

Thermal properties across the Chelungpu fault zone and evaluations of positive thermal anomaly on the slip zones: Are these residuals of heat from faulting?

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[1] Thermal properties of the fault zone materials are of fundamental importance for estimate of frictional heat generation during earthquake. The properties across the Chelungpu fault zone activated by 1999 Chi-Chi earthquake, Taiwan, are measured using the drilled core penetrating the fault zone at around 1100 m depth. The fault zone contains four distinct fracture zones, each of which includes thin slip zones. Thermal conductivity (K) lies between 1.0 and 3.0 $\text{Wm}^{-1}\text{K}^{-1}$ and shows the lowest value at the slip zones. Thermal diffusivity (α) varies between 0.8 to $2.0 \times 10^{-6} \text{ m}^2\text{s}^{-1}$, and is relatively low at the slip zones. Density (ρ) varies between 2200 to 2800 kgm^{-3} and shows the lowest values at a particular slip zone (1110 m depth). Specific heat (c) is calculated by using above data resulting in the values from 300 to 1000 $\text{Jkg}^{-1}\text{K}^{-1}$, and lowest values for slip zones. Using these data and spectral gamma ray logs, reported positive thermal anomalies at the slip zones are re-examined whether they are regarded as residual heat from friction by faulting. **Citation:** Tanaka, H., W. M. Chen, K. Kawabata, and N. Urata (2007), Thermal properties across the Chelungpu fault zone and evaluations of positive thermal anomaly on the slip zones: Are these residuals of heat from faulting?, *Geophys. Res. Lett.*, 34, L01309, doi:10.1029/2006GL028153.

1. Introduction

[2] Past studies of frictional heating by seismic slip have been performed mainly by theoretically or experimentally [McKenzie and Brune, 1972; Lachenbruch and Sass, 1980; Mase and Smith, 1987; Lockner and Okubo, 1983]. They have assumed thermal transport properties of the slip zone materials to estimate frictional heat generation. However, for natural fault zones, these properties are essential to estimate the frictional heat generated during an earthquake. Thermal diffusivity defines the transport distance of conductive heat from slip zone, and specific heat controls frictional temperature rise within a slip zone. Also, variation of thermal conductivity is directly related to the fluctuation of the geothermal gradient, which would lead a right decision whether thermal anomaly observed in the data is related to frictional heating. Until now, however, there have been few direct measurements of these properties from natural fault zone materials.

[3] Following the first drilling into the Chelungpu fault, Taiwan in 2000, which was activated by 1999 Chi-Chi earthquake (Mw 7.6) [Tanaka *et al.*, 2002], deep drilling project started in 2004 to penetrate the fault, and recover the materials of fault zone successively. The project provides unique opportunity to measure the thermal properties quantitatively across the active fault zone. The fault surface near the drilling site dips about 30°E with a slip vector of 66° during 1999 Chi-Chi earthquake [Ma *et al.*, 2001]. The slip zone of Chi-Chi earthquake is expected to be located at around 1100 to 1300 m depths in the drill hole based on the preliminary seismic reflection research [Wang *et al.*, 2007] (Figure 1) together with surface rupture observation. Results of drilled core observation clarify that several fracture zones exist at around 1100 to 1300 m (Figure 2). Here, we report the data sets of thermal properties across the active fault zones including that activated by 1999 Chi-Chi earthquake and discuss the effects of these properties to positive thermal anomaly, which are interpreted as residual heat from frictional faulting by Tanaka *et al.* [2006] and Kano *et al.* [2006].

2. Outline of Host and Fault Rock Materials

[4] The core materials recovered from borehole are first observed visually and lithological and structural descriptions are performed (Figures 2a and 2b). Crack density (numbers of cracks per meter) is measured regardless whether they are natural or artificial (Figure 2c). Host (undamaged) rocks are basically composed of sandstone, sandy siltstone and alternation of these two. Fracture zone is composed of proto breccia, fault breccia and fault gouge (Figure 2b). The proto breccia is simply defined as rocks with crack density $>20 \text{ m}^{-1}$, which can be comparable with rocks in “damage/process zone” [Chester *et al.*, 1993; Vermilye and Scholz, 1998]. Fault breccia and fault gouge are incohesive fault rocks, which contain visible rock fragments $>30\%$ and $<30\%$ respectively [Tanaka, 1992], and generally modified mineral composition by hydrothermal alteration [Tanaka *et al.*, 1995]. The proto breccia zone generally surrounds, or closely associated with zones of fault gouge and/or fault breccia. Fault gouge zones are generally 1 cm to 3 m in thickness (Figure 2d), which contain one to several thin planar zones, 5 to 100 mm in thickness, composed of soft, ultrafine-grained materials, in which few visible clasts are observed. This material is tentatively referred to as ultrafine gouge. The boundaries between thin zones of ultrafine gouge and fault gouge are very sharp and flat. We assume that the thin zones of ultrafine gouge as slip zones based on the similarities of occurrences reported by previous researches performed at

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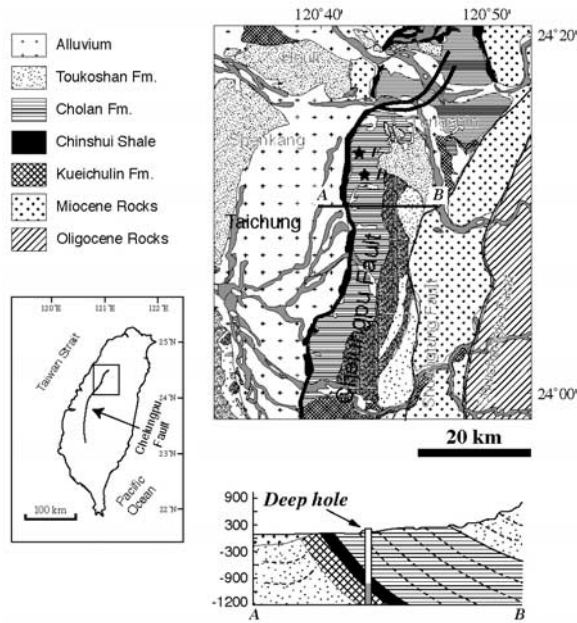


Figure 1. Locations of Fenguan (shallow, star with F) and Daken (deep, star with D) drill sites and geological cross section with outline trace of deep penetration drilling. Shaded zone in deep borehole corresponds to columnar sections in Figure 2.

active fault zones recovered from drilled core and exhumed fault zones [Chester et al., 1993; Tanaka et al., 2001; Sibson, 2003, Tanaka et al., 2006]. Four distinct fracture zones are observed at the depths of 1110 m, 1150 m, 1190 m and 1230 m (Figures 2a and 2b). For simplicity, these

fracture zones are referred to as A, B, C, and D zones, respectively. The thickness of these fracture zones ranges from 8.4 m (A zone) to 32.1 m (B zone).

3. Measurement Techniques

[5] After mesoscopic observations, we put the resin to fix the core surface especially for fragile parts, then split the core to be two halves parallel to the long axis. The cut-surfaces are checked visually and confirmed that surfaces are reasonably flat and the resin does not penetrate into the sample. The samples are 9.2 cm in diameter and 10 to 20 cm in length. Next, we measure the thermal properties on the cut-surface right after the splitting. The Transient Plane Source (TPS) techniques are used to measure thermal conductivity (K) and thermal diffusivity (α) of host and fault zone rocks. The apparatus is commercially available, which is known as Hot Disk System [Gustavsson, 1991; Gustavsson et al., 1995]. In this method, a thin plane sensor in the shape of double spiral resistance element, consisting of a number of concentric ring sources, was used. The sensor, 7 mm in diameter, is inserted between two flat surfaces of sample pieces, where it acts both as a heater and a resistance thermometer at the measurement. The samples are saturated in the water to put samples in conditions similar to natural underground and clamped two cut surfaces together tightly so that the sensor is in reasonable thermal contact with the sample. Repeated measurements of the standard sample (marble, K ; $4.70 \text{ Wm}^{-1}\text{K}^{-1}$, α ; $1.68 \times 10^{-6} \text{ m}^2 \text{ s}^{-1}$) confirms the error for thermal conductivity is less than 3% and for thermal diffusivity is less than 5%. Total 58 samples are measured across the fracture zones (Figures 2e and 2f).

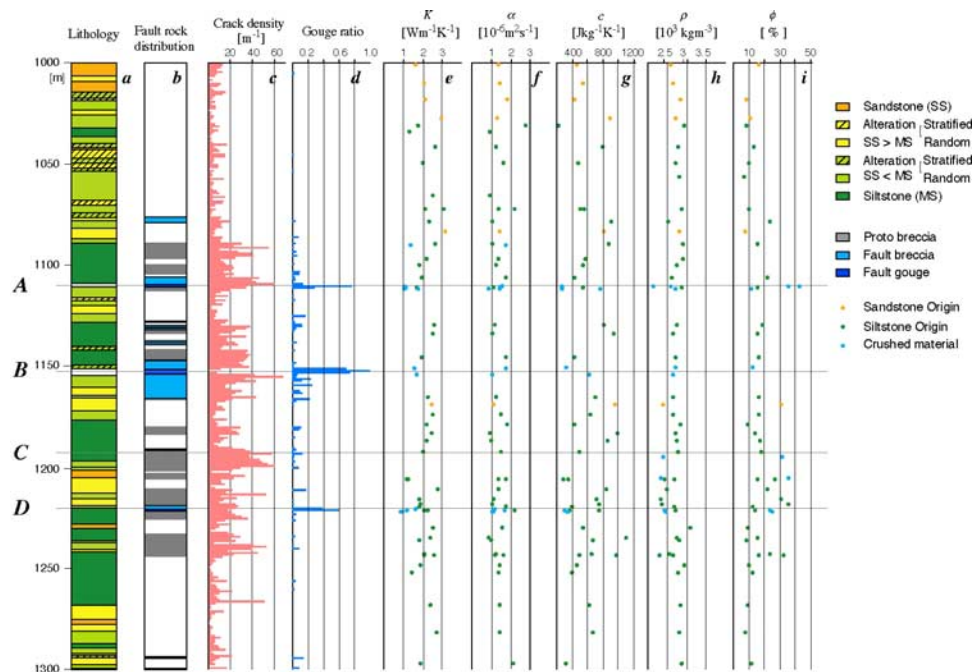


Figure 2. (a) Lithological and (b) structural columnar sections and distributions of (c) crack density, (d) gouge ratio (a ratio of gouge thickness to the unit length [1 m] of the core), (e) thermal conductivity K , (f) thermal diffusivity α , (g) specific heat c , (h) density ρ , and (i) porosity ϕ .

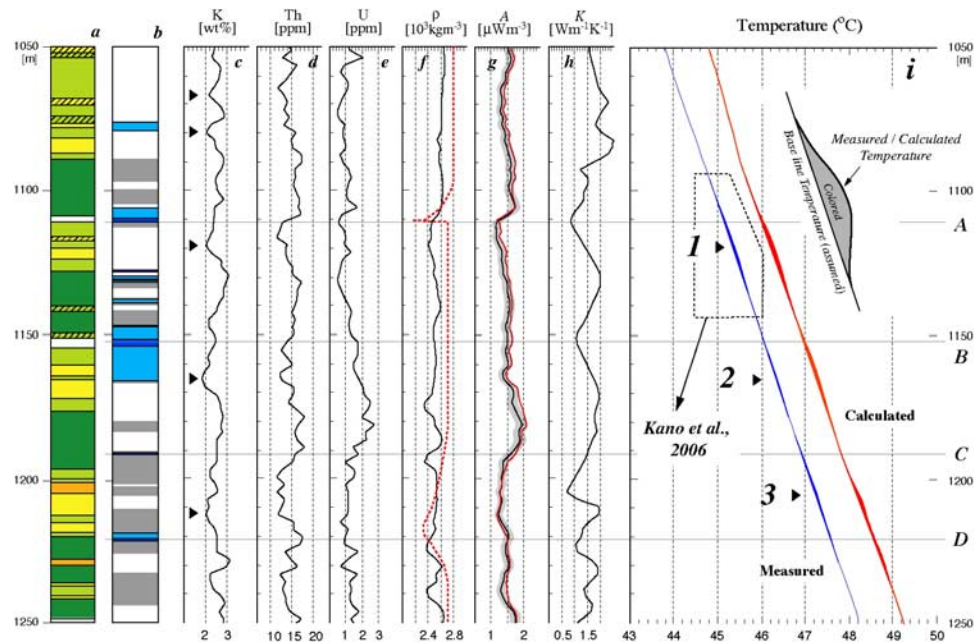


Figure 3. (a) Lithological and (b) structural columnar sections are compared with the results of spectral gamma ray logs for (c) K, (d) Th, (e) U, (f) two density models, (g) radiogenic heat production rate A , and (h) thermal conductivity model. (i) The calculated geothermal gradient (red line) is compared with temperature record across the Chelungpu fault zone (blue line). For easier comparison, measured data is subtracted by 1°C . Fluctuations of temperature are in good agreement between data and model. See text for explanations.

[6] The density (ρ) is measured for small chips of samples that are obtained from very close to the measurement points of thermal properties. We use He-gas type pycnometer (Accupic 1330, MicrometricsTM) for volume measurements and digital balance with accuracy to 0.0001 g was used to weigh the samples. A minimum weight of 10 g is used to obtain appropriate values of volume for recovered core materials. Samples are soaked into the water and evacuated 30 minutes by vacuum pump at 10 Pa. After evacuation, the weight and volume of wet sample are measured. Then these samples are dried at 105°C overnight, dry weight of sample is measured. Calibration is performed using crystal quartz of density 2.67 g cm^{-3} . The error for values of density is less than 2%. Porosity is calculated by $\phi = (M_w - M_d)/(\rho_w V_w)$ assuming water density (ρ_w) 1 g cm^{-3} , M_w and M_d are the mass of wet and dry state of the sample and V_w is volume of the saturated sample. Total 58 samples are measured (Figures 2h and 2i). Finally, we obtain 52 data of specific heat using the equation $c = K/\rho\alpha$ (Figure 2g). We obtain less numbers of specific heat because if the sampling point is more than 5 cm apart between thermal property and density, we treat these data as sampled from different depths.

4. Results

[7] General characteristics of thermal transport properties across the fault zone are as follows. Thermal conductivity data lie between 1.0 and $3.0\text{ Wm}^{-1}\text{K}^{-1}$ for host and fractured rocks and show the lowest values around $1.0\text{ Wm}^{-1}\text{K}^{-1}$ at the slip zones (Figure 2e). Thermal diffusivity data vary between 0.8 to $2.0 \times 10^{-6}\text{ m}^2\text{s}^{-1}$, and is relatively low at the slip zones ($1.0 \times 10^{-6}\text{ m}^2\text{s}^{-1}$, Figure 2f). Density data lie between 2200 kgm^{-3} and 2800 kgm^{-3} and

show the lowest value at the slip zones (Figure 2h). Specific heat shows low values from 300 to $1000\text{ Jkg}^{-1}\text{K}^{-1}$. The lowest values around $300\text{ Jkg}^{-1}\text{K}^{-1}$ are obtained from slip zones.

[8] Among four fracture zones, C zone is less well characterized for thermal properties due to heavy artificial fragmentation. On rest of three fracture zones (A, B and D), the thermal conductivity lowers from host rock towards the fault gouge zones (Figure 2e). The widths of transitions zones are about 20 m for A and B zones, and 8 m for D zone, which roughly corresponds to proto breccia zone (Figures 2b and 2e). The thermal conductivity is known to correlate roughly with density, porosity, grain size, degree of cementation and mineral composition [Brigaud and Vasseur, 1989; Kitajima et al., 2001; Maqsood et al., 2004; Hartmann et al., 2005]. The correlation between density and porosity and with conductivity shown in auxiliary material¹ Figure S1 may be related to the increase in pore water that lowers the effective thermal conductivity of fractured rocks as suggested by Matsubayashi et al. [2005]. However, as noted previously, change in mineral composition and grain size reduction in fault rocks may also result in lowering the thermal conductivity.

5. Re-Examination of Positive Thermal Anomalies on the Slip Zones

[9] There are some debates that which fracture zone is the one activated by 1999 Chi-Chi earthquake, although most of researchers related to this project would believe that A zone is the one based on abrupt change in resistivity and existence of a thin layer composed of ultra-fine black

¹Auxiliary materials are available in the HTML. doi:10.1029/2006GL028153.

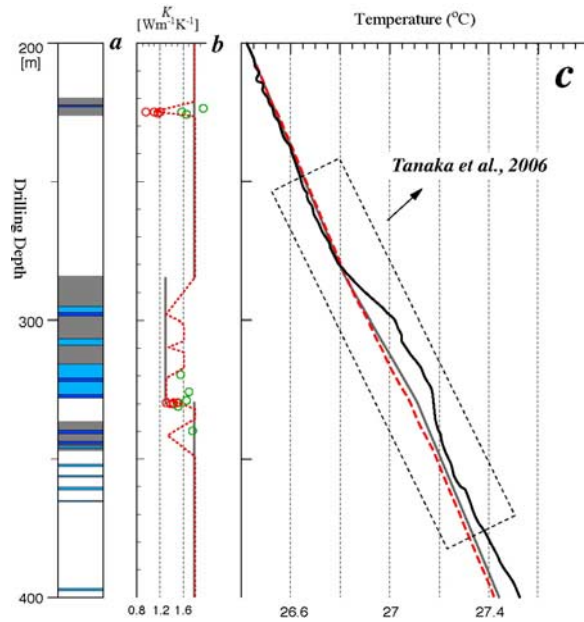


Figure 4. Structural columnar section (a) for shallow borehole, (b) with thermal conductivity data and thermal conductivity models (gray and red dotted lines). Green circles, thermal conductivity data from host rocks of sandstone and mudstone; red circles, those from fracture zone rocks including fault breccia and fault gouge. (c) Temperature record associated with model calculations of geothermal gradient. First and second models of conductivity are shown by the gray and red dotted lines. The positive thermal anomaly can still be observed beyond the fluctuations of geothermal gradient.

materials in the fracture zone. Temperature log presented by *Kano et al.* [2006] is shown in Figure 3i. They suggested that slip zone at 1111 m (A zone) shows the positive thermal anomaly about 0.06°C (peak 1 in Figure 3i). They used this value for estimate of frictional heating. However, we can observe two other thermal peaks (peaks 2 and 3 in Figure 3i) in addition to the peak 1. The peak 2 shows 0.03°C positive thermal anomaly, which is located roughly on the B zone. The peak 3 shows 0.04°C positive thermal anomaly and does not correlate with any slip zones. Since all of these thermal peaks are so small, temperature disturbance by variation of thermal conductivity across the fracture zone and that by radiogenic heat production should be considered, even though, as *Kano et al.* [2006] suggested, there is little disturbance of temperature distribution by fluid flow. Using thermal property data and spectral gamma ray logs for radioactive elements (K, Th, and U), we re-examine the temperature data presented by *Kano et al.* [2006].

[10] Assuming one-dimensional steady-state heat conduction with constant thermal conductivity and uniform heat production rate, heat transfer with volumetric heat generation can be expressed as $A = -K \frac{d^2 T}{dz^2}$ [Turcotte and Schubert, 1982]. The boundary conditions are, $K \frac{dT}{dz} = q_s$ and $T = T_s$ at the surface. Then the equation is integrated as:

$$T(z) = T_s + \frac{q_s}{K}z - \frac{A}{2K}z^2 \quad (1)$$

$$q(z) = q_s - Az \quad (2)$$

where T_s and q_s are temperature and heat flux at the surface, respectively, and z is depth. A is known to be estimated as;

$$A = 10^{-5} \rho (3.48C_k + 9.52C_u + 2.56C_{Th}) \quad (3)$$

where A 's units are μWm^{-3} , C_k , C_u and C_{Th} are concentrations of K (wt %), U and Th (ppm), respectively [Rybach, 1976; Kitajima et al., 2001].

[11] The spectral gamma ray logs for K, Th and U are presented in Figures 3c, 3d, and 3e, with lithology and fault rocks distributions (Figures 3a and 3b). Concentration of radioactive elements are more correlated with lithology than fault rock distribution, that is, K and Th are more contained in mudstones than sandy rocks, and lowest values of concentration of these two elements to the sandstones (black triangles in Figure 3c). U is increased in mudstone distributed just below shear zone A. These observations are in good agreement with previous research for sedimentary rocks [Stanley et al., 2006]. Two models of density are examined to estimate radiogenic heat production rate. For the first model, we simply use geophysical density logs (black line in Figure 3f). The second model is that the density is constant for host rocks (the value is the one averaged from host rocks) and lowers in the damage zone and is the lowest at slip zones, in which the trend roughly follows the measured data (red dotted line in Figure 3f). Errors of geophysical logs are assumed to be large, $\pm 10\%$, for our calculation. The estimated heat production rate A is around $1.6 \mu\text{Wm}^{-3}$ (Figure 3g), which is comparable with the previous study [Kitajima et al., 2001]. The results also show that difference between density models does not affect so much to the depth profile of heat production rate (Figure 3g). The radiogenic heat production rate lowers at the A and C zones, but no obvious change at B and D zones.

[12] In order to examine the effect of fluctuations of radiogenic heat production and that of thermal conductivity with depth to the geothermal gradient, we assume an infinitesimal thickness of a layer, which has uniform heat production rate and thermal conductivity, and variations of these two parameters with depth are regarded as variations of these two among the layers. Then change in temperature through the layers is integrated to evaluate fluctuation of geothermal gradient using equation (1) and following equations:

$$T_i = T_{i-1} + (q(z)/\bar{K}_i)\Delta z - (\bar{A}_i/2\bar{K}_i)\Delta z^2 \quad (4)$$

$$\bar{K}_i = \frac{\sum_{i=n}^{i+n} K_i}{2n+1}, \quad \bar{A}_i = \frac{\sum_{i=n}^{i+n} A_i}{2n+1} \quad (5)$$

$$\Delta z = z_i - z_{i-1} \quad (6)$$

where T_0 is temperature at the depth z_0 , z_0 is starting depth of temperature measurement, subscript i is the each layer for calculation and n is numbers of layers of summation.

Conductivity profile is modeled simply by interpolation of the data presented in Figure 2 (Figure 3h). The calculated geothermal gradient is parallel with the measured temperature data if we assume apparent heat flow of 45 mWm^{-2} (Figure 3i). The value of heat flow would be more or less lower than intrinsic value because temperature was measured half a year after drilling [Kano *et al.*, 2006], although recovery of temperature generally takes one to two years [Sass and Morgan, 1992]. The heat production rate is so small to affect the calculated geothermal gradient ($< 0.1\%$), indicating the fluctuations observed in calculated geothermal gradient are mostly resulted from fluctuations of thermal conductivity. Fluctuations of calculated geothermal gradient are observed at the depths around 1110, 1160 and 1200 m, which correspond well to the fluctuations of measured temperature data (peaks 1, 2 and 3 in Figure 3i). A good correlation of fluctuations between measured and calculated geothermal gradients indicate that fluctuations of measured temperature data are due to those of thermal conductivity, and thus difficulties arise if we estimate frictional heating from the positive thermal anomaly at 1111 m from temperature data. Kano *et al.* [2006] assumed a value of specific heat (c), $1700 \text{ J kg}^{-1}\text{K}^{-1}$ and a value of thermal diffusivity (α), $0.34 \times 10^{-6} \text{ m}^2\text{s}^{-1}$, for the calculation to fit the measured data to theoretical diffusion of temperature. However, the value of specific heat is too large compared to measured value for slip zones (300 to $500 \text{ J kg}^{-1}\text{K}^{-1}$, Figure 2g) and the value of thermal diffusivity is too small compared to measured value ($1.5 \times 10^{-6} \text{ m}^2\text{s}^{-1}$, Figure 2f) of the zones surrounding the slip zones. From these examinations, we conclude that the positive thermal anomaly presented by Kano *et al.* [2006] might be a result of thermal conductivity fluctuations, rather than a residual heat from frictional faulting.

[13] We examine next the effects of fluctuations of thermal conductivity to the positive thermal anomaly measured at the time of shallow drilling in 2000 [Tanaka *et al.*, 2006]. A thick fracture zone ranges from 285 m to 330 m depths. Numbers of thermal property data are limited for shallow core because we could not have intensive measurement of thermal properties across the fracture zone. The measurement method is basically the same as that for deep core. Thermal conductivity data are plotted with depth in Figure 4b with distribution of fault zone rocks Figure 4a. Host rock shows the values of thermal conductivity around $2.0 \text{ Wm}^{-1}\text{K}^{-1}$ and those of fault zone rocks around $1.3 \text{ Wm}^{-1}\text{K}^{-1}$. For careful evaluation, we examined two models of thermal conductivity profiles based on the measured data. For the first model, the thermal conductivity is assumed to be $2.0 \text{ Wm}^{-1}\text{K}^{-1}$ in host rock region and $1.3 \text{ Wm}^{-1}\text{K}^{-1}$ in fracture zone (gray line in Figure 4b). For the second model, gradual decrease of thermal conductivity to the several gouge zones in a fracture zone is assumed (red dotted line in Figure 4b), which is similar trend with the deep core (Figures 2e and 3h). The heat production from radioactive elements is ignored because of its little affection to geothermal gradient as described above. Calculated geothermal gradient is parallel with the measured temperature data if we assume apparent heat flow of 11 mWm^{-2} . The problem of low geothermal gradient is due to quick measurement of temperature after drilling and has already discussed by Tanaka *et al.* [2006]. For both models of

thermal conductivity, positive thermal anomaly is still recognizable on the calculated geothermal gradient at the fracture zone (Figure 4c). The possibility of thermal disturbance of fluid flow was already examined and excluded [Tanaka *et al.*, 2006]. We examined higher apparent heat flow (45 mWm^{-2}), which is same value assumed for the deep borehole. In this case, although fluctuations of thermal conductivity generate apparent temperature anomaly ($\sim 0.1^\circ\text{C}$), distribution and shape of apparent temperature anomaly is largely different from the measured anomaly, and the depth of temperature peak does not match with that of measured one. Thus, there are few possibilities that measured temperature anomaly was resulted from fluctuations of thermal conductivity under higher heat flow, and kept the distribution of apparent temperature anomaly even during fluid circulation by drilling. From these considerations, positive thermal anomaly observed in shallow borehole can be regarded as residuals of frictional heating.

6. Conclusion

[14] Heat production rate of 1 to $2 \mu\text{Wm}^{-3}$ for sedimentary rocks are estimated which are comparable with previous researches [Kitajima *et al.*, 2001; Rybach, 1976]. Thus, radiogenic heat production would very less contribute to fluctuation of geothermal gradient in such a short interval of examination for deep borehole (1050 to 1250 m) and also shallow borehole (200 to 400 m). However, for fracture zones at the depth of 1100 m, contrasts of thermal conductivity between host and fault zone rocks are higher, about $2.0 \text{ Wm}^{-1}\text{K}^{-1}$ (Figure 3h), than those from shallow core ($0.5 \text{ Wm}^{-1}\text{K}^{-1}$, Figure 4b), which generates larger fluctuations of geothermal gradient at the deeper fracture zones. As clearly observed in Figures 3i and 4c, fluctuations of temperature due to variations of thermal conductivity is around 0.1°C for deeper borehole and 0.03°C for shallow borehole. Residual heat from frictional faulting, if it exists, should be larger than the fluctuations of geothermal gradient. Thus, it would be necessary to measure the temperature before cooling down of the slip zones, within short periods after a large earthquake, for example, one to two years [Tanaka *et al.*, 2006], depending on maximum temperature achieved by frictional heating. However, it might be difficult to penetrate the deep fault zones within such a short period. In contrast, it is much easier to reach at the depth around 300 m in a short period after a large earthquake if a surface ruptured could be identified. From all of the discussions, shallow and quick penetration of slip zones after a large earthquake and measurements of thermal properties across the fracture zones would be worthy to detect the frictional heating and estimate the frictional energy of earthquake.

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