Seismicity in the northeast area of Izu Peninsula, Japan, comparing with three-dimensional velocity structure and temperature distribution of geothermal water

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Abstract

Seismicity in the northeast area of Izu Peninsula was compared with the three-dimensional P-wave velocity structure and the temperature distribution of geothermal water at 400 m below sea level. It is found that regional seismicity in the eastern margin of Hakone volcano is concentrated within a zone of both high velocity and low temperature (surrounded by a low-velocity zone) at depths of 10−15 km. Low seismic activity, however, seems to be associated with a low-velocity and high-temperature zone at depths of 5−10 km. A simple model for brittle−ductile transition was applied in order to examine the relationship between seismicity and the seismogenic layer in the region. Theoretical calculations suggest that the depth of brittle−ductile transition becomes gradually shallower with increasing surface temperature and/or heat flow. Accordingly, the seismogenic layer becomes gradually thinner from Sagami Bay to the land area of the northeast Izu Peninsula. This simple model can well explain the seismicity pattern in the study area. © 1999 Elsevier Science B.V. All rights reserved.

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1. Introduction

The geology and tectonics of the Izu Peninsula are characterized by a high degree of volcanic, geothermal and seismic activity. The Izu Peninsula itself is a huge volcanic landmass with eleven Quaternary volcanoes (Hoshino and Aoki, 1972), mainly along the volcanic front which runs through central Japan. This study area includes the Hakone, Yugawara and Taga volcanoes (Fig. 1). The Hakone is a triple volcano composed of two calderas, two sommas and seven post-caldera central cones. The eruption of the Hakone began about 400,000 years ago. Solfataric activity still persists at some places. The Yugawara volcano is a medium-sized stratovolcano. There is geothermal activity at the central part of the volcano, but the shape of stratovolcano, i.e. the central part, has been completely eroded. The Taga is a complex group of stratovolcanoes. The eastern half of the volcanic structure has been completely destroyed comparable with the Yugawara volcano (Oki and Hakamata, 1978). The Atami hot spring spa is associated with the Taga volcano.

About 40 geothermal fields are known in the Izu distinct, including the Hakone, Yugawara and...
Atami hot spa. The intense hydrothermal activity associated with Quaternary volcanoes is seen along the east coast of the Izu Peninsula. The total energy discharge from this area amounts to $11 \times 10^7$ cal/s, of which Hakone has 27%, discharging $3 \times 10^7$ cal/s (Oki and Hirano, 1974).

In addition, the Izu Peninsula and its vicinity have suffered from a series of remarkable earthquake phenomena: large $M_7$ class earthquakes and several earthquake swarms since the occurrence of the 1974 earthquake ($M_6.9$). In particular, earthquake swarms have intermittently been observed in the area off the east coast of the Izu Peninsula since 1978 (Tsumura, 1980). In July 1989, a submarine eruption occurred after an intensive earthquake swarm (Oshima et al., 1991).

The study area is located to the north of the swarms area and has repeatedly experienced large earthquakes with magnitudes of 7 to 8. This can be attributed to the fact that the Izu Peninsula is colliding with the Honshu Arc and that the Philippine Sea plate is subducting at the Sagami trough. According to historical data, earthquakes have occurred in a cycle of about 70 years (Ishibashi, 1988). Since the previous devastating earthquake, which was the Kanto earthquake of 1923, which caused the largest damage to Odawara and Hakone, it is pointed out that another disastrous earthquake can be expected in the near future. The Kanagawa prefectural government named this hypothetical $M_7$ class earthquake as the ‘Western Kanagawa Prefecture Earthquake’. The Hot Springs Research Institute of Kanagawa prefecture (HSRI) has carried out continuous monitoring of seismicity and crustal deformation (borehole tilt meters, GPS and EDM instruments, groundwater leveling, etc.) since 1989 to investigate the mechanism of the ‘Western Kanagawa Prefecture Earthquake’ (Ito et al., 1990).

In this paper, a detailed analysis of the relation between seismicity and geothermal structure in the region is made by (1) comparing the precise distribution of hypocenters with the 3D P-wave velocity structure and the result of hydrothermal analysis by Oki and Hirano (1974), and (2) estimating the depth to the base of the seismogenic layer using a simple brittle–ductile transition model.

2. Data

The hypocentral and first arrival time data used for the present analysis were from 16 stations. Fourteen of the 16 stations belong to the Hot Springs Research Institute of Kanagawa prefecture (HSRI), each of which is equipped with three-component velocity type seismometers adjusted to natural frequencies of 1 Hz and a damping coefficient 0.7. The other two stations belong to the National Research Institute for Earth Science and Disaster Prevention and the International Institute of Seismology and Earthquake Engineering. Seismic stations are deployed with a spacing of 10 km. This seismic network densely covers the study area (Fig. 1). All the data are telemetered to the HSRI in Odawara.

The accuracy of the determination of hypocenters and origin times of earthquakes in and around Kanagawa prefecture was investigated applying the method of prediction analysis (Tanada, 1998). This test supports that the error of hypocenter and focal depth is within less than 2–3 km in the study area.

The detectability of the HSRI seismic network is high and constant since 1989. The minimum magnitude of observable earthquakes is estimated to be 1.0 for the local magnitude scale (Tanada, 1998).

3. Recent seismicity

The northern Izu Peninsula is a tectonically unique area (Fig. 1). This area includes the Hakone and Yugawara volcanoes, and the Kozu–Matsuda fault system (KM on Fig. 1) which is considered as the convergent plate boundary. The tectonics in this area should be responsible for a complicated hypocenter distribution. Fig. 2a shows a seismicity map of this region from the HSRI catalog for the period of 1990–1997.

The Hakone volcano has a remarkable seismicity, where many earthquake swarms have occurred under the central cone (Hiraga, 1987). The focal depths of these swarms are much shallower than 6 km. Hiraga (1987) explained that the occurrence of earthquakes in Hakone was attributed to volcanic steam.

In 1994, an $M_4.8$ earthquake with focal depth of 6 km occurred at the southern part of the Hakone volcano where Kita–Izu fault systems cross the somma of Hakone volcano. The focal zone of this $M_4.8$
shock was clearly separated from the earthquake swarm area under the central cone mentioned above. Hypocenters with depths between 10 and 20 km along the eastern margin of the Hakone volcano are
aligned in a zone running north–south as shown in Fig. 2a. Seismicity seems to be dispersed and cannot be identified with a fault system (Research Group for Active Fault Japan, 1980). It is noted that there
Fig. 2. (a) Hypocentral distribution of earthquakes determined by the Hot Springs Research Institute of Kanagawa prefecture (HSRI) for the period of 1990–1997. (b) Hypocenter distribution from the Japan Meteorological Agency catalog for the period of 1960–1989.
exists a distinct seismic gap between this dispersed seismicity and the Hakone central swarm activities. A remarkable $M_{5.1}$ earthquake occurred in the central part of this dispersed seismicity area in 1990. The focal depths of main shock and aftershocks were around 16 km.

The most seismically active area in Fig. 2a is seen in the eastern part of Yamanashi prefecture, where the Philippine Sea plate has been colliding with Honshu Arc. The hypocenters are distributed in a range of 20–30 km, and are obviously deeper than the above described seismicity.

The earthquake catalog of the Japan Meteorological Agency (JMA) is used for a much longer-term seismicity study. Fig. 2b shows a seismicity map of the same area as indicated in Fig. 2a for the period 1960–1989. This map suggests that the long-term seismicity has the same features as in Fig. 2a although the accuracy of JMA hypocenters are less reliable than ours.

### 4. Three-dimensional velocity structure

Kantoh et al. (1996) determined a detailed 3D seismic velocity structure of the upper crust down to 15 km depth beneath the Hakone volcano and its surrounding area using local and explosion seismic data. They used 2082 P-wave arrival times from 192 local earthquakes and four explosions recorded by 16 seismic stations of HSRI. The inversion method covered a 30 (EW) $\times$ 35 (NS) $\times$ 15 (Depth) km region in this area. The model was parameterised by $6 \times 7 \times 5$ rectilinear grids, spacing 5 km horizontally and vertically (Fig. 3). Kantoh et al. used the method of Thurber (1983).

Analysis and interpretation of tomographic images allowed them to infer the following model: (1) the study area has remarkable lateral heterogeneity, and the velocity distribution of the surface layer corresponds well with geological features, and (2) volcanoes are located in the low-velocity zone.

Although several researchers have been studying 3D velocity structure including this region (e.g., Horie and Aki, 1982; Ishida, 1992; Lees and Ukawa, 1992), no previous study focused on a detailed 3D velocity structure in this area, because they modeled using large grid sizes (about a few tens of km).

### 5. The isothermal map of geothermal water

Fig. 4 shows an isothermal map of the Hakone and its adjacent areas at 400 m below sea-level and a north–south cross-section of the isothermal map (Oki and Hirano, 1974). This map is based on measurements of temperature in more than 300 deep wells. Their average depth is about 500 m, including several wells of about 1000 m deep.

It is noted that there are four high-temperature areas and these high-temperature centers line up from north to south located at the bottoms of the Hakone, Yugasawa and Taga (Atami) calderas. The isothermal contours in the calderas show concentric circles, whose diameter matches the size of volcano. For example, Hakone is an extremely large geothermal area of about 12 km in diameter, where a large amount of thermal water is discharged. The high temperatures are surrounded by a low-temperature area, where the temperature gradient is very small, sometimes less than 30 K km$^{-1}$ (Oki and Hirano, 1974). They pointed out that the combination between high- and low-temperature areas on the isothermal map seems to be related with the hydrothermal systems.

### 6. Comparison of earthquake distribution with 3D velocity structure and the isothermal map

Fig. 5 shows the comparison between the isothermal map at 400 m below sea-level and hypocenter distribution. As shown in Fig. 2a, the majority of the epicenters lie in the Hakone volcano. Seismicity except for the Hakone volcano shows dispersed patterns and is concentrated in depths greater than 10 km. Few epicenters exist within the high-temperature area (higher than 60$^\circ$C contour), in other words, most epicenters have been observed within the low-temperature area (lower than 60$^\circ$C contour).

Fig. 6 shows two sections of 3D velocity structure at depths of 10 (left figure) and 15 km (right figure), together with the distribution of earthquakes within 5–10 and 10–15 km depth, respectively. The isothermal contour lines from Fig. 4 are also shown. These figures suggest that few earthquakes occurred in the low-velocity zone of 5–10 km depth, which corresponds to the high-temperature area (higher than 60$^\circ$C) in the calderas of Quaternary volcanoes. Most
earthquakes have been observed in the high-velocity zone of 10–15 km depth limited within the low-temperature area (lower than 60°C).

7. Discussion

In general, the seismic–aseismic boundary zone correlates with the thermal structure in the crust and depends strongly on the brittle–ductile transition boundary of rock deformation (Kobayashi, 1977; Sibson, 1982; Doser and Kanamori, 1986; Ito, 1990). The relationship described in Section 5 shows the possibility that seismicity in the study area correlates with the thermal structure in the crust. In this section, we apply a simple model of the brittle–ductile transition of rock deformation (Sibson, 1982) to explain the regional or local variations of seismicity.

On the basis of Byerlee’s law (Byerlee, 1978) and the results of geological observations, Sibson (1974) calculated the shear resistance $\sigma$ in the brittle (seismogenic) layer as a function of depth for different fault types:

$$\sigma = \sigma_1 - \sigma_3 \geq b p (1 - l)$$

(1)

where $\sigma_1$ and $\sigma_3$ are the maximum and minimum compressive stresses, $b$ is a coefficient determined by the type of faulting, $p$ is lithostatic pressure,
and $l$ represents the pore-fluid factor. As seen from Byerlee’s law and Eq. 1, shear resistance in a brittle regime does not depend on temperature.

In the ductile (aseismic) layer, dominance of creep is assumed, so that the approximate constitutive law is:

$$
\dot{\varepsilon} = A\sigma^n \exp(-V/RT)
$$

where $\dot{\varepsilon}$ is strain rate, $R$ gas constant, $T$ absolute temperature, $V$ activation energy, with $A$ and $n$ being material constants. Ductile behaviour is sensitive to temperature, strain rate and rock mineralogy.

It is assumed in the same manner as Ito (1990) that temperature $T$ is a function of surface heat flow $Q$ and depth $z$:

$$
T = T_0 + \frac{Qz}{\kappa} + B\frac{z^2}{2\kappa}
$$

where $T_0$ is the average temperature at the surface, $\kappa$ conductivity, and $B$ radioactive heat production.

Fig. 7 shows an example of shear resistance for $\kappa = 2.7$ W/(mK), $B = 2.3$ mW/m$^2$, $\dot{\varepsilon} = 10^{-14}$, $A = 1.26 \times 10^{-9}$ MPa$^{-n}$/s, $n = 3.0$, $V = 123$ kJ/mol, $T_0 = 20^\circ$C and $Q = 60$ mW/m$^2$. In Fig. 7, shear resistance in a brittle regime is shown for three
different types of faulting. Most earthquakes are strike-slip or dip-slip type in this study area (Ukawa, 1991).

Generally, we assume a sharp transition from a purely brittle regime to a ductile regime in rheological modelling. The depth to this transition zone can be derived by equating shear resistances at brittle and ductile regimes. The base of the seismogenic layer corresponds closely to the top of this brittle-ductile transition zone. Within the uncertainty of parameters, this figure coincides with the result of Bodri and Iizuka (1993), who calculated thermal models and rheological profiles in central Japan and showed that the brittle-ductile transition occurred at a depth of around 10 km in the Izu Peninsula including this study area.

As for the geothermal effect, we calculated the shear resistance varying two parameters, that is, the surface temperature ($T_0$) and heat flow ($Q$) in Eq. 3. The other parameters are the same as adopted in Fig. 7. Fig. 8a shows the shear resistance versus depth relation varying the surface temperature from 20° to 80°C, but with a constant $Q$ of 60 mW/m². This variation is based on the value on the isothermal map. It is a matter of course that the isothermal map at 400 m below sea-level does not always indicate directly temperatures at depth in the crust. However, we adopt surface temperatures to be in the range
from 20° to 80°C, considering the geological setting that the high-temperature areas just coincide with the Hakone, Yugawara and Taga calderas and that reflects on the geothermal effects of volcanoes.

Fig. 6. P wave velocity structure at a level of 10 km depth (left) and at 15 km depth (right). The isothermal contour lines and the hypocenters within the zone of 5–10 km depth or 10–15 km depth are plotted on each figure, respectively.

Fig. 7. Shear resistance versus depth relation for a simple brittle–ductile transition model in the case of granite.

Fig. 8b shows the shear resistance versus depth relation with variation of the heat flow (Q) from 60 to 120 mW/m² in the case of a constant T₀ of 20°C. This variation of heat flow values was taken from the generalized heat flow map of Li et al. (1989) as illustrated in Fig. 1c. It is clearly shown that the heat flow values in the study area increase abruptly from 60 mW/m² in Sagami Bay to 120 mW/m² in the land area of Izu Peninsula.

It can be seen from Fig. 8a and b that the depth of the brittle–ductile transition boundary becomes gradually shallower with increasing surface temperature and/or heat flow. Heat flow variation seems to give a stronger effect rather than the variation of surface temperature in our model calculation. Accordingly the seismogenic layer becomes gradually thinner from Sagami Bay to the land area of the northeast Izu Peninsula due to increasing geothermal effects.

At present, the spatial distribution of temperature in the crust is not precisely known in this region. Therefore, 3D P-wave velocity structure is used instead of the temperature distribution. If a low-velocity layer presumably corresponds to a high-temper-
At a first glance of the present results, seismicity in the study area seems to correlate with the geothermal structure in the crust. Both the low seismicity within the high-temperature area of the isothermal map at ~400 m and the low-velocity zone in the range of 5–10 km depth might be attributed to geothermal effects, that is, the shallower depth of the brittle–ductile transition boundary, so that it is difficult to accumulate sufficient stress for an earthquake occurrence. In contrast, the depth of the brittle–ductile boundary is relatively deeper in the low-temperature area and/or high-velocity zone (surrounded by low-velocity zone) than the boundary depth in the high-temperature area and/or low-velocity zone. Consequently many earthquakes occur in such a region (within the thicker seismogenic layer).

An alternative interpretation must be possible for the seismicity pattern in the region. For example, such factors as strain rate, rock mineralogy and fluids should influence shear resistance. But we have no measured stress data in the study area, unlike in the case of Willis-Richards (1990). Then the evaluation for rock mineralogy is too complex to extrapolate laboratory results to in situ geological conditions.

Fluids and magma are also important factors that reduce strength in all rheological domains. Fluid pressure gradients greater than the hydrostatic gradient decrease strength in the brittle regime than in the ductile regime. In consequence, the depth of the brittle–ductile boundary increases apparently. Hill (1977) showed that magma and other fluids are a possible cause of earthquake swarms. The fact that a few volcanic earthquakes occur in the high-temperature zone of the Hakone caldera can be interpreted as due to the existence of volcanic steam and geothermal water circulation within the caldera (Hiraga, 1987).

On the other hand, the 1930 Kita–Izu earthquake (M 7.3) occurred at the west edge of a low-temperature area. Furthermore, the northeastern region of the Izu Peninsula has been damaged repeatedly by large earthquakes. Ishibashi (1988) suggested that the next M 7 class earthquake named the ‘Western Kanagawa Prefecture Earthquake’ might occur in the near future. Ito (1990) pointed out that large earthquakes occurred at the zone of abrupt change in depth of the seismogenic layer around volcanoes by investigating the regional variations of the seismic–aseismic boundary.
in Japan. Hasegawa et al. (1991) indicated using tomographic images that large crustal earthquakes occurred in the periphery of low-velocity zones beneath active volcanoes in northeast Japan. The present result, in which the depth change of the seismogenic layer should be attributed to geothermal effects provides an important constraint on the fault model of the ‘Western Kanagawa Prefecture Earthquake’.

8. Conclusions

The comparative study of seismicity to the three-dimensional velocity structure and the isothermal map of geothermal water in the northeast area of Izu Peninsula has revealed the following results:

(1) In the eastern margin of the Hakone volcano, seismicity concentrates within the high-velocity zone of 10–15 km depth, corresponding to the low-temperature zone in this area.

(2) Low seismic activity is associated with the low-velocity zone as well as the high-temperature zone of 5–10 km depth.

(3) Theoretical consideration of geothermal effects suggests that the depth of the brittle–ductile transition boundary in the low-temperature and/or high-temperature zones should be relatively deeper than the depth of the boundary in the high-temperature and/or low-velocity zones, where the seismogenic layer is thick enough to accumulate the necessary stress to an earthquake occurrence.

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