Journal of Nepal Geological Society, 2003, Vol. 28, pp. 101–111

Heat flow and vertical groundwater flux in deep fractured basement rock in Nara Basin, southwest Japan

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ABSTRACT

The temperature measurement in a borehole that intersects the flow zone is one of the effective methods to detect the vertical flow of water. Geothermal gradients, thermal conductivity, and heat flow have been assessed for the crystalline basement and the overlying sedimentary cover in the study area. The geothermal gradient in sedimentary cover ranges from 0.83-3.87 °C/100 m, whereas in the basement rock it ranges from 2.49-3.38 °C/100 m. The heat flow values in sedimentary cover and basement rocks lies between 13.46 and 56.87, and 79.89 and 108.63 mW/m^2, respectively. Vertical specific discharge of groundwater flow in the fractured basement rock is estimated for different depth intervals. Groundwater in the unconsolidated sediment and at most places in fractured rocks is moving vertically down. However, at some locations, water is moving upward especially at greater depth intervals. The calculated vertical specific discharges reflect the relative degree of fracturing in the basement rocks at the corresponding depth intervals.

INTRODUCTION

Temperature measurements in boreholes are generally used to determine the terrestrial heat flux. This principle of determination is based on the assumption that the heat flow being measured is entirely conductive, and that steady condition exists. Under such condition, heat flux (Q, W/m^2) is given by:

$$Q = -k\,(dt/dz) \qquad (1)$$

where, k and (dt/dz) are thermal conductivity (W/m°C) and thermal gradient (°C/m), respectively, and depth (z) is positive downwards. The negative sign on the right hand side indicates that heat flows down the temperature gradient.

Stallman (1963) for the first time proposed an equation for the simultaneous transfer of heat and water within the earth. This led to a mean to calculate the rates of groundwater movement by measuring the subsurface temperature. Bredhoeft and Papadopulos (1965) developed an analytical solution by means of type curve to describe vertical steady flow of groundwater and heat through an isotropic, homogenous and fully saturated semi-confining layer. Further, several authors (e.g. Cartwright 1970; Mansure and Reiter 1979; Boyle and Saleem 1979; Drury and Jessop 1982) modified the method to assess the movement of water in faults, cracks and permeable rocks. All these works are based on the fact that hydrodynamics contribute to formation of temperature anomalies. In the Nara Basin, Inagaki and Taniguchi (1994) and Taniguchi (1994 and 1995) have utilized the transient groundwater temperature observed in the wells to estimate hydraulic conductivity and recharge rates for shallow depths.

The Nara Basin is located in southwest Japan. The basin is about 770 km^2 in areal extent. The average elevation of the basin is around 60 m above sea level, while the height of the surrounding mountains ranges from 470 m to 900 m. The highest peak lies in southern part of the study area. The mean annual precipitation in the area is about 1325 mm. There are many streams reaching the valley from the surrounding mountains to form the Yamato River, which flows almost through central part of the basin towards the west (Fig. 1).

There are several water wells reaching depths of 250 m. However, few deep wells which penetrates the granitic basement are drilled (to the depth of 1,500 m) in order to utilise the Hot Spring for commercial purpose. The downward change in temperature and hydrogeology is poorly known from such a deep level. The present study is focused on estimating vertical specific discharge of water by calculating the heat flux at different depths in 12 Hot Spring wells that have temperature-logging data. Such information had not been available to date for the area. Further, understanding the direction of vertical flow of groundwater will help in understanding the interaction of groundwater between fractured basement rocks and sedimentary cover.

THEORY

In order to determine the vertical discharge of groundwater, several techniques dealing with temperature profile, thermal gradient and heat flow are available. However, all the methods have the same principal based on the general differential equation for simultaneous non-steady flow of heat and fluid through fully saturated, isotropic, and homogeneous porous medium as proposed by Stallman (1963).

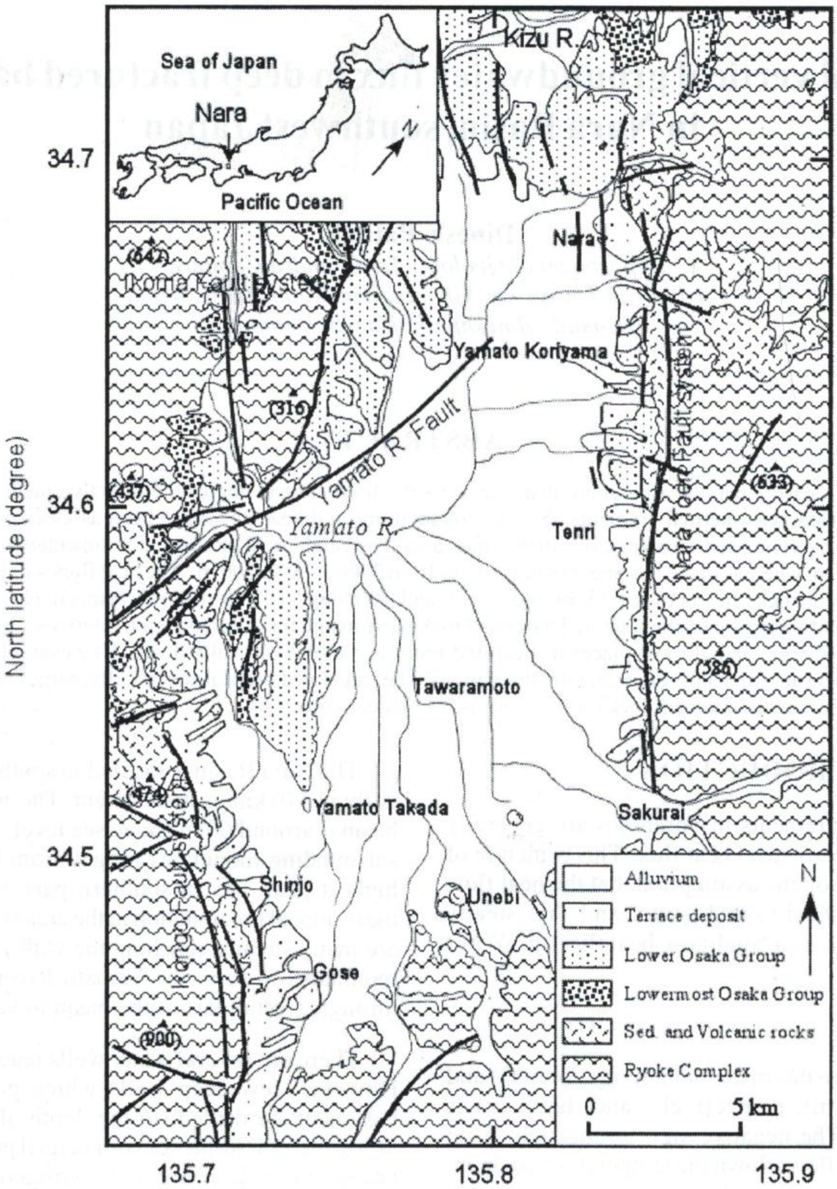

Fig. 1. Geological map of Nara Basin and surrounding area (Modified after Itihara et al. 1993) The solid lines are faults and the location of Nara Basin is given in the inset.

The equation states:

$$\frac{\partial^2 T}{\partial x^2}+\frac{\partial^2 T}{\partial y^2}+\frac{\partial^2 T}{\partial z^2}-\frac{c_w\rho_w}{k}\left[\frac{\partial(v_xT)}{\partial x}+\frac{\partial(v_yT)}{\partial y}+\frac{\partial(v_zT)}{\partial z}\right]=\frac{c\rho}{k}\frac{\partial T}{\partial t} \qquad (2)$$

where,

v = vertical velocity of fluid in x, y, and z direction

T = temperature at any point at time t_1

t = time since flow started

c = specific heat of rock/fluid complex

c_w = specific heat of fluid

ρ = density of rock fluid complex

ρ_w = density of fluid

k = thermal conductivity of solid-fluid complex

x, y, z = Cartesian coordinates

Several authors have proposed different methods to estimate the vertical flow of water by means of temperature measurement in the borehole. These methods are briefly reviewed below.

Bredhoeft and Papadopulos (1965) developed type curves to calculate the vertical one-dimensional flow of heat and fluid in steady state. The authors modified equation (2) to the following form:

$$k\frac{\partial^2 T}{\partial z^2} - v_z c_w \rho_w \frac{\partial T}{\partial z} = 0 \quad (3)$$

For a zone of height L over which specific discharge remained constant and T_1 and T_2 are the temperature measurements at upper and lower boundary of length L, and T is the temperature measurement at any depth z. Equation (3) reduces to:

$$\frac{(T - T_1)}{(T_2 - T_1)} = f(\beta, z/L) \quad (4)$$

$$f(\beta(z/L) = \frac{[\exp(\beta z/L) - 1]}{[\exp(\beta) - 1]} \quad (5)$$

and

$$\beta = \frac{c_w \rho_w v_z L}{k} \quad (6)$$

Here, β is a dimensionless parameter that is positive or negative depending on whether v_z is downward or upward. The authors presented type curves (z/L ~ f(β, z/L)) for the solution of equation (6). This needs to plot $(T_2 - T_1)/(T_L - T_1)$ against z/L and superimpose on the type curve, interpreting the value of β that best matches the field data curve to get vertical specific discharge.

Mansure and Reiter (1979) proposed a method to determine the vertical movement of water by means of temperature versus temperature gradient plot, which is based on the principle of conservation of energy. If there is no water movement, the plot will yield a horizontal line. Their modified equation is as below:

$$\frac{dT}{dZ} = (\beta/L)(T - T') - E/k \quad (7)$$

Where, E is the total energy flux, dT/dZ is the temperature gradient, and T' is the temperature at which the internal heat energy is taken to be zero. The plots of dT/dZ versus T will yield straight lines when β is constant, thus allowing estimates of discharge. The slope of the plot will yield (β/L). The following equation has been proposed to calculate the specific discharge of the fluid movement:

$$v_z = (\beta/L)(k/c\rho) \quad (8)$$

The above method is based on the change in temperature gradient with depth where thermal conductivity values are relatively constant within the depth interval under consideration. However, for the sedimentary layer that is consisting of different layers with varying thermal conductivity values, the result may be different. In order to overcome this problem, Reiter et al. (1989) suggested a relationship to determine the vertical groundwater flow from temperature vs. heat flow plot. Modifying equation (7),

$$k\frac{dT}{dz} = c\rho v_z (T - T') - E \quad (9)$$

Rewriting equation (9),

$$Q = c\rho v_z (T - T') - E \quad (10)$$

Where, Q is the heat flow. Thus, plot of Q vs. T will have slopes of $c\rho v_z$.

Therefore:

$$m_z = dQ/dT = c\rho v_z \quad (11)$$

Integrating equation (10) across the layer of constant specific discharge,

$$Q_{z2} - Q_{z1} = c\rho v (T_{z2} - T_{z1}) = c\rho v_z (T_2 - T_1) \quad (12)$$

Equation (12) states that the change in conductive heat flow equals the negative of the change in advected heat flow across the layer of constant specific discharge (the positive Q and v are oppositely directed). The slope m_z can be obtained from the slope of the Q vs. T plot, fluid density (ρ) and heat capacity per unit mass (c) have constant values, the vertical discharge (v_z) across an interval of interest can be calculated using equation (11). In the present study, the specific vertical discharges in the wells are calculated using this method.

GEOLOGICAL SETTING OF THE AREA

The geology of Nara Basin can be broadly subdivided from top to bottom into the Alluvium (Holocene), Terrace deposits (Upper Pleistocene), Osaka Group (Late Pliocene-Upper Pleistocene), volcanic and sedimentary rocks (Miocene) and the Cretaceous granitic basement of Ryoke complex (Fig. 1). Alluvial deposits mostly cover the surface of the Nara Basin. The terrace deposits lie between the hills and the Plain area. Likewise, the Osaka Group sediments can be observed along the foothill, and the Miocene beds (volcanic and sedimentary rocks) are sporadically exposed in some localities. In the higher elevation areas, the granitic basement is well exposed, mostly forming higher hills surrounding the basin.

Alluvial deposits consist of mainly unconsolidated clay, sand and sandy gravel with the average thickness of about 3 m. The central part of the basin has the maximum thickness of about 8 m. The terrace deposits has the average thickness of about 12 m and consists of alternating beds of clay, sand, gravel, and their mixtures. The Osaka Group sediments are further subdivided into the Upper, Lower- and Lowermost parts. The upper and lower parts consist of alternating beds of marine clay with sand and gravel and non-marine sandy gravel, sand, and clay. The Lowermost part is consisting of non-marine clay, sand, and gravel. In the Nara Basin, the upper part of Osaka Group, is not exposed

(Itihara et al. 1991). The available deep boring data shows that the maximum thickness of the Osaka Group in the basin is around 600 m, lying in the northern part.

The main geological structures of the Nara Basin were formed by a series of crustal movements, which caused compression in the north-south and east-west direction, developing several faults in the basin and surrounding mountainous regions. These faults are trending both in the north-south and east-west direction, the former being predominant. The major fault systems in the area are the north-south running Nara-Toen fault system in the northeastern margin of the basin, Kongo fault system in the south western margin of the basin and Ikoma fault system in the northwestern part lying in the mountainous terrain (Research group for active faults in Japan 1991). Yamato River Fault is NE-SW trending fault that passes more or less parallel to the Yamato River, in the central part of the basin. This is one of the major faults that is traversing the main part of the basin. Besides, Median Tectonic Line, a regional structure that trends NE-SW, is passing through further south of the study area.

HYDROGEOLOGIC FRAMEWORK OF THE STUDY AREA

Groundwater in the Nara Basin is available in the basement rock as well as in the overlying unconsolidated sedimentary sequences belonging to alluvial, terrace and Osaka Group. In the basement rocks, groundwater occurs mainly in the fractures. The aquifer materials in the overlying sediments are sand and sandy gravel and the aquifers are multi layered. Unconfined aquifers occur mainly on the topmost shallow depth belonging to alluvium and terrace deposits and some confined aquifer horizons exist below. The major confined aquifer horizons belong to the Osaka Group sediments. These confined aquifers can be subdivided either as shallow confined (<50 m depth) or deep confined (>50 m deep) depending upon their depth of occurrence. The shallow confined aquifers are considered semi-confined in nature. The recharge rate for the shallow depth in the Nara Basin is estimated to be 460 mm/year from the measurement of transient temperature-depth profile (Taniguchi 1994).

The Nara Basin is traversed by many rivers originating from the surrounding mountains and flowing from both north and south towards the center of the basin to form the Yamato River, which flows towards Osaka Basin in the west (Fig. 1). These rivers are recharging to or from the shallow groundwater system depending upon the topographic surface and water table elevation of the location. Generally, groundwater flows towards the center of the basin from the surrounding areas. The yield, specific capacity and hydraulic conductivity values of the aquifers in unconsolidated sediments are more in northern part of the basin in comparison to that of the southern part. The hydraulic conductivity values reaches up to around 25 m/day in the northern part. Analysis of pumping test data in some of the wells considered in this study shows that generally the hydraulic conductivity values of the fractured and deep aquifers are low (Table 1). The three-dimensional

Table 1: Thermal gradient (TG) and Heat flow (Q) calculated for basement and overlying sedimentary cover

Well ID	Well depth (m)	Altitude (m)	Basement Depth (m)	Water level (m asl)	Hydraulic conductivity (10^{-2} m/day)	Basement		Sediment	
						TG (°C/100 m)	Q (mW/m²)	TG (°C/100 m)	Q (mW/m²)
HS5	800	62.5	646	37.0	35.0	-	-	3.87	56.87
HS9	1500	76.0	552	-	-	3.01	96.68	-	-
HS11	813	94.0	30	-	-	3.38	108.63	-	-
HS18	700	70.0	472	-	-	3.02	96.99	3.82	54.59
HS43	625	54	617	13.3	81.0	-	-	2.26	35.41
HS53	1100	86	200	54.0	1.0	2.55	81.71	-	-
HS54	650	38	18	21.1	0.7	2.60	83.53	-	-
HS60	800	111	400	77.5	2.3	2.55	81.87	2.56	38.87
HS104	1100	62.5	610	-3.0	0.3	2.49	79.89	1.70	22.99
HS123	1100	273	10	272.9	1.7	2.80	89.88	-	-
HS128	1021	70	268	63.2	1.4	2.52	80.92	0.83	13.46
New	1300	70	519	-	-	2.56	82.09	2.24	34.67

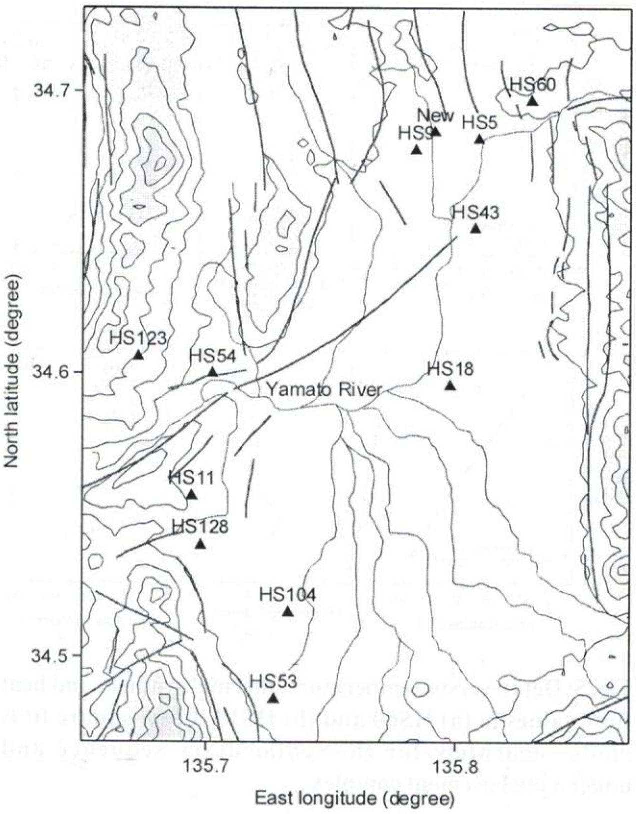

Fig. 2: Location of the wells in which temperature logging has been carried out. The solid lines are the faults.

groundwater flow model of the basin shows that groundwater is flowing towards the central part of the basin from the surrounding areas and the flux is high in the northern part of the basin (Pathak 2002). The discharge of groundwater water outside the basin takes place through the Yamato River.

SUBSURFACE TEMPERATURE AND GEOTHERMAL GRADIENT

The temperature logging and other hydrogeological data from 12 deep wells have been used to calculate the heat flux, which is considered to estimate the vertical groundwater flow in selected sites in the study area (Fig. 2). These wells are reaching deep into the basement and the lithologic log, pumping test, and temperature logging data have been used during this study. The temperatures are measured at an interval of 50 m.

The downward change in temperature in boreholes falls within narrow zones for most of the wells (Fig. 3). However, the subsurface temperature in well HS123 is lower than that of others. This well lies at higher altitude than other wells and belongs to the recharge zone. Hence, the temperature in this well is low in comparison to other wells.

The depth wise changes in temperature, thermal gradient and heat flow in some representative well sites are plotted in Figs. 4, 5, 6, and 7. In some wells, the temperature logging is carried out only in a portion of drilled depth. Sharp change in temperature is clearly observed at the boundaries of sediments and underlying basement rocks in sites HS104, HS128, HS18, and HS9 (Figs. 4, 5b, and 7a). The temperature at the base of sediments range from 27.1 °C to 40.0 °C, respectively in HS128 and HS5, which is directly related to the depth of the sediment-basement interface (Figs. 4b and 6a).

Slopes of least square fitted straight lines to temperature-depth plots were used to determine the thermal gradient for basement and sedimentary cover. Very high correlation coefficients can be seen in the least square fit. The thermal gradient in basement varies from 2.49 °C/100 m to 3.38 °C/100 m while in the sedimentary cover it ranges from 0.83 °C/100 m to 3.87 °C/100 m (Table 1). The geothermal gradient is comparatively low in the southern part of the study area. The lowering of geothermal gradient may be due to the lateral heat transport, driven by groundwater circulation that caused decrease in temperature (Clauser 1999). Majorowicz and Jessop (1981) described that low geothermal gradient areas coincide with water recharge areas and high hydraulic head distribution regions. Because most of the well sites considered in this study lie on or near the mountainous area (recharge area), a good comparison of the difference in temperature gradient between recharge and discharge area could not be achieved. However, the well HS18 that lies in the central east of the basin shows comparatively higher thermal gradient. This well site is interpreted to be lying in the discharge zone because groundwater is moving towards the basin from the surrounding mountainous areas.

THERMAL CONDUCTIVITY ESTIMATION

Thermal conductivity values of the sediments and basement rocks of the area could not be measured because no boring core samples were available from the wells. However, the thermal conductivity of sediments and basement rocks of the Nara Basin was estimated based on the thermal conductivity data available from the adjacent Osaka Basin. The geology of both areas is similar. Nakagawa and Komatsu (1979) measured thermal conductivity values of different sedimentary facies and that of the basement rocks of Osaka Basin. The thermal conductivity values of the sedimentary facies were shown mostly depending on the porosity of the material. The porosity of clay, silt and sand, in average, can be considered to be 45-55%, 40-50%, and 35-40% (Todd 1980). These estimated porosity values for Nara Basin fall well within the measured porosity values in Osaka Basin (Nakagawa and Komatsu 1979). The measured thermal conductivity values vary negligibly within the above porosity range. The thermal conductivity values are estimated to be 1.18, 1.58, 1.67 and 3.21 W/m.°C for clay, silt, sand and granite, respectively. This method of thermal conductivity estimation is appropriate because thermal conductivity of geologic materials varies over a much narrower range than does hydraulic conductivity (Clark 1966; Freeze and Cherry 1979).

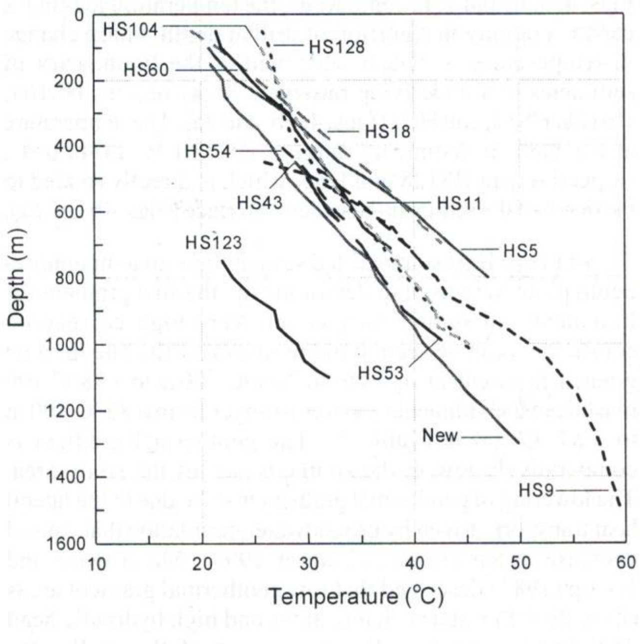

Fig. 3: Plot of depth versus temperature in the boreholes

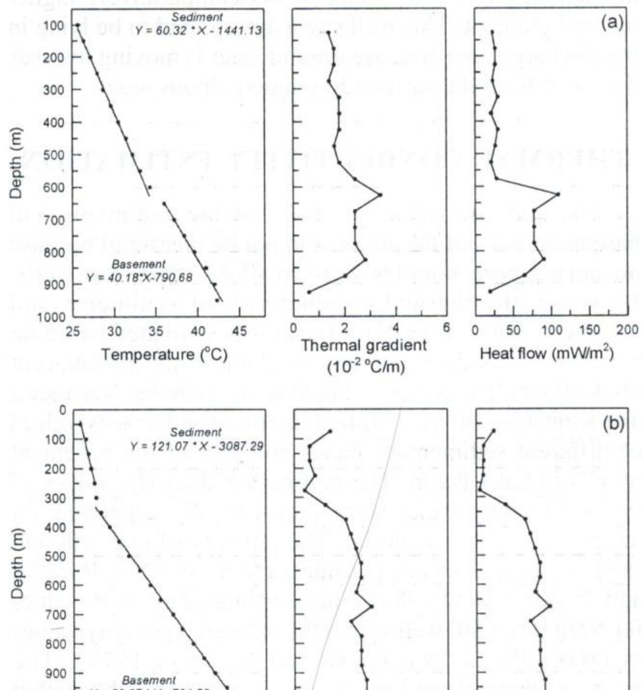

Fig. 4: Depth versus temperature, thermal gradient and heat flow values in (a) HS104 and (b) HS128. Least square fit is made separately for the sedimentary sequence and underlying basement complex.

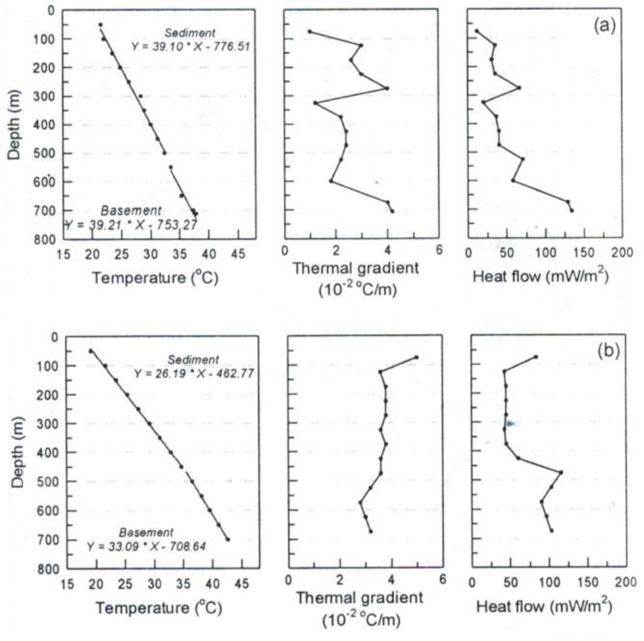

Fig. 5: Depth versus temperature, thermal gradient and heat flow values in (a) HS60 and (b) HS18. Least square fit is made separately for the sedimentary sequence and underlying basement complex.

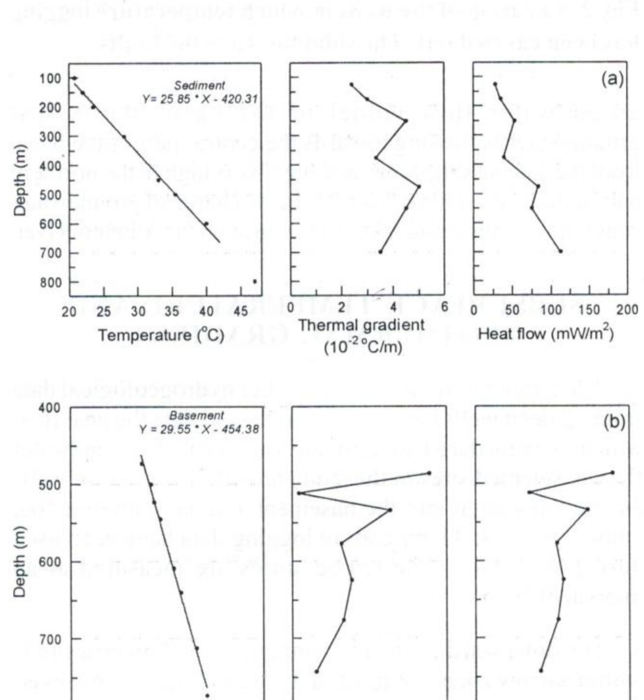

Fig. 6: Depth versus temperature, thermal gradient and heat flow values in (a) HS5 and (b) HS11

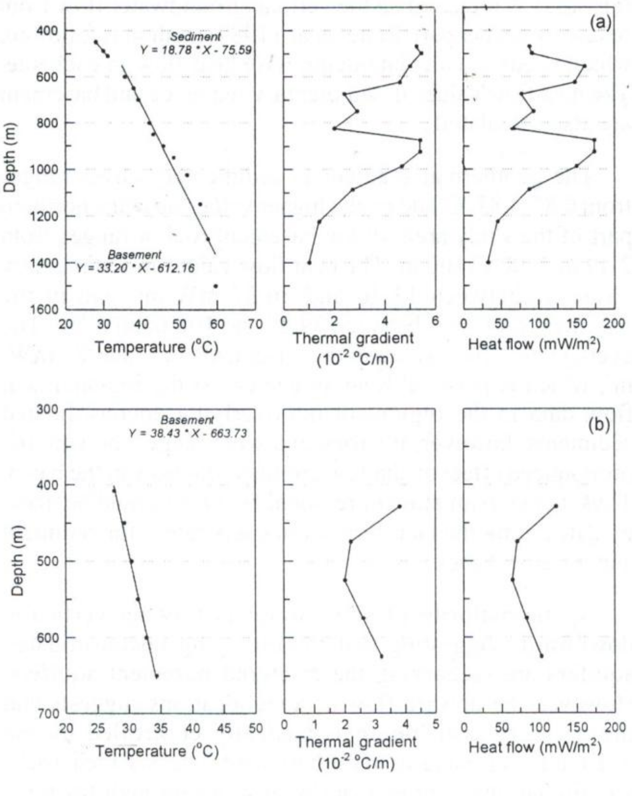

Fig. 7: Depth versus temperature, thermal gradient and heat flow values in (a) HS9 and (b) HS54

The effective heat conductivity k for the sedimentary layer has been calculated using the following relationship (Majorowicz et al. 1986):

$$k = \frac{\sum_{i=1}^{n} l_i}{\sum_{i=1}^{n} (l_i / k_i)} \qquad (13)$$

Where, the quantities k_1, k_2, ..., k_n are estimated heat conductivities based on rock types for discrete layers of thickness l_1, l_2,......, l_n. The effective heat conductivity is especially important to calculate the heat flow in sedimentary layer where different layers of varying heat conductivity values lies within length of calculated thermal gradient. Equation (13) is not applied in case of the granitic basement where the conductivity value is supposed to be constant throughout the depth of interest.

HEAT FLOW

The thermal gradient obtained from least square fit was multiplied by the estimated thermal conductivity of the basement rock (granite) and the estimated effective thermal conductivity for the sedimentary cover to calculate the heat flow (equation 1). Depth wise heat flow is calculated for each interval of temperature measurement (Figs. 4, 5, 6 and 7). The similar trend of thermal gradient and heat flow in the basement is because the thermal conductivity is considered constant for the entire granitic basement. However, in case of sedimentary formation, the effective thermal conductivity values are used to calculate the intervals of temperature measurement. Comparatively higher heat fluxes are observed in the basement rocks (Figs. 4 and 5). The heat flow values in the basement lie between 79.89 mW/m^2 and 83.53 mW/m^2 in the majority of sites (Fig. 8, Table 1). However, in three sites HS9, HS11 and HS18, it is comparatively higher, reaching up to 108.6 mW/m^2. The values in majority of the sites are very close to the one calculated at Nakagun, Wakayama Prefecture (75 mW/m^2), lying about 50 km southwest of the basin (Clark 1966). In the overlying sedimentary cover, the heat flow values ranges from 34.67 mW/m^2 to 56.87 mW/m^2 with the exception of two sites (HS128 and HS104) where it is 13.46 mW/m^2 and 22.99 mW/m^2, respectively.

The heat flow values in sedimentary sequence as calculated in this study is in good agreement to that of Osaka Basin (Nakagawa and Komatsu 1979). In the Osaka Basin, the heat flow values ranges from 32.2 mW/m^2 to 46.1 mW/m^2. The low heat flow values in the overlying unconsolidated sediments show the pronounced effect of subsurface water flowing through the aquifers.

VERTICAL SPECIFIC FLOW OF GROUNDWATER

Vertical flux of groundwater is calculated using equation (11) as proposed by Reiter et al. (1989) in eight well sites (Figs. 9, 10, 11 and 12). This method is quite effective to identify the depth interval in which water is flowing vertically. The slopes of Q-T (heat flow-temperature) plots are obtained for the region where vertical flow of water is evidenced and the values of constant c (4.184 kJ kg^{-1} $°C^{-1}$) and ρ (10^3 kgm^{-3}) were used to calculate vertical specific discharges.

The positive slope, indicating the downward flow of water, ranges from 2.328 mW/m^2°C in HS5 to 12.628 mW/m^2°C in HS60 (Table 2, Figs. 11a and 10a). Likewise, the negative slope (upward groundwater flow) lies between –6.527 mW/m^2x°C in HS11 and –11.178 mW/m^2x°C in HS9 (Figs. 11a and 12a). Normally, downward flow is observed in many sites, especially within the depth range of 300 m - 600 m. In the sediments or sediment-basement zone, the downward flow is as high as 3.019×10^{-9} m/sec (HS60). In the southern part, two sites (HS11 and HS104) have upward discharge of 1.561×10^{-9} m/sec and 1.899×10^{-9} m/sec through the deeper part of the basement. In the northern part, the upward flow is observed in only one well site (HS9). There is good agreement between the screened zone in the deep wells and the depth interval of flow region obtained from present study. The screen locations are on the fractured zone. This suggests that the vertical flow in basement is controlled by the distribution and degree of fracturing.

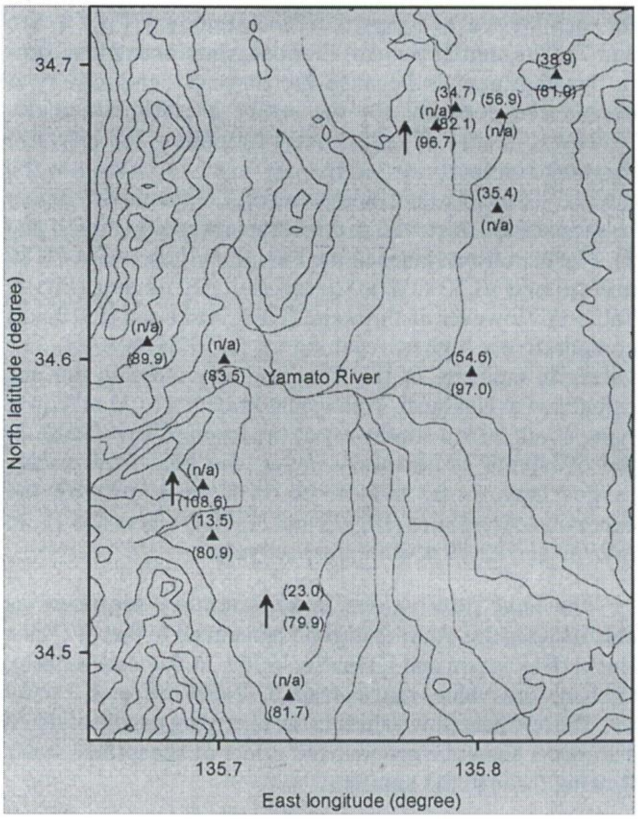

Fig. 8: Heat flow values (mW/m²) in sediment and in basements are given above and below the symbols, respectively. The values are calculated from least square corrected thermal gradient. The arrows show the locations having vertical (upward) groundwater flows.

DISCUSSION AND CONCLUSION

Understanding the direction of vertical flow of water is very important to understand the hydrogeologic regime of a groundwater basin. Determination of vertical flow is very difficult because it is necessary to have water level data from wells at different depths should be available. Though such data could be available for the shallow depth, it is practically rare for the very deep aquifer, especially in the basement. There is no hydraulic test method to measure vertical specific discharge and vertical hydraulic conductivity from a single well. In order to overcome this difficulty, the temperature log data from several deep wells in the Nara Basin have been used to get an idea about the interaction of groundwater between the basement aquifers and overlying sedimentary aquifers. The temperature log data may be either one time or transient measurement. From the transient temperature measurement in borehole, Taniguchi (1993) showed that the monthly change in temperature takes place only within some tens of meters of depth, below which it is almost constant round the year. In this study, temperature measurement data from single observation is used because this study is focussed on the vertical groundwater flow from relatively deeper part. Reiter et al. (1989) method is followed, which needs to calculate depth wise heat flow in each site. The heat flow values in sedimentary sequence and basement are also calculated.

The geothermal gradient in sedimentary cover ranges from 0.83-3.87 °C/100 m, the higher values lie in the northern part of the study area. In the basement rock it ranges from 2.49 to 3.38 °C/100 m. The heat flow values in sedimentary cover lie between 13.46 and 56.87 mW/m², and in the basement rocks it is between 79.89 and 108.63 mW/m². The average heat flow value in the basement is around 85 mW/m², which is reasonably in agreement to the regional heat flow data in the region. In the overlying unconsolidated sediments, however, it varies in a wide range, showing the pronounced effect of shallow groundwater flow in the basin. Thus, better estimation of regional heat flow could be made by calculating the heat flow values separately for sediment and basement rock.

At the majority of sites, water is flowing vertically downward, suggesting that the overlying unconsolidated aquifers are recharging the fractured basement aquifers. However, the upward flow at some locations suggests that the pressure distribution is basically controlled by the distribution of fractures in the basement. The basement rocks are discharging a large quantity of water through fractures at different depths as observed during the very deep drilling. This study could locate the flow regions in fractured basement rocks, where sufficient quantity of water is flowing vertically and the lithology do not vary with depth. In order to locate the flow regions and calculate the vertical specific discharges in shallow unconsolidated formations, precise temperature measurement is necessary at narrow interval. Though the 50 m interval temperature logging data, as in present study, could detect the flow regions in fractured basement, this interval is not enough for shallow depth where the sedimentary facies changes within some meters of depth. Thus, heat flow method can be a good tool to confirm the location of fracture zones and to calculate the vertical direction of groundwater flow.

ACKNOWLEDGEMENTS

The author is thankful to Prof. K. Nakagawa, Osaka City University, Japan for providing laboratory facilities and his constant encouragement and valuable suggestions during various stages of the study. Mr. M. Sato, Tokushu Plant Industry Co. Ltd., Osaka, Japan, provided valuable data of deep borings. I am thankful to Dr. Marshall Reiter, New Mexico Bureau of Geology and Mineral Resources, USA for providing valuable comments and suggestions on the manuscript. Dr. H. Nemoto, Lecturer, Osaka City University has helped during various stages of this study. The author is thankful to the Government of Japan for supporting the research work through Monbukagakusho scholarship.

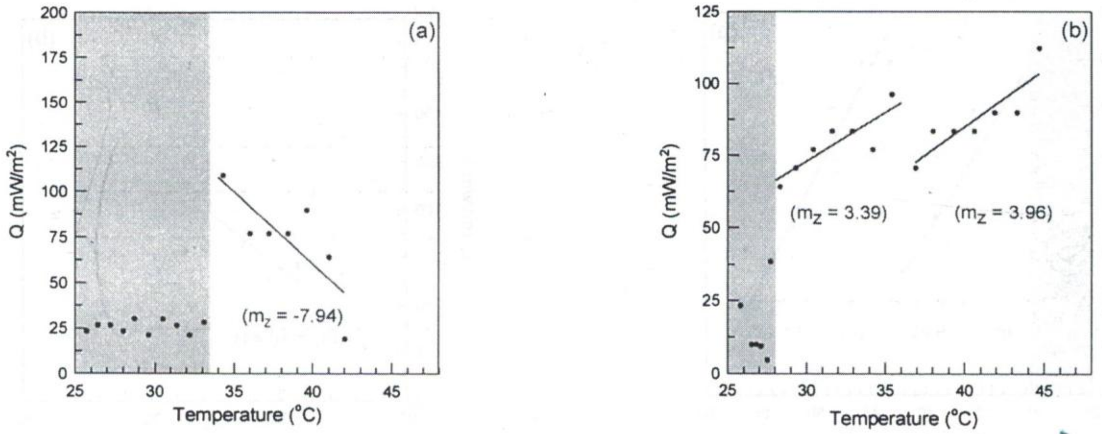

Fig. 9: Plot of heat flow versus temperature (Q-T). (a) One vertical flow region between 600 m and 950 m is observed in HS104. (b) two zones between 350 m and 650 m, and 700 m and 1,000 m are identified in HS128. m_z represents slope of the least square fitted line. Shaded area represents zone of no vertical groundwater flow.

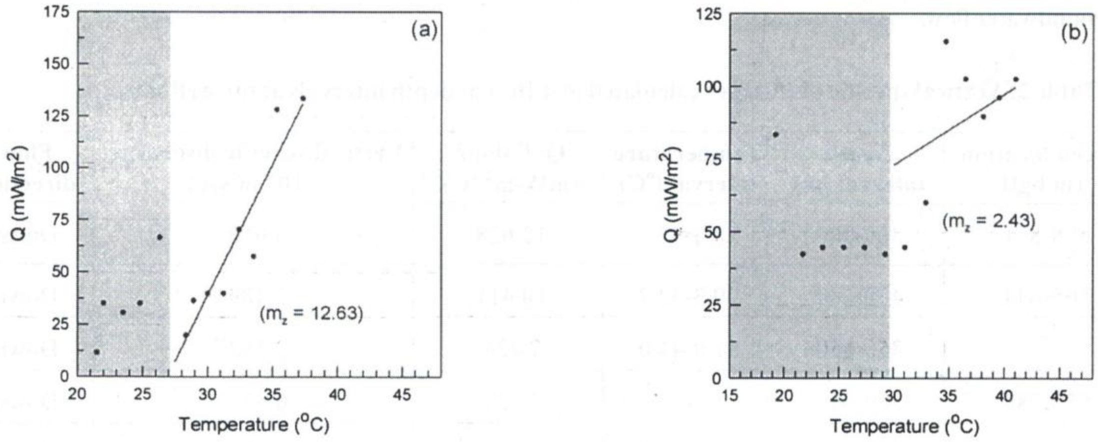

Fig. 10: Plot of heat flow versus temperature (Q-T). Flow region is observed between (a) 300 m-700 m in HS60 and (b) 350 m-650 m in HS18. m_z represents slope of the least square fitted line. Shaded area represents zone of no vertical groundwater flow.

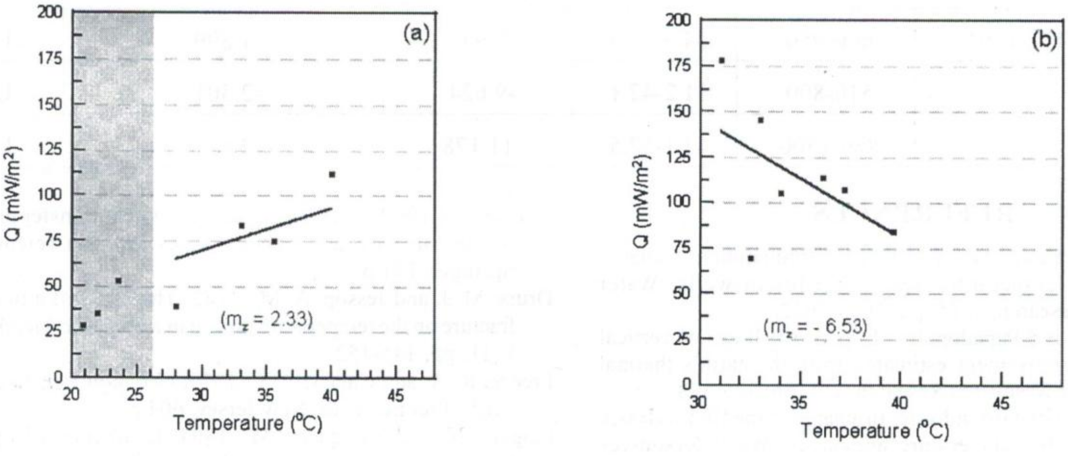

Fig. 11: Plot of heat flow versus temperature (Q-T). The flow regions are between (a) 300 m - 600 m in HS5 and (b) 475 m - 712 m in HS11. m_z represents slope of the least square fitted line. Shaded area in 11(a) represents zone of no vertical groundwater flow.

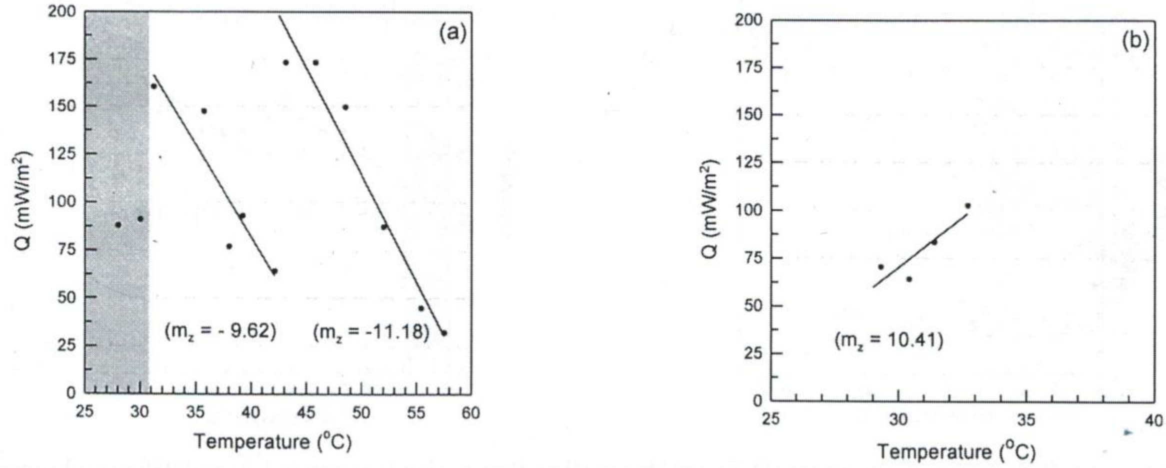

Fig. 12: Plot of heat flow versus temperature (Q-T). The flow zones are between (a) 510 m-800 m, and 850 m-1300 m in HS9, and (b) 450 m-600 m in HS54. m_z represents slope of the least square fitted line. The shaded area in 12 (a) represents zone of no vertical groundwater flow.

Table 2: Vertical specific discharges calculated at different depth intervals at the well sites.

Well ID	Screen location (m bgl)	Depth interval (m)	Temperature interval (°C)	Q-T slope (mW/m^2 °C)	Vertical specific discharge (10^{-9} m/sec)	Flow direction
HS60	698-801	300-700	28.3-37.3	12.628	3.019	Down
HS54	505-644	450-600	29.3-32.7	10.411	2.489	Down
HS18	-	350-650	31.0-41.0	2.434	0.582	Down
HS5	580-788	300-600	28.0-40.0	2.328	0.557	Down
HS128	751-1015	350-650	28.3-35.4	3.392	0.811	Down
		700-1000	36.9-44.7	3.958	0.946	Down
HS11	-	475-712	31.0-39.6	-6.527	-1.561	Up
HS104	477-1094	600-950	34.3-42.3	-7.943	-1.899	Up
HS9	-	510-800	31.2-42.1	-9.624	-2.301	Up
		850-1300	43.1-57.5	-11.178	-2.673	Up

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