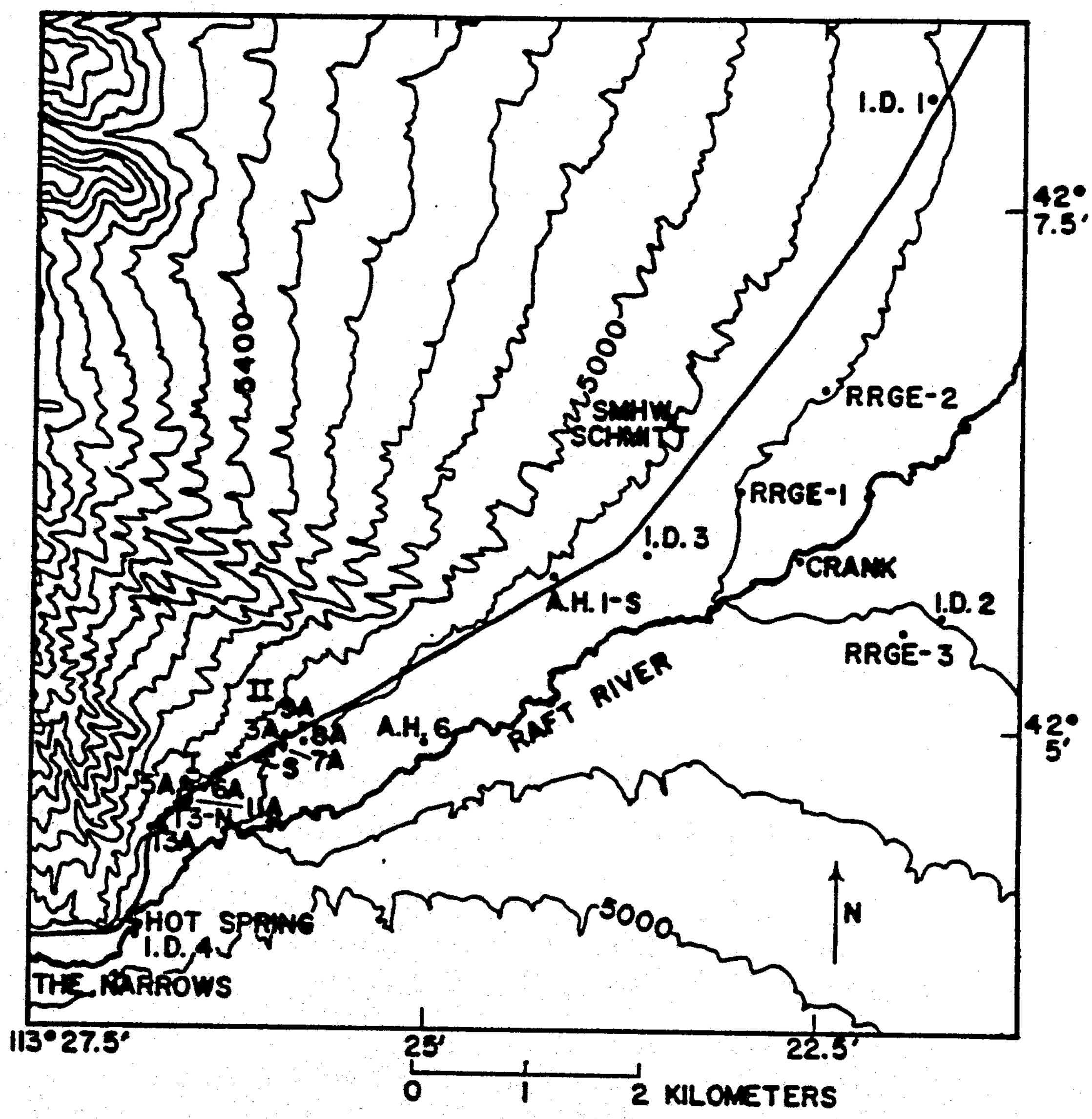


FIG. 5. Topographic map (80-foot contour interval) showing locations of auger holes (A.H.), intermediate-depth drill holes (I.D.), Raft River geothermal exploration wells (RRGE), drill hole SMHW, and Schmitt and Crank hot wells. I and II denote groups of auger holes. See Figures 1 and 2 for general location of map.

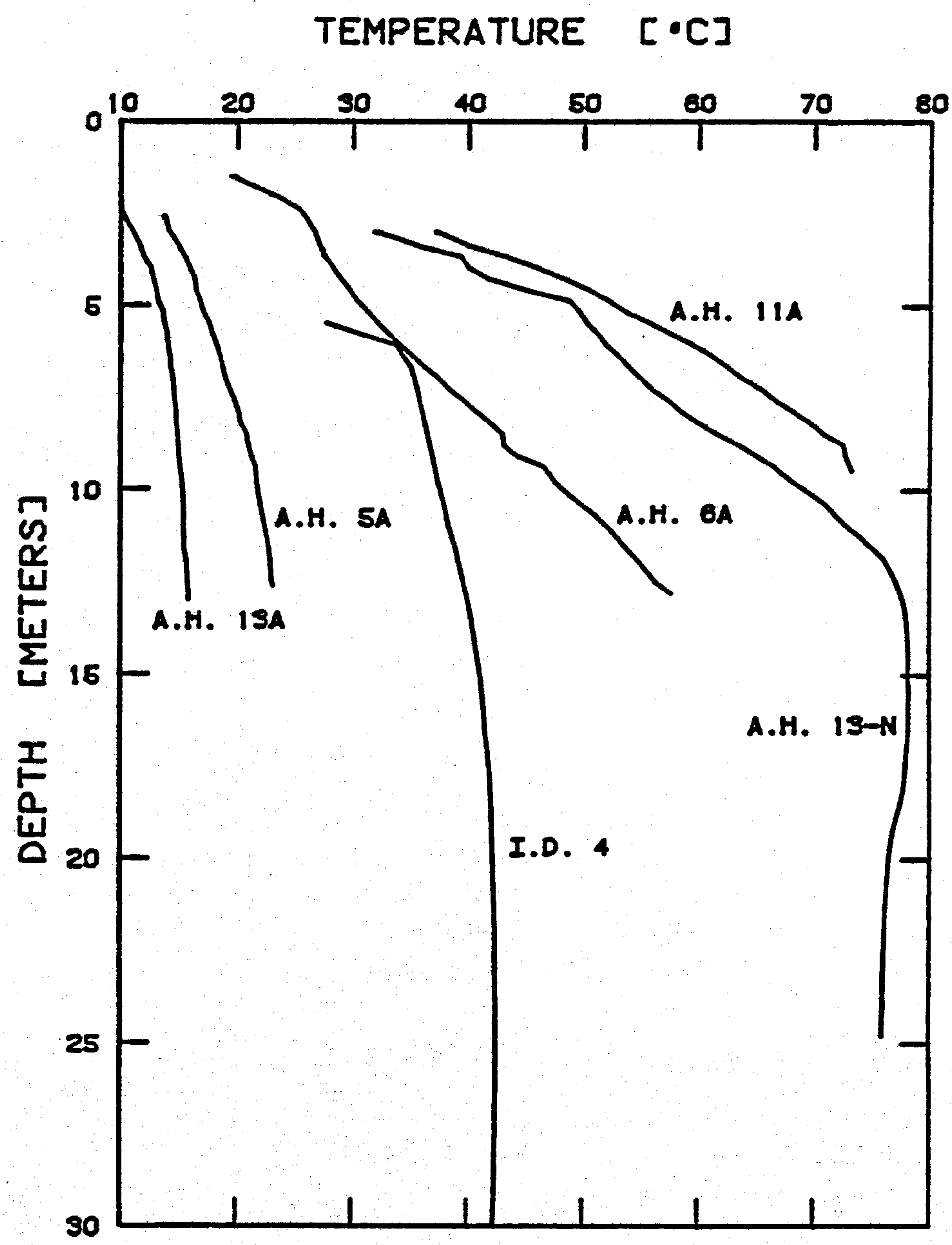


FIG. 6. Temperature logs of group I auger holes, I.D. 4, and A.H. 13A; see Figure 5 for locations. Temperatures measured on 16 January 1976 (A.H. 11A), 17 January 1976 (A.H. 13A, 13-N), 9 February 1976 (A.H. 5A), 8 August 1976 (I.D. 4), and 16 August 1976 (A.H. 6A).

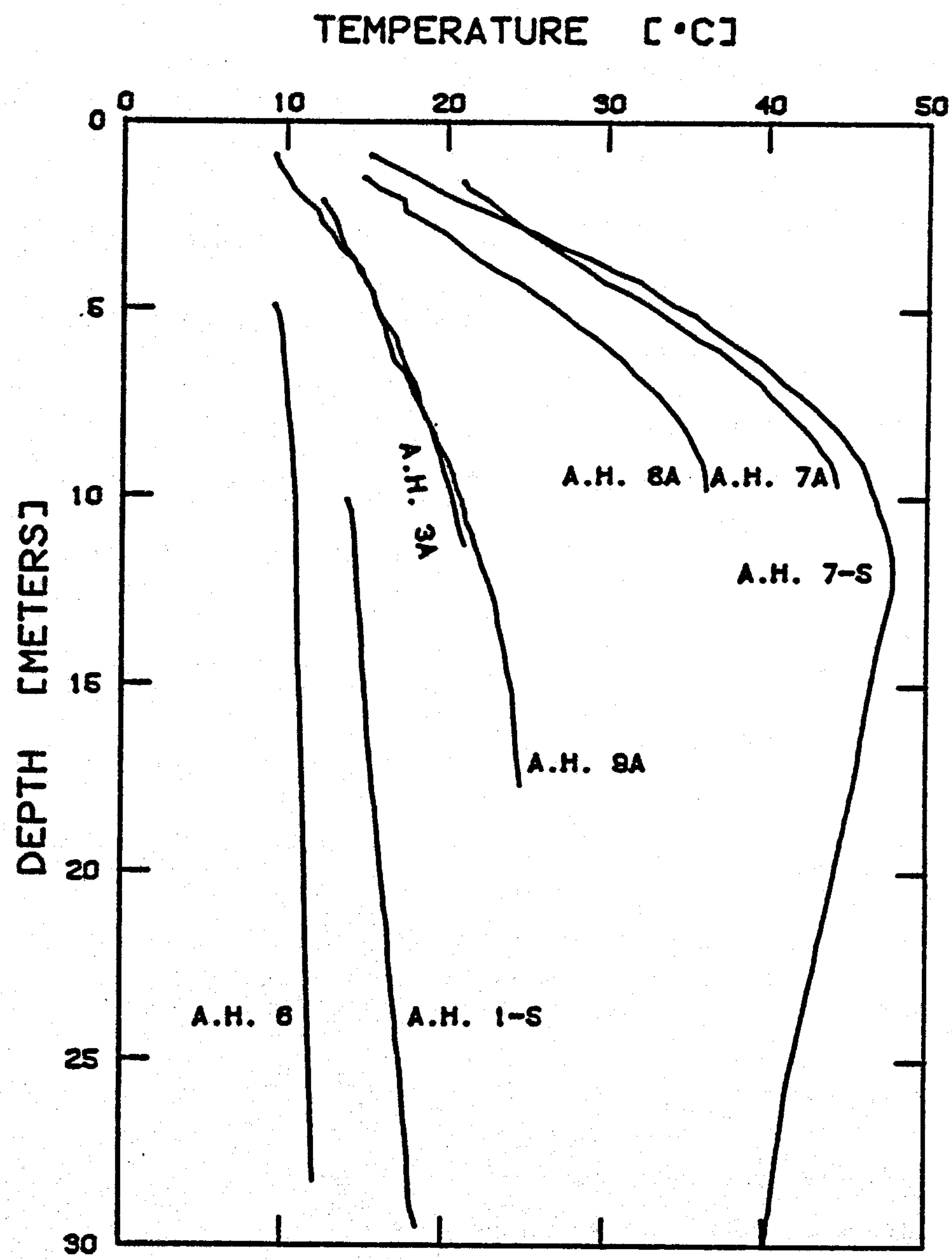


FIG. 7. Temperature logs of group II auger holes, A.H. 6, and A.H. 1-S; see Figure 5 for locations. Temperatures measured on 16 January 1976 (A.H. 3A, 7A, 7-S, 9A) and 9 February 1976 (A.H. 1-S, 6, 8A).

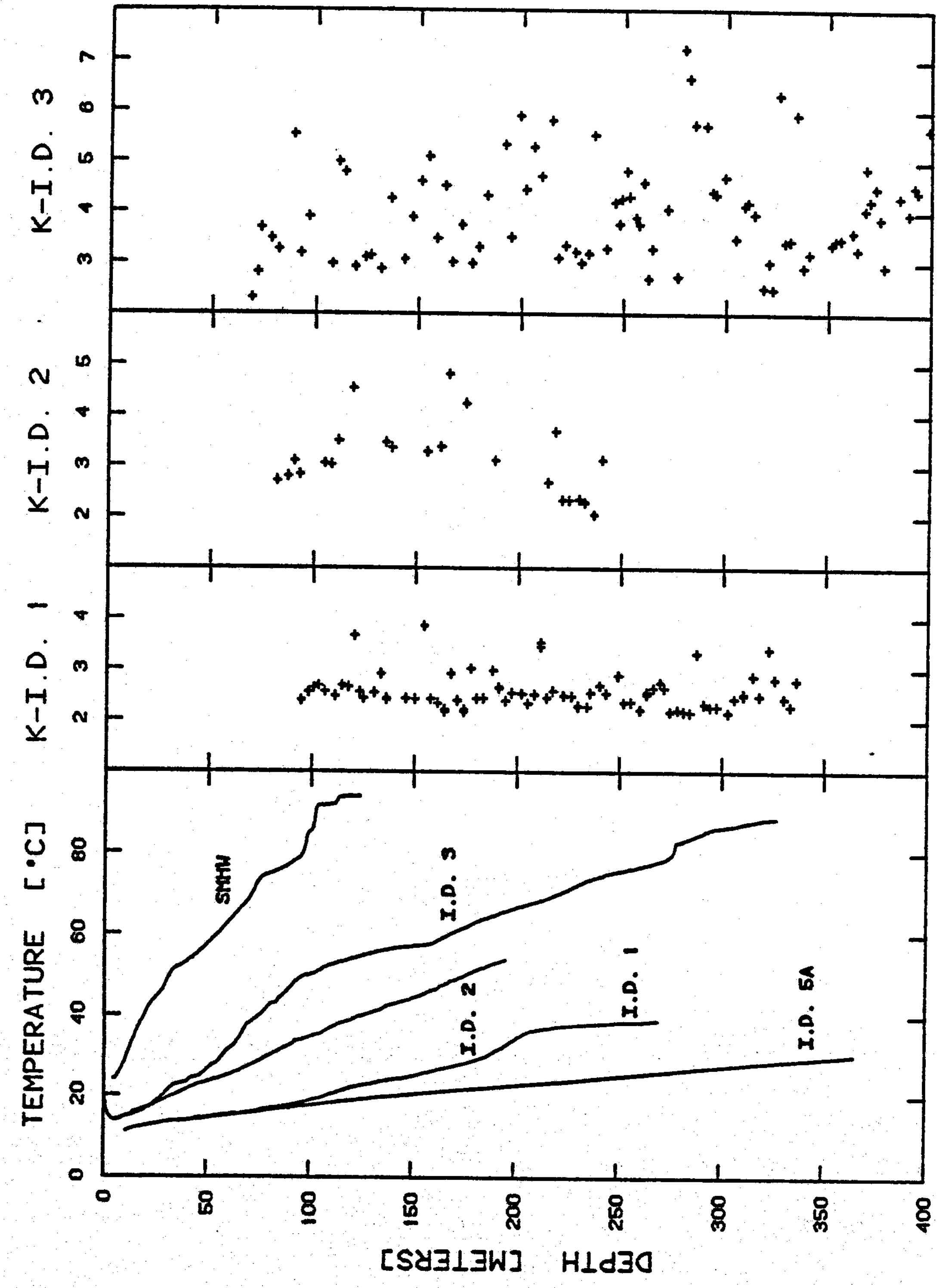


FIG. 8. Temperature logs of drill holes I.D. 1, 2, 3, and 5A, and SMHW; see Figures 1 and 5 for locations. Temperatures measured on 15 January 1976 (I.D. 1, 2), 6 August 1976 (I.D. 5A), 11 August 1976 (I.D. 3), and 14 August 1976 (SMHW). Wellhead pressure of 0.9 bar gauge at ground level in I.D. 3 during logging with no flow. Thermal conductivities are in mcal/cm sec °C.

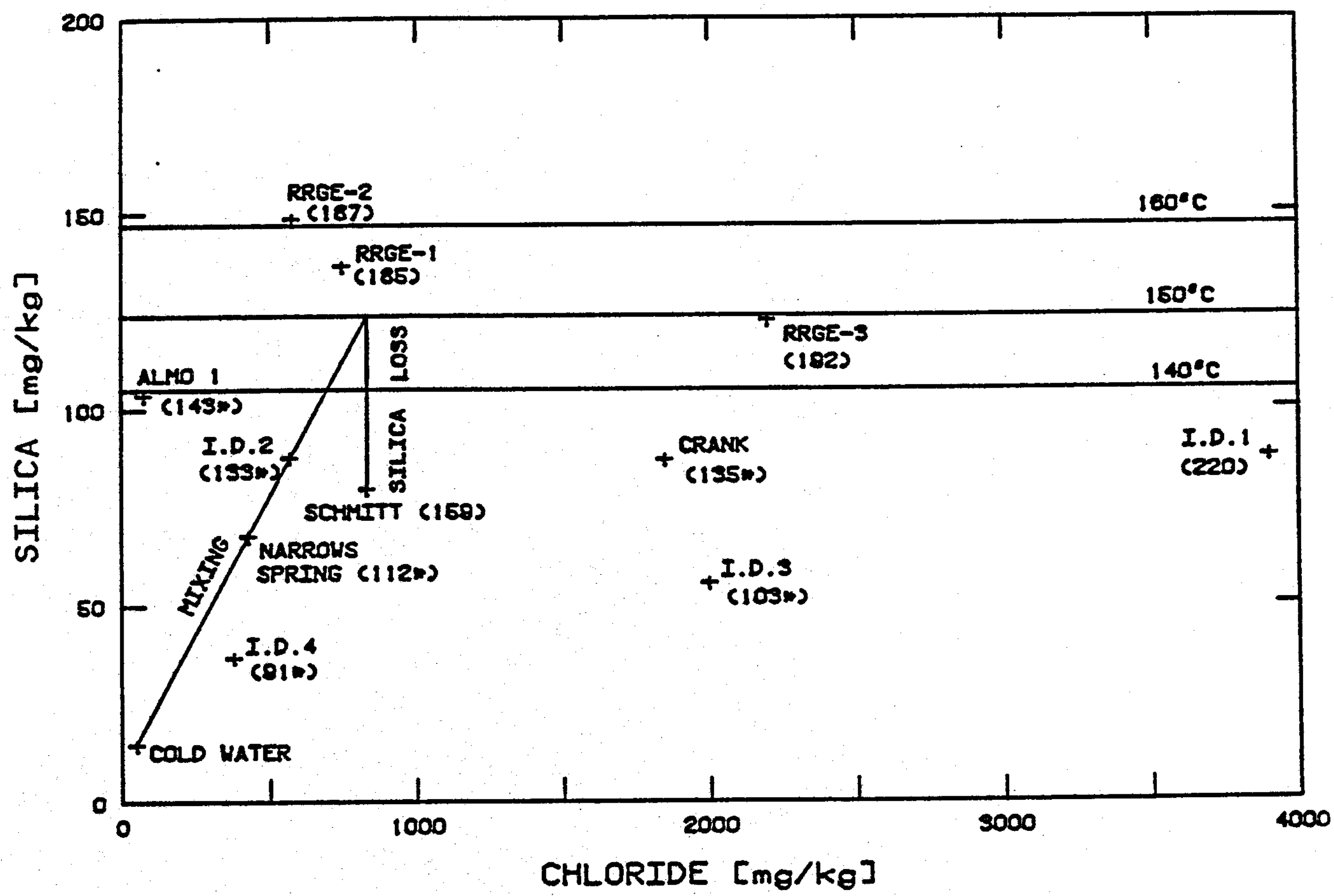


FIG. 9. Silica versus chloride concentrations in water samples from southern Raft River area; see Figures 1 and 5 for locations. Numbers in parentheses are Na-K-Ca geothermometer temperatures. Asterisk denotes a water sample for which Na-K-Ca and quartz geothermometers give nearly the same temperature. Horizontal lines drawn at silica concentrations that give the temperatures noted when used in the quartz geothermometer.

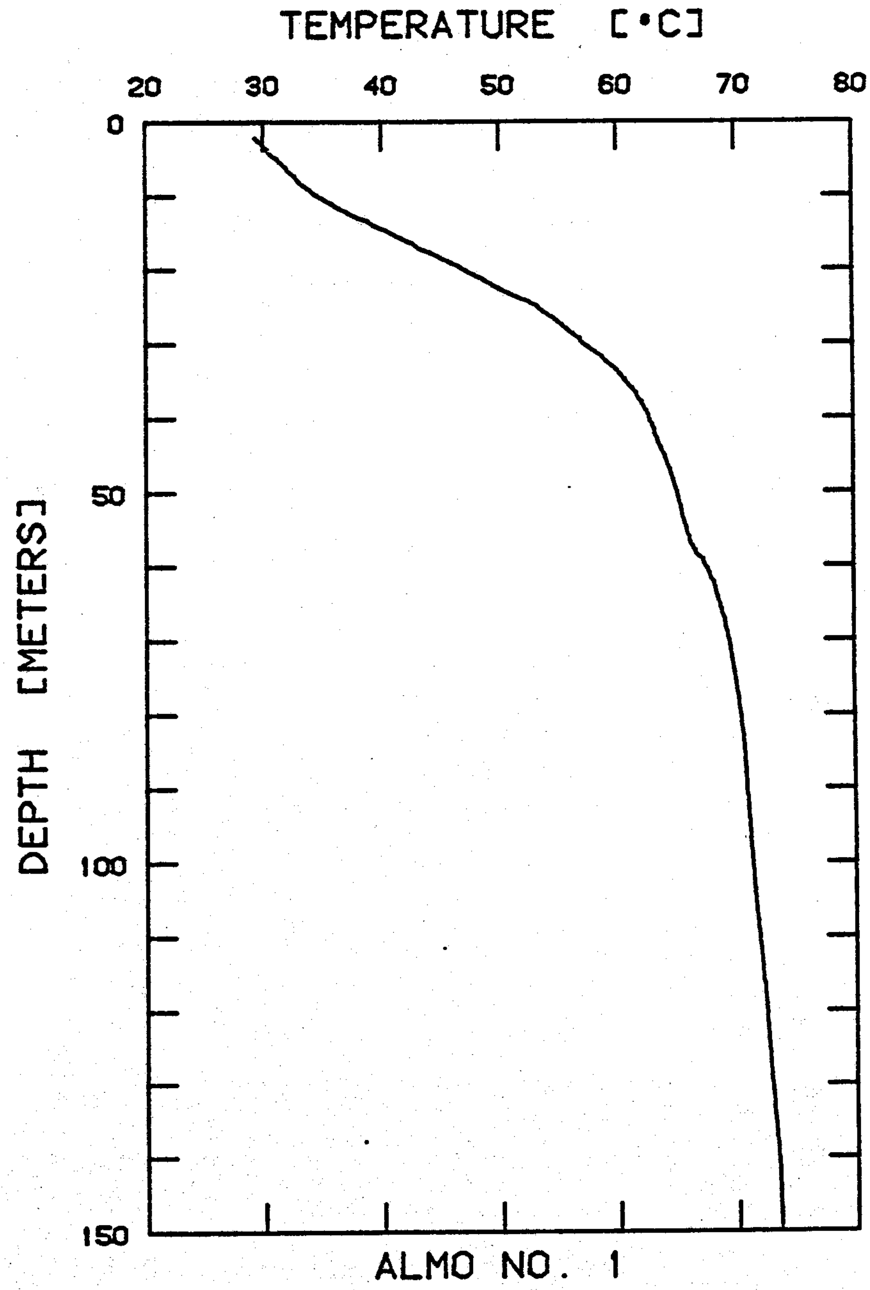
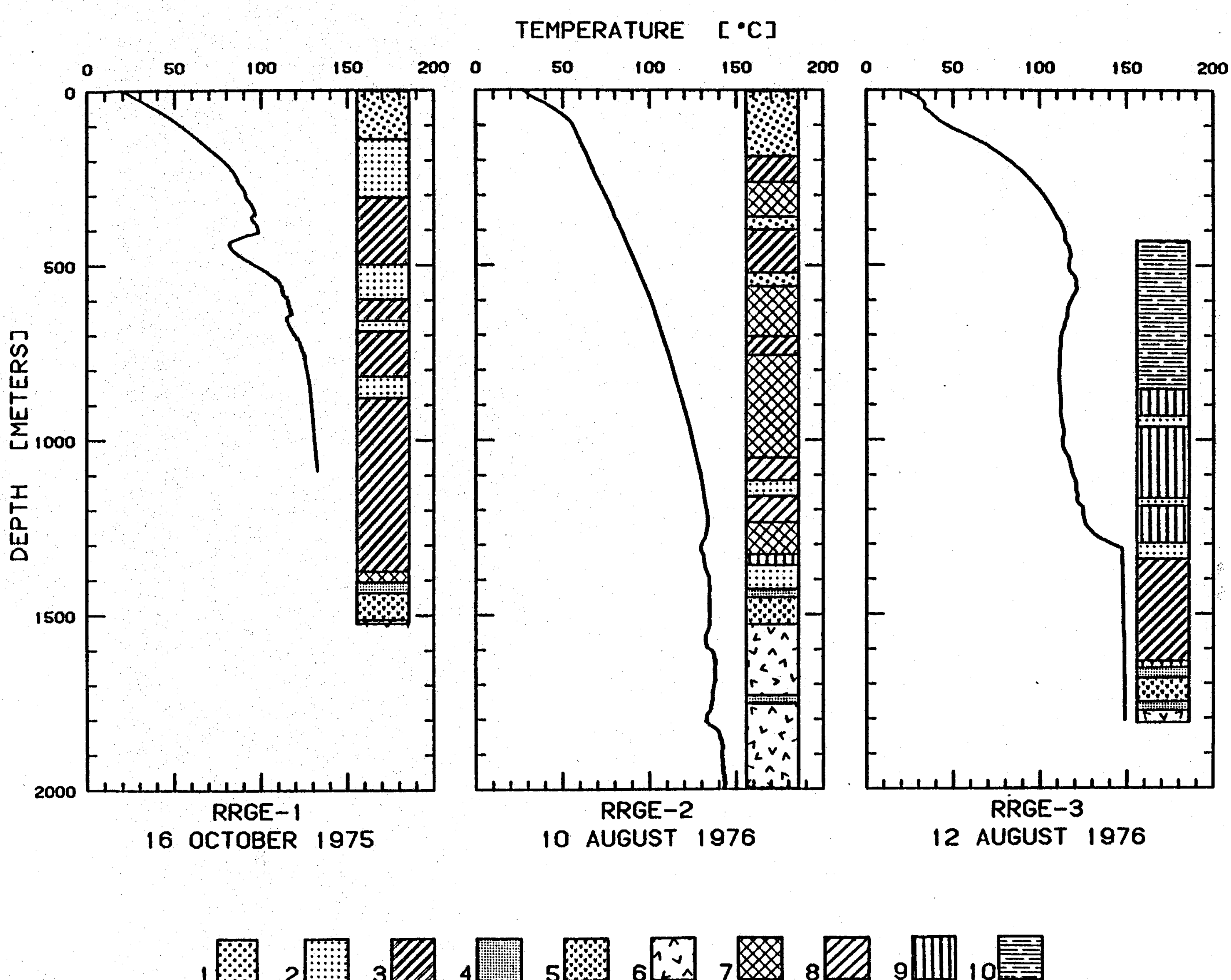


FIG. 10. Temperature log of drill hole Almo 1, obtained on 7 August 1976. Wellhead pressure of 0.79 bar gauge at 0.5 m above ground level. Flowing 3 mL/sec prior to logging.

FIG. 11. Temperature logs of RRGE - 1, 2, and 3; see Figures 1 and 5 for locations. Wellhead pressures are approximately 10.0, 8.7, and 8.2 bar, respectively. No flow during logging. Geology generalized from Covington (1977a, b, c). Symbols shown for lithology are: 1) sand and gravel, 2) sandstone, 3) tuff and siltstone, 4) schist, 5) quartzite, 6) quartz monzonite, 7) siltstone and sandstone, 8) siltstone, 9) tuff, and 10) siltstone and tuff.


TEMPERATURE [°C]
0 50 100 150 200
DEPTH [METERS]
0
500
1000
1500
2000
RRGE-1
16 OCTOBER 1975
RRGE-2
10 AUGUST 1976
RRGE-3
12 AUGUST 1976
1 2 3 4 5 6 7 8 9 10