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TRUE-TEMPERATURE DETERMINATION OF GEOTHERMAL RESERVOIRS

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ABSTRACT

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Parameters governing the resistivity in geothermal areas are analyzed. A method for the calculation of the true temperature of geothermal reservoirs is explained, and the effectiveness of the method is evidenced.

INTRODUCTION

Recently geothermal energy as a major potential energy source has drawn considerable attention due to the realization of the limited amount of hydrocarbon energy resources, of the advantages of the terrestrial heat over other forms of energy sources, and of the fact that economically accessible geothermal energy is more widespread than was believed only a few years ago (Armstead et al., 1974).

The utilization of geothermal energy is relatively free from danger of human life and environmental pollution, and requires neither advanced technology, as in the case of nuclear energy, nor massive construction, as in the case of hydroelectric power. From a study of the economics of geothermal power in Italy, New Zealand, and the United States, Facca and Ten Dam (McNitt, 1965) concluded that geothermal power is the cheapest among various sources of energy available at the time of McNitt's report. Cost analyses reported by Kaufman (1970), using the standard evaluation procedures suggested by the U.S. Federal Power Commission, reveal that of various energy sources geothermal energy will be least in total cost for the generation of electricity under favorable conditions, still inexpensive and competitive with others under less favorable conditions. His cost comparison was made with assumption of a base-load plant with a capacity of 1500 MW, privately financed, having either variable load factors for the various fuels or a constant 90% load factor for all fuels over the plant life of 30 years except for hydropower which was estimated at 50 years. In the case of geothermal energy, well life was estimated at 20 years.

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The principal use of geothermal fluid is for the generation of electric power. The power is generated by tapping the natural thermal fluid underground and guiding it to drive a turbine of the generator system. The geothermal fluid is also utilized for space heating, agriculture, industries, hydrotherapy, extraction of chemicals which it may contain, and for fresh water.

For the exploration of geothermal resources, geophysical techniques such as electrical resistivity, gravity, magnetic, seismic, and heat flow methods as well as geological and geochemical methods have been employed in Italy, New Zealand, Iceland, the United States, Japan, Mexico, and the Soviet Union, among others.

Of the various geophysical methods, the electrical resistivity method is considered to be and has been one of the most effective tools for geothermal energy prospecting. Keller (1970) has reported more effective aspects of electromagnetic induction methods over direct current methods in measuring resistivity in geothermal areas. Readers who are interested in the induction methods are referred to his well presented paper. The resistivity method has advantages of easy operation, low expenses, and wide range of applicability. One of its unique aspects may be that it is very useful in estimating both the quantity and quality of the ground water in geothermal provinces.

In the following, parameters which are essential in the resistivity method are analyzed and a method, with evidences for its effectiveness, of calculation of the true temperature of geothermal reservoirs is presented. Although the nature of, and the relationship between, those parameters are known to a limited number of people, the fact that the true temperature of geothermal reservoirs can be determined, has not yet, to this author's knowledge, been pointed out clearly to broad readers by using an explicit formula.

THEORY

In the following, physical and chemical theories which govern the electrical resistivity of earth materials are explained concisely. For most earth materials near the surface of the earth, the conduction of electricity is almost entirely through the water contained in the pores of the materials. In this paper, consideration is given to geothermal reservoirs whose rock possesses appreciable porosity and does not contain a high percentage of electricity conducting minerals such as graphite, pyrite, magnetite, specular hematite, chalcopyrite and pyrrhotite. Therefore in geothermal reservoirs considered in this paper, the electric conduction is assumed to be liquid-electrolytic, with the adjective "liquid" meaning the phase of the electrolyte contained in the pores of the solid rock at moderate temperatures rather than the phase of the rock itself in fused state at very high temperatures.

The resistivity of water-bearing rocks largely depends on the amount of water they contain, the chemical and physical, including thermal, properties of the water, and on the manner in which the water is distributed in them.

Porosity — qu

The resistivity content. In a v porosity, while (Archie, 1942)

$$\rho_{100} = a \rho_w \phi^{-m}$$

where ρ_{100} is the resistivity of the water and a and m are constants of one type of formation. In this discussion. Although sedimentary rocks on this subject in (Archie, 1942) has been observed (Archie, 1970).

From eq. 1, With a and m known, the amount of water can be determined from the equation

$$V_w = V \left(a \frac{\rho_w}{\rho_{100}} \right)^{\frac{1}{m}}$$

In eq. 2, V and ρ_w can be measured.

Salinity — quality

The electrical resistivity of the same terms of an electrolyte solution is the salinity or conductivity.

$$\frac{1}{\rho} = F \sum_{i=1}^n c_i m_i v_i$$

where ρ is the resistivity (96,490 C/mole) (moles/cm³), m is the equivalent weight (cm²/V sec), v is the number of kinds of ions, c is the resistivity of a solution in turn the viscosity in turn the viscosity below.

Porosity — quantity of water

The resistivity of a water-bearing rock decreases with increasing water content. In a water-saturated rock, the amount of water may be equated with porosity, while the porosity is related to the resistivity by the Archie's law (Archie, 1942) modified by Winsauer and others (Winsauer et al., 1952):

$$\rho_{100} = a\rho_w\phi^{-m} \quad (1)$$

where ρ_{100} is the bulk resistivity of the water-saturated rock, ρ_w is the resistivity of the water contained in the pore structure, ϕ is the porosity of the rock, and a and m are empirical constants. It is noted here that, for simplicity, only one type of formation water is assumed to exist in the pores of the rock under discussion. Although the amount of data on igneous rocks is much less than on sedimentary rocks, chiefly due to the historic development of the research on this subject in connection with petroleum exploration, the above relationship has been observed for measurements made on various igneous rocks, too (Keller, 1970).

From eq. 1, the amount of water stored in a reservoir can be determined. With a and m for a particular aquifer or formation already established, the amount of water, V_w , which an aquifer of volume V contains can be calculated from the equation:

$$V_w = V \left(a \frac{\rho_w}{\rho_{100}} \right)^{1/m} \quad (2)$$

In eq. 2, V and ρ_{100} can be calculated from geophysical field survey data and ρ_w can be measured in the laboratory with water sample taken from the field.

Salinity — quality of water

The electrical properties of a water-bearing rock should be describable in the same terms as the electrical properties of an electrolyte. The conductivity of an electrolyte contained in a rock can be expressed as a linear function of the salinity or concentration of the electrolyte (Milazzo, 1963):

$$\frac{1}{\rho} = F \sum_{i=1}^n c_i m_i v_i \quad (3)$$

where ρ is the resistivity of the electrolyte (Ω cm), F is Faraday's number (96,490 C/mole), c is the concentration of each species of ion in solution (moles/cm³), m is the electrical mobility of each species of ion in solution (cm²/V sec), v is the ionic valence of the individual ionic species, and n is the number of kinds of ions present in the solution. Eq. 3 tells us that the resistivity of a rock is inversely proportional to the salinity and mobility of ions in the solution. The mobility of ions increases with decrease in viscosity, in turn the viscosity decreases with increase in temperature which is considered below.

Temperature

Contrary to metal, in water-bearing earth materials, resistivity decreases with increase in temperature. In either an electrolyte or a rock saturated with an electrolyte, increased temperature lowers the viscosity of the electrolyte and the decreased viscosity helps the ions move freely.

At moderate temperatures, the dependence of resistivity on temperature for a rock saturated with an electrolyte is defined by the equation (Glasstone, 1937):

$$\rho_t = \frac{\rho_{15}}{1 + \alpha_t (t-15)} \quad (4)$$

where ρ_t is the resistivity at the ambient temperature $t^\circ\text{C}$, ρ_{15} is the resistivity measured at a reference temperature of 15°C (any other reference temperature may be used), and α_t is the temperature coefficient of resistivity.

Degree of saturation

The pore space of a rock may or may not be filled with an electrolyte; above the water table, part of the pore space is filled with air. The amount of water in the zone of aeration changes with time as meteorological and hydrological processes such as precipitation, drainage, evaporation on the soil surface and transpiration through plants take place.

A quantitative relationship between resistivity and the degree of saturation in a rock has been observed to hold (Keller, 1953):

$$\rho = \begin{cases} \rho_{100} S^{-n_1}, & S > S_c, \\ b\rho_{100} S^{-n_2}, & S < S_c \end{cases}$$

where ρ is the bulk resistivity of a partially saturated rock, ρ_{100} is, as defined previously, the resistivity of the same rock when it is completely saturated with the same electrolyte, S is the degree of saturation which is expressed as the ratio of pore volume filled with electrolyte to the total pore volume, S_c is the critical water saturation, which is the least saturation for which there is a continuous film of water over all the surfaces in the rock, and b , n_1 , and n_2 are experimental parameters.

Texture

The textural factors such as the particle-size distribution of the constituent particles, the shape of the particles, the orientation of the particles in space and with respect to each other, and the forces tending to bind the particles together, play a significant role in the resistivity of the rock. Pore space must be interconnected and filled with water in order that a rock may conduct electricity. Pores may be classified as interconnected and isolated ones,

depending on the pores may consist of the storage of water expected from a to storage pores in stones and volcanic

A mathematical rock in its resistivity follows (Ward and

$$\rho = F\rho_w$$

and:

$$F = \frac{L_w}{L} \left(\frac{A}{A_w} \right)$$

where ρ and ρ_w are efficient (Pirson, 1953) which actual effect of the diminution of cross-sectional area to that of the electrode

Both the tortuosity and shape of the rock are interconnected. For $a\phi^{-m}$, which can be better. It is noted that the porosity decreases due to liquid electrolyte discussions which will to formations with conduction is predominant

Interaction between

It is known that the resistivity is always the same as that of the rock. According to the independent of ρ_w formation factor of a rock when it is with a high concept of surface conductivity negatively charged substances and systematic under an applied electric

depending on the connection of pore space with others. The interconnected pores may consist of two parts; the larger voids, which are major space for the storage of water, and the finer connecting channels. A high resistivity is expected from a rock which has very small volume ratio of connecting pores to storage pores even when it possesses great porosity as in some shales, limestones and volcanic rocks.

A mathematical expression which accounts for the role of the texture of a rock in its resistivity may be shown by use of the formation factor, F , as follows (Ward and Fraser, 1967):

$$\rho = F\rho_w \quad (5a)$$

and:

$$F = \frac{L_w}{L} \left(\frac{A}{A_w} \right) \quad (5b)$$

where ρ and ρ_w are the same as defined previously; L_w/L , the tortuosity coefficient (Pirson, 1958), is the ratio of the length of the electrolyte path via which actual electric current flows, to that of the rock; A/A_w , the reciprocal of the diminution coefficient (Ward and Fraser, 1967), is the ratio of the cross-sectional area of the rock perpendicular to the direction of electric current, to that of the electrolyte path, which is in an arbitrary direction.

Both the tortuosity and diminution coefficients are controlled by the size and shape of the rock pores and the manner in which they are disposed and interconnected. For the formation factor, the expression in eq. 5b, instead of $a\phi^{-m}$, which can be derived from eq. 1, has been employed here, since the author believes that eq. 5b explains the characteristics of the formation factor better. It is noted from both the expressions for the formation factor that as the porosity decreases the formation factor increases. For a rock of extremely small porosity, the formation factor becomes so great that the ionic conduction due to liquid electrolyte in the rock becomes insignificant. Therefore the above discussions which were based on electrolytic conduction of rocks apply only to formations with large enough porosities in which liquid electrolytic conduction is predominant.

Interaction between rock and electrolyte

It is known that the conductivity of the water extracted from a rock is not always the same as that which the water would have before it is removed from the rock. According to eq. 1 or 5a, the formation factor, ρ/ρ_w , should be independent of ρ_w for a given rock. However, it is usually found that the formation factor of a rock is less when it is saturated with a dilute solution than when it is with a highly saline solution. This phenomenon is explained by the concept of surface conduction. The surface conduction is attributed to the negatively charged surface of clay materials, which is balanced by adsorbed cations and systematically oriented water molecules, which are free to move under an applied electric field.

Rock-forming minerals usually fracture in such a way that one species of ion in the crystal is commonly closer to the surface than others. In the silica—alumina units, which are the most abundant constituents of the crust, the oxygen ions are usually closest to the surface, resulting in the negatively charged surface of the clay materials (Keller and Frischknecht, 1966). Substitution within the lattice structure of clay materials of cations of lower valence for those of higher valence would also result in a negatively charged surface (Grim, 1968). This negative charge of the surface would be balanced by adsorption of cations in the electrolyte. The adsorbed cations are free to move under an applied electric field, therefore increase the electric conductivity. Since, in a high salinity solution, the increased ionic pressure in the adsorbed layers increases the viscosity of the electrolyte and decreases the mobility of ions (Keller and Frischknecht, 1966), the effect of this added conductivity is greater in a rock saturated with a dilute solution than one with a high salinity solution.

A similar effect takes place with water molecules, which are polar. When an electrolyte is in contact with a negatively charged rock surface, the water molecules will become adsorbed to the surface in such a way that the positive hydrogen atom side is closer to the surface than the negative oxygen atom side. The conductivity of water in this systematically oriented phase is higher than that of free water (Keller and Frischknecht, 1966). The thickness of the oriented adsorbed water molecule layer will be greater in a rock saturated with a dilute solution than one with highly saline solution, since the abundant cations, in a high salinity solution, adsorbed to the surface earlier than the water molecules due to their charges being stronger than those of water molecules, will neutralize the rock surface to some extent.

APPLICATION

The resistivity of a rock formation decreases with increase of temperature so that it is possible to map a geothermal area by resistivity observations. Moreover, it is very interesting and important to note that the actual temperature of the thermally anomalous underground can be determined by the resistivity values measured at the surface.

Utilizing eq. 4, the following formula which explicitly gives the temperature of the geothermal reservoir is easily deduced:

$$t_h = t_c + \frac{1}{\alpha} \left(\frac{\rho_c}{\rho_h} - 1 \right) \quad (6)$$

where t_h and ρ_h are, respectively, the temperature and resistivity of the hot portion of the earth, that is, geothermal reservoir; t_c and ρ_c are, respectively, the temperature, which is around 15°C, and the resistivity of the cold portion of the earth; and α is the temperature coefficient of resistivity of the ground water in the area under question. ρ_c and ρ_h are calculated from resistivity field survey data and α is determined by laboratory measurements. In applying

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eq. 6 to the geothermal problems, however, the following should be considered.

Eq. 4 from which eq. 6 was derived has been established only for the temperatures ranging from slightly higher than freezing point to slightly lower than boiling point of the water. Therefore laboratory examination is necessary in order to determine whether eq. 6 is valid at temperatures up to 388°C, which is the highest reported reservoir temperature (Mercado, 1969), or higher. If eq. 6 is found to be invalid at higher temperatures, a new function should be formulated. However, eq. 6 agrees reasonably well with the observations made at high temperatures. Breusse (McNitt, 1965) has reported that ground heated from 17°C to 150°C decreased in resistivity by a factor of 5, and if heated from 17°C to 280°C, its resistivity decreased by a factor of 9. Meanwhile, the calculations by eq. 6 for the same temperature increments as those in Breusse's observations yield resistivity decrements of 4.33 and 7.58 times, respectively, which are in rather good agreement with those measured by Breusse. In this calculation, 0.025/°C was used for α . This is the average value of the temperature coefficient of resistivity of ground water (Keller and Frischknecht, 1966). The use of the particular value of α measured of the samples from the Breusse's study area could bring the resistivity changes closer to Breusse's observations. Whenever desired, the exact value of the temperature coefficient of resistivity for individual geothermal areas may be obtained by laboratory measurements of the outcrop of the water, that is, hot spring or body of water derived from the pertinent geothermal area.

The effectiveness of the resistivity method in geothermal areas is also corroborated by the results of other successful workers. Studt (1958) has reported a successful resistivity survey in Kawerau, New Zealand. He wrote: "Electrical surveying extended the hot water zone", and further described the results of his observation: "The best contrast between hot and cold water country is recorded with an electrode separation of 100 feet; at this spacing, the mean values in hot and cold country were 60 and 600 Ω m respectively. A value of 150 Ω m may be taken as an approximate dividing line between hot and cold water."

Another example of a successful resistivity survey has been described by McNitt (1965): "Electric resistivity surveys, in conjunction with gravity surveys, have been very helpful in locating major faults in the Italian steam fields. The electrically resistant anhydrite and carbonate rocks, which are the steam reservoirs, are overlain by an impermeable, conductive shale. The resistivity method is successful because the thickness of the shale cover is characteristic of earth fault block, so that the resistivity survey shows the position of the faults as well as the comparative thickness of the shale, and therefore the depth to the reservoir."

A word of warning is appropriate at this point. Lower resistivity of an area, in either lateral or vertical sense, than in the surroundings, can be attributed to anomalously high temperature only where both the high and low resistivity provinces are geologically or geoelectrically the same or similar except for their different temperatures. The problem of differentiation between various causes

of the low resistivity should be solvable by incorporating the resistivity data with geologic, geochemical and other geophysical findings of the area. An excellent example of non-thermal province with lower resistivity than the adjacent geothermal reservoir is provided in the report by Zohdy and others (1973), where geothermal reservoir of about 30 Ω m is bounded by non-thermal lacustrine deposits with resistivity of 10–12 Ω m.

It should also be admitted here that the above examples are only some of successful results of reported investigations carried out under somewhat favorable conditions. However, it is, in the main, the aim of this paper to present the theoretical capability of the determination of the true temperature of the geothermal reservoirs. It is only a matter of improving our technique in order for us to be able to utilize the theory not only under favorable conditions but also under many unfavorable conditions as well. We have witnessed that only theoretically possible things turned out to be reality, as our techniques advance, in other disciplines of geophysics in our time.

CONCLUSION

The electrical resistivity method has been very useful in exploration of geothermal fields. The resistivity method is effective not only for estimating the amount of geothermal fluid but also for determining the true temperature of the thermal fluid reservoirs and in that point this method is unique.

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