Underground Temperatures and Heat Flow in the East Tintic District, Utah

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By T. S. LOVERING *and* H. T. MORRIS

SHORTER CONTRIBUTIONS TO GENERAL GEOLOGY

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Techniques and results of a geothermal study of an area having substantial temperature and heat-flow anomalies related to subterranean hot springs and geologic structure

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UNDERGROUND TEMPERATURES AND HEAT FLOW IN THE EAST TINTIC DISTRICT, UTAH

By T. S. Lovering and H. T. Morris

ABSTRACT

The deep water table in the East Tintic district slopes very gently to the east at approximately 4,550 feet above mean sea level and is from 1,000 to 1,500 feet below the surface. The temperature of the water at this elevation differs greatly from place to place, and measured temperatures range from approximately 80°F in the western and northwestern parts of the district to 140°F in the south-central part. Rock temperatures were measured in drill holes, ranging from 33 to 2,400 feet deep, and at many places in the extensive mine workings present in the central and southern parts of the district. Thermal conductivities were determined in the laboratory on many individual specimens, and on formations in the field by various methods. From these data, temperature gradients, heat loss, and mean annual temperatures at the surface were calculated. These data also enabled us to calculate the temperature at the level of ground water even where points at which temperatures were measured are well above or below water level, provided that the conductivities of the rocks between the point of temperature measurement and water level were known. Temperatures either measured or calculated were plotted for the water table over a wide area and a sufficient number of such points are available to show the general thermal pattern on maps. The theoretical basis for these calculations and the techniques used for measuring temperatures are described, and the results are shown on maps by means of isotherms representing mean annual temperatures at the surface, temperatures at water level, and isograms showing anomalous heat flow in microcalories per square centimeter per second in the East Tintic thermal area.

The geographic distribution and geologic relations of the thermal waters indicate that a steep northward-trending fracture zone in the footwall of the East Tintic thrust fault acts as a conduit for rising hot water which in part follows the overlying thrust fault and in part moves upward through short-circuiting fissures in the hanging wall. This water rises through much-fractured rock that is saturated with ground water, which in general is moving toward the east or northeast. The subterranean hot springs, however, apparently cause local reversals in the direction of ground-water flow.

Although the highest temperature measured at water level is 140°F, the calculated temperature at water level below the southernmost drill hole entered (EP 2) is 163°F. This hole is in the Latite Ridge thermal area, an area that to date (1964) has been inadequately explored, and therefore most of the quantitative estimates of temperature and heat flow are given for the main East Tintic thermal area which lies north of the Apex Standard thermal trough. However, the temperature rises southeastward at a uniform rate of 2.5°F per 100 feet horizontally through a distance of 2,000 feet (to ET 2) and if this rate held for another quarter of a mile the ground water there would be at the boiling point.

In the main East Tintic thermal area, the average temperature of the anomalously hot water is about 104°F, and an estimate of the approximate annual recharge of water at a temperature of 80°F allows calculation of the amount of water, 100 gpm (gallons per minute), at 143°F, which is the highest temperature calculated for ground-water level, required to raise its heat content to that of water at the average temperature of 104°F. The heat loss from the rock cover can also be calculated and is equivalent in heat energy to about 120 gpm of water at 143°F. The sum of these two waterflow rates suggests that subterranean hot springs in the East Tintic district contributes about 200 gpm of water if the temperature of the hot water in the conduit underlying the East Tintic thrust fault approximates 143°F.

The hottest waters sampled are saline and apparently belong to a family of saline hot springs that are found at several places in Utah. The ratios of SiO_2:solids, Br:Cl, K:Na, and Li:Na all suggest a magmatic contribution, and the deuterium content of the water clearly precludes its being hot connate water.

The evidence now available indicates either that the heat and mineral content of the hot saline water is of deep volcanic origin or that meteoric water moving through Jurassic salt beds below two or more major thrust faults is mixed with volcanic emanations at depth. The high gradients in the Latite Ridge thermal area, together with a seemingly favorable geological structure, suggest that the area should be explored as a possible source of geothermal power, but the thermal data now available (1964) are inadequate for an appraisal of this possibility.

INTRODUCTION

The East Tintic district, which is about 60 miles south of Salt Lake City, Utah, is well known to mining men not only for the rich blind ore bodies that have been exploited there but also for the great range in temperatures in the mine workings, most of which are above the permanent ground-water level. The ground-water level, which is at an elevation of approximately 4,550 feet, is from 1,000 to 1,500 feet below the surface throughout much of the district, and the mine workings above the water level are dominantly dry and well suited for measuring temperatures that are undisturbed by moving ground water except quite locally. During the period 1943–63 we were able to measure underground and surface temperatures at many places while studying the geology of the district. The information on temperatures gathered during this time is summarized on the maps that accompany this paper and is the basis for the conclusions reached concerning heat flow, heat sources, and the underground temperature distribution in the East Tintic district.

The study of the surface temperatures, which was an essential preliminary to the present report, has been described in another paper (Lovering and Goode, 1963). Drill-hole and other underground temperatures were recorded by many members of the field parties who participated in the geological study of the district from 1943 to 1950, but we assume the responsibility for the accuracy of all measurements reported here and regret that individual credit cannot be given for each measurement used. However, we wish to thank Robert H. Roy for the detailed temperature logs of holes BC 5 and BC 90 (see also Roy[1]) and the Bear Creek Mining Co. for temperatures from several of their deep drill holes.

MEASUREMENT OF TEMPERATURES

SURFACE TEMPERATURES

The method of ascertaining mean annual surface temperatures, which has been described in detail (Lovering and Goode, 1963), depends primarily on extrapolating to the surface the temperature gradient determined in shallow or deep holes. Direct measurements of temperature gradients at depths of less than 100 feet are in general unreliable unless the temperatures measured are recalculated to include the effect of the annual temperature wave at the surface. At depths greater than 100 feet, two or more temperature measurements should indicate the rate of change of temperature with depth; if the temperature gradient so ascertained is then extrapolated to the surface, the resulting figures should approximate closely the mean annual temperature at the surface unless the conductivity of the rocks changes drastically through a substantial part of the section above the points where the gradient was measured. In few places in the East Tintic district does the lower conductivity of the weathered layer near the surface suffice to change by as much as 1°F. the temperature extrapolated from below a depth of 100 feet.

Surface temperatures are determined almost entirely by climatic, geographic, and vegetational features and show virtually no relation to deep underground heat sources. The exposure of slopes to sun and wind are major factors governing the mean annual temperature, but the vegetational cover and the snow cover are also of importance. A dry bare, south-facing slope may have a mean annual temperature several degrees Fahrenheit higher than that of an adjacent north-facing slope covered with pinyon pine or brush. The surface temperatures as determined from many shallow and deep drill holes, together with the inferred isotherms representing mean annual temperatures at the surface, are shown on plate 1.

TEMPERATURE GRADIENTS IN DEEP HOLES

The measurement of temperatures in deep drill holes appears deceptively simple. The most common major error in such measurements is caused by the assumption that if the probe has been held at a point long enough to come to the temperature of the water or air surrounding it, the temperature measured by the probe is the true rock temperature. A thermometer in a cold metal case that occupies a substantial part of the drill hole at the point of measurement may greatly lower the temperature of the air or water surrounding it. A difference of only a few degrees Fahrenheit between the case and the surroundings has little effect on the recorded measurement, whereas a difference of several tens of degrees, such as may exist between the temperature of the equipment at the surface on a cool day and the temperature measured underground, would materially lower the temperature reached by the maximum thermometer. A probe lowered rapidly on a cold February morning to the bottom of one deep drill hole, some 625 feet below ground-water level, and left there for 45 minutes gave a reading that was 20°F too low. Unless the probe can be left in place for several days, in-hole measurement at given depths should always be a two-step procedure; the probe should first be held a short distance—15 to 30 feet—above the point of measurement long enough for the probe to come reasonably close to temperature equilibrium with the surrounding air or water, and then should be lowered to the point of temperature measurement. In water, the time required to reach equilibrium commonly ranges from 3 to 15 minutes, whereas in air the time required is generally at least 1 or 2 hours, and a halt of several hours is desirable. The actual time required depends on the accuracy sought, on the kind of probe employed, and on the material used to protect or enclose the probe. The requisite time for satisfactory results with a give a given type of probe should be determined by trial in the field.

The probable error inherent in measuring a high temperature underground with cold equipment has just been noted; the converse is also true. If drill-hole temperatures are being measured on a warm summer day,

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[1] Roy, R. F., 1963, Heat flow measurements in the United States: Harvard Univ. unpub. Ph.D. thesis.

the equipment must be prechilled, and a maximum thermometer must start its downward journey showing a temperature well below that which is measured underground. We used a Dewar flask filled with ice and water to provide cold water for chilling the rag-wrapped thermometer and then whirled the thermometer at the end of a 3-foot string to lower the mercury column to the desired level. The thermometer was then quickly placed in a chilled case and lowered rapidly for the first 50 feet below the surface. Each time the thermometer was brought to the surface to be read, the thermometer case was quickly immersed in cold water and opened, and the thermometer itself was withdrawn only far enough to read—before the summer heat could affect it.

Where an accuracy of ±0.1°F is sufficient, well-calibrated maximum mercury thermometers in watertight probes are quite satisfactory, though somewhat difficult to use in hot weather.

Most of the drill-hole temperatures in the East Tintic district were first measured with maximum thermometers calibrated by the Bureau of Standards and believed accurate to ±0.1°F. The equipment used was that devised by C. E. Van Orstrand while he was Chief Physicist of the U.S. Geological Survey (Van Orstrand, 1924). A bronze tube about a foot long, with watertight screw cap, held one or more thermometers and was attached to piano wire that ran over a grooved calibrated wheel meshed with a counter that showed the depth of the thermometer in feet. A small sash weight was attached to the bottom of the brass tube to provide tension on the line at all times and to prevent the wire from snarling when lowered into an an open hole with irregular sides. This weight also signaled the presence of obstructions or the bottom of a hole by the sudden drop in tension on the line.

For accuracy greater than 0.1°F, perhaps the two best devices are (1) the bathythermograph and (2) the resistance thermometer (of the type that utilizes a small thermistor designed for the range of temperatures expected). The bathythermograph, being of larger diameter than either the maximum or resistance thermometer, has the disadvantage of requiring a hole at least 4 inches in diameter; it is most useful where air temperatures are to be measured and no pressure corrections are required. If the bathythermograph is well calibrated, it is a very satisfactory instrument but requires more time to reach equilibrium than do either the maximum thermometor or the resistance thermometer. We used a bathythermograph somewhat similar to those designed by A. F. Spilhaus for oceanographic work (Spilhaus, 1938).

In many ways a resistance thermometer having a well-calibrated thermistor is the most satisfactory probe for measuring temperatures in a drill hole. In deep drill holes, however, the weight of the electric cable causes a considerable stretch in a new line, and the length of wire going into a deep hole should be checked against the length that comes out, so that a correction factor can be applied to compensate for the stretch. It is, of course, essential to calibrate the thermistor accurately through the range of temperature to be measured; one should also be aware that an extrapolation beyond the points of calibration may introduce substantial errors in the measurement. As with the maximum thermometer, it is advisable to hold the resistance element a few feet above or below the point to be measured before recording the temperature. The current used in some sensitive single thermistors if applied steadily for a half minute may raise the temperature of an element in air by as much as 0.2°F and therefore should be used only momentarily. The heating effect, however, depends on the thermistor and the current necessary to get the required sensitivity with the monitoring eqiupment. No heating is observable with some of the more sophisticated equipment now available.

In our work with a resistance thermometer, it was customary to measure the temperature at set points during the probe's descent, leave the probe at the bottom for an appreciable period of time, and then repeat the measurements on the way up; at each check point we waited until equilibrium had been established. The two temperatures rarely differed by as much as 0.1°F; the first temperature was influenced by the introduction of an element cooler than the surroundings and the second by the heat given off by an element warmer than the surroundings. The mean of two such measurements is assumed to be close to the actual temperature of the walls of the drill hole. If the resistance is measured at 5- or 10-minute intervals when the probe is in air and the readings are plotted against time, the equilibrium value is usually apparent well before it is reached.

Temperatures were also measured with thermocouples but this technique required that a cold junction be attached to the line no more than 20 feet from the thermocouple and lowered into the hole with it. If the cold junction was kept at the surface, the response became so sluggish when the hot junction was lowered a few hundred feet as to be unreadable even with the excellent portable potentiometer used. The volume and mass of the container used for the cold junction also discouraged the use of this device and led to the adoption of thermistors where comparable sensitivity was sought.

Although the preceding discussion suggests many of the difficulties that arise in measuring temperatures in drill holes, one should realize that the temperature measured may not be the temperature sought. If the drill hole contains a steel casing, the high conductivity of the steel will do much to average the temperatures of the poorly conducting rocks through which the hole passes; furthermore the steel will carry the mean annual temperature wave much deeper into the hole than will the poorly conducting rocks. In cased drill holes in the East Tintic district, temperature perturbations caused by the annual wave were most perceptible in late summer and early spring.

Convection currents rarely caused appreciable temperature changes in deep drill holes in the area under study, but sharp changes in barometric pressure did cause temperature changes corresponding to the compression or expansion of the air column in the hole. Where the water table is more than a thousand feet below the surface, compression of the column of air between it and the surface by a change in barometric pressure can cause a movement of as much as 25 feet in the column of air. A cloud shadow and the attendant cooling breeze passing over the collar of the drill hole one hot summer day was observed to cause a drop in temperature corresponding to such a fluctuation of the air column during temperature measurements made with a thermocouple at hole TS 14.

It is generally assumed (Hales, 1937) that convection occurs in a water-filled tube if the water below a given point is less dense than the water at or above the point, except as inhibited by friction. Above 40°F, water expands with increasing temperature. The decrease in density downward due to the temperature gradient is opposed by the increasing compression caused by the weight of the overlying column of water. When the two opposing factors are equal, a critical gradient is said to exist. This critical gradient has been calculated by Hales (1937), but he pointed out that the basic assumptions must be in error because the calculated critical gradient is many orders of magnitude below gradients measured in geysers that erupt periodically. Hales suggested that the use of the molecular coefficients for viscosity and thermal diffusivity in the final equation should be replaced by corresponding "eddy coefficients" which have been found to range from 1 to 400 as compared to a general order of magnitude of 10^{-3} for the molecular coefficients.

Temperatures measured in water-filled holes having gradients in excess of 8°F per 100 feet clearly show that either convection does not take place or, if it does occur, it must be in many superposed cells of such short length that the temperature perturbations are to be measured in hundredths rather than in tenths of a degree or more.

The minimum height of a cell is at least equal to the width of the tube and may be several times as great. (Hales (1937) showed that the tendency toward convection varies with the fourth power of the diameter of the tube.) The drill holes in the East Tintic district for which temperature data have been obtained range in diameter from 2½ to 6 inches and the tendency toward convection therefore is 33 times greater in the 6-inch hole than in the 2½-inch hole. For these holes the temperature differences for the minimum cell would be $\frac{\Gamma d}{1200 \text{ in.}}$, where Γ is the gradient in degrees Fahrenheit per 100 feet and d is the diameter of the drill hole; the differences are from 0.002 to 0.005 times the geothermal gradient. For a 6-inch hole with $\Gamma = 10°$ per 100 feet, the minimum convection cell should then have a perturbation of 10×0.005 or 0.05° F. Perturbations that approximate this figure have been observed in some drill holes when the probe was first positioned but they commonly attenuated and disappeared after about an hour. An individual cycle of 0.05° F usually takes less than 10 minutes and the average of two cycles is within $\pm 0.01°$ of the temperature ascertained by measurements prolonged over many hours.

Rock temperatures that are accurately measured in a hole still may not give the true gradient. The temperature changes caused by drilling the hole diminish slowly, and the longer the time spent in drilling the hole, the longer the time before the normal preexisting temperatures are regained. The true gradient is closely approximated long before the temperatures have returned to the predrilling condition. The minimum length of time that should elapse after drilling is completed and before the temperature is measured should approximate the duration of drilling, but a period three times that long is desirable.

Lachenbruch and Brewer (1959), reporting on an observation drill hole at Point Barrow, Alaska, stated that 67 days after drilling was completed (duration of drilling, 58 days) the gradient measured was only 5 percent greater than the probable equilibrium value, although the individual temperatures in this frozen terrain were about 2°C greater than the probable equilibrium values. These authors have devised a simple formula for computing the equilibrium temperature at any given depth when measurements are made at time intervals (Lachenbruch and Brewer, 1959, p. 105): temperatures are plotted as ordinates and the values of $\log_e \left(\frac{t}{t-s}\right)$ are plotted as abscissae, where t is time since drilling stopped—in either days or seconds—and s is the duration of drilling in the same units. Extrapolation of the nearly straight-line curve to a zero value for this expression gives the equilibrium value of the temperatures at the depth in question. The temperatures predicted by this method for the Point Barrow hole after

$t=3s$ were within 0.05°C of those predicted by the use of 6 years' data, when $t=35s$.

The temperature at the bottom of an advancing drill hole is less disturbed by the drilling fluid than is that of the wall of the hole above. For this reason an excellent temperature gradient can usually be obtained if the bottom-hole temperature is measured about once a week during the time a hole is drilled, but the temperature should not be taken until after a 24-hour shutdown.

TEMPERATURES IN MINE WORKINGS

The best time to get a temperature that approximates the temperature of the undisturbed rock in a mine is when the opening is first made; a side hole 4 to 5 feet deep drilled at the same time that the blastholes are made to advance the heading can be used to the best advantage if the temperature at the bottom of the side hole is measured within 24 hours. The high temperature of the blast is only momentary and has no effect on the temperature of the rock 2 feet from the walls of the newly created opening, but the air used to ventilate the heading may have a perceptible effect within a few days. As such prompt rock-temperature measurements are rarely available, the perturbations caused by mine openings must be recognized. Ventilation usually cools the rocks because of cooler temperature of introduced air and evaporational effects where rocks are damp or wet. Where freshly opened porous masses of sulfide are exposed to moving air, heating from oxidation of the sulfides usually proceeds more rapidly than cooling from ventilation and may cause a notable increase in the temperature of the rocks. The longer the period during which an opening is ventilated, the farther from the opening is the perturbation of temperatures perceptible.

The difficult problem of calculating the original rock temperature has been treated mathematically by Jaeger and Le Marne (1963) for the case where the temperature of the walls of the mine opening is held constant for substantial time intervals and the ventilation history of the opening is known. Such information was lacking for the mines in the East Tintic district. The calculations and measurements made by Jaeger and Le Marne show that in rock having moderately high diffusivity ($\alpha=0.014$), ventilation for about 5 years had affected the rock temperatures to a distance of about 120 feet from the wall of the drift.

Where heating due to oxidation may be neglected, ventilation and moisture are the chief factors that change the temperature in the mine opening from that of the undisturbed rock. It is axiomatic therefore that temperature measurements should never be taken in a drift where air is moving unless this fact is recorded; temperatures so taken rarely correspond to those of the rock before the opening was made. Where mine openings disturb the pattern of subsurface drainage, a concomitant change in rock temperatures is to be expected. Data from places in the East Tintic mines that were suspect were omitted.

In mine openings that have been unventilated longer than they were ventilated, the rock temperatures a few feet from the opening may approximate those of the undisturbed rock; the conditions for equilibrium are similar to those discussed for drill holes (p. F2, F3). Most drifts are floored with rubble commonly 6 inches or more deep on which the tracks were originally laid and such rubble, if dry, makes an excellent insulating medium. The conductivity of such material is usually only a fraction of that of the solid rock nearby and 6 inches of rubble may thus correspond in its insulating qualities to 2 to 3 feet of solid rock of the same composition. A thermometer placed on the bedrock under the dry rubble in the bottom of a drift usually gives a temperature corresponding to that at the end of a hole 2 to 3 feet deep in the side of the drift.

When mapping underground in the East Tintic district, U.S.G.S. geologists customarily excavated a hole to the bottom of the rubble in a drift, placed a thermometer at the contact of the rubble and the bedrock, covered the thermometer with the material excavated from the bottom of the rubble, and then left the thermometer until it had reached a constant temperature. This temperature was then recorded on the mine plan with the notation as to whether or not there was an air current—"live air" or "dead air". At any place in a drift where there was moisture, the temperatures were abnormally low because of evaporation even where there was no ventilation. Only temperatures taken in drifts that had long been unventilated are shown on the maps (pls. 3, 5).

At the time of our study, all the accessible mine workings were above water table and were mostly unaffected by water circulation. Some temperatures from below water level in the Eureka Standard mine were measured earlier by engineers of the Tintic Standard Mining Co. and may represent temperatures appreciably disturbed by ground water moving into the mine openings.

In the mine workings where oxidizing sulfides are generating heat, the additional heat source is usually evident from the sulfurous odor in the mine workings as well as from the anomalous temperatures and gradients that are found near such areas. Temperatures recorded there rarely represent those characteristic of the undisturbed rock and ore, and must be used with great caution.

TEMPERATURE GRADIENTS, CONDUCTIVITY, AND HEAT FLOW

RELATION OF THERMAL CONDUCTIVITY, TEMPERATURE GRADIENT, AND HEAT FLOW

The quantity of heat flowing through the rock is determined by both conductivity and gradient, and identical gradients can result in very different quantities of heat moving through the rocks. In some areas in the East Tintic district the temperatures at water level are the same but, because of different conductivities, the rocks above water level have markedly different values for the heat flow (q), as is shown on the map (pl. 1), southwest of the Eureka Standard shaft.

The relation of the quantity of heat (q) passing through a uniform body of length (L) and area (a) for the time (t) is shown in equation 1,

$$q=\frac{k(T_1-T_2)ta}{L}, \tag{1}$$

where k is the thermal conductivity and T_1 and T_2 are the temperatures at two points in the body separated by the distance L.

The average temperature gradient Γ, through the distance L, is

$$\Gamma=\frac{T_1-T_2}{L}. \tag{2}$$

The amount of heat per unit area per unit time is therefore:

$$q=k\Gamma. \tag{3}$$

In the United States, temperature gradients are generally specified in degrees Fahrenheit per 100 feet, but the conductivity of rocks is customarily given in cgs units (calories per centimeter per second per degree centigrade). Converting the gradient measured in degrees Fahrenheit per 100 feet to degrees Centigrade per centimeter requires use of the conversion factor shown in equation 4,

$$q=k1.823\times10^{-4}\Gamma_f, \tag{4}$$

where q is in cgs units and Γ_f is the gradient in degrees Fahrenheit per 100 feet.

From equation 4, it is apparent that k in the cgs system (cal cm^{-1} °C^{-1}) is

$$k=q\times10^4/1.823\Gamma_f, \tag{5}$$

where q represents calories per second per square centimeter.

If materials of different conductivities lie in horizontal layers and the lower and upper faces of the stack of layers are held at constant and different temperatures, we have conditions analogous to a simplified geological structure. The temperature at the surface of the earth is assumed constant (equivalent to the mean annual temperature) and the temperature deep within the earth is also assumed constant. If no heat source (such as oxidizing sulfides) and no heat sink (such as moving ground water) are present, the heat flow must also be constant, and the gradients in adjacent layers will be inversely proportional to their conductivities:

$$q=k_1\Gamma_1=k_2\Gamma_2=k_n\Gamma_n \tag{6}$$

where $k_1=k$ of layer 1, and Γ_1=gradient in layer 1, and similarly for layers 2, 3 . . . n; hence

$$\frac{\Gamma_1}{\Gamma_2}=\frac{k_2}{k_1}. \tag{7}$$

Where the conductivity of one rock layer is known, the conductivity of another layer of rock can be determined if the gradients in the two rocks have been measured. Where the layers are inclined rather than horizontal, the calculation of the conductivity may be less simple; but unless there is a steep dip together with a pronounced difference in the conductivities of the inclined beds, the conductivities determined by measuring gradients and using equation 7 give satisfactory values for k. (See pl. 2.)

Roy (1963, p. 18; see footnote 1, p. F2) has derived an equation that permits calculation of the vertical gradient through inclined layers if the requisite parameters are known. Roy assumed that a dipping tabular body of different k from that of the surrounding media refracts the lines of heat flow:

$$\frac{\tan a_1}{\tan a_2}=\frac{k_1}{k_2}, \tag{8.1}$$

where a_1 and a_2 are the angles with the normal of the incident and refracted heat-flux vectors, and k_1 and k_2 are the conductivities. If end effects are neglected and the isotherms in the underlying medium are assumed to be horizontal, then

$$\sin^2 a_1=\left(1-\frac{q''}{q_1}\Big/1-\frac{k_2}{k_1}\right), \tag{8.2}$$

where a_1 is the angle of dip, q'' is the vertical component of heat flow in the inclined tabular body, and q_1 is the heat flow in the underlying medium. The ratio of flux densities is

$$\frac{q_1}{q_2}=\cos a_2/\cos a_1. \tag{8.3}$$

For most places in the East Tintic district the calculation of corrections for dip would be a meaningless refinement of the gross picture presented by calculations based on the simplified structure assumed, and therefore no such correction was attempted.

In our study of the East Tintic district, the conductivities of many rocks were determined experimentally in the laboratory. The conductivities of many other rocks and formations were calculated by measuring gradients (see p. F10) and assuming that the conductivities determined in the laboratory were appropriate for a particular rock type or geologic unit present in part of the section under study and represented by a sample on which laboratory work had been done.

In many places the gradient at the boundaries of different units could not be determined, but only the average gradient for the inhomogeneous stack of layers. If the conductivities of the layers are known, however, it is possible to derive a figure for the average conductivity of these rocks that is appropriate to use with the average gradient.

Equation 6 may be written:

$$q=\frac{k_1T'_1}{L_1}=\frac{k_2T'_2}{L_2}=\dots\frac{k_nT'_n}{L_n}, \tag{9}$$

where T'_1, T'_2, . . . T'_n are the differences in temperature at the boundaries of the layers having conductivities k_1, k_2, . . . k_n.

It is evident that

$$k_1=\frac{qL_1}{T'_1}, \tag{10}$$

and

$$T'_1=\frac{qL_1}{k_1}, \tag{11}$$

similar results being obtained for the other layers.

The average conductivity ($\bar{k}$) of a multiple-layered rock is (by analogy with eq 10)

$$\bar{k}=qd/(T'_1+T'_2+\dots T'_n)$$

where d is the total thickness of all the layers, or (from eq 11)

$$\bar{k}=\frac{qd}{\frac{qL_1}{k_1}+\frac{qL_2}{k_2}+\dots+\frac{qL_n}{k_n}},$$

or

$$\bar{k}=\frac{d}{\frac{L_1}{k_1}+\frac{L_2}{k_2}+\dots\frac{L_n}{k_n}}; \tag{12}$$

because the dimensions of L and d are the same, no conversion factor is needed if L and D are measured in feet and k in cgs units.

The problem of multiple layers is common where a gradient is known between two widely separated points, as at the surface and in a mine opening below, and it is desirable to know the temperature some distance below the point where the deep temperature was measured. If the strata in the section at the locality are known to be horizontal or to dip gently and if reasonable conductivities and thicknesses can be assigned to rock layers above and below the deep point (x), the temperature at a greater depth (y) is readily calculated from the heat flow and average gradient Γ_x. By use of equation 8 the deeper gradient Γ_y is calculated from equation 14 (below) and the deeper temperature difference (T'_y) is found and added to the measured temperature T_x.

$$q=\bar{k}_x\Gamma_x=\bar{k}_y\Gamma_y, \tag{13}$$

where $\bar{k}_x$ is average conductivity above point x and $\bar{k}_y$ is average conductivity between points x and y.

$$\Gamma_y=\frac{\bar{k}_x\Gamma_x}{\bar{k}_y}, \tag{14}$$

$$T'_y=\Gamma_yL_y, \tag{15}$$

and

$$T_y=T'_y+T_x \tag{16}$$

where T_x is the actual temperature measured at the deep point, T_y is the calculated temperature sought, and L_y is the separation of y from x.

Where temperatures could be checked at depth by measurement, the temperatures predicted by the use of equations 12, 14, 15, and 16 were found to be in very satisfactory agreement with actual temperatures if the conductivities given in table 1 were used and if the geologic conditions approximated the simple configurations implied in the equations. In areas of complex geologic structure the results may be less satisfactory, and average gradients determined in steeply dipping rocks of very different conductivities give misleading results if the dip is unknown and assumed to be nearly horizontal.

HEAT FLOW AND THERMAL GRADIENTS IN UNIFORM ROCKS, CONSTANT NONUNIFORM TEMPERATURE AT DEPTH

Using the term "constant" only for time and the term "uniform" only for space, we note that wherever there is a constant nonuniform temperature distribution along a deep horizontal plane and the temperatures at the surface of the earth are essentially constant, the isotherms between the surface and the plane are subparallel to those of the surface at shallow depth and to those of the deep plane at deep levels. Where the horizontal gradient on this plane is nearly equal to the vertical gradient, the heat flux has a direction at an angle to the vertical, and the amount of heat moving vertically differs with depth below the surface. The quantity of heat flowing vertically through the rock is almost the maximum flux if the horizontal gradient at the deep level is small.

The maximum natural horizontal gradient observed in the East Tintic district is about 7°F per 100 feet in the Burgin mine and approximates the vertical gradient. By use of an electronic analog model simulator, as described by Karplus (1958), the approximate dimensions of the region affected by a given distribution of isotherms at depth can be found. Assuming uniform conductivity, we first solved a comparatively simple problem: that of a change from a constant temperature of 140°F uniformly distributed north of a given line in a horizontal plane to 120°F uniformly distributed south of a transition zone 250 feet wide in the same plane, which is assumed to be 1,050 feet below the cooling surface (at 55°F). (See fig. 1.)

The zone of appreciably disturbed isotherms above the temperature-transition area of the deep plane extends about 300 feet vertically and about twice this distance horizontally. The direction of heat flow in the homogeneous assumed medium would be perpendicular to the isotherms, except that above the temperature-transition area of the deep plane and within 200 feet of it horizontally, the direction of maximum heat flow would depart from the vertical by less than 15°; the error introduced by assuming a uniform vertical heat flow thus would be less than 3.5 percent for a depth of 850 feet, and would diminish to less than 1 percent above a depth of 600 feet where the isotherms have a maximum deviation from the the horizontal of about 7°.

Assuming vertical heat flow in areas containing appreciable horizonal gradients has inherent errors. The general effect of ignoring these errors is to show transition zones that may be wider than those that actually exist at water level and to average or minimize sharp thermal valleys and ridges. Although the general pattern of temperatures at water level is revealed by calculation based on gradients measured in the upper part of the section, details of the pattern in complex thermal areas can be determined only by measurements close to water level.

In the second problem (fig. 2) solved by the electronic analog model simulator, the temperature profile approximates one that crosses the area of maximum temperature contrast known in the East Tintic district. A tongue of relatively cool water lies between two broad areas of hot water in the western part of the Burgin mine; even there the distortion of isotherms a few hundred feet above the ground-water level is small, and the narrow thermal trough representing the tongue of cool water would not be apparent from gradients measured more than 300 feet above the water table. The heat flow, however, appears to be approximately uniform along any vertical line except close to areas of sharp temperature changes at water level.

Where evidence to the contrary is lacking, it has been assumed that the quantity of heat moving through the rocks above the deep source is constant for a vertical column a few feet in diameter, and that within a single vertical drill hole the quantity of heat moving up through the walls of the rock is constant unless the heat flow is modified by a current of air or water. Such conditions are usually indicated by the temperature measurements. In the absence of such indications, differences in thermal gradients in a given drill hole can ordinarily be used with confidence to give conductivity ratios by use of equation 7 if due consideration is given to the depth of the hole, ground-water level, and possible horizontal temperature variations at water level.

CONDUCTIVITY OF ROCKS

DIRECT DETERMINATION OF CONDUCTIVITY

The thermal conductivity of many rocks was determined by the method devised by Birch and Clark (1940). The amounts of heat required to maintain certain gradients in an insulated disk are accurately determined, and from these data the conductivity is calculated. Such measurements were made on many specimens from the East Tintic district (table 1). The laboratory measurements of "air-dry" samples made in Washington, D.C., in the early spring (1962) probably represent humidity of about 75 percent saturation and are closer to the measured conductivities of saturated specimens than to those of desiccated ones. The "air-dry" conductivities agree closely with the "in-place" conductivities for gradients measured above the water table; for gradients measured below the water table, however, laboratory values for saturated specimens are clearly appropriate. Apparently the laboratory values for "air-dry" samples closely approximate those of the rocks in their natural damp, but unsaturated, environment above the water level.

INDIRECT DETERMINATION OF CONDUCTIVITY

The conductivity of rocks can also be determined from measurement of other thermal properties such as diffusivity (α):

$$\alpha=\frac{k}{\sigma\rho}, \qquad (17)$$

where σ is specific heat and ρ is specific gravity. Methods of determining the diffusivity of rocks in place

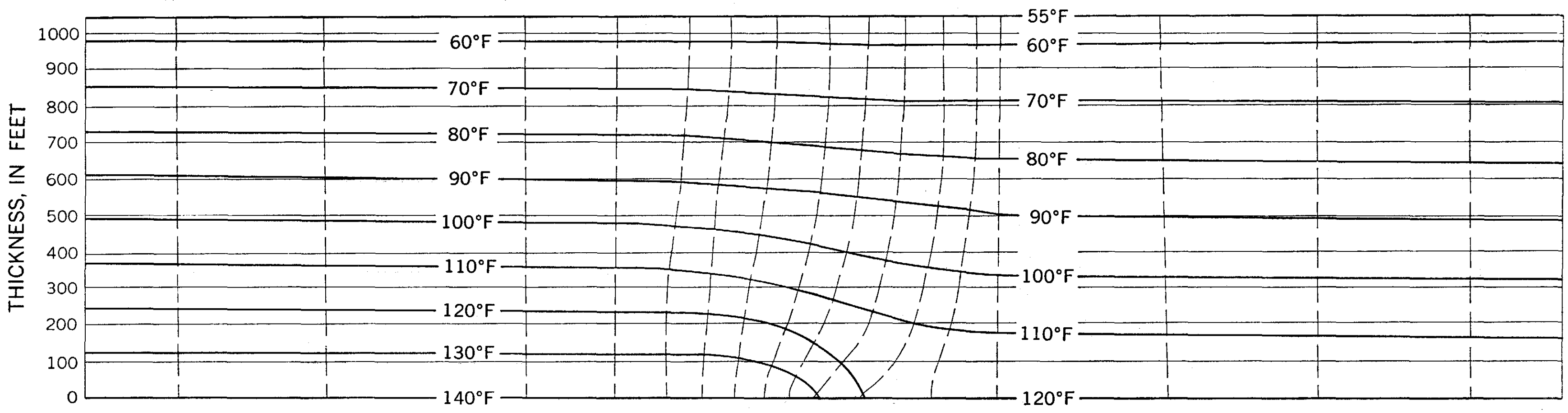

FIGURE 1.—Isotherms in homogeneous rock layers; upper surface maintained at 55°F, bottom surface maintained at 140°F left of temperature transition zone and at 120°F to right of it. Rock layer assumed 1,050 feet thick and transition zone 250 feet wide. Direction of heat flow shown by dashed lines, spaced closely where greatest deviation from vertical occurs.

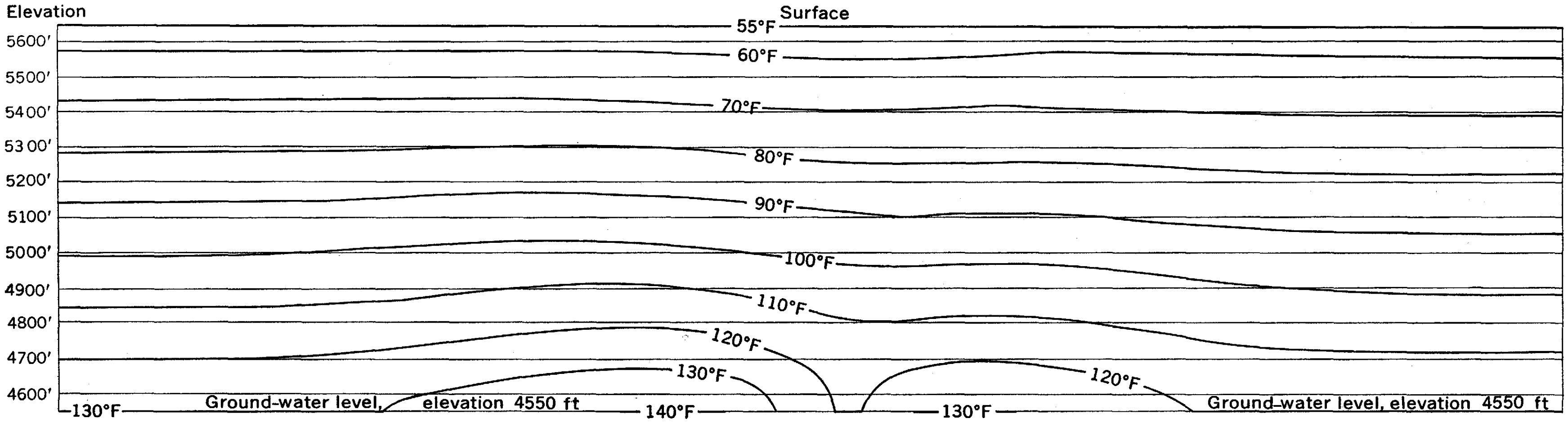

FIGURE 2.—Analog-model-simulator graph showing distribution of isotherms (heavy lines) in homogeneous rock above hot and cool zones. Temperature conditions shown approximate those in northwest-trending section centered near 31,000 N., 26,000 E. (pl. 3).

from temperature measurements in shallow drill holes have been described by Lovering and Goode (1963). Given an accurate value for the diffusivity, the conductivity may then be determined by laboratory measurements of thermal capacity and density. Conductivities of some rocks in the East Tintic district were determined by measuring the specific heat and density in the laboratory and determining the diffusivity of the rocks in place by the shallow-hole temperature method described by Lovering and Goode (1963); this method requires two or more temperature profiles to a depth of at least 35 feet and taken more than a month apart. Where conductivities are known for certain rock units present in an uncased drill hole, the variations in temperature gradient within the hole make possible the calculation, by use of equation 7, of conductivities of other rock units present (see pl. 2.)

CONDUCTIVITY OF ROCKS AND FORMATIONS IN THE EAST TINTIC DISTRICT

The "in-place" conductivity of the rocks in the East Tintic district ranges from about 1.7×10^{-3} for dry tuff to 13×10^{-3} for wet quartzite. The presence of water in porous materials greatly changes the conductivity of the rocks, but in most of the areas investigated we were concerned chiefly with the conductivity of unsaturated rocks above the water table. The data for conductivity obtained by various methods for the rocks and most of the formations shown on the geologic map of the East Tintic district (Lovering and others, 1960) are shown in table 1.

TABLE 1.—*Thermal conductivity* (k) *of rocks and formations of the East Tintic district as determined by various methods*

Stratigraphic-unit symbols, after Lovering and others (1960):
Quaternary units: Qal, alluvium; Qg, gravel.
Tertiary units: Tpr, Packard Quartz Latite, Packard Rhyolite of Lindgren and Loughlin (1919); Tprt, basal tuff, rhyolite tuff of Lindgren and Loughlin (1919); Tac, Apex Conglomerate.
Devonian unit: Dv, Victoria Formation.
Devonian, Silurian, and Ordovician unit: DSOb, Bluebell Dolomite.
Ordovician units: Of, Fish Haven Dolomite; Oop, Opohonga Limestone.
Cambrian units: Ꞓa, Ajax Dolomite; Ꞓop, Opex Formation; Ꞓcc, Cole Canyon Dolomite; Ꞓh, Herkimer Limestone; Ꞓd, Dagmar Dolomite; Ꞓte, Teutonic Limestone Ꞓo, Ophir Formation; Ꞓt, Tintic Quartzite.

Method or calculation used to determine *k*: Sh hl, *k* calculated from diffusivities determined in shallow holes, according to method of Lovering and Goode (1963). Lab, determined experimentally. Γ_p/Γ_x, *k* calculated from ratio of gradients, *k* of one rock known; Γ_x, gradient in rock named in column 3 (eq 7); Γ_p/Γ_x, in pyritized rock; Γ_d/Γ_x, in dolomite; $\Gamma_{\mathrm{Tpr}}/\Gamma_x$, in Packard Quartz Latite. Est, *k* estimated from comparison of lithology with that of rocks of known conductivity. Calc, *k* of formation calculated from conductivities and proportions of rocks comprising it.

$k\times10^3$: Number in parentheses shows number of separate specimens used in laboratory determination of *k*, or number of separate shallow holes in given material to obtain value of average *k*. None of the averages is more than ±0.0003 from the maximum or minimum value of the group.

[All laboratory determinations of conductivity are by E. C. Walker and William Huff, U.S. Geol. Survey, on wafers of rock about 5mm thick cut from drill cores supplied by the Bear Creek Mining Co. Variation of *k* of different wet samples is such that the third digit is meaningless; therefore it is omitted]

Stratigraphic unit		Rock, alteration, weathering, and fracturing	Method or calculation used to determine *k*	Porosity (percent)	$k\times10^3$, in cal cm^{-1} °C^{-1} sec^{-1}			Value of $k\times10^3$ used	
Symbol	Thickness (feet)				Dry	Air dry [1]	Wet [1]	Air dry [1]	Wet [1]
Qal	5-50	Alluvium and colluvium (sand and silt)	Sh hl			2.3 (4)		2.3	
Qg	5-50	Gravel (boulders, pebbles, sand, and silt)	Sh hl			2.5 (3)		2.5	
Tpr	2500+	Quartz latite porphyry, unaltered	Sh hl			4.8 (2)		4.8	
		do	Lab	3.2	4.49	4.81 (1)	5.00	4.8	5.0
		Quartz latite porphyry, plagioclase altered to calcite	Lab	>3.9	4.08	4.82 (1)	4.97	4.8	5.0
		do	Sh hl			4.7 (2)		4.8	
		Quartz latite porphyry, pyritized, unweathered	Sh hl			4.6 (1)		4.5	5.0
		Quartz latite porphyry, pyritized, slightly oxidized	Lab	3.5	4.24	4.50 (1)	4.93	4.5	4.9
	100	do	Γ_p/Γ_x				4.89	4.5	4.9
		Quartz latite porphyry, slightly argillized and iron-stained, weathered.	Lab	9.7	4.08	4.36 (2)	4.50	4.5	5.0
		Quartz latite prophyry, slightly argillized and pyritized, weathered.	Lab	4.5		4.47	[2]	4.5	5.0
		Quartz latite prophyry, moderately sericitized, moderately weathered.	Lab	4.8	4.33	4.25	6.29	4.25	6.3
		Quartz latite porphyry, pyritized and argillized, moderately weathered.	Γ_d/Γ_x			3.98		4.0	
		Quartz latite porphyry, moderately argillized, iron-stained, moderately weathered.	Γ_p/Γ_x			3.75		3.75	
		Quartz latite porphyry, pyritized, argillized, strongly weathered.	Γ_d/Γ_x			3.49		3.5	
		Do	Sh hl			3.1 (5)		3.1	
		Quartz latite porphyry, strongly argillized, strongly weathered.	$\Gamma_{\mathrm{Tpr}[c]}/\Gamma_x$			3.09		3.1	
Tprt	25-250	Quartz latite tuff and breccia, calcitized, pyritized, and argillized.	Γ_d/Γ_x			1.63		1.6	
		Quartz latite tuff, pyritized	Γ_d/Γ_x			2.23		2.2	
Tac	10-150	Pre-lava colluvium, shale, and minor quartzite	$\Gamma_{\mathrm{Tpr}}/\Gamma_x$			2.21		2.2	
		Pre-lava colluvium, quartzite, and minor shale	$\Gamma_{\mathrm{Tpr}}/\Gamma_x$			4.56		4.6	
Dv	280	Dolomite, limy dolomite, and minor sandstone, moderately fractured.	Γ_p/Γ_x			7.39	11.89	7.4	11.9
	100	Dolomite and limy dolomite	Γ_p/Γ_x			8.50		8.5	11.9
DSob	500	Dolomite, fresh	Est—					10.0	11.9
		Dolomite, leached ("sanded")	$\Gamma_{\mathrm{Tpr}}\Gamma/_x$			4.00		4.0	
Of	300	Cherty dolomite	Est					10.6	
Oop	800	Shaly limestone	Est—					5.6	6.0
Ꞓa	600	Cherty and sandy dolomite	Est—					10.0	
		Dolomite, slightly leached and sandy	$\Gamma_{\mathrm{Tpr}}\Gamma/_x$			7.2		7.2	

See footnotes at end of table.

TABLE 1.—*Thermal conductivity* (k) *of rocks and formations of the East Tintic district as determined by various methods*—Continued

Stratigraphic unit		Rock, alteration, weathering, and fracturing	Method or calculation used to determine *k*	Porosity (percent)	$k \times 10^3$, in cal cm^{-1} °C^{-1} sec^{-1}			Value of $k \times 10^3$ used	
Symbol	Thickness (feet)				Dry	Air dry[1]	Wet[1]	Air dry[1]	Wet[1]
Ɛop	250	Shaly limestone and dolomite	Est.					5.6	
Ɛcc	800	Dolomite, fine-grained	Est.					10.0	
		Dolomite, slightly leached and "sanded"	Γ_{Tpr}/Γ_x			6.9		7.0	
Ɛh	400	Limestone, 91 percent; shale, 9 percent	Calc.			5.94		6.0	
Ɛd	80	Argillaceous dolomite	Est.	2.74				7.5	
Ɛte	400	Limestone	Lab.	1.66		6.56 (1)	6.76 (1)	6.5	6.8
		Hydrothermally dolomitized Ɛteequivalent to Ɛte limestone tested.	Lab.	1.75	10.58	10.63 (2)	11.85 (2)	10.6	11.85
Ɛo		Shale	Lab.	4.0	2.88 (1)	3.19 (2)	3.58 (1)	3.2	3.6
Ɛo[3]	80	Limestone, 13 percent; sandstone, 12 percent; shale, 75 percent.	Calc.			3.61		3.6	
Ɛo[4]	145	Limestone, argillaceous	Sh hl.			5.6		5.6	
		Limestone, 63 percent; shale, 37 percent	Calc.			4.38		4.4	
Ɛt[5]	175	Dolomite, 6 percent; sandstone, 17 percent; shale, 77 percent.	Calc.			3.62		3.6	
Ɛt		Quartzite, slightly pyritized	Lab.	1.35	11.83	12.25 (2)	12.93 (2)	12.25	13.0
Ɛt[6]		Quartzite, 90 percent; shale, 10 percent	Calc.			9.56		9.6	
Ɛt	3000	Quartzite, fractured	Γ_{Shale}/Γ_x			10–11		10–11	

[1] Includes laboratory measurements made on air-dried samples, humidity about 75 percent, temperature about 80°F, and in-place conductivities of unsaturated rocks above the water table.

[2] Sample disintegrated during saturation.

[3] Upper part of Ophir Formation.

[4] Middle part of Ophir Formation.

[5] Lower part of Ophir Formation.

[6] Upper part of Tintic Quartzite.

CALCULATION OF TEMPERATURES AT DEPTH

The problems of establishing the temperature of a given point underground are of three kinds. If the temperature has actually been measured at the point in question, the uncertainty is reduced to assessing the actual technique of the temperature measurement and the extent to which the measured temperature represents underground temperatures undisturbed by mining or drilling. A second type of problem occurs when the point for which the temperature is required lies well below points at which temperatures have been measured in different rocks and for which a temperature gradient has been calculated. Here the temperature gradient appropriate to the rocks where the temperatures were measured must be modified for any change in conductivity of the rocks at greater depth. A third type of problem exists where the temperature is known at only one point underground and a calculated temperature is desired for a point at some specified elevation vertically above or below the place at which the temperature is known. This third problem can be solved with a fair degree of success if the conductivities of the rocks are known and if the mean annual temperature at the surface can be approximated. For many places in the East Tintic district where underground temperatures at only a single elevation above the water table have been obtained, the requisite quantities are known with sufficient accuracy to allow calculation of temperatures at the water table to within ±1°F.

GEOLOGIC FACTORS

Inasmuch as the calculation of temperatures below the point where a temperature has been measured depends not only on the gradients above this point but also on conductivity of the overlying and underlying materials, the geology must be interpreted correctly if the structure and type of material lying between the two points are to be correctly predicted. It is also important to know the physical and chemical changes that have occurred in the rocks. The conductivity of a formation may change greatly with alteration—or may be very little affected. The replacement of feldspars in quartz latite by calcite (calcitic alteration) has no measurable effect on the conductivity of the "air-dry" rock; if, however, the feldspar is altered to clay, the conductivity of the rock may be decreased by 20 to 30 percent. Conversely, the change of a limestone to a dolomite may nearly double its conductivity. Although structural discontinuities such as faults and shear zones usually decrease the conductivity of a dry formation markedly, such zones when moist may have a much greater conductivity than the unbroken dry formation. In a section more than a thousand feet thick, however, such changes in conductivity have little effect on the average *k* unless the fault zones are unusually thick.

In a region of complex geology such as is found in the East Tintic district, the calculation of temperatures at a given level must utilize the best geologic information available. Fortunately, the geologic maps and reports prepared by the mining companies working in the East Tintic district were made available for our study; these data, together with the experience we have gained in 20 years of detailed geologic mapping and petrographic study of the district, gave us, we believe, reasonably adequate understanding of the geologic factors in the places where the temperatures have been calculated at the water table.

CALCULATION OF TEMPERATURES AT THE WATER TABLE

As noted in the preceding section, the conductivities of rock types and formations have been measured, calculated, or estimated for the many rock units present in the East Tintic district, and correct calculation of temperature at water level as well as of heat loss depends on assigning correct conductivities to the formations present. The methods of calculating temperatures at various points are perhaps most clearly explained by the following examples that represent the three types of problems just noted.

1. Temperature measured at the water table in an abandoned drift: A temperature of 88°F was measured on the 1,450 (bottom) level of the Tintic Standard mine in the Tintic Standard-North Lily Development Unit at coordinates 31,890 N., 21,240 E. Ventilation had been stopped in this drift 2 years earlier, but the levels above it had not been worked for about 15 years. The elevation of the drift is 4,555 feet, about that of the water table; even though the temperature had been measured with a well-calibrated mercury maximum thermometer, the figure obtained was nevertheless suspect because it was appreciably lower than the temperatures measured in stub drifts nearby. The temperature of 98°F measured in a raise 100 feet above the level indicated a reverse gradient. Temperature measurements were made nearby at higher elevations in the mine and the following temperatures were obtained: 67°F at an elevation of 5,604 feet, 98°F at 4,806 feet, and 98°F at 4,637 feet. The readings in mine workings above 4,555 feet probably approximated the rock temperatures except where they reflected the ventilation history of the drift at 4,555 feet. The reverse gradient between the top of the raise and the bottom level reflects the cooling caused by air circulating in the lower drift for several years prior to the terminating of exploration, which happened 2 years before our measurement was made. The problem then became one of projecting the apparently "natural" gradient from the upper point (elev. 4,806 ft), where a temperature of 98°F had been obtained, downward to the water level at 4,550 feet, the pertinent geologic conditions (see example 2) being considered. The temperature so calculated for water level (102°) was in harmony with those measured in the stub drifts a few hundred feet away.
2. Temperature calculated at depth from a gradient measured for a short distance below the surface: The Chief Consolidated Mining Co. diamond drill hole CC 71, about 2,800 feet northwest of the North Lily shaft, was drilled in porphyry of the Packard Quartz Latite many years before our temperature measurements were made, and presumably the temperature gradient represents the gradient in the surrounding rocks unaffected by the opening. The hole was caved at a depth of 370 feet, and only the gradient in the wet unaltered porphyry of the Packard Quartz Latite, above that depth, could be measured. However, the geology of this area is quite well known from mine workings in the vicinity and from other drill holes. The manner in which data were selected and calculations performed to solve this type of problem is illustrated in the following example.

For calculation of temperature at water level (elev 4,550 ft) directly below drill hole CC 71 at 36,285 N., 18,540 E., the following information was available: Collar at 6,300 feet, bottom in porphyry of the Packard Quartz Latite at an elevation of 5,365 feet; water saturated from surface to depth of 367 feet (elev, 5,933 ft), where the hole was closed by heavy ground. Data for calculation follow (values of k are times 10^3; abbreviations as in table 1):

$$\Gamma\ (\text{measured})=\frac{T_2-T_1}{x_1-x_2}=\frac{58.2-56.0}{6050-5933}=\frac{2.2}{117}=1.897;$$

thus, $\Gamma \approx 1.90$°F per 100 ft and, taking k from table 1, the values are—

Tpr (wet), $k=5.0$, $L_1=6300-5933=367$ feet, $\Gamma=1.9°$,
Tpr (dry), $k=4.8$, $L_2=5933-5270=663$ ft (from surface and subsurface data),
Tprt (dry) $k=1.7$, $L_3=5270-5220=50$ ft (from thickness nearby),
limestone (dry), $k=6.5$, $L_4=5220-5120=100$ ft (geology from mine level at 5,105 ft),
quartzite (dry), $k=12.25$, $L_5=5120-4550=570$ ft (geology from mine level at 5,105 ft), T at 5,933 ft $=58.2°$ (measured).

From equation 12, the average conductivity projected from 5,933 feet to water table at 4,550 feet is

$$\bar{k}\times 10^{-3}=\frac{5933-4550}{\frac{663}{4.8}+\frac{50}{1.7}+\frac{100}{6.5}+\frac{570}{12.25}}=6.03,$$

and

$$\bar{k}=6.03\times 10^{-3}\ \text{cal sec}^{-1}\ \text{cm}^{-1}\ °\text{C}^{-1}.$$

From equation 4, $q=5.0\times 10^{-3}\times 1.9\times 1.82\times 10^{-4}=1.74$ (from measured gradient in wet quartz latite porphyry), where q is calories per square centimeter per second. The average gradient between elevations of 5,933 and 4,550 feet is given by

$$\bar{\Gamma}=\frac{q}{1.82\times10^{-4}\bar{k}}=\frac{1.743\times10^{-6}}{6.03\times10^{-3}\times1.823\times10^{-4}}$$
$$=1.583°F/100 \text{ ft.}$$

T at 4,550 ft$=58.2°+(1.58°\times13.83)=80.12°$;

$\therefore T_w$, the temperature at water level (4,550 ft), $\approx$ 80°F.

3. Temperature at the water table calculated from only a single temperature measured above it in a mine opening: The water-level temperature was calculated for a position about 600 feet south-southeast of drill hole CC 71 discussed above. Temperatures measured at 50 feet, 150 feet, and 250 feet below the collar of CC 71 are respectively 53.7°F, 55.0°F, and 56.0°F, or $\Gamma\approx1.15°$ per 100 feet; by extrapolation of this gradient to the surface, the mean annual surface temperature at the drill hole is $56.0°F(-1.15\times25.0)=53.1°F$. The isotherms shown on the map (pl. 1) south of the collar of CC 71 suggest a slightly higher mean annual surface temperature, and an assumed mean annual temperature of 54°F at the surface (elev., 6,315 ft), over the point of temperature measurement underground, should not be in error by more than a half degree. The underground temperature of 79.5°F was measured on the 900 level (elev., 5,110 ft) of the North Lily mine at coordinates 35,715 N., 18,850 E. (fig. 1). The rocks above this point on the 900 level are limestone, 70 feet ($k=6.5$); dolomite, 580 feet ($k=7.5$); and sericitized quartz latite, 555 feet ($k=4.25$). Below the 900 level the rock is quartzite to the water table 560 feet beneath. The average conductivity of the rocks above the level is given by

$$\bar{k}\times10^3=\frac{6315-5110}{\frac{555}{4.25}+\frac{580}{7.5}+\frac{70}{6.5}}=\frac{1205}{130.6+77.3+10}=5.53,$$

and

$$\bar{k}=5.53\times10^{-3} \text{ cal sec}^{-1} \text{ cm}^{-1} \text{ °C}^{-1}.$$

The average gradient is

$$(79.5-54)/12.05=2.116°F \text{ per 100 feet.}$$

$q=2.12\times5.5\times10^{-3}\times1.8\times10^{-4}=2.13$ mc sec^{-1}cm^{-2} where mc=microcalories (10^{-6} cal). The gradient in the quartzite is

$$\Gamma=\frac{q}{1.8\times10^{-4}k}=\frac{2.13\times10^{-6}}{12.25\times10^{-3}\times1.82\times10^{-4}}$$
$$=0.95°F \text{ per 100 ft.}$$

The temperature at water level T_w is calculated to be:

$$T_w=79.5+(0.95\times5.60)=84.8, \text{ and } T_w\approx85°F.$$

TEMPERATURE OF GROUND WATER

GENERAL FEATURES

As shown by the map (pl. 3), there is a great variation in the temperature of ground water which seems closely related to major structural elements such as the Eureka Lilly fault and the East Tintic thrust fault. The temperature of ground water at the water table in the main Tintic district has been measured in only a few localities. Commonly it is consistent with a temperature gradient of 1.5 to 1.8°F per 100 feet, and with mean annual temperatures at the surface in the 40's or lower 50's. For example, the temperature in the Chief mine at water level (1,780 ft below the surface) at an elevation of 4,800 feet is only 71°F; a gradient of 1.5°F per 100 feet would require a surface temperature there of 44°F. On the eastern side of Tintic Valley, which lies just west of the East Tintic Mountains, the temperature in Mintintic diamond-drill hole 4, 2½ miles south-southwest of Mammoth, is 74.1° at an elevation of 4,900 feet, 1,200 feet below the surface; the temperature gradient in this hole averages 1.67° per 100 feet and the surface temperature is 54°F. However, in Mintintic diamond-drill hole 2A, about half a mile farther south, the temperature was much higher, reaching 98.2° at an elevation of 5,190 feet, 700 feet below the surface. In hole 2A, however, the water was under strong artesian pressure and when tapped at a depth of 455 feet below the collar, it rose almost at once to within 20 feet of the surface.

None of the mines in the main Tintic district are known to have had unusually warm workings even at the greatest depths attained. We may say with confidence that the temperature of ground water is less than 80°F in the region west of the Eureka Lilly fault zone and west of its projected course south from the Eureka Standard fault system; the temperature at the water table is in general that appropriate to its depth where the temperature gradient is 1.5° to 2°F per 100 feet. East of the Eureka Lilly fault zone, the temperatures at the water table are much higher than to the west and reach or exceed 140°F in several places.

As shown on plate 4, the water table in the area of the monzonite stock south of Mammoth stands close to the surface or within a few hundred feet of it; commonly it ranges in elevation from about 6,100 feet to 6,400 feet within the fractured stock but drops abruptly to about 4,900 feet in the Paleozoic rocks just north of the stock and then slopes northward to about 4,800 feet in the vicinity of Eureka. The ground-water surface also slopes both east and west from this north-trending ground-water high, which underlies Eureka. To the east the water table falls at the rate of 45 feet per mile as far as the Eureka Lilly fault zone, beyond which its slope is much less, probably no more than 5 feet per mile for the next 2 or 3 miles. The configuration of the ground-water surface indicates that water is moving from the East Tintic Mountains westward to Tintic Valley and eastward toward Goshen Valley where the ground-water level—as indicated by Lake Utah, a large permanent body of water about 9 miles northeast of the East Tintic district—is at an elevation of 4,490 feet, only 60 feet below the water level in the Burgin shaft. The marked flattening of the water table east of the Eureka Lilly fault zone is in part caused by the greater fracturing and higher permeability of the rocks to the east but also may reflect the change in hydraulic gradient caused by the addition of water from underground springs east of the fault zone.

EAST TINTIC THERMAL AREA

The temperatures in the extreme southeastern part of the East Tintic district are higher than those found farther north (see pl. 3), but the thermal center to which they are related cannot as yet be defined. We will therefore attempt to describe the heat-flow quantitatively only for the thermal area north of the Apex Standard thermal trough, which extends northeastward through the district a short distance southeast of the Apex Standard fault. This thermal trough crosses the Apex Standard workings about a thousand feet southeast of the Apex Standard No. 1 shaft. North of this thermal trough, and generally east of the Eureka Lilly fault zone, an area of at least 20 sq km (square kilometers) has a ground-water temperature at the water table above 80°F, but the position of the 80° isotherm in the eastern part of the district is not known. The average temperature at the water table for an area of 13 sq km, which includes the 80° isotherm to the west, is 104°F. For the area of 10 sq km that includes only temperatures above 90°F, the average temperature is 109°F. As shown on the map (pl. 3), the thermal maxima show two distinct trends: a general north-south alinement along a north-trending thermal ridge, and northeast and southwest extensions from centers along it. Another striking feature of the thermal pattern is the subparallelism of cold-water areas to the northeast- or southwest-trending tongues of hot water.

The linear thermal trends are strongly influenced by the geologic structures of the area. The northeast-trending tongues of hot water are evidently related to major northeast-trending mineralized fractures, and the transition from cool water west of the Eureka Lilly fault into warm or hot water a short distance east is closely related to this major fracture zone. Not all the northeast-trending mineralized fractures, however, are marked by abnormally high temperatures. Both the Eureka Lilly fault zone and the North Lily shear zone are well mineralized in many places, but the ground water along these channels is not appreciably warmer in mineralized ground than elsewhere. It is also noteworthy that the north-trending thermal maximum (the Greyhound thermal ridge) which extends for about a mile northward from the Apex Standard No. 1 shaft through the Greyhound mining claims does not coincide with any structure observable in mine workings or drill holes in this area. The north trend, however, is parallel to the strike of the vertical or overturned beds and strike faults east of the East Tintic thrust fault and probably reflects permeable beds or fractures below the hanging wall of the East Tintic thrust fault, where the footwall of the thrust plate underlies the Greyhound thermal ridge. The highly permeable fault breccia of the East Tintic thrust fault also exerts a marked influence on the distribution of the hot and cold water, for in most places there is a sharp drop in temperature to the east of the thrust fault at water level.

HEAT LOSS

The heat loss at the surface above the abnormally hot ground water depends mainly on the conductivity of the rocks, the mean annual temperature, and distance to the surface. It may, therefore, differ from place to place even where surface temperatures and ground-water temperatures are uniform. The general areas of greatest heat loss, nevertheless, do correspond with the areas of abnormally high temperatures at depth. Isograms through points of equal heat loss have been drawn for intervals corresponding to differences of 1 $\mu c\ cm^{-2}\ sec^{-1}$ (microcalorie per square centimeter per second) and are shown on the map (pl. 5). Inasmuch as the heat loss at the surface in the areas having ground-water temperatures less than 80°F corresponds in general to a measured loss of 1.5 to 1.9 $\mu c\ cm^{-2}\ sec^{-1}$, we assume that any heat loss of more than 2 $\mu c\ cm^{-2}\ sec^{-1}$ is anomalous. On the basis of this assumption, the total anomalous heat loss in the East Tintic thermal area has been computed (table 2) and is found to approximate 265,000 calories per second for an area of 14.7 sq km. This heat loss amounts to 8.4 trillion (8.4×10^{12}) calories per year.

SOURCES OF HEAT

Only two sources of heat for the abnormally high temperatures at water level seem possible: oxidizing

TABLE 2.—*Heat flow in excess of 2 mc cm^{-2} sec^{-1} in East Tintic area north of Apex Standard fault zone*

Region (mc)	Area (cm^2)	Mean excess heat flow (mc cm^{-2} sec^{-1})	Rate of excess heat flow (mc sec^{-1})
2–3	44.6×10^9	0.5	22.3×10^9
3–4	47.1×10^9	1.5	70.6×10^9
4–5	28.0×10^9	2.5	70.0×10^9
5–6	18.3×10^9	3.5	64.0×10^9
6–7	8.35×10^9	4.5	37.6×10^9
>7	$.178\times10^9$	5.0	$.89\times10^9$
Total	$147\times10^9=14.7$ km^2		$265\times10^9=265,000$ cal per sec

sulfides and subterranean hot springs. During the early years of the present investigation, the heat generated by sulfides in the Tintic Standard mine was very noticeable and a few years earlier had caused a mine fire which was brought under control only with difficulty. The temperature of the rocks was a matter of great interest for mining and ventilation engineers because of the practical problems of operating the mine; this interest and the possibility that abnormal geothermal gradients surrounding oxidizing sulfide bodies could be used in prospecting for blind ore bodies prompted the U.S. Geological Survey to begin its broader thermal studies. As the temperature data were accumulated over the years, however, they showed less and less direct relation to ore bodies, but until the Bear Creek Mining Co.'s extensive drilling campaign and their development of the Burgin mine, the temperature data were insufficient to show a regional pattern. By the late 1950's, however, the theory that areas of high heat flow always indicate the presence of sulfides at depth was becoming increasingly untenable. The hot saline water discovered on June 15, 1961, in the Burgin mine eventually provided the evidence necessary to establish the presence of subterranean hot springs; indeed, the first analysis of this saline water led Donald E. White, who had specialized in the study of thermal waters, to point out that this saline water was typical of a family of widely distributed surface saline hot springs in Utah, which had been sampled and analyzed in conjunction with his project. Several of the analyses shown in table 3 come from Mr. White's project and are published here for the first time.

The oxidation of sulfides in the main Tintic district causes no appreciable rise in temperature at ground-water level, and temperatures measured in the Chief mine in the vicinity of oxidizing ore bodies in limestone above the water table showed a maximum increase of only 3°F above the normal for the level at which temperatures were measured. Similarly in the North Lily mine, temperatures of 1° to 2° above normal were found in partly worked stopes around small ore bodies adjacent to the Eureka Lilly fault zone and well north of the North Lily shear zone. Oxidation of sulfides in the San Manuel district in Arizona where the ore is undisturbed by mining caused an increase in the thermal gradient of 0.6°F or less per 100 feet (Lovering, 1948); if applied to the East Tintic district, such a change in gradient would increase the temperature at water level by less than 10°F, not enough to cause the temperatures of more than 100°F at ground-water level over large areas in the East Tintic district. Where sulfides have been disturbed by mining, the rate of oxidation may be sufficient to greatly increase the temperatures locally, but this increase could not possibly cause the pattern of thermal maxima in the vicinity of the Burgin and Apex Standard mines. No ore had been extracted from the Burgin mine at the time the temperature measurements were made and very little had been taken from Apex Standard mine when it was active 15 years before we measured the temperatures in it. Oxidizing sulfides, though capable of raising nearby temperatures a few degrees where undisturbed by mining, are completely incapable of producing the high temperatures found in the East Tintic thermal area.

If oxidizing sulfides cannot supply the heat, then it follows that most of the heat must be supplied by subterranean hot springs. This conclusion is also supported by the composition of the water and the pattern of hot and cold tongues of water. Where the water table was first penetrated by an inclined winze in the Burgin mine, the water had a temperature of 139°F; a sample of this water contained 6,050 ppm (parts per million) of total solids. This sample was taken on December 26, 1960, shortly after the winze reached water level. Heavy pumping was necessary through the ensuing months while the winze was slowly advanced. Water samples taken during the next several months showed a gradual increase in total salinity, which reached 7,480 ppm by November 1961 when the temperature was 138°F. During this period the chloride content increased from 3,090 to 3,670 ppm. The increase in total solids with heavy pumping strongly suggests a subterranean hot spring source, and the slight difference in temperature reflects only a slight perturbation probably caused by the cooling of the rock in the area of drawdown near the ventilated opening through which the water moved on its way to the sump. A sample from a drill hole 200 feet southwest of the winze sample showed a temperature of 128°F and contained only 72 ppm of chloride. Other samples in the tongue of cool water ranged from 780 to 1,560 ppm of chloride, whereas samples from the tongue of hot water to the south proved it was also a high-chloride water which contained more than 3,500 ppm of chloride.

The analyses show that the hot water in the Burgin mine is a saline water similar in composition to several surface hot springs in Utah (tables 3, 4). The striking similarity of the Burgin water to these hot-springs waters is especially evident when the ratios of some of the major elements are compared (table 3). The ratios are nearly independent of dilution by fresh water or of concentration by evaporation, and thus emphasize the chemical relations that characterize this family of hot springs. There seems little doubt that the surface hot springs—represented by analyses 8 to 18—belong to a family characterized by high chloride, low to moderate sulfate, and low to moderate bicarbonate. The contrast between waters of this family and either

TABLE 3.—*Chemical analyses of water from Eureka city well, Burgin mine,*

[Analyses in parts

Analysis	Source	Date of collection	Constituents																		
			SiO_2	Al	Fe	Mn	Ca	Mg	Sr	Ba	Na	K	Li	As	Pb	Zn	[3] Ra	U	NH_4	HCO_3	SO_4
1	Eureka city well.	Dec. 31, 1940	44.6	-----	None	-----	59.2	34.2	--------	--------	7.7	-----	-----	-----	-----	-----	--------	--------	-----	-------	51.3
2	North Lily mine.	Nov. 14, 1942	190	35	-----	-----	340	112	--------	--------	--------	-----	-----	-----	-----	-----	--------	--------	-----	-------	936
3	Burgin mine	Dec. 22, 1960	[1]30	[1].3	7	[1]0.2	480	[2]240	[1] 0.36	--------	[1]1,815	[1]121	[1]12.1	[1]1.2	[1]0.2	[1]0.4	--------	--------	-----	-------	380
4	Burgin mine	May 25, 1961	30	-----	-----	-----	315	62	--------	--------	1,700	-----	-----	-----	-----	-----	--------	--------	-----	644	356
5	Burgin mine	June 15, 1961	43	.16	.00	.7	256	66	[1]3.1–0.31	[1]0.3–0.03	2,000	182	5.0	[1].6	-----	-----	--------	--------	1.9	695	360
6	Burgin mine	Aug. 28, 1961	-----	.16	.00	.00	396	39	--------	--------	--------	-----	-----	[1].6	-----	-----	--------	--------	.00	689	389
7	Burgin mine	Oct. 1, 1961	48	.00	1	.64	375	53	15	[1].4–.04	2,270	228	5.5	.58	-----	-----	--------	2.4×10⁻³	.24	749	391
8	Stinking Springs.	June 8, 1954	48	1.05	.03	.00	946	297	31	4.1	12,600	571	6.9	-----	-----	-----	--------	--------	.40	324	111
9	Cooper Hot Springs.	Prior to 1906	23.5	0	0	-----	267.7	35.1	--------	--------	591.4	37.5	.6	-----	-----	-----	--------	--------	-----	-------	167.6
10	Roosevelt Hot Spring.	Sept. 11, 1957	313	<.04	<.04	.0	22	.0	0	0	2,500	488	27	2.0	.00	.00	--------	--------	.0	156	73
11	Crystal Hot Springs.	Oct. 27, 1951	31	-----	.61	.00	803	219	--------	--------	14,700	763	8	-----	-----	-----	--------	--------	-----	465	466
12	Utah Hot Springs.	Apr. 5, 1958	38	.46	.42	1.9	1,140	70	--------	--------	7,030	901	9.9	.00	.01	.00	39×10⁻⁹	.3×10⁻³	5.4	192	189
13	El Monte (Ogden) Hot Springs.	Nov. 3, 1951	53	-----	.08	.89	337	8	--------	--------	2,740	407	6	-----	-----	-----	--------	--------	-----	200	100
14	Hooper Hot Spring.	Nov. 3, 1951	28	.4	.01	1.7	523	118	--------	--------	2,390	283	2	-----	-----	-----	100×10⁻⁹	.4×10⁻³	-----	245	36
15	Joseph Hot Spring.	Oct. 11, 1957	84	.1	.56	.41	264	44	--------	--------	1,380	45	1.5	-----	-----	.00	2.7×10⁻⁹	1.9×10⁻³	-----	412	1,250
16	Red Hill Hot Spring.	Sept 10, 1957	54	.0	.38	.1	288	33	--------	--------	555	67	1.1	.19	-----	.10	--------	--------	.4	416	833
17	Abraham Hot Springs	Oct. 12, 1957	75	.0	.0	.75	352	49	0	0	770	54	.0	.06	-----	-----	--------	--------	.1	142	704
18	Becks Hot Springs	May 19, 1942	32	8	-----	-----	653	134	--------	--------	4,045	444	-----	-----	-----	-----	--------	--------	-----	119	875
19	Becks Hot Springs.	Nov. 3, 1951	36	-----	.02	.10	720	125	--------	--------	4,050	262	3	-----	-----	-----	--------	--------	-----	227	879

[1] Calculated from spectrographic analyses by Kennecott Copper Corp.
[2] Stearns, Stearns, and Waring (1937, p. 183).
[3] Concentration of radium∼micromicrocuries per liter.

1. Eureka city drinking-water supply; composite sample collected 280 ft from dug wells 2½ miles northeast of Eureka. Collected by H. J. Hansen; analyzed by Utah State Board of Health.
2. Utah County, face of 1517 crosscut, 1500 level. Analyzed by International Smelting and Refining Co.; trace of free oxygen reported.
3. Utah County, 269 winze. Collected by J. D. Bush and Roger Banghart, Bear Creek Mining Co.; analyzed by Kennecott Copper Corp. laboratories.
4. Utah County, winze 269, 400 ft down incline from 1050 level (70 ft below water level). Collected by T. S. Lovering and H. T. Morris; analyzed by U.S. Geol. Survey Salt Lake City laboratory; lab. No. 23665.
5–7. Utah County, winze 269, 230 ft down incline from 1050 level (20 ft below water level). Collected by J. D. Bush and Roger Banghart, Bear Creek Mining Co. Chemical analyses by H. C. Whitehead (lab. Nos. 1517, 1536, 1525, respectively); spectrographic analyses by Harry Bastron and Robert Mays (lab. Nos. 62M11, 62M13, 62M12, respectively).
8. Located 6.8 miles northwest of Corinne, Box Elder County (White, Hem, and Waring, 1963, table 15, analysis 5).

fresh shallow ground water or normal mine water is clearly shown by analyses 1 and 2 in table 3. Analysis 1 represents water from the perched water table in the gravels and the upper part of the lavas just northwest of Homansville Canyon in the area from which Eureka derives its water supply. The contrast in its ratios of Na:Ca and of Cl:SO_4 and Cl:HCO_3 marks it as completely unrelated to the Burgin water in origin. Although the water from the North Lily mine (table 3, anal. 2) is dilute as compared to the chloride water of the Burgin mine, it contains far more sulfate and very little chloride and presumably represents water of the sort shown in analysis 1 after it had passed through oxidizing sulfides. The water of the North Lily mine is nevertheless relatively cool (about 80°F) in contrast to that of the Burgin mine.

North Lily mine, and of saline hot springs of Utah having temperatures above 115°F

per million]

Constituents—Continued										pH	Temperature		Discharge	Ratios, by weight										
Cl	F	Br	I	NO_2	NO_3	PO_4	B	H_2S	Total Solids		°F	°C	(gal per min)	$\frac{Ca}{Na}$	$\frac{Mg}{Ca}$	$\frac{K}{Na}$	$\frac{Li}{Na}$	$\frac{SiO_2}{\text{total solids}}$	$\frac{HCO_3}{Cl}$	$\frac{SO_4}{Cl}$	$\frac{F}{Cl}$	$\frac{Br}{Cl}$	$\frac{B}{Cl}$	$\frac{I}{Cl}$
53.0	0.1	----	----	----	4.0	----	----	----	384	7.6	----	----	----	7.69	0.578	----	----	0.116	----	0.968	----	----	----	----
70	----	----	----	----	----	----	----	----	1,578	(acid)	----	----	----	----	.318	----	----	.12	----	13.4	----	----	----	----
3,090	----	----	----	----	----	----	[1]6	----	6,050	6.65	139	59.4	----	.26	.50	0.07	0.006	.005	----	.12	----	----	0.002	----
2,720	----	----	----	----	2.1	----	----	----	5,500	6.6	127	53.9	----	----	.20	----	----	.005	0.24	.13	----	----	----	----
3,100	1.5	----	----	0.23	.0	0.31	6.0	----	6,310	6.3	140	60	----	.13	.26	.09	.0025	.007	.22	.12	0.0005	----	.002	----
3,590	----	----	----	.00	.0	----	----	0.0	----	6.4	138	58.9	----	----	.102	----	----	----	.19	.11	----	----	----	----
3,670	1.7	4.1	0.3	.00	3.1	.54	5.6	.3	7,480	6.1	130	54.4	----	.165	.141	.100	.0024	.006	.20	.11	.0005	0.001	.0015	8×10^{-5}
21,600	1.9	15	1.3	.00	.0	3.6	----	60	36,600	6.7	[2] 122	48	5	.075	.31	.045	.00055	.0013	.013	.0051	.00009	.0007	.00017	6×10^{-5}
634.6	----	----	----	----	----	----	----	----	1,794	----	144–156	62.2–68.8	[2] 100	.45	.13	.06	.001	.01	----	.26	----	----	----	----
4,240	7.5	3.3	.3	.01	11	2.2	38	----	7,800	7.9	131	55	[2] 3	.009	<.04	.20	.011	.04	.04	.02	.002	.0008	.900	7×10^{-5}
25,000	.0	----	----	----	----	----	4.5	----	42,200	6.5	132.4	55.8	1,320	.05	.37	.052	.0005	.0007	.02	.02	$<4\times10^{-5}$	----	.0002	----
13,300	3.2	8.2	.2	.00	.0	.00	5.1	.0	22,800	7.3	134	56.7	[2] 20–25	.162	.061	.138	.0014	.0027	.01	.01	.0002	.0006	.0004	1.5×10^{-5}
5,060	3.4	----	----	----	----	----	3.6	----	8,820	7.4	136.6	58.1	30±	.12	.024	.15	.002	.006	.04	.02	.0007	----	.0007	----
5,100	.6	----	----	----	----	.08	1.2	----	8,600	6.8	139.8	60	30	.22	.23	.12	.0008	.003	.05	.007	.0001	----	.0002	----
1,690	6.0	----	----	----	.0	.04	4.8	----	4,970	6.6	147	64	[2]5–30	.19	.17	.03	.001	.02	.24	.74	.003	----	.003	----
660	3.0	0.3	.0	----	.0	.29	93.9	----	2,705	6.4	168.8	76	50	.52	.12	.12	.002	.02	.63	1.27	.005	.0005	.006	$<10^{-4}$
1,480	4.5	1.8	.1	.13	2.5	.06	.9	----	3,560	6.6	179.6	82	25±	.46	.14	.07	$<10^{-4}$	.02	.10	.47	.003	.001	.0006	7×10^{-5}
7,668	----	----	----	----	----	----	----	----	13,872	6.5	----	----	----	.16	.21	.11	----	.002	.02	.11	----	----	----	----
7,260	2.3	----	----	----	----	----	2.6	7.5	13,500	----	132	55.5	----	.18	.17	.06	.0007	.003	.03	.12	.0003	----	.0004	----

9. Located ½ mile east of Monroe, Sevier County (Richardson, 1907, p. 37).

10. Located 15 miles northeast of Milford on western slope of Mineral Mountains, Beaver County. Collected by D. E. White; analyzed by H. C. Whitehead; lab. No. 420.

11. Located 12 miles north of Brigham City, Box Elder County. Analyzed by U.S. Geol. Survey Salt Lake City laboratory; lab. No. 7683.

12. Northernmost of 4 springs, 8 miles north of Ogden near boundary between Weber and Box Elder Counties. (White, Hem, and Waring, 1963, table 16, analysis 3).

13. Mouth of Ogden Canyon, Weber County. Analyzed by U.S. Geol. Survey Salt Lake City laboratory; lab. No. 7679.

14. Located 3 miles southwest of Hooper, Davis County. Analyzed by U.S. Geol. Survey Salt Lake City laboratory; lab. No. 7681.

15. Located 1 mile southeast of Joseph, Sevier County. Collected by D. E. White; analyzed by H. C. Whitehead and J. P. Schuch; lab. No. 440.

16. Red Hill Spring, Monroe Hot Springs, 1 mile northeast of Monroe, Sevier County. Collected by D. E. White; analyzed by H. C. Whitehead; lab. No. 419.

17. Near road north-northwest of Delta, Juab County. Collected by D. E. Wite, analyzed by J. P. Schuch; lab. No. 421.

18–19. 4 miles north of Salt Lake City, Salt Lake County. 18, collected by Leonard Tanner; analyzed by N. E. McLachlan. 19, analyzed by U.S. Geol. Survey Salt Lake City laboratory; lab. No. 7677.

TABLE 4.—*Spectrographic analyses of residues of water samples from Burgin mine and Abraham Hot Springs*

[Analyses in weight percent; M, major constituent]

Element	Burgin mine, winze 269					Abraham Hot Springs
	Field sample (anal., table 3; date collected)					
	[1] 4121 (3; 2–22–60)	[1] 4625 (5–11–61)	[2] 1517 (5; 6–15–61)	[2] 1536 (6; 8–28–61)	[2] 1525 (7; 10–1–61)	[3] D 3 (17; 1963)
Ag	0. 0001	0. 0005	0	0	0	<0. 0001
Al	. 005	0. 1	<. 0005	<. 0005	<. 0005	0
As	. 02	. 10	0	0	0	. 3
B	. 1	. 1	. 005–. 05	. 05	. 005–. 05	. 001
Ba	. 008	0	. 0005–. 005	. 0005–. 005	. 0005–. 005	>. 5
Be	. 00008	0	0	0	0	. 005
Ca	5	22	M	M	M	3
Cb	. 001	0	---	---	---	0
Cd	. 005	0	0	0	0	<. 002
Co	. 000	. 0001	0	0	0	<. 0005
Cr	. 0005	. 0007	0	0	0	<. 0005
Cu	. 002	. 0004	<. 0005	. 0005	<. 0005	. 0005
Fe	. 02	. 3	. 0005	. 0005	. 0005	>20
Ga	. 0004	0	0	0	0	. 003
Ge	. 0001	0	0	0	0	<. 002
In	. 00007	0	0	0	0	0
K	2	2	---	---	---	0
La	0	0	---	---	---	. 005
Li	. 2	. 5	0	0	0	---
Mg	1	4	. 05–. 5	. 05–. 5	. 05–5	. 15
Mn	. 003	0	<. 0005	. 005	<. 0005	>. 5
Mo	. 00005	0	0	0	0	<. 0002
Na	30	18	---	---	---	0
Ni	. 00008	. 0000	0	0	0	<. 0005
P	. 05	0	0	0	0	0
Pb	. 003	. 007	0	0	0	>. 1
Pt	0	0	Tr	<. 05	0	0
Rb	. 006	0	---	---	---	---
S	4	4	---	---	---	---
Sb	. 02	0	0	0	0	. 15
Sc	0	0	0	0	0	<. 0005
Si	. 5	. 7	. 05–. 5	. 05–. 5	. 05–. 5	0
Sn	. 0003	0	0	0	0	<. 001
Sr	. 06	. 05	. 05–5	. 05–. 5	. 05–. 5	. 5
Ti	. 0007	0	0	0	0	. 02
V	. 0006	. 0008	0	0	0	. 003
W	0	0	0	0	0	. 5
Y	. 002	. 002	---	---	---	. 0015
Zn	. 007	. 1	0	0	0	. 02
Zr	. 001	0	0	0	0	<. 002

[1] Analyst, R. E. Word, Kennecott Copper Corp. Looked for, but not detected: Au, Bi, Ce, Dy, Er, Eu, Hf, Hg, Ho, Lu, Nd, Os, Pd, Re, Rh, Ru, Se, Sm, Ta, Tb, Te, Th, Tl, Tm, U, Yb.

[2] Analysts, Harry Bastron and Robert Mays, U.S. Geol. Survey. Looked for, but not detected: Au, Bi, Hf, Ir, Nb, Pd, Ta, Tl.

[3] Analyst, Maurice De Valliere, U.S. Geol. Survey. Sample of black ferruginous precipitate collected by W. R. Griffitts; lab. No. 63–1082–S.

Table 5 shows the chloride content and temperatures of nine pairs of water samples taken from the same points in the Burgin mine about a month apart; although most pairs of samples are nearly the same in both temperature and composition, a few show striking differences which must be attributed to the movement of chloride-rich waters through dilute meteoric water. The decrease in temperature of the water samples taken from the inclined winze may be due to the cooling of the rock during mining operations, but the increase in chloride content can only be explained by the introduction of more concentrated brine. The heavy pumping as the winze progressed probably caused deep hot-springs water to move through the fractured rocks that had previously been in equilibrium with the tongue of cold water just south of the winze. (See fig. 4).

All the evidence accumulated by the end of 1963 supports the conclusion that the centers of highest temperature reflect the circulation of thermal water which rises along a north-trending fracture zone in the footwall of the East Tintic thrust plate and spreads northeastward or southwestward along northeast-trending fractures in its hanging wall.

APPROXIMATE FLOW OF SUBTERRANEAN HOT SPRINGS

Although the amount of hot water supplied by the hot springs cannot be directly measured, the flow can

TABLE 5.—*Variation in chloride content, temperature, and relative deuterium content of water samples from Burgin mine*

[Water samples collected April 11 (1) and May 16 (2) 1962 from drill holes from 1,050 level]

USGS field No.		Location coordinates	Drill-hole No. or location	Chloride (ppm)		Temperature (°F)		Deuterium analyses (May 16 samples)	
1	2			[1]1	[2]2	1	2	Lab. No.	D:H (percent)[3]
401-TL-62a	401-TL-62m	30,392 N., 26,170 E	ET 71	564	510	120	120		
402	402	31,160 N., 26,040 E	B 29	1,380	1,390	139	138		
403	403	30,900 N., 26,720 E	269 winze	3,560	[4]3,480	139	119	I.F. 3160-17	−13.1
404	404	31,100 N., 27,433 E	ET 26	3,440	3,200	110	110		
405	405	30,425 N., 26,900 E	B 8	1,560	2,040	125	123		
406	406	30,717 N., 26,485 E	ET 80	73	80	128	128	I.F. 3160-16	−10.9
407	407	30,535 N., 26,697 E	B 33	2,290	2,430	126	127		
408	408	30,713 N., 27,715 E	B 32	3,520	3,590	135	137	I.F. 3160-19	−11.0
409	409	30,825 N., 25,862 E	B 24	780	4,070	139	139	I.F. 3160-18	−12.4

[1] Analysis by K. E. Edwards, U.S. Geol. Survey.

[2] Analysis by Kennecott Research Lab., Salt Lake City, Utah.

[3] Isotopic analyses by Irving Friedman, U.S. Geol. Survey; the figures show relative deuterium content of the sample compared to that of the standard mean ocean water (SMOW):

$$\left(\frac{D}{H}\right)\text{ percent SMOW}=\left[\frac{\left(\frac{D}{H}\right)_{\text{sample}}-\left(\frac{D}{P}\right)_{\text{smow}}}{\left(\frac{D}{H}\right)}\right]\times 100$$

[4] Sample taken May 11, 1962.

be estimated within certain limits if we know the amount of heat escaping at the surface and the amount of hot-springs water required to raise the temperature of the meteoric water to that which has been found. As noted earlier, the heat loss for much of the East Tintic thermal area can be computed from the data shown in figure 6, and in turn, the heat loss that is above normal can be equated with the amount of hot water necessary to supply this heat if the difference in temperature between the subterranean thermal water and the cool meteoric water is known. Assuming that the average temperature of "normal" ground water at the water table is about 80°F (26.67°C) and that the maximum temperature at the water table of the rising thermal water is 143°F (61.65°C), we have a temperature difference of 63°F (35°C). The total anomalous heat flow at the surface in the area studied (pl. 5) is approximately 265×10^3 calories per second (table 2); it represents heat from a quantity of water 35°C hotter than the normal water-table temperature. This water has 35 calories per gram more than does water at 80°F (26.67°C), and the total excess heat flow is thus equivalent to an influx of water at 143°F of about $\frac{265\times10^3}{35}$ grams per second, or 7.6×10^3 grams per second—approximately 2 gallons per second.

The amount of hot water required to heat the cool ground water is more difficult to estimate but is a comparable quantity. An approximation of the amount of ground water moving through the thermal area must first be made; the general slope of the water table and the control of ground-water movement by major fractures are known. (See pl. 4.) Because of the accelerated evaporation and limited rainfall during most of the hot summer months (June–September), no appreciable amount of precipitation reaches the water table; however, during the winter and early spring when runoff, evaporation, and transpiration of water by vegetation are minimal, an appreciable part of the total precipitation probably moves down to the water table and slowly migrates away from the hydrologic divide that passes under Mammoth and Eureka. The precipitation at Eureka, just west of the East Tintic area, and at Elberta, a few miles to the east, averages about 15 inches and 10½ inches, respectively. (See table 6.)

Quantitative data on rate of ground-water recharge from precipitation in areas near the East Tintic district are lacking. Recently, however, Gates (1963) estimated water loss in the Oquirrh Mountains, which are just north of the Tintic Mountains. Evapotranspiration is the greatest single factor in water loss. Although Gates did not have exact figures for transpiration losses in the Oquirrh Mountains, he utilized the quantitative figures obtained by Croft and Monninger (1953) in the Wasatch Mountains east of Farmington, Utah, where 44 percent of the total precipitation is transpired in areas covered by aspen and herbaceous vegetation. Unfortunately these figures apply chiefly to steep mountain watersheds between 7,000 and 10,000 feet in altitude and are not directly applicable to the Oquirrh Mountains. Gates, however, noted that the percentage of precipitation represented by evapotranspiration increases as total precipitation decreases and as average temperatures rise. He therefore concluded that two-thirds, or 67 percent, of the precipitation is lost by evapotranspiration in the Middle Canyon area of the Oquirrh Mountains. Adding to this figure the measured water losses from surface runoff, channel underflow, springs, drains, and wells, Gates found that approximately 16 percent of the total precipitation is unaccounted for and can be presumed to be leakage to permanent ground water. The average precipitation in the Tintic area is about half that in the drainage basin of Middle Canyon, and the proportion of water lost by evapotranspiration thus should be substantially greater there than in the Oquirrh Mountains. Because

TABLE 6.—*Precipitation and ground-water recharge in the East Tintic district*

[Data in inches; precipitation data from U.S. Weather Bur. (1952, p. 189)]

Month	Elberta			Eureka			Ground-water recharge, East Tintic thermal area	
	20-yr average precipitation	1950 precipitation	Departure from 20-yr average precipitation	20-yr average precipitation	1950 precipitation	Departure from 20-yr average precipitation	In Paleozoic bedrock [1]	In lava bedrock [2]
Jan	0. 81	1. 61	+0. 80	1. 36	2. 32	+0. 96	0. 136	0. 068
Feb	1. 02	. 42	—. 55	1. 52	. 70	—. 82	. 152	. 076
Mar	1. 01	. 38	—. 73	1. 47	. 66	—. 81	. 147	. 073
Apr	1. 02	. 47	—. 55	1. 34	. 17	—1. 17	. 134	. 067
May	1. 15	1. 02	—. 13	1. 36	. 75	—. 61	. 068	. 034
June	. 59	. 09	—. 50	. 92	. 33	—. 59	----------	----------
July	. 83	. 88	+. 05	. 98	1. 48	+. 50	----------	----------
Aug	. 79	. 04	—. 75	1. 38	. 12	—1. 26	----------	----------
Sept	. 71	. 64	—. 07	. 62	. 53	—. 09	----------	----------
Oct	. 99	. 38	—. 61	1. 59	. 33	—1. 26	. 080	. 040
Nov	. 73	1. 02	+. 29	. 96	1. 30	+. 34	. 096	. 048
Dec	. 74	. 48	—. 26	1. 58	1. 02	—. 56	. 158	. 079
Total	10. 39	7. 48	—2. 01	15. 08	9. 71	—5. 37	. 971	. 485

[1] Estimated from 20-yr average at 10 percent of precipitation during spring, late fall, and winter and at 5 percent during May and October.

[2] Estimated as half that in Paleozoic bedrock.

the East Tintic area has higher temperatures and less total rainfall, it seems reasonable to assume that evapotranspiration there is closer to 75 percent than to 67 percent and that, accordingly, no more than 5 to 10 percent of the total precipitation reaches the water table.

If we assume that 10 percent of the total rainfall during the late fall, winter, and spring reaches the water table in the East Tintic Mountains, we may assign no more than about 1 inch (2.5 cm) of the precipitation near Eureka to ground-water recharge. Much more of the water falling on the Paleozoic bedrock surfaces is available for deep water recharge, however, than is available from the rainfall on the lava terrane. Perched water tables under gravel are common on lava but are rare in the Paleozoic rocks. The lava is relatively unfractured, but the Paleozoic bedrock is highly fractured, folded, and stratified and thus is well suited for conducting moisture to depths; it is assumed arbitrarily that in the lava terrane half as much moisture reaches the water table as in the areas of Paleozoic bedrock. The map (pl. 4) shows the areas of Paleozoic bedrock and lava and also the approximate elevation of the water table. It is further assumed that recharged ground water moving through the East Tintic thermal area is derived from precipitation in the area that lies east of the hydrologic divide, west of the eastern part of the Greyhound thermal ridge, and between the Apex Standard and the Centennial-Homansville Canyon fault zones, and that the ground water moves generally toward the northeast and east. The area assumed to supply ground water is shown by pattern on the map (pl. 4).

The recharge area includes 11.2 sq km of Paleozoic bedrock and 19.4 sq km of lava terrane. Computation—$((2.5\times 11.2\times 10^{10})+(1.25\times 19.4\times 10^{10}))/10^{3}$—shows that the annual increment of water available from precipitation in these areas is in the order of 52×10^{7} liters per year, equivalent to about 16 liters per second. Of the recharge area, 13.2 sq km is underlain by ground water that has a temperature in excess of 80°F; the average temperature at the water table for this ground water is computed as 104°F. To raise 16 liters per second of water having a temperature of 80°F to an average temperature of 104°F by addition of water having an initial temperature of 143°F requires about 6 liters per second or approximately 100 gallons per minute. If the initial temperature of hot-springs water was higher or the recharge was less than assumed, the volume of hot water required would diminish proportionately.

The heat loss from the rock cover and the heat required to raise the temperature of the incoming ground water would be supplied by the sum of the two quantities of water at 143°F whose equivalence in heat energy has been calculated (above and p. F19); the sum is 3.7 gallons per second or 220 gallons per minute. After allowing for the uncertainties entering into the calculations, we can say that the subterranean hot springs in the area north of the Apex Standard thermal trough must have a flow of no more than a few hundred gallons per minute.

ORIGIN OF THERMAL WATERS

The isotopic analyses of the chloride brines (table 5) by Irving Friedman show deuterium-hydrogen ratios

(D:H) that are completely different from those characteristic of ratios in undiluted connate waters. The D:H ratios for normal connate water are commonly between 0.0 and 4 percent less than for standard mean ocean water (SMOW). The ratios of the Burgin water samples range from −10.9 percent for the most dilute water (73 ppm Cl), to −13.1 percent for the most concentrated brackish water analyzed for deuterium (3,560 ppm Cl) in the sample from the inclined winze; such ratios are appropriate to either magmatic water or meteoric water, but would also be found in any water that had been greatly diluted by either magmatic or meteoric water. Tenfold dilution of a connate water of a 0.0 D:H percent would change the ratio of meteoric water having −12.5 D:H percent only to −11.2 D:H percent. Although the determinations do not suggest that the negative values of D:H percent increase with dilution, the variations are erratic and could represent mixed waters or the effect of local evaporation.

Tritium analyses of some typical waters of the area furnish additional data bearing on the origin and length of residence of the ground water. Background information: tritium has a half life of 12.5 years; the average for the Mississippi River before the first hydrogen bomb (1954) was 5 T.U. (1 T.U.=1 tritium atom per 10^{18} hydrogen atoms, $\sim$3.2 picocuries per liter), in 1955 the average rose to 44 T.U., and in 1959 to 119 T.U. (Junge, 1963). Samples were collected in June 1964 and analyzed for ΔD/H (SMOW) by Irving Friedman and for tritium by G. L. Stewart 6 months later. A sample from the hottest of the Abraham Hot Springs, 50 miles west of Eureka, temperature 185.5°F, contained 14.8 T.U. (ΔD/H=13.8). In contrast, the cool Aperdue Spring 4,000 feet southwest of the Apex Standard No. 1 shaft, temperature 60°F, contained 445±45 T.U. (but ΔD/H=12.6) and probably represents shallow perched ground water fed by the current annual precipitation. Deeper perched water tables tapped by the Newmont No. 2 shaft between 190 and 700 feet below the surface, flowing at 20 gpm, temperature 86°F, contained only 12±3 T.U. (ΔD/H=12.6). A sample of acid mine water from a winze used in mining a few months earlier—the 261 winze of the Burgin 1,050 level—had a temperature of 104° F, and contained 25±4 T.U. (D/H=12.5). The deep drill hole at coordinates 30,645 N, 26,598 E, about 20 feet west of the No. 2 vertical winze on the same level of the mine represents ground water undisturbed by mining or drilling for at least 2 years; its temperature was 133°F and it contained 7.5 T.U. (D/H=13.2). A general decrease in tritium content and therefore an increasing age is evident with increasing temperature and depth but, unfortunately, adequate samples of the hot water (140°F) first tapped by the inclined winze of the Burgin Mine were not available.

The chemical character of the waters of many thermal springs yields more persuasive evidence of their ultimate origin. Much attention has been given to this problem by White, Brannock, and Murata (1956), White (1957), and White, Hem and Waring (1963), but here it is only necessary to focus our attention on a few components of the saline water.

The ratio of bromine to chlorine in brines furnishes one of the best indices of marine or connate water as contrasted with meteoric waters that have dissolved marine evaporites, but the ratio does not adequately distinguish between magmatic water and saline lake water of closed basins. Valyashko (1959) made an excellent study of the theoretical and practical aspects of the distribution of bromine in brines and of the salts that crystallize from them. He found that the distribution coefficient between solid and solution for bromine included in the crystal lattice of halite separating from a bromine-bearing brine had a ratio of nearly 30:1; the bromine content of the salt crystallizing from the brine was thus only a small fraction of the amount left in the brine. Some selected figures from tables prepared by Valyashko are given in our table 7, which shows that the ratio of bromine to chlorine in normal marine connate brines is greater than 0.0032, whereas in the halite the ratio is less than 0.0032 by at least one order of magnitude; the Br:Cl ratio for the brine of Great Salt

TABLE 7.—*Chlorine and bromine data for evaporating sea water and the resulting solid phases*

[Data from Valyashko (1959, tables 2, 5)]

Specific gravity	Brine (percent chlorine and bromine)		Ratio $\left(\frac{Br \times 10^3}{Cl}\right)$ of bromine to chlorine in—			
	Cl	Br	Brine	Halite	Sylvite	Carnallite
1.01	1.01	0.0035	3.48			
1.199	14.33	.047	3.28			
1.220	15.72	.051	3.24	0.11		
1.300	15.77	.226	14.30			
1.308	15.93	.236	14.80	.44	2.9	
1.323	19.10	.334	17.50	.61	4.00	
1.325	19.23	.342	17.80	.60	4.20	8.3

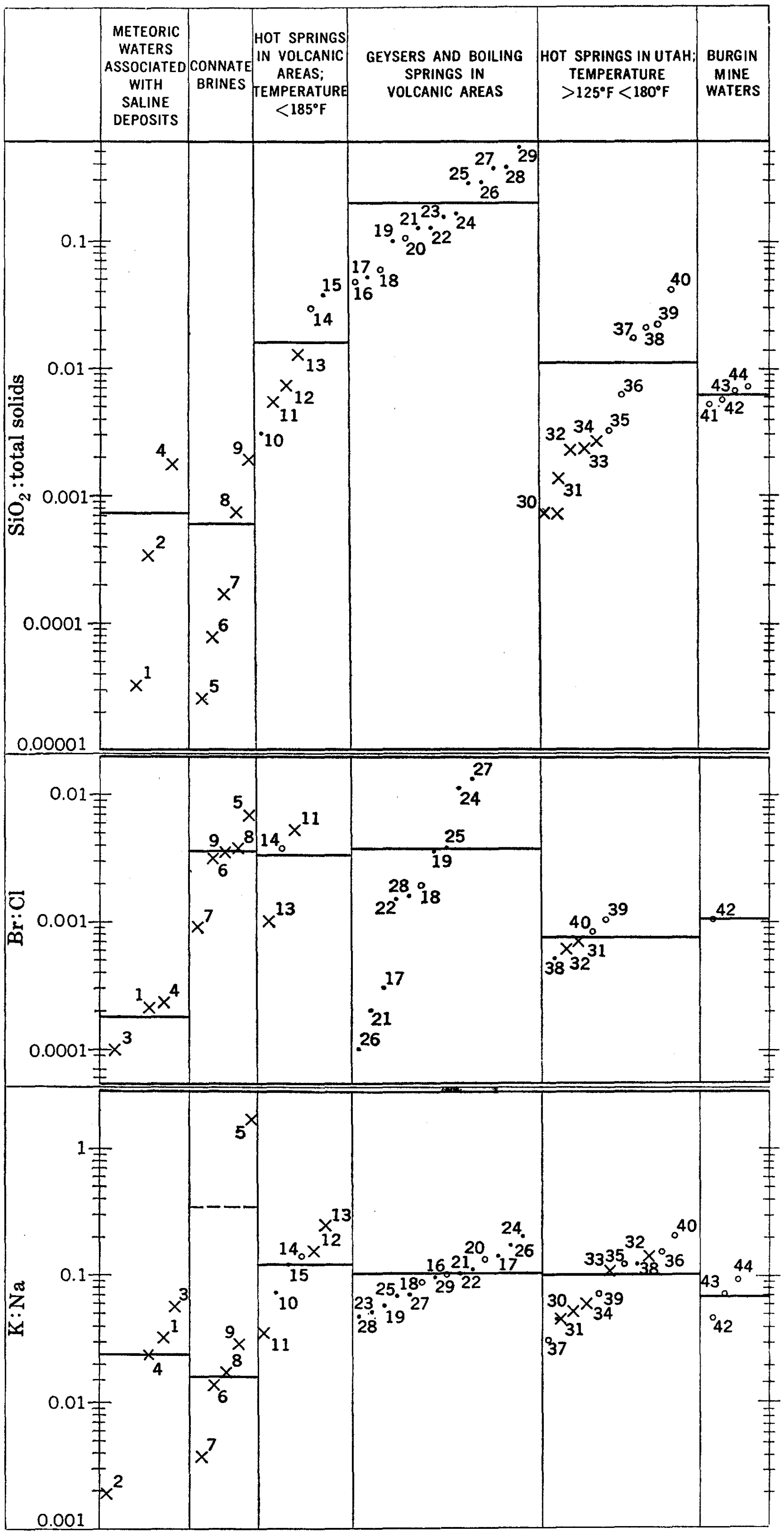

FIGURE 3.—Ratios of SiO_2 to total solids, Br:Cl, and K:Na in subsurface waters. Heavy line shows mean of group; dashed line shows mean of group including sample 5, which was collected from a potash salt mine. ●, fresh water (less than 3,000 ppm solids); ○, brackish water (3,000–10,000 ppm solids); ×, brine (more than 10,000 ppm solids).

Samples shown in figure 3

Sample	Locality	Temperature °C	Temperature °F	Reference	Analysis
1	Brine seep near Malaga, Eddy County, N. Mex.	37	98.6	White, Hem, and Waring (1963, table 27).	9
2	Salt spring, 7 miles north of Newcastle, Wyo. Water from the Minnelusa Formation.	9	48.2	Gott and Schnabel (1963, table 2)	2211
3	Auger holes in Great Salt Lake Desert, Tooele County, Utah; composite of 126 brine samples. Brine formed by meteoric water that leached saline clay and sand seams a few feet deep.	Unknown	Unknown	White, Hem, and Waring (1963, table 27).	4
4	Glenwood Springs, Garfield County, Colo.	52	125.6	do	6
5	Brine seep in National Potash Co. mine, Lea County, N. Mex.	Unknown	Unknown	do	7
6	Gulf Oil Corp. well 28-E, West Bay oil field, Plaquemines Parish, La.	87.5 (bottom of hole); 42 (discharge).	189.5 (bottom of hole); 107.6 (discharge).	White, Hem, and Waring (1963, table 13).	3
7	Gulf Oil Corp. well PP-19, Timbalier Bay oil field, Lafourche Parish, La.	70	158	White, Hem, and Waring (1963, table 12).	4
8	Seaboard Oil Co. well S.T.U. 305-13, Fresno County, Calif.	Unknown	Unknown	White, Hem, and Waring (1963, table 13).	1
9	Amerada Petroleum Corp. Peters gas well, Maine Prairie, Solano County, Calif.	Unknown	Unknown	White, Hem, and Waring (1963, table 12).	3
10	Well, near Niland, Imperial County, Calif.	40	104	White, Hem, and Waring (1963, table 18).	2
11	Spring, Kuan-Tsu-Ling, northern part of Taiwan	77	170.6	do	8
12	Springs, Cerro Prieto, Baja California, Mexico	82	179.6	do	5
13	Sprudel spring, 20 miles west of Mount Vesuvius, Italy	70.5	158.9	do	6
14	Spring 1, Nalachevskie group, southeast Kamchatka, U.S.S.R.	73	163.4	do	7
15	Spring near Jemez, Sandoval County, N. Mex.	65.5	149.9	do	4
16	Taumatapuhipuhi Geyser, Tokaanu, North Island, New Zealand.	Boiling	Boiling	White, Hem, and Waring (1963, table 17).	13
17	Growler Spring, Morgan, Tehama County, Calif.	95.4	203.7	do	5
18	Paryashchii 1, Pauzhetsk, Kamchatka, U.S.S.R.	100.6	213.1	do	12
19	"Geyser" spring, Mono County, Calif.	93	199.4	White, Hem, and Waring (1963, table 18).	1
20	Drill hole 4, Wairakei, North Island, New Zealand	228 (at 970-ft depth)	442.4 (at 970-ft depth)	White, Hem, and Waring (1963, table 17).	14
21	Spring 8, Steamboat Springs, Washoe County, Nev.	172 (at 379-ft depth)	341.6 (at 379-ft depth)	do	3
22	Velikan Geyser, Shumhaya, Kamchatka, U.S.S.R.	98.9	210.2	do	11
23	Geyser H-1, Umnak Island, Alaska	101.5	214.7	do	6
24	Sugar-bowl Geyser, Kamchatka, U.S.S.R.	98	208.4	do	10
25	Spring, Upper Basin, Yellowstone National Park, Wyo.	94	201.2	do	1
26	Spring, Norris Basin, Yellowstone National Park, Wyo.	84.5 (average for basin).	184.1 (average for basin).	do	2
27	Spring, Beowawe Geysers, Eureka County, Nev.	96	204.8	do	4
28	Sisjodandi spring, south of Geysir, east-northeast of Reykjavik, Iceland.	100	212	do	7
29	Blahver, Hveravellir, west-central Iceland	90.5	194.9	do	9
30	Crystal Hot Springs, Box Elder County, Utah	55.8	132.4	D. E. White, in this paper, table 3	11
31	Stinking Springs, 6.8 miles northwest of Corinne, Utah	48	118.4	White, Hem, and Waring (1963, table 15).	5
32	Utah Hot Springs, Weber-Box Elder County line, Utah	56.7	134	This paper, table 3	12
33	Becks Hot Springs, Salt Lake County, Utah	Unknown	Unknown	do	18
34	do	55.5	131.9	D. E. White (this paper, table 3)	19
35	Hooper Hot Spring, Davis County, Utah	60	140	do	14
36	El Monte (Ogden) Hot Springs, Weber County, Utah	58.1	136.6	do	13
37	Joseph Hot Spring, Sevier County, Utah	64	147.2	do	15
38	Red Hill Hot Spring, Sevier County, Utah	76	168.8	do	16
39	Abraham Hot Springs, Juab County, Utah	82	179.6	do	17
40	Roosevelt Hot Spring, Beaver County, Utah	55	131	do	10
41	Burgin mine, Utah County, Utah	53.9	127	This paper, table 3	4
42	do	54.4	130	do	7
43	do	Unknown	Unknown	do	3
44	do	60	140	do	5

Lake, however, is extremely small (<0.0001), and presumably the Br:Cl ratio for a salt in equilibrium with it would be at least an order of magnitude less.

Fresh water dissolving such halite certainly would also have a similar bromine to chlorine ratio, and the sharp difference between Br:Cl ratios in the two types of water (as shown in fig. 3) is striking confirmation of this fact. Inasmuch as the Br:Cl ratio of the Burgin mine water is approximately 0.001, the ratio is in harmony with our conclusions reached from study of the D:H isotopes, and we conclude also that the Burgin brine is neither a diluted marine connate water nor yet a diluted inland connate water of the type represented by Great Salt Lake.

White (1957), among others, stressed the high content of silica characteristic of waters from geysers and boiling springs associated with volcanic areas and, conversely, the low content of silica in connate or meteoric water. The silica content of meteoric water, however, has a substantial range, as does the silica content of thermal-spring water that issues some distance from shallow sources of volcanic heat. We have found that the ratio of silica to total solids in saline waters is appreciably different in three general groups: (1) water associated with geysers and boiling springs, (2) thermal water associated with volcanic areas but reaching the surface at temperatures less than 85°C, and (3) thermal or cool chloride water of meteoric—or connate—origin. Ratios for such waters are shown on the graph (fig. 3), which reveals that connate water and chloride water of meteoric origin have extremely low ratios of silica to total solids; the boiling springs and geyser waters have the highest ratios of silica to solids, for they are relatively dilute, but their silica content is mainly a temperature-dependent phenomenon caused by saturation or supersaturation of silica in very hot water.

The waters from thermal springs associated with volcanism but issuing at temperatures well below the boiling point have ratios of silica to total solids that overlap the range of ratios shown by boiling springs, and although their average is much below that of the boiling springs, their ratios are an order of magnitude greater than those of connate brines and brines formed by meteoric water in contact with salt beds. The Burgin mine water and most of the hot springs of Utah have ratios of silica to total solids that are clearly in the intermediate range, and even the lowest ratios among them are well above the ratios in connate brines or brines of meteoric origin. It might be argued, however, that the silica content in the Burgin water—about 48 ppm—suggests a meteoric origin; nevertheless, dilution of hot siliceous brines of volcanic origin by meteoric water having the composition of local meteoric water (table 3, anal. 1) could yield a brine with approximately the same silica content as the cool diluent if the amount present represented saturation at the cooler temperature.

The siliceous chloride waters of many boiling springs and geysers have several hundred parts per million silica in solution where they first appear at the surface orifice, but the waters precipitate some of their silica load in the outlet close by and soon attain a content appropriate to the equilibrium between opal or amorphous silica and hot water. As is well known, amorphous silica gel is much more soluble than quartz. According to Morey, Fournier, and Rowe (1962) the solubility of quartz in pure water is very low at room temperature and rises appreciably with increasing temperature: 6 ppm at 25°C, 19 ppm at 63°C, 45 ppm at 94°C, 57 ppm at 100°C, and 190 ppm at 180°C; in one experiment these workers found unequivocal evidence that quartz was deposited from a dilute but super saturated solution (80 ppm) at 25°C after 1 year. The solubility of amorphous silica is much greater, and according to these workers is about 135 ppm at 25°C and about 380 ppm at 100°C.

The effect of a large content of chloride ions on the solubility and rate of reaction of quartz may be great, however, and recent work by Van Lier and others (1960) is relevant. They reported that quartz has a well-defined solubility in water at temperatures below 100°C, that the solubility ranges from about 30 ppm at 60°C to about 75 ppm at 100°C, and furthermore that sodium chloride solutions greatly accelerate the rates of dissolution and of crystallization of quartz. Their experiments showed that equilibrium was attained in a few days in chloride solutions whereas in pure water several weeks were required. Although they concluded that the solubility of quartz in pure water is almost the same as in solutions containing less than 0.1 N sodium chloride (<5,850 ppm NaCl), they found some evidence of increased solubility of quartz in salt solutions of higher concentration. Inasmuch as the hot chloride water of the Burgin mine approximates 0.1*N* sodium chloride when collected, it is probable that they have an even higher chloride content at greater depth and that the temperatures are also higher than those measured. The value of 45 ppm of silica found in this mine water corresponds with the solubility of quartz either at 94°C (201°F) according to Morey and others (1962) or at 80°C (176°F) according to the data graphed in figure 6 of Van Lier and others (1960). It thus seems that the amount of silica in solution in the hot sodium chloride water of the Bergin mine is appropriate to a chloride solution that had attained equilibrium with quartz at temperatures near the boiling point. The solutions presumably pass through the thick basement of quartzite on their way to groundwater level, and equilibrium with quartz should be effected quickly as a result of the high salt content of the water.

Silica gel and opal precipitate on hot-spring aprons, but quartz apparently does not; it is rarely deposited from solution in the laboratory below 100°C. Quartz readily precipitates from solution at elevated temperatures, however, and if hot siliceous chloride waters deposited quartz as the stable phase at depth, the silica content of the solution would drop to a fairly low figure. Such a mechanism may explain the low silica content (72 ppm) of spring 21 of the Steamboat Springs area, noted by White, Brannock, and Murata (1956, p. 49).

Still another mechanism would give the low ratio of silica to total solids: If magmatic vapor rose into an evaporite series and there condensed on contact with either connate water or brine derived from meteoric water, the salt content might increase relative to the magmatic contribution while the silica content decreased. It may be significant, however, that, although many of the Utah hot springs having low ratios of silica to total solids are associated with modern evaporite basins, salt deposits have not yet been reported from the nearby Tertiary and pre-Bonneville Pleistocene deposits.

The K:Na ratio is also helpful in distinguishing between waters of meteoric origin and those of volcanic origin. As shown on the graph (fig. 3), the K:Na ratios in brines of meteoric origin range from 0.022 to 0.045 and average about 0.03; in contrast, ratios in water from volcanic sources are much higher, ranging from 0.04 to 0.25. The average of the K:Na ratios of hot springs associated with volcanic activity is 0.10, which is almost exactly the K:Na ratio of the hot saline Burgin mine water. This ratio is so far above that of the average brine formed by meteoric water (0.03) that it alone would strongly suggest a volcanic origin for the water. The relatively high K:Na ratios of boiling or hot saline springs in volcanic areas has often been ascribed to a temperature dependence—the hotter the water the greater the proportion of potassium. No such relation is evident in the family of Utah hot springs to which the Burgin water belongs. As shown in table 3, the K:Na ratios of eight springs having temperatures of 130 to 139°F range from 0.05 to 0.20, of three having temperatures of 140 to 149°F range from 0.03 to 0.09, and of two at temperatures at 169°F and 179°F are 0.12 and 0.07. No relation of K:Na ratios to temperature are discerned.

Finally, because of its twofold significance, the ratio of Li:Na should be considered. White (1957) has stressed the fact that, relative to other components, lithium is higher in thermal chloride water of volcanic association than in any other known type of natural water; he suggests that lithium and other alkali metals were transported as soluble alkali halides in a dense vapor phase from a magma to the base of the zone saturated by ground water, where the vapor condensed to form a brine unusually high in lithium. The ratios of Li:Cl shown in table 8 bear out his observation that thermal chloride waters associated with volcanic areas have a much higher Li:Na ratio than do brines formed by meteoric water in contact with salt. The Li:Na ratio of the Burgin chloride water fits that of the group of hot chloride waters of volcanic association and is thirteen times greater than that of the average of connate water and eighteen times greater than that of meteoric water in contact with salt deposits.

TABLE 8.—*Lithium to sodium ratios of various chloride-type waters*

[Ratios computed from analyses used for fig. 3, in which lithium is reported]

Type of chloride water	Ratio of Li:Na		
	Average	Median	Range
Geysers and boiling springs	0.00940	0.00900	0.00010–0.01900
Hot springs in volcanic areas	.00440	.00220	.00050– .01100
Burgin mine, chloride waters	.00430	.00400	.00220– .00660
Hot springs in Utah, T>125°	.00220	.00110	.00050– .01100
Connate waters	.00033	.00025	.00011– .00074
Meteoric waters in contact with salt	.00024	.00021	.00011– .00055

White and Brannock (1950) argued that an outstanding characteristic of a system of thermal springs deriving its heat from a normal temperature increase with depth is a lower gradient than that of the surrounding region. In the East Tintic thermal area where the heat flow above the water table is abnormally high, the heat flow is also high for the few holes in which gradients have been measured below the water table. Neither is there any evidence of abnormally low heat flow in the surrounding region. In as much as hot ground water moves eastward to Utah Valley, the heat flow measured above the water table to the east must be well above average, but no data are available for deep heat flow there. To the west in the main Tintic and Mintintic areas, the heat flow seems about average except for a few places where it is abnormally high. Although data for regional heat-flow isograms are lacking, there is no evidence that any area has given up heat at depth to deeply circulating ground water. The heat flow (according to our measurements) in the Mintintic area on the west side of the range is greater than 2 mc cm^{-2} sec^{-1}, and the heat flow on the eastern slope in Government Canyon, 5 miles south of the map area of figure 6 is 1.9 ± 0.3 mc cm^{-2} sec^{-1} according to Roy (1963, p. 49, see footnote 1, p. F2). The heat-deficient areas, required by the assumption that deep meteoric water moving laterally has abstracted heat at depth and has then risen again in the East Tintic thermal area, are at present (1964) unknown. Lacking evidence of these areas, we can only say that the three-dimensional heat-flow pattern as currently known, as well as the composition of the brine, argues for a volcanic source of heat.

Tertiary nonmarine evaporites have been suggested as the source of salts in the hot chloride spring waters of Utah, but the lack of any known salt-bearing evaporites of Tertiary age anywhere in Utah argues strongly against this interpretation. No salt is known in the thick Salt Lake Formation of Tertiary age present in the valleys adjacent to the East Tintic Mountains. Furthermore, the Br:Cl and Li:Na ratios of recent brines in the Great Salt Lake Desert, which should simulate those in a local Tertiary closed-basin evaporite, suggest a different interpretation. The ratios in brines that represent meteoric water in contact with salt deposits of the desiccated Lake Bonneville, as represented by a composite of many samples taken from holes a few feet deep in the Great Salt Lake Desert are totally unlike the ratios in the Burgin water (fig. 3 No. 3).

Meteoric water that has passed through either marine or continental evaporites seems unlikely to have the composition or the temperature required to explain the hot chloride water of the Burgin mine. If Burgin water is a composite of magmatic water and deep meteoric water that circulated through pre-Tertiary evaporite beds, the meteoric water must have picked up its additional load of salt from evaporites far below the surface; yet no such evaporites are known near the East Tintic district. Evaporites of Jurassic age are present some 25 miles to the southeast and may lie below another thrust plate well below the thick Paleozoic section in the footwall of the East Tintic thrust. If so, meteoric water moving down to great depths, passing through the evaporites, mixing with magmatic emanations, as at Larderello, Italy (Elizondo, 1964; Facca and Tonani, 1964), and rising through thousands of feet of sedimentary rocks in two thrust plates might eventually issue as a hot saline spring in the East Tintic thermal area. This possibility is assuredly worthy of consideration and is discussed further in the following pages. The best alternative to this explanation assumes a magmatic emanation which follows deep fractures and heats the ground water to the temperatures observed. Geophysical work or deep drilling may solve the problem of the origin of these thermal springs more satisfactorily, but at the present time we favor the hypothesis that both the heat and mineral content of the Burgin brine are chiefly of volcanic origin.

LATITE RIDGE THERMAL AREA

South of the Apex Standard thermal trough the temperature at elevation 4,550 feet increases steadily through a distance of 2,000 feet to the southeast at a rate of 2.5°F per 100 feet horizontally; it reaches 163°F at drill hole EP–2, the southeasternmost drill hole for which thermal data are available. If this lateral gradient persists for another 1,600 feet, the temperature would reach the boiling point of water at the elevation of the ground-water surface. Such temperatures suggest that at depths of a few thousand feet temperatures may be in the range of geothermal power requirements.

Although high temperature is the first essential of geothermal power, it is becoming increasingly evident that several other factors are significant. Ideally a geothermal power supply is the geologic analog of a steam dome on a boiler which has an ample supply of water and heat. Such a combination may not be as rare as it might at first seem. The geologic equivalent of the steam dome and boiler does in fact exist at Larderello, Italy. There, drill holes about 2,000 feet deep tap steam that produces the cheapest electrical power now available in the world—about 2.4 billion kilowatt hours per year from generating plants with a total installed capacity of 300,000 kw (Electrical World, 1963). The cost is 2.55 mills per kilowatt hour for the most modern plants (Facca and Ten Dam, 1963, p. 12). The steam wells are used for about 20 years before the accumulation of material deposited from the steam makes it necessary to drill another hole. This area has no known volcanic activity other than the steam field (Bozza, 1961).

A review of the geology of the Larderello field is instructive. Steam at high pressures and temperatures (200° to 250°C according to Facca and Ten Dam, 1963) is found in a somewhat broken horstlike fault block that is structurally high; the steam reservoir is in gypsiferous Jurassic evaporites under the Argille scagliose, a thrust plate of impervious broken claystone, serpentinized rock, limestone, and sandstone. For 6,000 feet below the thrust plate, convection of hot saline ground water maintains nearly constant temperatures as the water carries heat upward from its deep and presumably volcanic source. The broken horst simulates a regional steam dome, and to a geologist the analogy between the steam trap and a geologic trap for natural gas or petroleum is striking.

Oil seeps or gas leaks (as at Baku, U.S.S.R.) are common over many gas and oil traps and have led to discovery by drilling at the site of the seep. Usually the seeps are found eccentric to the productive structure, however, and the position of the trap must be found by geological and geophysical studies.

At Larderello the "geothermal seeps" of steam and hot springs occur at the surface in the broken horst where a few fairly tight fractures in the confining thrust plate permit the slow escape of the steam from the cavernous evaporite beds below. The analogy of steam trap and gas trap is apparent.

Geothermal exploration throughout the world has been confined almost solely to areas of boiling springs and geysers, but as the geologic factors that favor geothermal power production become better understood, hot springs probably will be regarded more and more as "geothermal seeps" which will serve as the locus of geologic studies and geothermal surveys that use techniques of the sort described earlier in this report (but see also Facca and Ten Dam, 1963).

The possibility that the Latite Ridge area has geothermal power potential should be given consideration. Additional drill holes in the area to the southeast are needed, but the geothermal gradient of about 10°F per 100 feet in EP–2 demonstrates the presence of a heat source worthy of consideration, if a favorable structure is present.

The region southeast of the Apex Standard thermal trough may have some similarity to the Larderello structure and its geology is therefore of considerable interest. Much of the East Tintic Mountains south of the Hansen fault is covered by lava, but the drill holes southeast of the Apex Standard mine and a few windows through the lava on the western side of the range show the presence of beds low in the stratigraphic sequence and indicate that the area is geologically high. The western side of the range is bordered by a Basin and Range fault which probably has a displacement of about 6,000 feet, the western or basinward side being faulted down.

At the southern end of the mountains the beds dip south or southeastward and the lower part of the Paleozoic section is exposed near the southern edge of the lava field. Here, too evidence is persuasive that the bedrock is structurally high. It is very probable that all these beds are in the upper plate of a major thrust fault, but the position of the thrust beneath the lavas on the eastern side of the East Tintic Mountains is uncertain.

The northern limit of this structurally high unit is probably a northeasterly shear zone that is parallel to the major shear faults of the East Tintic Mountains. Pyritic alteration in the lava southeast of the Inez and Hansen faults suggests that several northeasterly fractures underlie the lava within a few thousand feet of the Inez fault. North of the Hansen fault, beds of middle and late Paleozoic age are present in the footwall of the East Tintic thrust, which apparently is cut off by the northeastward-trending Inez shear zone. To the southeast, however, Tintic Quartzite or the immediately overlying Ophir Formation lies directly below the lava.

A northeastward-trending shear zone that separates Pennsylvanian rocks on the northwest from Cambrian rocks on the southeast cuts through West Mountain at the southeastern side of Lake Utah 10 miles to the northeast of the place where the Inez fault would reach the eastern edge of the East Tintic range. This major shear zone is believed to be related to the Inez fault and it probably represents a persistent regional tear fault (Morris and Shepard, 1964). Hole ET-89 (pl. 3, coordinates 29,245 N., 29,650 E.), which is southeast of the Inez fault line, cuts quartzite below the lava and cuts a sandstone and shaly dolomite of unknown age below a braccia zone that may represent the footwall of a thrust lying below the footwall of the East Tintic thrust to the north.

Although the Inez fault may be a tear fault affecting only a warped upper plate of the East Tintic thrust, the logs of diamond-drill holes suggest that it is a major shear zone that cuts deeper and reaches a lower thrust fault of large displacement. Southwest of the Apex Standard mine there are many minor shear zones and faults; as a result, progressively older formations crop out to the southeast. The drill logs and surface exposures clearly show that the northwest sides of the Hansen and Inez faults are downdropped relative to the southeast sides. The evidence available at this time (1964) indicates that the northeasterly zone of shearing along the Inez fault has affected the footwall as well as the hanging wall of the East Tintic thrust fault.

If the sandstone, shale, and dolomite found below the quartzite in drill hole ET 89 were present at depth southeast of hole EP 2 and had an average thermal conductivity of 6.0×10^{-3}, the temperature would increase vertically at a rate of about 6.5 °F per 100 feet. This gradient, if maintained, would give a temperature of approximately 460 °F (242 °C) at sea-level elevation under drill hole EP 2; saturated steam at this temperature has a pressure of 450 pounds per square inch. The gradient, however, would not persist where convection became a major factor, as beneath a thrust plate that covered Jurassic evaporites. Inasmuch as the gradient seems to be increasing to the southeast, higher temperatures should be found at shallower depths.

The evidence now available indicates that the terrane underlying the lava in the East Tintic Mountains southeast of the Inez fault zone is structurally high compared with that to the northwest. The great thrust fault which underlies the southern Wasatch Mountains has brought Cambrian to Pennsylvanian rocks over gypsiferous Jurassic rocks; if this fault underlies the East Tintic Mountains also, it is very possible that the upfaulted blocks southeast of the Inez and Hansen faults are underlain by this thrust fault at a depth accessible to the drill. Surveys of the type described in this report and in the paper by Facca and Ten Dam (1963) are fairly cheap and could be used to evaluate quickly the potential of possible geothermal areas. The marked thermal gradient southeast of the Inez fault and the possibility of structures favorable for development of geothermal power in the uplifted area to the southeast should offer encouragement for further exploration and geothermal surveys in that area.

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