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Regional Variations of Cutoff-Depth of Seismicity in the Crust and Their
Relation to Heat Flow and Large Inland Earthquakes

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More than 8000 earthquakes are relocated to derive the regional variations of seismic-aseismic boundary in the mid-crust in the northern Kinki district, Japan. The depths to the boundary are estimated as 13-15 and 18-20km in southwestern and northeastern parts of the study area, respectively. The relationship between the cutoff depth of seismicity and the cause of large intra-plate earthquakes is studied from the present observations and other available data of seismicity and surface heat flow, on the basis of the brittle-ductile transition of rock deformation. The regional variations of the cutoff depth of seismicity seem to be well correlated with thermal structure of the crust. The cutoff depths of seismicity in various heat flow provinces in Japan and other countries are found to be inversely proportional to the surface heat flow values, the depths roughly corresponding to isotherms of 200-400 °C.

The shape of high-quality depth-frequency distribution of earthquakes is quite similar to that of the shear resistance calculated from a simple brittle-ductile transition model. Large intra-plate earthquakes appear to originate at the peak or just below the peak of the depth-frequency distribution, which corresponds also to the deepest portion of the seismogenic layer. Furthermore, at the source area of the large earthquakes, rupture seems to start at the steeply changing portions of the cutoff depth of seismicity. Thus, the seismic-aseismic boundary is closely related to large intra-plate earthquakes and eventually to tectonics of island arcs.

1. Introduction

Seismic activity in the crust is mostly confined to the upper crust and only a few earthquakes occur in the lower crust. This fact was noticed in late 1960's in California (i.e. Eaton *et al.*, 1970). In Japan the restriction of

seismicity to the upper crust was first reported in eastern Chugoku and northern Kinki districts (Hashizume, 1970a, b), then it has been recognized in many regions by the construction and improvement of microearthquake observation networks under the Earthquake Prediction Project (Oike, 1975; Takagi and Hasegawa, 1977; Takagi and Matsuzawa, 1988). Although many papers have been published so far for seismic activity with reference to the distribution of active faults, extensive studies for the depth distribution have been quite a few in Japan.

Kobayashi (1977) first interpreted the cutoff depth distribution of seismicity as brittle-ductile boundary of rock deformation, indicating that surface heat flow values are related to the depth of seismicity in southwestern Japan. The idea is mainly based on the results of laboratory experiments of rock deformation. He also suggested a possibility that the crust is not made of two layers, 'granitic' and 'basaltic' layers, but of a rock such as granodiorite. However, the seismicity data he used are not so accurate that further studies have not been developed for a long time.

On the other hand, the seismic-aseismic boundary was clearly noticed along the San Andreas fault (Eaton *et al.*, 1970) and the causes of the boundary have been studied in relation to the results of laboratory experiments for rock deformation (Byerlee, 1968; Brace and Byerlee, 1970). On the basis of the accumulated data of seismicity and rock experiments, a conceptional rheologic modeling of earthquake occurrence in the crust has been presented by Brace and Kohlstedt (1980) and Sibson (1982). The model has been applied to the depth distribution of seismicity at several regions in California, Europe and other countries (Meissner and Strehlau, 1982; Chen and Molnar, 1983; Doser and Kanamori, 1986; Mikumo *et al.*, 1989).

Sibson (1984) reported that the cutoff depth of seismicity varies along the San Andreas fault and that large shocks are likely to occur at the bottom

of the seismogenic layer. Moreover, tectonic features of the continents and island arcs have been discussed on the basis of the strength of the lithosphere derived from the thickness of a brittle layer (Vink *et al.*, 1984, Shimamoto, 1989). Therefore, a precise determination of hypocenters is considered very useful to reveal their relation to the occurrence of large intra-plate earthquakes and consequently to the tectonics.

In this paper, we relocate hypocenters of a large number of earthquakes in order to examine regional variations of a seismic-aseismic boundary in the northern Kinki district where a dense network has been monitoring seismic activity. Then its relation to the thermal structure and large intra-plate earthquakes is also examined with reference to the relation between the depth to the seismic-aseismic boundary and surface heat flow data for well determined hypocenters in Japan and other countries. From these data, the occurrence of earthquakes in the crust is discussed on the basis of the rheologic model of rock deformation. We present other evidence which supports the model such as variations of the seismic-aseismic boundary near active volcanoes where thermal structures change abruptly. Further, we show the depth to the seismic-aseismic boundary is closely related to the occurrences of large earthquakes or the tectonics of the inner zone of the Japanese islands.

2. Earthquake Data and Hypocenter Determination in Northern Kinki District

Seismic activity in the northern Kinki district has been monitored since 1963 by the dense network of the Abuyama Seismological Observatory, Kyoto University (Okano and Hirano, 1968). The network was improved and equipped with a telemetered system by using telephone lines and radio links in 1975 (Kuroiso and Watanabe, 1977), employing an automatic data processing for phase

detection and determination of hypocenters and magnitudes (Watanabe and Kuroiso, 1977). Since then, both automatic and manually processed data have been stored for more than ten years. The network stations of which data are used in this study are shown in Fig. 1.

Although the automatically processed data are accurate enough to real-time monitoring of seismic activity, high-quality data are necessary for precise determination of focal depths. More than 21,000 earthquakes were located manually and their phase data were stored during 1976 and 1987. In the present study we relocated all the earthquakes that meet the conditions described later by the use of the manually picked P and S wave data. The uncertainty in the readings used for this study is within 0.02s(A) and 0.05s(B) for P wave arrivals, which are used for relocating hypocenters. S wave arrivals are used only to estimate a starting value of origin time.

Hypocenter and origin time are relocated from P wave arrival times to minimize the standard error of P wave residuals between observed and calculated travel times for a horizontally layered velocity structure. The crustal structure used consists of three layers in which P wave velocities are 5.5, 6.0 and 6.7km/s with thicknesses of 5.0, 20.0 and 10.0km, respectively. The model is based on the results of refraction surveys (Yoshii *et al.*, 1974; Ito and Murakami, 1979). Since the study area is 100x100km wide and the focal depths of earthquakes range between 5-20km, most of the rays from hypocenters to stations estimated as direct waves. Therefore, the upper part of the crustal structure is mainly used for the calculation of travel times. The corrections for station heights are made in calculating the travel times by the use of a P wave velocity of 4.5km/s for surface layer.

The uncertainty in the determination of hypocenters for the data from the

present network has been estimated by Ito and Kuroiso (1988). The station corrections and the thickness of the surface layer have considerable effects in estimating the focal depths. The absolute focal depths have an error of about 2km which is caused by the difference of 2km in the thickness of the surface layer. Ito and Kuroiso(1988), however, indicated that the uncertainty of the relative focal depths determined from the data of this network is one tenth of it regardless of the velocity structure, comparing the hypocenters to those determined by using master event location technique. Simultaneous determination of hypocenters and three-dimensional velocity structure is also carried out using inversion method to confirm the regional variations of focal depths. Eventually the hypocenters are accurate enough to derive regional variations of focal depths over the whole study area.

We selected the earthquakes that meet the following criteria; P wave arrivals for an earthquake are recorded at more than seven stations and the standard error of P-wave residuals is within 0.1s. It is important for precise focal depth determination that the epicentral distance of the nearest station is comparable to focal depth and that the range of epicentral distance is sufficiently large. These criteria are satisfied for most of the relocated shocks with the using of P arrivals of more than seven stations, because the relocated focal depths range between 5 and 20 km and the mean spacing of the stations is about 15km for the network. Thus we obtain 8431 relocated high-quality hypocenters with the magnitude range of 1.8 to 5.3 on the magnitude scale of Japan Meteorological Agency (JMA) (Fig.2). Actually we used Watanabe's formula (1971), which is based on JMA scale, for the determination of magnitude of small earthquakes.

3. Distribution of the Depth of Seismic-Aseismic Boundary in the Northern Kinki

District

Fig.2 shows all relocated high-quality hypocenters in the northern Kinki district. This figure shows the background seismicity in this district for more than ten years. Although the seismic activity varies from place to place and is concentrated near some active faults, we see a great number of microearthquakes taking place all over the area, except in a region southwest of Lake Biwa. Therefore, we can estimate the distribution of focal depths over the whole district.

Figs.3(a) and (b) show the focal depths of earthquakes in several profiles with a width of 20km projected on vertical cross sections along the northeast-southwest and northwest-southeast directions, respectively. In all the profiles, earthquakes are restricted in the upper 13-20km of the crust and a seismic-aseismic boundary is clearly defined with only a few shocks in the lower crust. Moreover regional variations of the boundary are clearly seen in Fig.3. In particular the boundary is deepest in the northeastern part of the study area as compared with that in the other areas.

The epicenters are plotted at every 1km depth interval from 13 to 20km in Fig.4 to draw depth contours of the seismic-aseismic boundary. The contours indicating a depth boundary of each focal depth are drawn as shown in Fig.4, neglecting a few earthquakes. The contours shown here are not uniquely determined but the general tendency of depth variations would not be changed by different choices of contours. The resultant contour map is shown in Fig.5.

Roughness or irregularity of the base of the seismogenic layer was reported by Sibson(1984) for earthquakes along the San Andreas fault, where the wavelength of undulations is of the order of 10-100km along the fault with a change of its depth of 5km. The regional variations of the seismic-aseismic boundary shown in Fig.5 have wavelength of 10-50km with the maximum

change of 6km in depth. This is nearly the same as those along the San Andreas fault, but the obtained variations are in three-dimensional, different from that in two-dimensional along the San Andreas fault.

4. Relationship Between the Seismic-Aseismic Boundary and Heat Flow

Although we have only limited data of the thermal gradient and surface heat flow in the northern Kinki district. Fig. 6 shows surface heat flow values around this district (Nagao, personal communication). The low heat flow values in Lake Biwa area (Fujisawa *et al.*, 1985) are consistent with deep cutoff depths of seismic activity there.

Fig. 7 shows the Curie point depth map derived from aeromagnetic survey (Okubo, 1984; Okubo *et al.*, 1985, 1989). Since the Curie point depth distribution in the Japanese islands is correlated with surface heat flow, the distribution is thought to indicate the thermal structure of the crust (Okubo *et al.*, 1989). It should be noticed that the absolute depths of the Curie points in Fig. 7 are shallower than the seismic-aseismic boundary shown in Fig. 5 by about 8-9km in the same area, though the variations of the Curie point depths appear to roughly agree with the depth variations of the focal depths. Thus the general tendency of the regional variations of the cutoff depth of seismicity seems to be governed by the thermal structure of the crust.

In order to reveal the relation between the depth of seismic-aseismic boundary and surface heat flow, all published data of well defined cutoff depths and surface heat flow are plotted in Fig. 8, taking from Sibson (1982) and Meissner and Strehlau (1982) together with some available data in Japan. We used the data sets that have both depth-frequency distribution of high-quality hypocenters and heat flow values in a geothermal province. The upper

side of rectangles in Fig.8 denotes the depth above which 90% of earthquakes occur and the lower side the cutoff depth of seismicity. We have a small number of complete sets of well determined focal depths and surface heat flow values in the inner zone of the Japanese islands except for geothermal areas where many kinds of surveys have been conducted for the development of geothermal energy. The data in Kuju geothermal area (NEDO,1983) is plotted as a representative of the data from hot regions (a in Fig.8). The data of cutoff depth of seismicity in western Nagano Prefecture (b in Fig.8) is obtained from the joint observation of seismic activity by several universities (The Group for the Seismological Research in western Nagano Prefecture, 1989). The data in Tottori region (c in Fig.8) is after Tsukuda(1978) and the data in Lake Biwa (d in Fig.8) is taken from the present study. As shown in Fig.8 the depths to the seismic-aseismic boundaries are nearly inversely proportional to the surface heat flow. This indicates that the temperature in the crust is a primary factor that determines the cutoff depth of seismicity.

If the thermal gradient is a constant, that is the simplest case of thermal structure, temperature T at depth z is given by

$$T = T_0 + Qz/\kappa,$$

where T_0 is the temperature at surface and about 15 °C in Kinki district, Q is the surface heat flow and κ is the thermal conductivity. The equation is transformed as

$$Qz = \kappa (T - T_0),$$

we get rough values of Qz as $Qz=500-1500$ W/m from the data in Fig.8. If we assume $\kappa=2.7$ W/m K (Smithon and Decker,1974), $T-T_0=180-550$ °C. For the average value of $Qz=1000$ W/m, $T-T_0=370$ °C. The isotherms of 300 and 350 °C are also drawn in the Fig.8 calculated by the same method as that of Chapman(1986). In

this case more realistic relation is assumed for temperature increase with depth (Sibson, 1982).

The temperature over which brittle-ductile transition occurs is about 300 °C (Kobayashi, 1977). In geothermal areas the cutoff depth of seismicity is as shallow as 2-4km which corresponds to temperatures of 200-400 °C (NEDO, 1983). These values suggest that 200-400 °C is a rough estimation of the temperature at the seismic-aseismic boundary in the middle crust.

5. A Rheologic Model of the Seismic-Aseismic Boundary

There are several explanations for the possible cause of the seismic-aseismic boundary. The explanations include a change from stick-slip to stable sliding by increasing temperature with depth (Brace and Byerlee, 1970) or by the 'granitic' and 'gabbroic' crustal layers (Scholz, *et al.*, 1969), or the suppression of brittle behavior of rocks by increasing confining stress with depth (Byerlee, 1968), or brittle-ductile transition by increasing temperature with depth (Kobayashi, 1977; Brace and Kohlstedt, 1980; Sibson, 1982). Recently other possibilities for the base of the seismogenic layer have been pointed out: one model is a frictional sliding constitutive law indicating transition from velocity-weakening to velocity-strengthening in the brittle regime (e.g. Tse and Rice, 1986), and another model is unstable ductile instability in the ductile regime (Hobbs *et al.*, 1986). These are both temperature-dependent models based on rather complicated constitutive law.

Since the seismic-aseismic boundary is closely related to the thermal structure as described in the previous section, the temperature dependent models are plausible. In this section a model based on simple brittle-ductile transition of rock deformation (Sibson, 1982) is applied to explain the cutoff depth of seismicity. Other temperature dependent models are too complicated

and with too many experimentally-undetermined parameters to estimate the strength of the crust.

In the seismogenic layer frictional shear failure of pre-existing fractures is governed by Byerlee's law(1978).

$$\tau = \mu \sigma_n',$$

where τ and σ_n' are respectively shear and normal stresses acting on the fault surface, and the coefficient of static friction $\mu=0.75$. The effective normal stress is assumed to be

$$\sigma_n' = \sigma_n - P = \sigma_n - \lambda_v \sigma_v.$$

P is the fluid pressure, λ_v is the pore fluid factor and σ_v is the vertical stress taken as equal to the lithospheric load. Under hydrostatic fluid pressure $\lambda_v=0.36$ for an average crustal density of 2.75g/cm^3 (Sibson, 1982).

On the other hand, in the ductile regime the dominance of creep is assumed so that the approximate constitutive law is

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

where $\dot{\epsilon}$ is the strain rate, R is the gas constant, T is absolute temperature, V is the activation energy and A and n are constants of material. We assume the formula for temperature T as a function of the surface heat flow Q,

$$T = T_0 + Qz/\kappa + Bz^2/2\kappa,$$

where T_0 is the average temperature at the surface, κ is the conductivity and B is the radioactive heat production. When we get T as a function of depth z using above equation, we get shear resistance under the assumption of other parameters.

We calculated shear resistance as a function of depth for many combinations of the parameters. Fig.9 shows an example of the shear resistance calculated for $\kappa=2.7\text{ W/m K}$ and $B=2.3\text{ nW/m}^2$ for temperature distribution and $\dot{\epsilon}=10^{-14}$, $A=1.26\times 10^{-9}\text{ MPa}^{-n}/\text{s}$, $n=3.0$ and $V=123\text{ kJ/mol}$ in

ductile regime. The average temperature at the surface is taken as $T_s=15^{\circ}\text{C}$ and heat flow, $Q=60\text{mW/m}^2$. The pore pressure and temperature have considerable effects on the depth of the brittle-ductile transition (Sibson, 1984; Weissner and Strehlau, 1982). However, the parameters involved in the power law also affects the brittle-ductile transition depth.

In Fig. 9 the shear resistance in brittle regime is shown for three different types of fault movement (Sibson, 1974). In northern Kinki district most of the earthquakes are of strike-slip or dip-slip type (Ito and Watanabe, 1977). Consequently, we can estimate an appropriate depth of the brittle-ductile transition that coincides with the observed cutoff depth of seismicity in the northern Kinki district within the uncertainty of the parameters.

Fig. 10 shows a depth-frequency distribution of earthquakes in the southern part of the study area by the use of master event location method. It is outstanding that the frequency of earthquakes steeply decreases in the depth range of 12-15km. When focal depths are not very accurate, resultant depth-frequency distribution is a normal distribution with a mean at the peak of the distribution and in this case a asymmetrical sharp decrease of the frequency in the deeper portion cannot be not seen. The abrupt decreases of the seismicity below the peak of depth-frequency distribution has been detected in many regions for high-quality focal depths. Some examples are seen in Weissner and Strehlau (1982) and Sibson(1984).

The general shape of the distribution in Fig. 10 is similar to that of the shear resistance shown in Fig. 9. In particular, the steep decrease of earthquakes with increasing focal depths is quite similar to the decrease of the shear resistance in the ductile regime. This similarity suggests that the

shear resistance in the crust has strong effects on the frequency distribution of focal depths of earthquakes in the crust (Meissner and Strehlau, 1982).

If the peak of the frequency distribution of focal depths corresponds to that of the shear resistance, the brittle-ductile boundary would not coincide exactly with the seismic-aseismic boundary, and earthquakes could occur even below the peak of the shear resistance or in the ductile regime. The seismic-aseismic boundary in that case may occur at a certain level of the shear resistance in the ductile regime. When a rock contains more than two kinds of minerals with different strengths such as quartz and feldspar, the brittle-ductile boundary may have a certain depth range of the transition zone (Strehlau, 1986). The zone with steep decrease of the shear resistance might correspond to the transition zone.

The relation between the seismic-aseismic boundary and the shear resistance is not completely understood in terms of the mechanical property of rocks. However, the transition zone between the brittle and ductile layers, if any, may not be so thick because the frequency distribution of the focal depths is quite similar to the shear resistance with increasing depth. Therefore, the simple brittle-ductile transition model seems to well explain the depth distribution of seismicity in the crust.

P wave velocity in the lower crust is much faster than that in the upper crust in the northern Kinki district (Yoshii *et al.*, 1974; Ito and Murakami, 1979). The lower crust may therefore be made of different rocks from that of the upper crust. Hence the shear resistance with depth is calculated, taking plagioclase as a representative constituent of rock for the lower crust, where the parameters in the power creep law are $A=2.0 \times 10^{-4}$ MPa/s, $n=3.2$ and $V=238$ kJ/mol (Shimamoto, 1989). The shear resistance thus calculated in the lower crust is shown by a dashed line in Fig. 9. In this case the lower

crust is still aseismic.

6. The Seismic-Aseismic Boundary and the Velocity Discontinuity

The seismic-aseismic boundary has long been believed as a velocity boundary between the upper and lower crusts, or the Conrad discontinuity (Takagi and Hasegawa, 1977; Hashizume and Matsui, 1979). Although the seismic-aseismic boundary happens to coincide with the Conrad discontinuity in some regions, it is not generally true. In particular, in geothermal areas and near volcanoes, the seismic-aseismic boundaries are as shallow as 2-8km (e.g. NEDO, 1983) which is too shallow for the Conrad discontinuity. Moreover, no velocity discontinuity seems to correspond to the cutoff depth of seismicity there.

The Conrad discontinuity has not been clearly determined in southwestern Japan from seismic refraction experiments, because the first arrivals through the lower crust are observed only in a short epicentral range and because reflected waves from the Conrad discontinuity have been poorly recorded.

The available data of travel times in the northern Kinki district indicate that the P-wave velocities in the upper and lower crust are about 6km/s and 6.4-6.8km/s, respectively. The depth to the velocity boundary is as deep as about 21-24km in the northeastern part of the study area (Yoshii et al., 1974; Hasegawa, 1983), on the other hand, it is rather shallow about 18-20km in its southwestern part. (Ito and Murakami, 1979). The difference in the depth to the velocity boundary correlates with regional variations of the cutoff depth of seismicity. However, the absolute depth to the velocity discontinuity is deeper than the seismic-aseismic boundary by about 3-5km, which seems to exceed the uncertainty in focal depths and velocity discontinuity. This suggests that there is a transition layer between the

cutoff depth of seismicity and the lower crust with a high P-wave velocity, although the detailed characteristics of the layer are not clearly obtained.

Okano and Kuroiso(1986) presented a P wave crustal structure with a velocity increase in the middle and lower crusts from the analysis of travel times from earthquakes in the northern Kinki district. In the model P wave velocity increases from 6.0 to 6.4 km/s in the depth range of 12-20km and also increases gradually from 6.4 to 6.7km/s in the lower crust. This intermediate layer may correspond to that between the cutoff depth of seismicity and the velocity boundary derived from refraction seismology.

Klemperer(1987) indicated that the depth to the top of the reflective lower crust is related to surface heat flow values in England, the United States and other continents. The top of the reflective lower crust is shallow at high heat flow provinces. In northern Kinki district there seems to be relationship among the cutoff depth of seismicity, heat flow and velocity discontinuity. In addition, Poisson's ratio derived from velocity ratio, V_p/V_s , is larger in the lower crust than those in the upper crust in the Chugoku, northern Kinki and Chubu districts (Hashizume, 1970a; Ukawa and Fukao, 1981, 1982).

The above evidence leads to a hypothesis that reflective boundaries may be growing by ductile flow in the lower crust in tectonically active regions such as in the Japanese islands (Pavlenkova, 1984; Smith and Bruhn, 1984; Kuszniir and Park, 1986; Meissner and Wever, 1986; Mooney and Brochier, 1987). This flow in the lower crust may cause stress concentration near the base of the seismogenic layer.

7. Regional Variations of the Seismic-Aseismic Boundary and Large Intra-Plate

Earthquakes

Fig.11 shows the distribution of historical large earthquakes ($M \geq 6$) (Usami, 1987) superposed on the contour map of the seismic-aseismic boundary shown in Fig.5. The large events with magnitudes greater than 7 appear to have occurred either at the deepest portion of the seismogenic layer or at the steeply changing portions of the seismic-aseismic boundary. The epicenters of the historical earthquakes are not so accurate as compared with those of instrumentally determined events in recent years. However the locations of historical earthquakes in the northern Kinki district are considerably accurate in view of many historical documents on earthquake damage. Even if the epicenters shift by about 10km, their locations relative to contours do not change greatly as compared with the depth variations of the seismic-aseismic boundary.

In the central part of the Tohoku district the seismic-aseismic boundary beneath the source region of the 1896 Rikuu earthquake ($M=7.5$), which is one of the largest inland earthquakes there, is about 5km deeper than those in the surrounding area (Faculty of Science, Tohoku University, 1987). In addition, the depth to the boundary is well correlated with the change of Curie point depths (Hasegawa, 1990). This seems to be a piece of evidence that the Rikuu earthquake occurred in a deep portion of the seismogenic layer in the Tohoku district. Thus, the focal depths of large intra-plate earthquakes are related to the thickness of the seismogenic layer. Honkura *et al.* (1988) reported that earthquake magnitudes of large earthquakes in the Japanese islands are roughly correlated with the Curie point depth beneath the source region, and tried to make a zoning of the possible largest earthquakes based on the Curie point depths.

Sibson (1982) reported that most of the main shocks of moderate to large

earthquakes initiate near the base of aftershock areas which nearly coincides with the seismic-aseismic boundary of background seismicity. Hamada(1987) relocated aftershocks of large inland earthquakes in Japan and indicated that most of the main shocks, especially those with dip-slip fault mechanisms, started near the base of aftershock areas.

Fig.12(b) shows the relation between focal depths and the cutoff depths of seismicity for the hypocenters of small to moderate-sized earthquakes ($M \geq 3.8$) occurred in the northern Kinki district and accompanied with aftershocks. Epicenters are also shown in Fig.12(a). These indicate that most of the earthquakes occurred near the base of the seismogenic layer, where the shear resistance is inferred to be large.

Recently focal depths of large earthquakes have become so accurate that we can obtain not only the relation of the main shock to its aftershock area but also the regional variations of the base of aftershock area of large earthquakes. Fig.13 shows, for example, the hypocenter distribution of the main shock and aftershocks of the 1984 Western Nagano Prefecture earthquake ($M 6.9$) (Oida *et al.*, 1990). These aftershocks define a clear depth boundary of the seismogenic layer dipping from east-northeast to west-southwest direction, with the maximum difference in the depth of the boundary about 6km. It has been revealed that the largest aftershock of $M6.2$ as well as the main shock occurred near the bottom of the aftershock area.

Other well-located focal depths of aftershocks of large earthquakes also show some regional variations of their cutoff depth, for example, the 1978 Oshima-Kinkai earthquake of $M6.9$ (Association for the Development of Earthquake Prediction Research, 1983) and the 1983 Nihonkai-Chubu earthquake of $M7.7$ (Kosuga *et al.*, 1987). Both of the main shocks started at steep

portions of the base of aftershock areas, which seem to correspond to steep boundary of the seismic-aseismic transition. A model to explain these cases is shown schematically in Fig.14. In addition, it has been recognized that the aftershock area of a large earthquake consists of several segments corresponding to different rupture modes of the main shock. This suggests that the change of the mode during the rupture process of the main shock may be caused by the regional variations in the depth of the seismogenic layer.

Guodong and Zhaoxun(1990) and Loo *et al.* (1990) reported that the 1975 Heicheng and the 1976 Tangshan earthquakes occurred in the areas where the heat flow and the crustal structure change abruptly from surrounding areas. Earthquake generating stress is supplied by plate motions on global scale, but an additional factor that causes the regional or local stress concentration is necessary for the recurrence of large earthquakes in the same region in inland areas located far from plate boundaries. They considered that upheaval of the uppermost mantle causes some additional stress near a steep boundary of the crustal structure. Thermal stress may also play an additional role in order to trigger earthquakes. This is an important suggestion for better understanding of the cause of intra-plate earthquakes.

8. Large Earthquakes near Active Volcanoes

We have another evidence that seems to indicate a clear relationship between the depth to the seismogenic layer and earthquake magnitudes. Fig.15 shows the cross section of the depth distribution of earthquakes near Izu-Oshima volcano (Yamaoka *et al.*,1988). The cutoff depth of seismicity just below Izu-Oshima volcano is as shallow as 5km, while it is 8-10km deep below the flanks of the volcano.

The accuracy of hypocenter determination of earthquakes near some active

volcanoes in Japan have been greatly improved in recent years by the wide-area seismic networks of volcanological observations for the prediction of volcanic eruptions. The depth sections of seismicity similar to that of Izu-Oshima volcano have been obtained by the observation networks for several active volcanoes, for example Sakurajima (Nishi, 1978), Aso (Sudo *et al.*, 1984), Unzen (Matsuo, 1985), and Kirishima (Iida *et al.*, 1986). Smith *et al.* (1977) reported similar features of focal depth distribution and the thermal structure in the Yellowstone hot spot. Shallow cutoff depths of seismicity can also be seen in many of the geothermal areas (e.g. NEDO, 1983).

In Fig. 16 earthquake magnitudes are plotted as a function of distance between the epicenters and the crater of Izu-Oshima volcano for the earthquakes that occurred during 1975-1987 taken from the Seismological Bulletin of JMA. This shows that large earthquakes with magnitude equal to or greater than 6 occur more than 10 km away from the center of the volcano. Kubotera (1988) listed large earthquakes with magnitude greater than 6 that occurred near volcanoes in Japan. Similar results are derived for most of active volcanoes in the list. Consequently, large and moderate earthquakes near active volcanoes in Japan occur more than about 10 km away from the center of the volcanoes as shown in Fig. 17.

The largest earthquake accompanied with the 1986 fissure eruption of Izu-Oshima volcano has a magnitude of 5.1. Okada (1983) indicates that the magnitudes of earthquakes directly related to volcanic eruptions in the world are about 5 and that no larger shocks occur at the same time of eruptions. It may be inferred from these observations that the seismogenic layer is so thin immediately close to craters that the accumulated stress is too small to cause a large earthquake, whereas at the foot or outside the volcano the seismogenic layer is thick enough to accumulate high stress

leading a large earthquake.

Mizoue *et al.* (1986) classified the inner zone of central Japan into three zones by the type of earthquake occurrence. They are aseismic zones in and near the volcanic front, earthquake swarm zones in the neighboring inland area and large earthquake zones in inland area. They interpreted these by lateral thermal variations from high-temperature zone in the volcanic front to low-temperature inland area, referring to the depth distribution of Curie points (Okubo, 1984). This fact can be explained by the change in the thickness of a brittle layer due to temperature variations in the same way as that for the zones near volcanoes. As described above only small earthquakes occur in the volcanic front zone where the seismogenic layer is thin. Earthquake swarms likely to occur in the neighboring inland region, where the brittle layer is not thick enough for large earthquakes. Thermal stress may also contribute to the swarms. Large earthquakes occur only in the inland zone where the seismogenic layer is thick.

Since volcanic region is considered as a weak area in the crust, active faults may be nucleated at the volcano during tectonic deformation process (Nakamura, 1987). For further details, however, large earthquakes are known to occur more than 10km away from the center of volcanoes as described above, and hence active faults should have originated below the flank of volcanoes. Thus the variations of the seismogenic layer and thermal structure have close relations to the magnitudes and nucleation of large inland earthquakes.

9. Discussion

The Curie-point depth distribution shows appreciable regional variations of thermal structure over the wavelength of 10-100km (Okubo *et al.*, 1985). The existence of many hot springs also indicates that the thermal structure varies

not only in volcanic and geothermal areas but also all over the Japanese islands. It may therefore be concluded that the regional variations of the seismic-aseismic boundary are possibly attributed to those of the thermal structure in the Japanese islands.

Intra-plate earthquakes are generally thought to be caused by the movement of active faults. However, Matsuda(1975) indicated that only 50% of inland earthquakes with magnitude equal to or greater than 6 correspond to active faults. Therefore we can consider that active faults are results of repeated large earthquakes at the same area. If the large earthquakes tend to occur along the steeply changing portions of seismic-aseismic boundary, active faults should be formed at the steep brittle-ductile boundary. Since the stress field at the boundary will remain almost same conditions for a sufficiently long time, we can understand that the repeated earthquakes have nearly the same type of fault mechanism in the same area. The boundary between geological structures or old fractures with the same direction as that of steep seismic-aseismic boundary also possibly plays an important role for the formation of the faults in the area.

Sibson(1984) suggested that large earthquakes are likely to nucleate at the base of the deepest portion of seismogenic layer. However as described in the previous section, many large earthquakes started at the base of the steeply changing portion of the seismogenic layer. This reason is not necessarily obvious. Generally it may be easier for the rupture to nucleate at the rough boundary of the base of the seismogenic layer with different mechanical property. The undulation of the seismic-aseismic boundary may also be a possible source of complex rupture process of large earthquakes due to asperities or barriers. It is plausible that an earthquake become large when

the deepest portion of the seismogenic layer is ruptured. We need more high quality data for the variations in the base of aftershock area to make clear the relationship among undulations of the base of the seismogenic layer, nucleation of an earthquake and rupture process of the main shock.

Matsuda(1976) reported that the dominant type of active faults in the Tohoku district is dip-slip type, which differs from strike-slip type in southwestern Japan. This has been explained by the combination of fault strikes and the direction of the tectonic stress. In the Tohoku district the tectonic force resulting from plate subduction is trending east-west and the strikes of faults are oriented north-south, being perpendicular to each other, therefore dip-slip movement of the faults is predominant. In southwestern Japan, on the other hand, as fault strikes are oblique to the tectonic stress, strike-slip movement is predominant for the faults. This is also understood by replacing the strikes of active faults by the direction of weak zone of the seismogenic layer, or by the direction of steep change of the seismic-aseismic boundary. The condition whether the weak zone is perpendicular or oblique to the tectonic stress specifies the predominant fault type.

Although the regional variations of the cutoff depth of the seismicity is closely related to the thermal structure, the local change in the depth of the seismic-aseismic boundary can be caused by the change of the structure, or the change of the kind of rock materials (Doser and Kanamori, 1986). In particular an abrupt change of the boundary may be related to the crustal structure. As shown in Fig.5, contours of the seismic-aseismic boundary change abruptly in the southern part of Lake Biwa, which may correspond to the change of the structure. The relationships among the seismic-aseismic boundary, thermal structure and velocity structure have not yet been sufficiently known in the northern Kinki district. We need more precise determination of the velocity

structure and more detailed surveys of heat flow to reveal the relationships.

A major part of the crustal strength in the inner zone of island arcs is supported by a brittle layer in the crust, because ductile deformation of rock is predominant in the lower crust and the upper mantle due to high temperature (Shimamoto, 1989). In this case the depth variations of thermal structure is important for tectonics of island arcs, because the thermal structure may be the primary cause of the regional variations in the thicknesses of seismogenic layer. However, heat flow measurements on land areas are very hard and cost too much. Hence the precise determination of focal depths is useful to estimate the strength of the crust and the upper mantle.

The parameters involved in the rheologic model still have large uncertainty to derive a definite model of the shear resistance in the crust, especially of the depth to the brittle-ductile boundary. Therefore, rheologic modeling is still a conceptional one to explain the cutoff depth of seismicity. However the model is a useful guide to reasonable explanations for the nucleation of intra-plate earthquakes and tectonics, because it relates seismicity with plausible rheologic deformations of rocks and with mechanical properties such as shear resistance, with reference to the thermal structure. We need more precise estimates of various parameters of rock deformation by laboratory experiments as well as more precise survey of depth distribution of seismicity for further development of the model. Direct boring to the cutoff depth of seismicity is also expected for in situ measurements of rock properties including pore pressures in the crust.

10. Concluding Remarks

The depth variations of a seismic-aseismic boundary in the mid-crust have been studied here from precise determination of focal depths of more than 8000

small to moderate-sized shocks in the northern Kinki district. The wavelength of undulation of the seismic-aseismic boundary is found to be about 10-50km and the change of depth to the boundary is about 6km.

The depth to the seismic-aseismic boundary is inversely proportional to the surface heat flow value, and the regional variations of the seismic-aseismic boundary seem to be well correlated with the thermal structure of the crust, where temperature at the cutoff depth of seismicity in the crust is estimated about 200-400 °C.

The shape of frequency distribution of the focal depths is quite similar to the depth distribution of the shear resistance calculated from a simple rheologic model which consists of the upper crust in a brittle regime and the lower crust in a ductile regime. This shows that the rheologic model is appropriate to account for the cutoff depth of seismicity.

Large intra-plate earthquakes seem to nucleate at steep changing portions of the seismic-aseismic boundary in the crust, and large inland earthquakes occur in a deep portion of the seismogenic layer. Thus the undulations of the cutoff depth of seismicity is closely related to the depth of large earthquakes in inland areas of island arcs.

The regional variations of the seismogenic layer play an important role to tectonics in island arcs because the strength of the inner zone of island arcs depends largely on the thicknesses of a brittle layer in the crust.

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Figure captions

Fig.1 Map showing seismograph stations of the Abuyama Seismological

Observatory, Kyoto University in the northern Kinki district, Japan.

Fig.2 Distribution of epicenters of well-determined earthquakes used to draw

contours of seismic-aseismic boundaries. Triangles show recording

stations. Size of the plus sign denotes a magnitude.

Fig.3 Cross sections of hypocenters in profiles with 20km in width projected

on planes along NE-SW direction (A-B) (a) and along NW-SE direction (a-f)(b).

Station names attached to the top of each figure show the stations

in each profile.

Fig.4 Distributions of well-determined epicenters at a depth interval of 1km

and contours of seismic-aseismic boundary. Thick lines indicate seismic-

aseismic boundary. Triangles and plus signs denote stations and

epicenters, respectively.

Fig. 5. Contours of the seismic-aseismic boundary in the northern Kinki district.

Fig. 6. Surface heat flow data (Nagao personal communication) in the northern

Kinki district.

Fig. 7. Curie point depth distribution after Okubo (1984). Contour interval is

1 km.

Fig. 8. Depth of seismic-aseismic boundaries and surface heat flow for well

determined focal depths. The upper side of the rectangle indicates the depth above which 90% of shocks occur and lower side is the cutoff depth of seismicity. A-D are after Sibson(1982), E-G are after Weissner and Strhela(1982) and a-d (hatched rectangles) denote data in Japan; a, Kuju geothermal area; b, western part of Nagano prefecture; c, Tottori region; d, Lake Biwa region, see text for reference. Isotherm of 300 and 350 °C are calculated by the same method as Chapman(1985).

Fig. 9. Shear resistance with depth for a simple brittle-ductile transition model for granite (solid lines) and plagioclase (dashed line), see text for parameters.

Fig. 10. Depth-frequency distribution of earthquakes determined by the use of master event location technique in the southern part of the study area.

Fig. 11. Contours of the seismic-aseismic boundary and historical large ($M \geq 6$) intra-plate earthquakes in the northern Kinki district. Epicenters are determined by Usami (1987).

Fig. 12. (a) Epicenters of small and moderate earthquakes during 1976-1987 in the northern Kinki district superposed on the contours of seismic-aseismic boundary in Fig. 5. (b) The seismic-aseismic boundary beneath an earthquake is plotted as a function of the focal depth of the earthquake. Solid line shows the cutoff-depth of seismicity. Magnitude scale is the

same in both figures.

Fig.13. Hypocenter distribution of the main and aftershocks of the 1984

Western Nagano Prefecture earthquake modified from Ooida *et al.*,1990.

Double circles denote the main shock, the largest aftershock (M 6.2) and the secondly largest aftershock (M 5.3).

Fig.14. (a) A schematic model of the nucleation of a large inland earthquake.

(b) Shear strength with increasing depth based on a simple rheologic model.

Fig.15. Depth section of hypocenters near Izu-Oshima volcano and the cutoff

depth of seismicity (solid line) revised from Yamaoka *et al.* (1988). The 1986 fissure eruptions occurred in the horizontal distance range of 0-5km in northwest direction.

Fig.16. Earthquake magnitudes as a function of distance between the epicenters

of the earthquakes by JMA and the crater of Izu-Oshima volcano. It is noticed that no large earthquakes occur near the volcano.

Fig.17. Magnitude of large earthquakes near active volcanoes in Japan as a

function of the distance from volcano. The large earthquakes with magnitude equal to or greater than 6 are based on the list by Kubotera (1988).

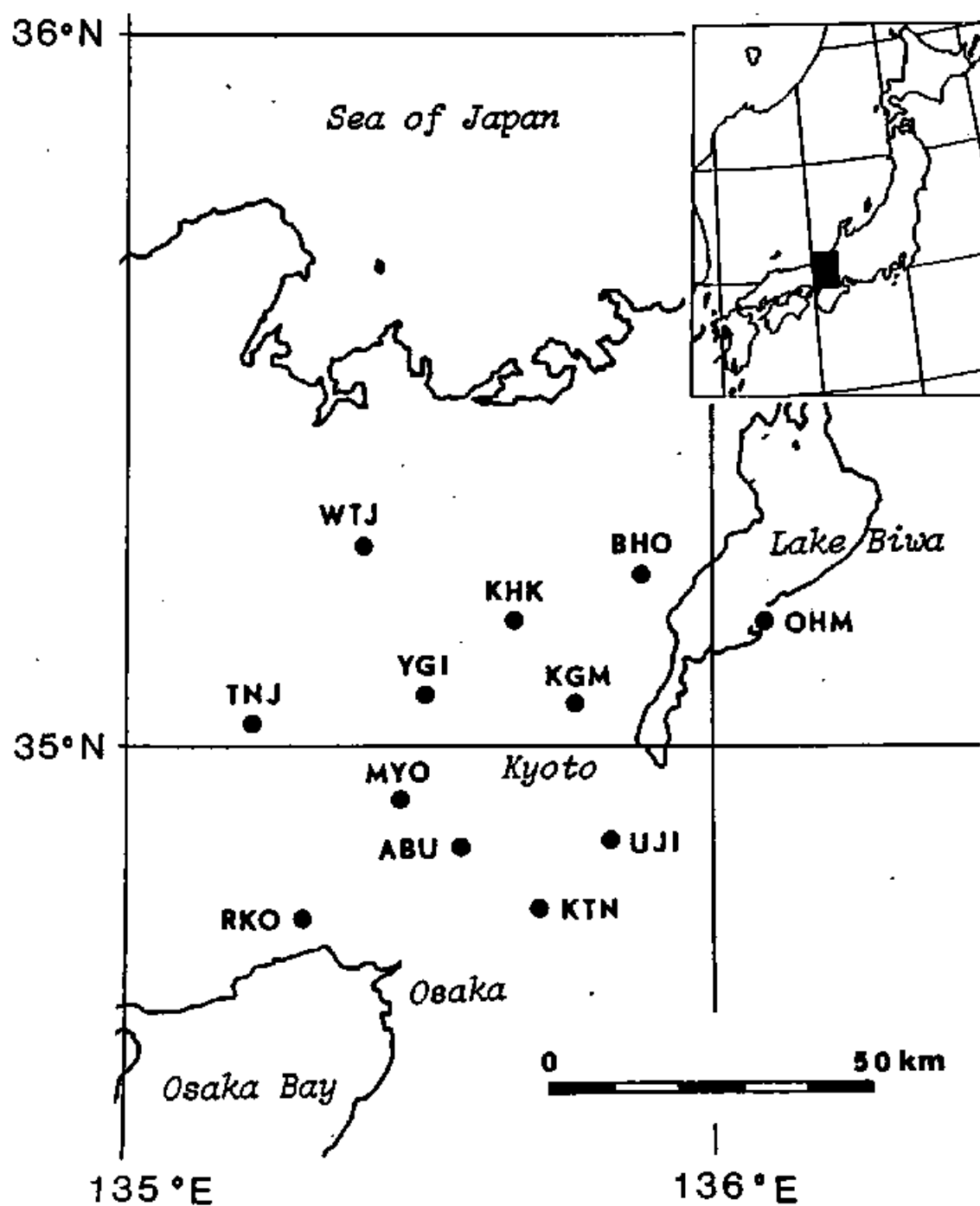


Fig. 1.

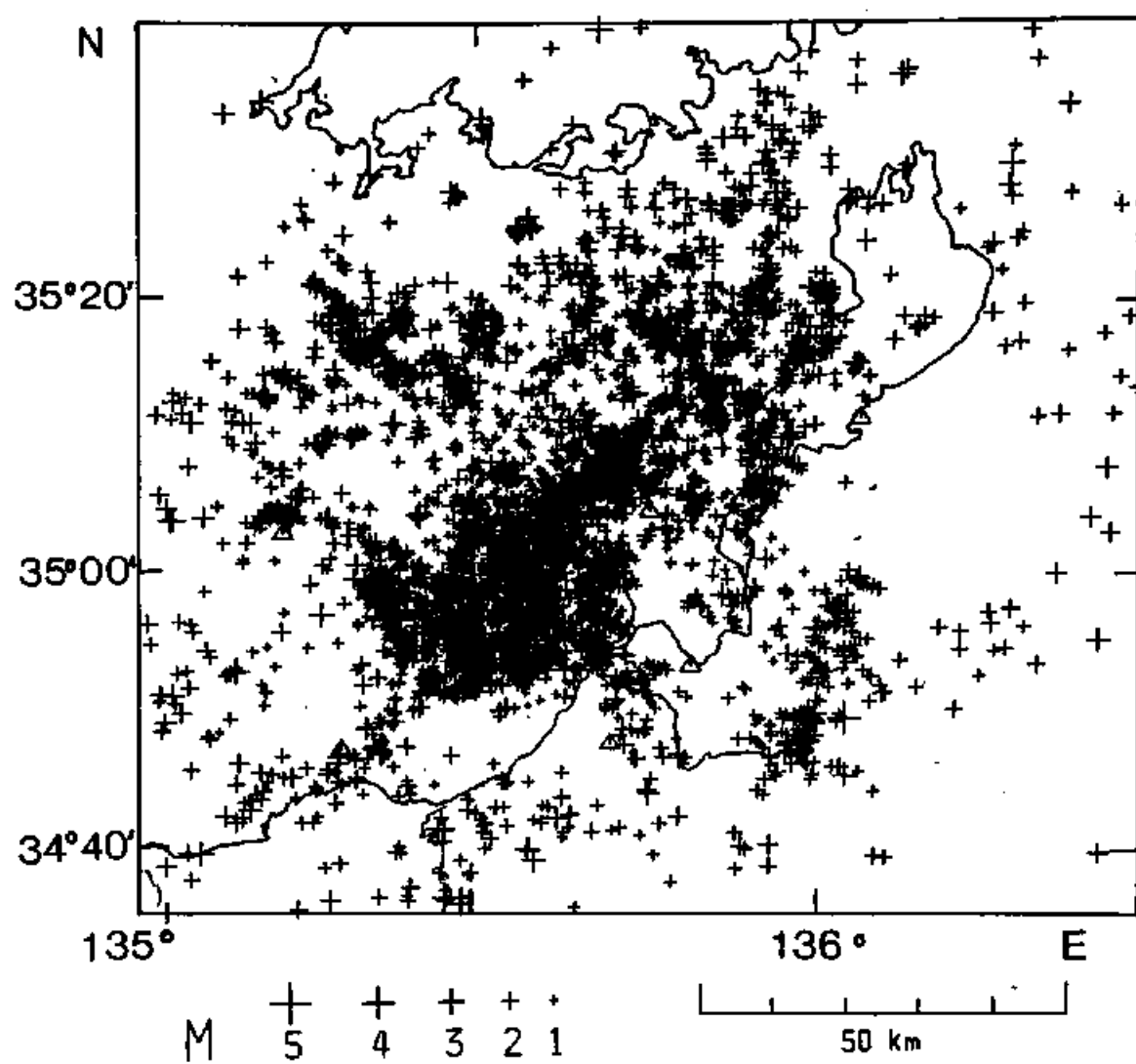


Fig. 2.

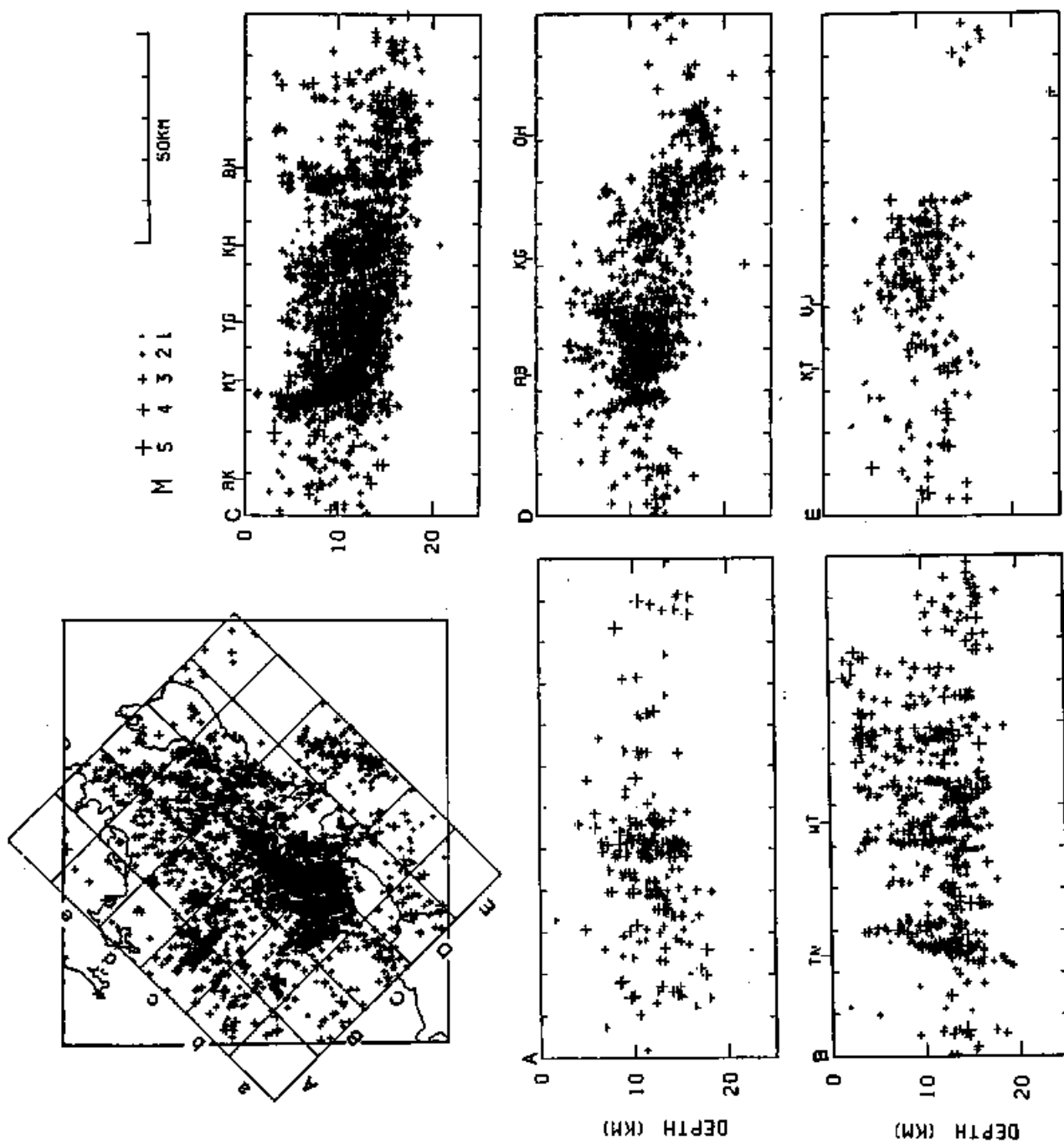


Fig. 3a.

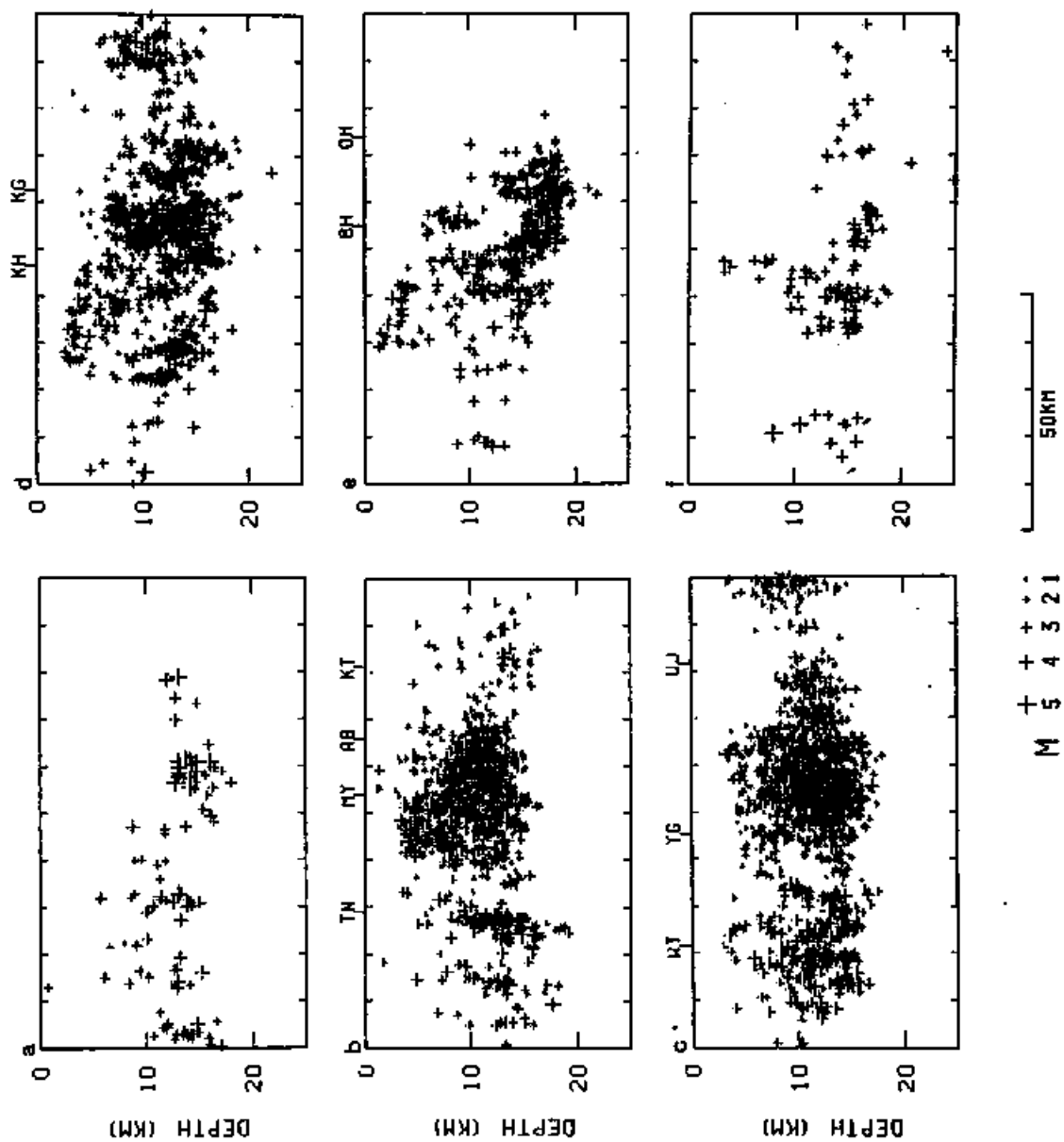
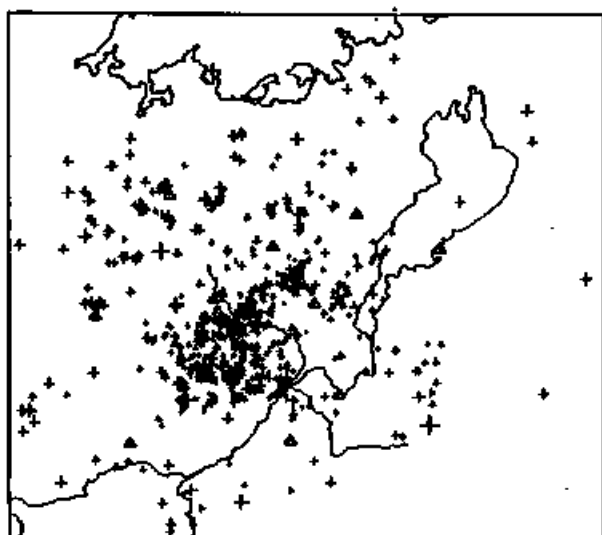
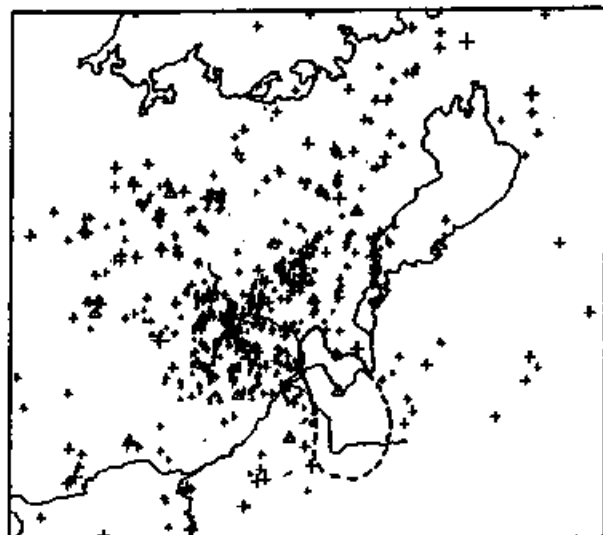


Fig.3b

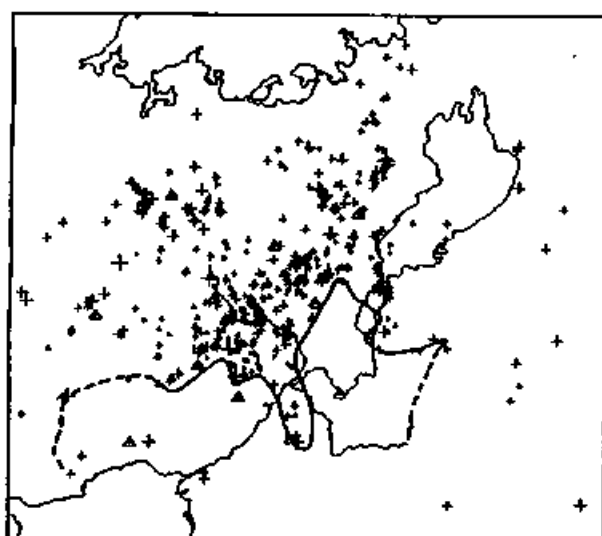
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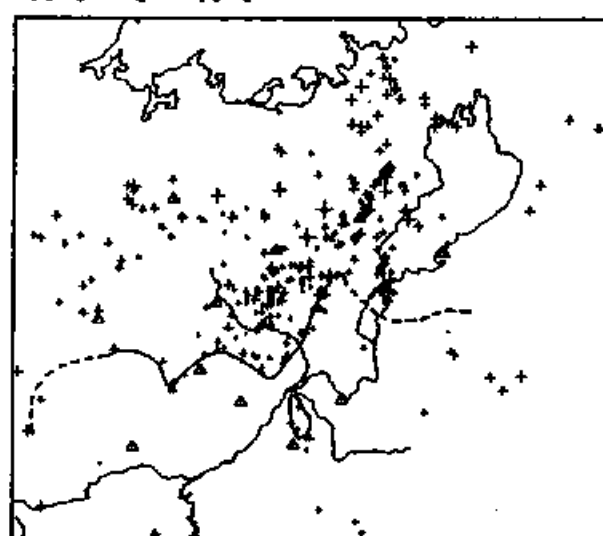
13.0 < Z < 14.0



14.0 < Z < 15.0



15.0 < Z < 16.0

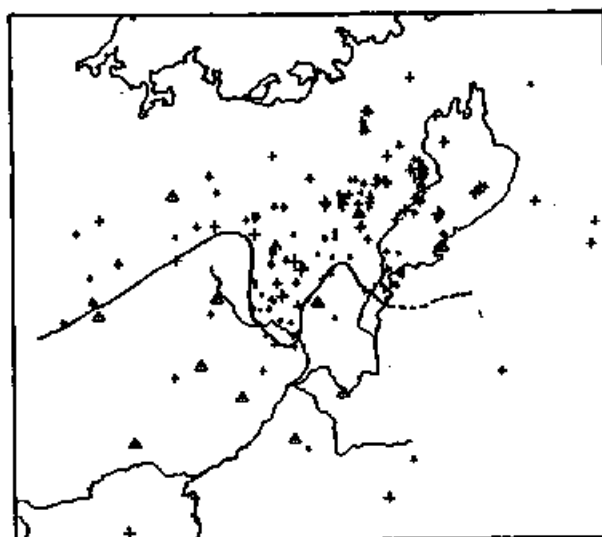


+++++
M 5 4 3 2 1

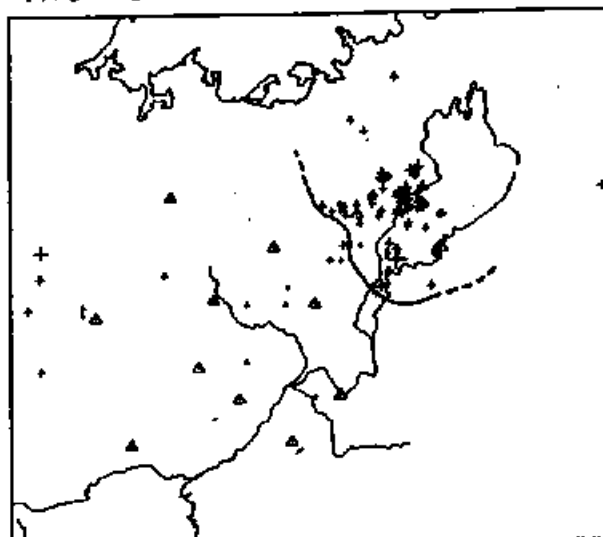
50km

Fig. 4a.

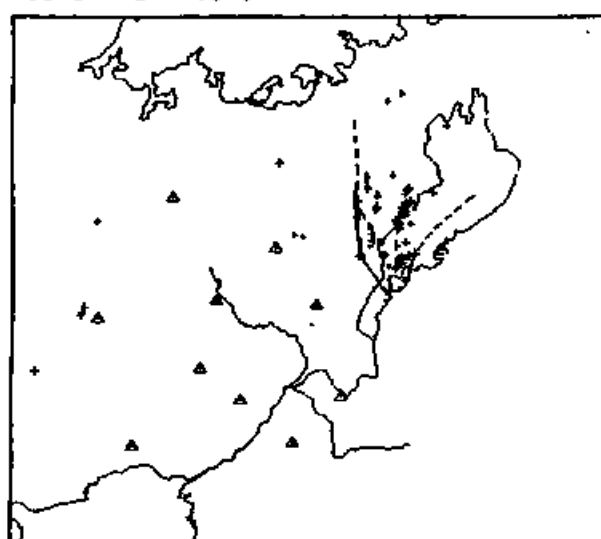
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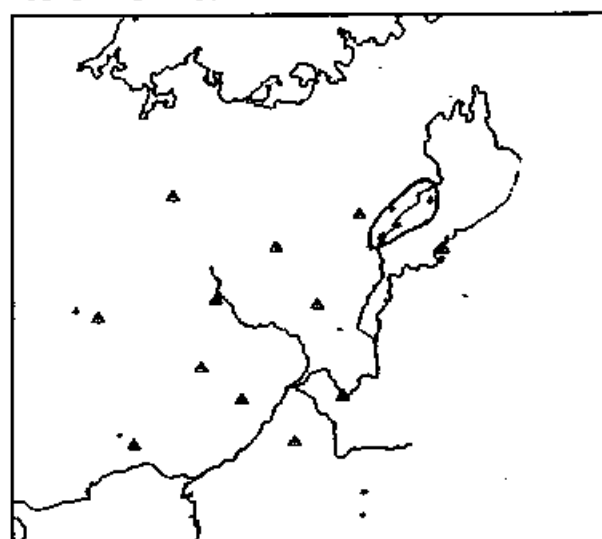
17.0 < Z < 18.0



18.0 < Z < 19.0



19.0 < Z < 20.0



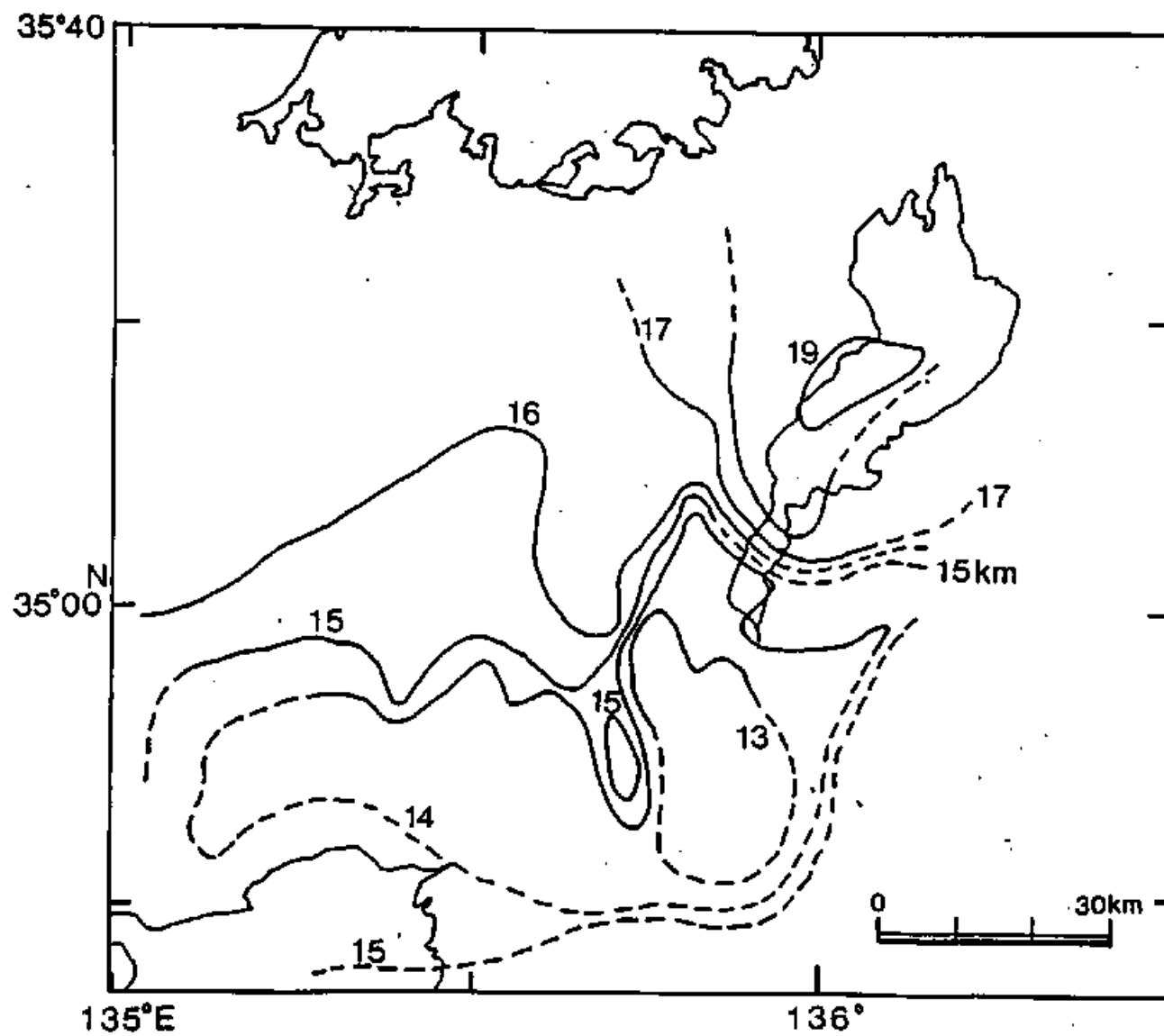
$\begin{matrix} + & + & + & + & + \\ M & 5 & 4 & 3 & 2 & 1 \end{matrix}$

30KM

Fig. 4b.

Fig. 5

Fig. 5.



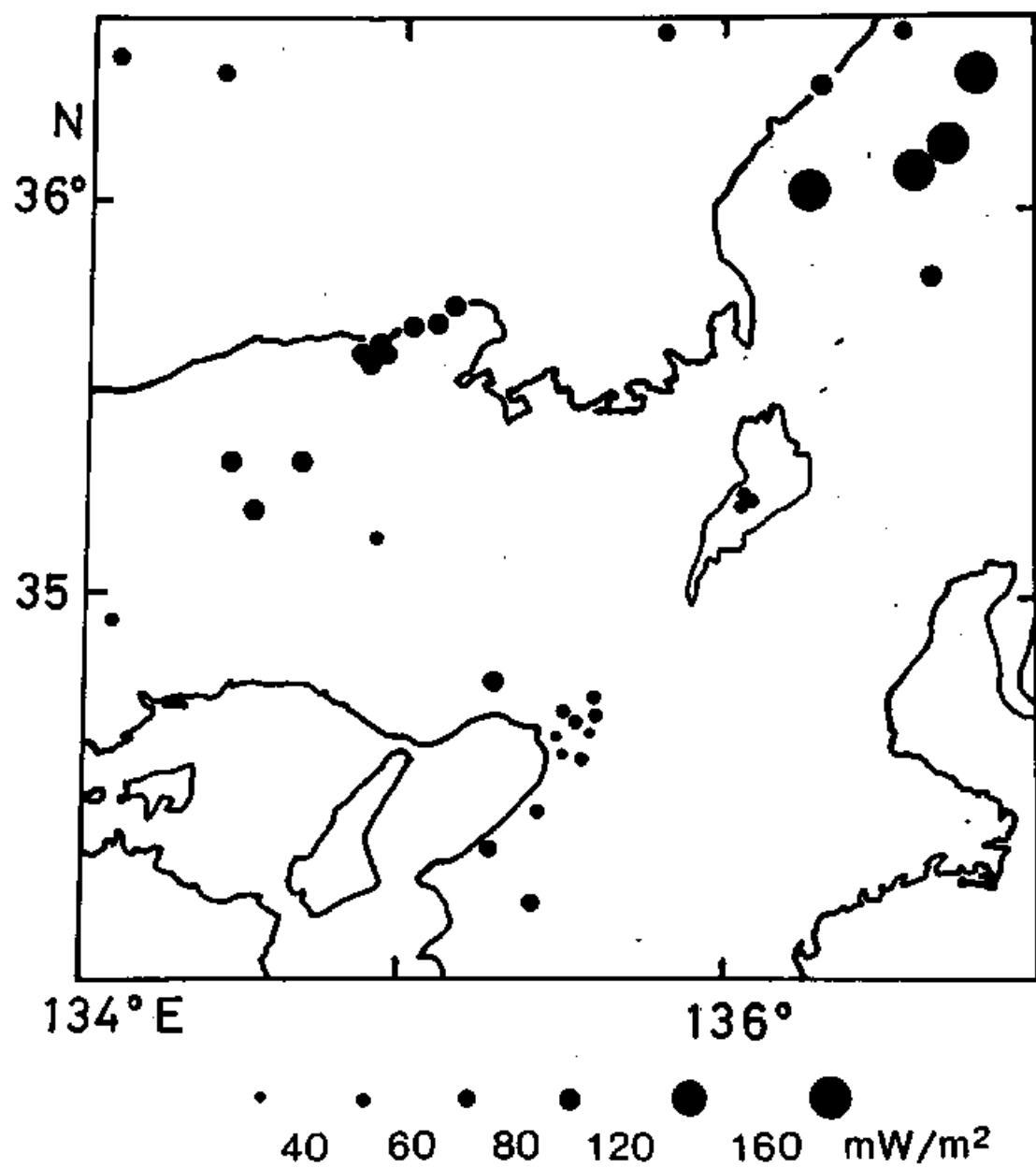


Fig. 6.

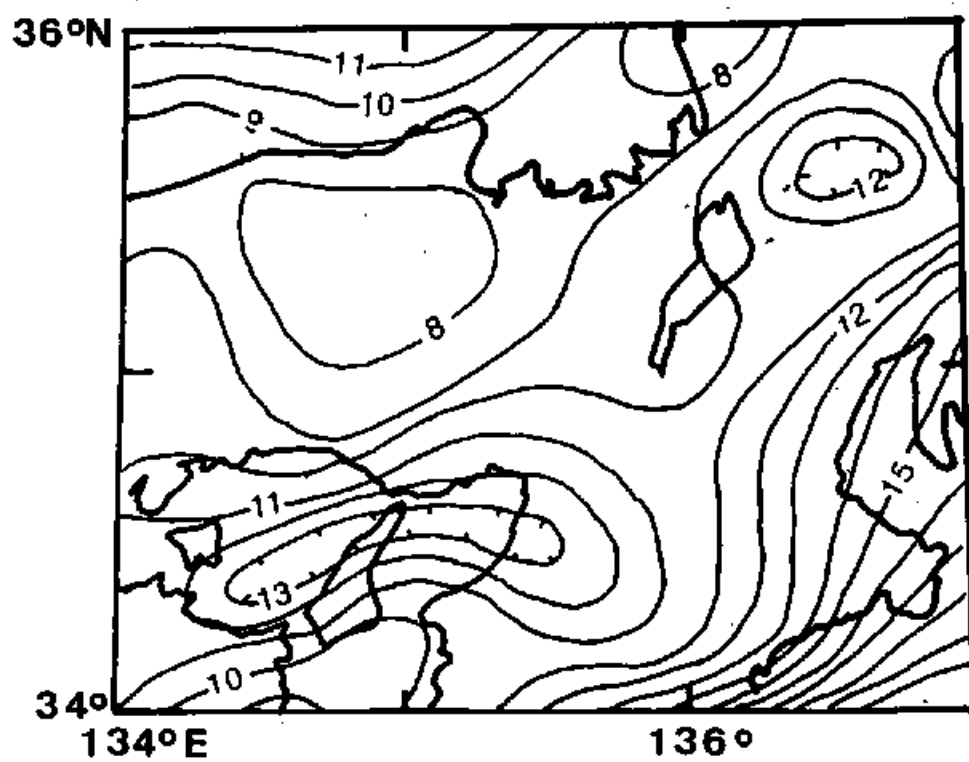


Fig. 7.

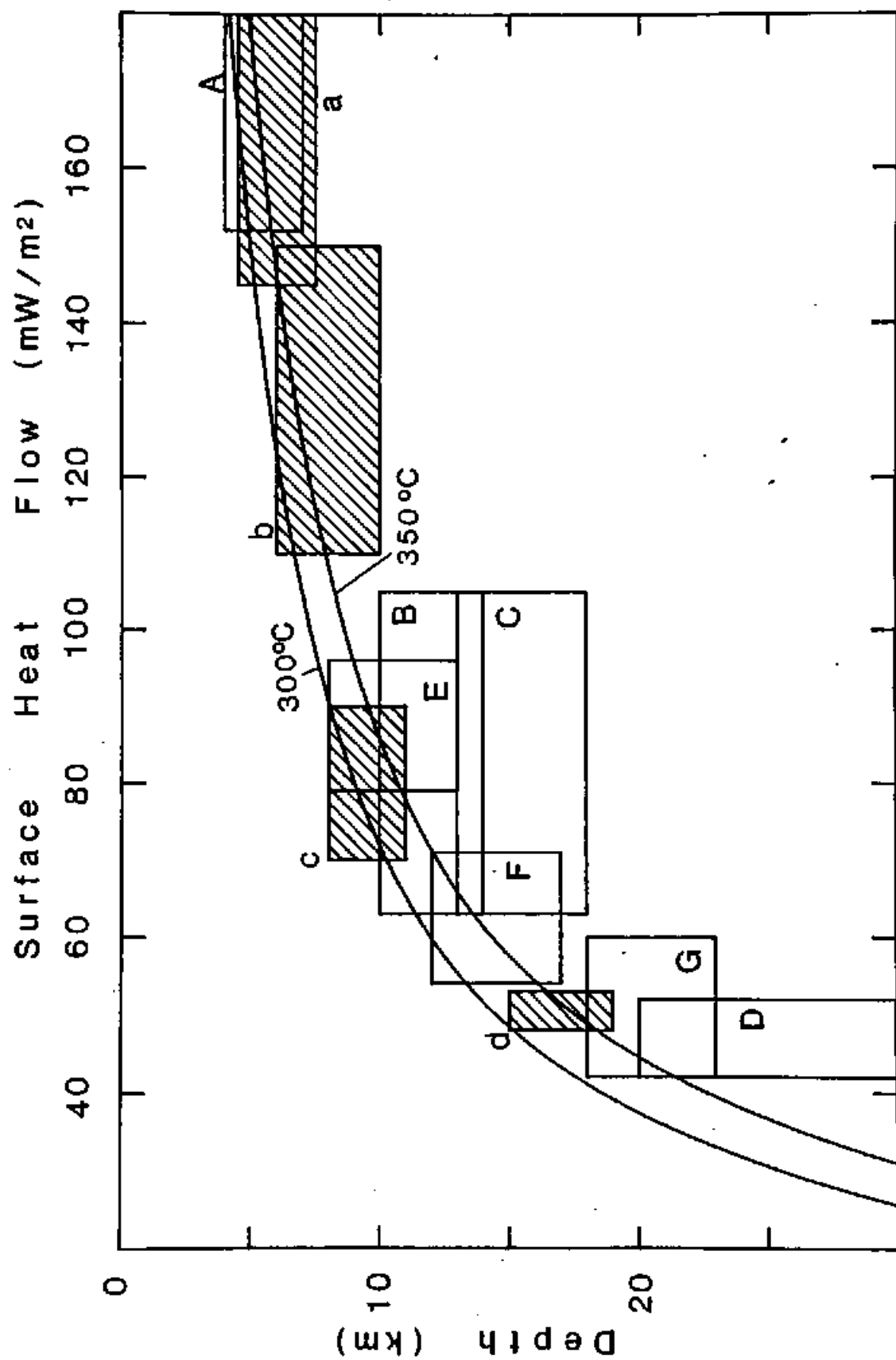


Fig. 8.

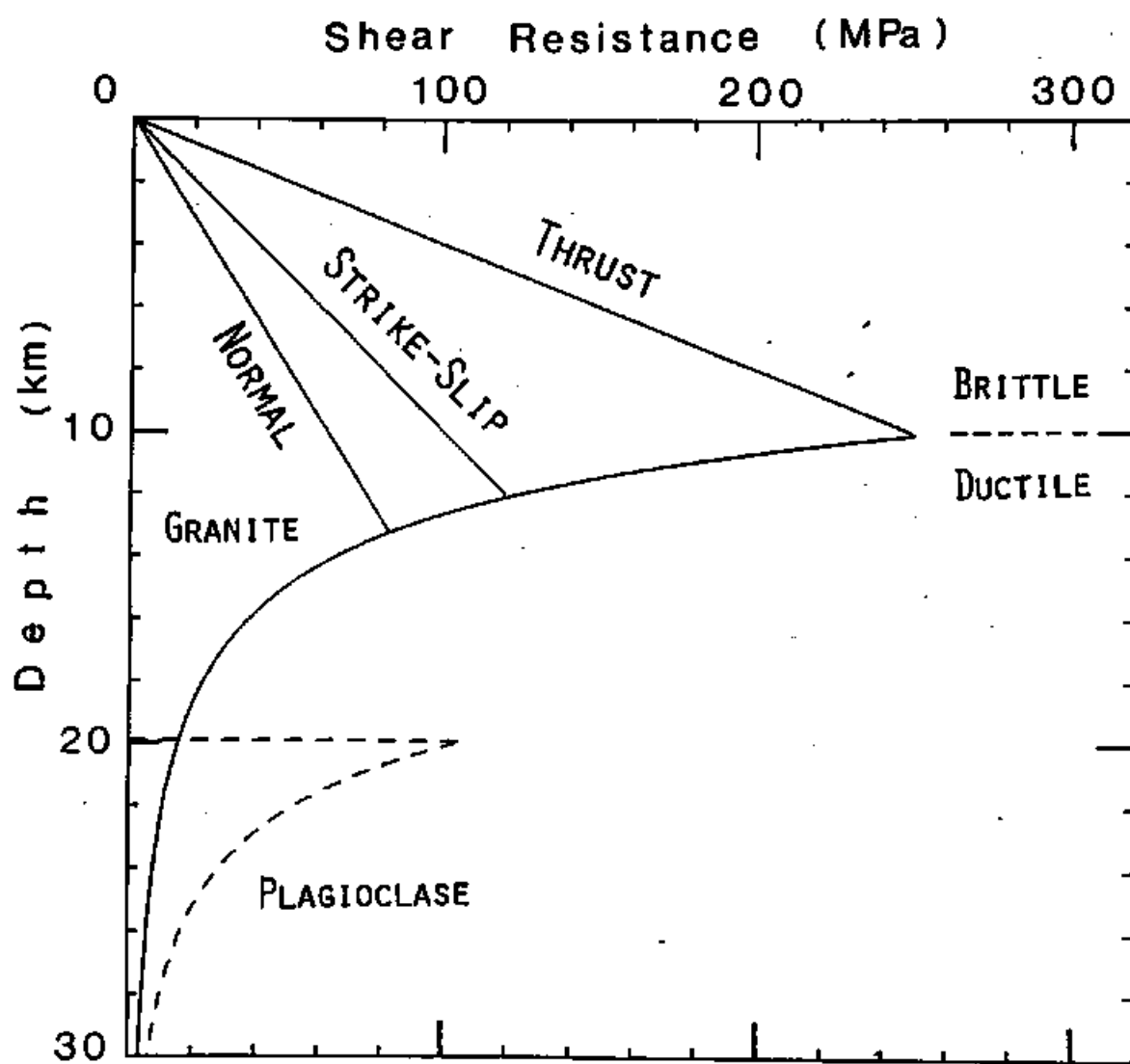


Fig.9.

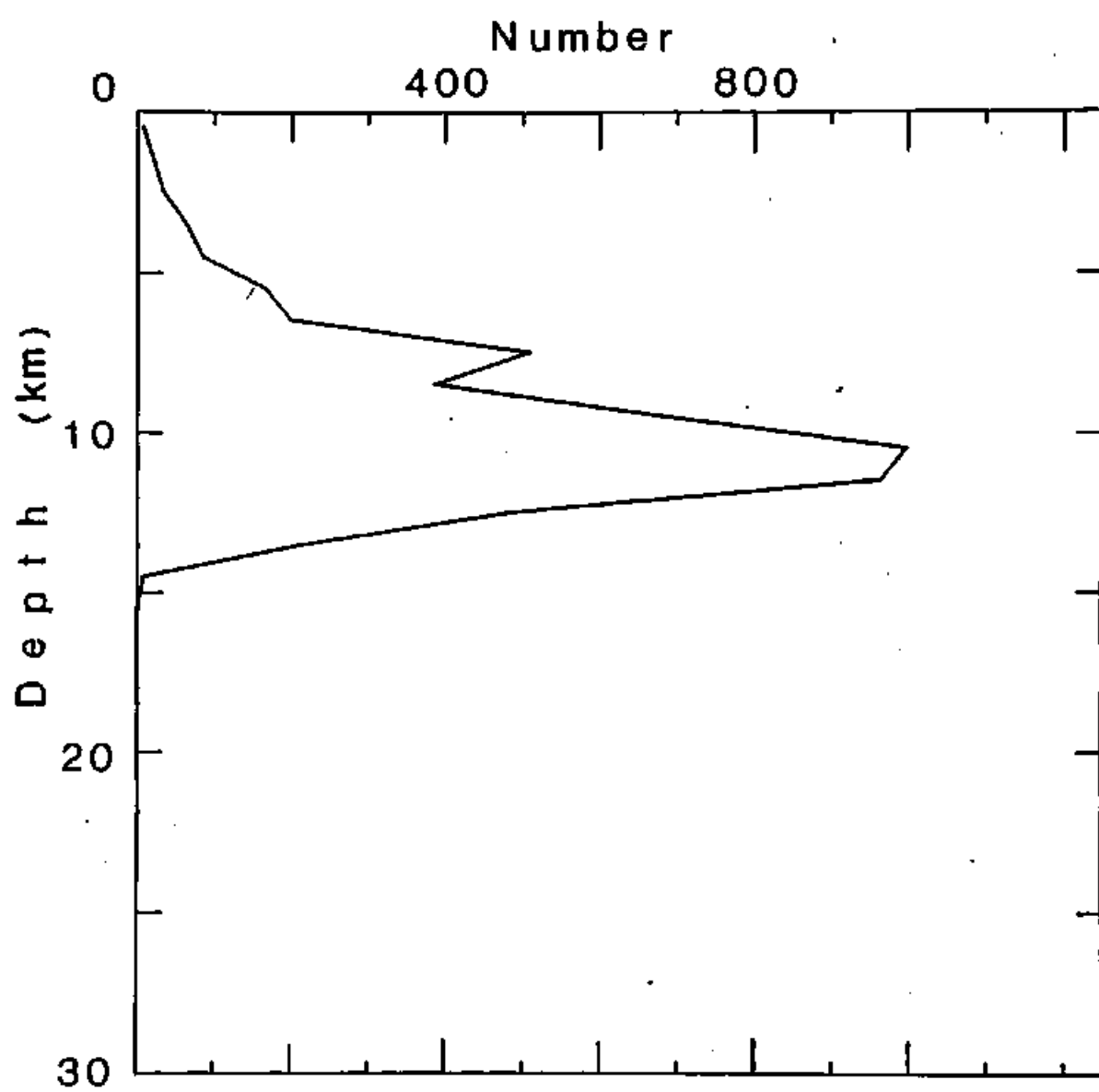


Fig. 10.

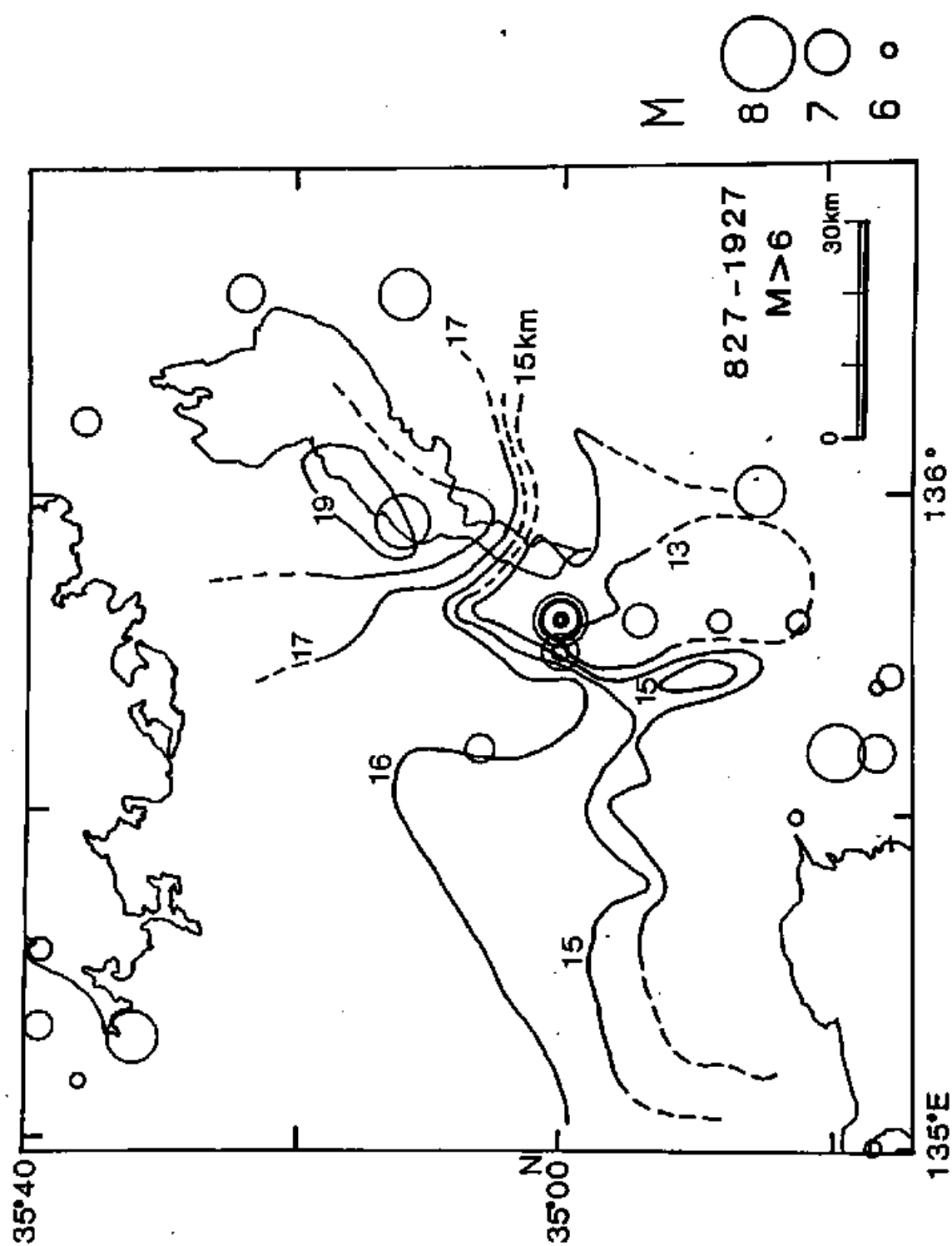


Fig. 11.

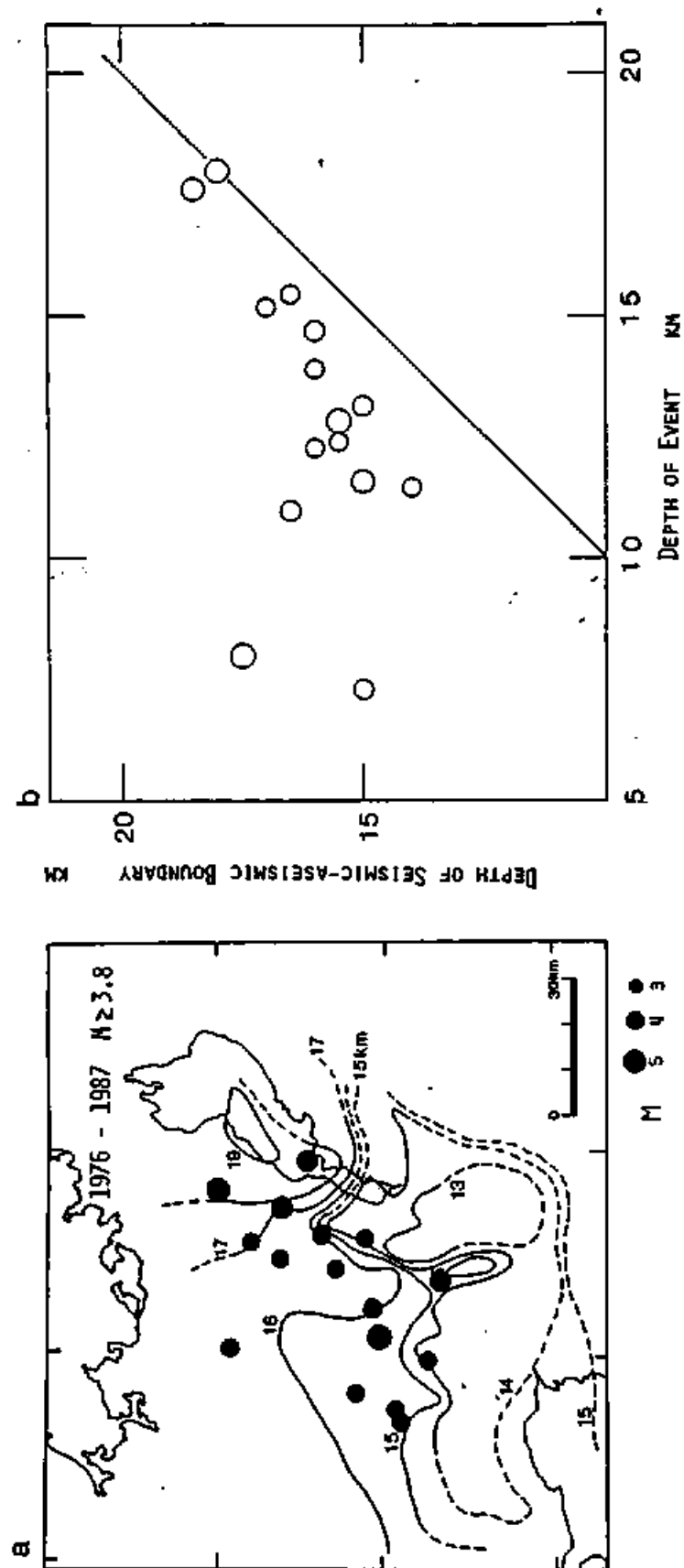


Fig. 12.

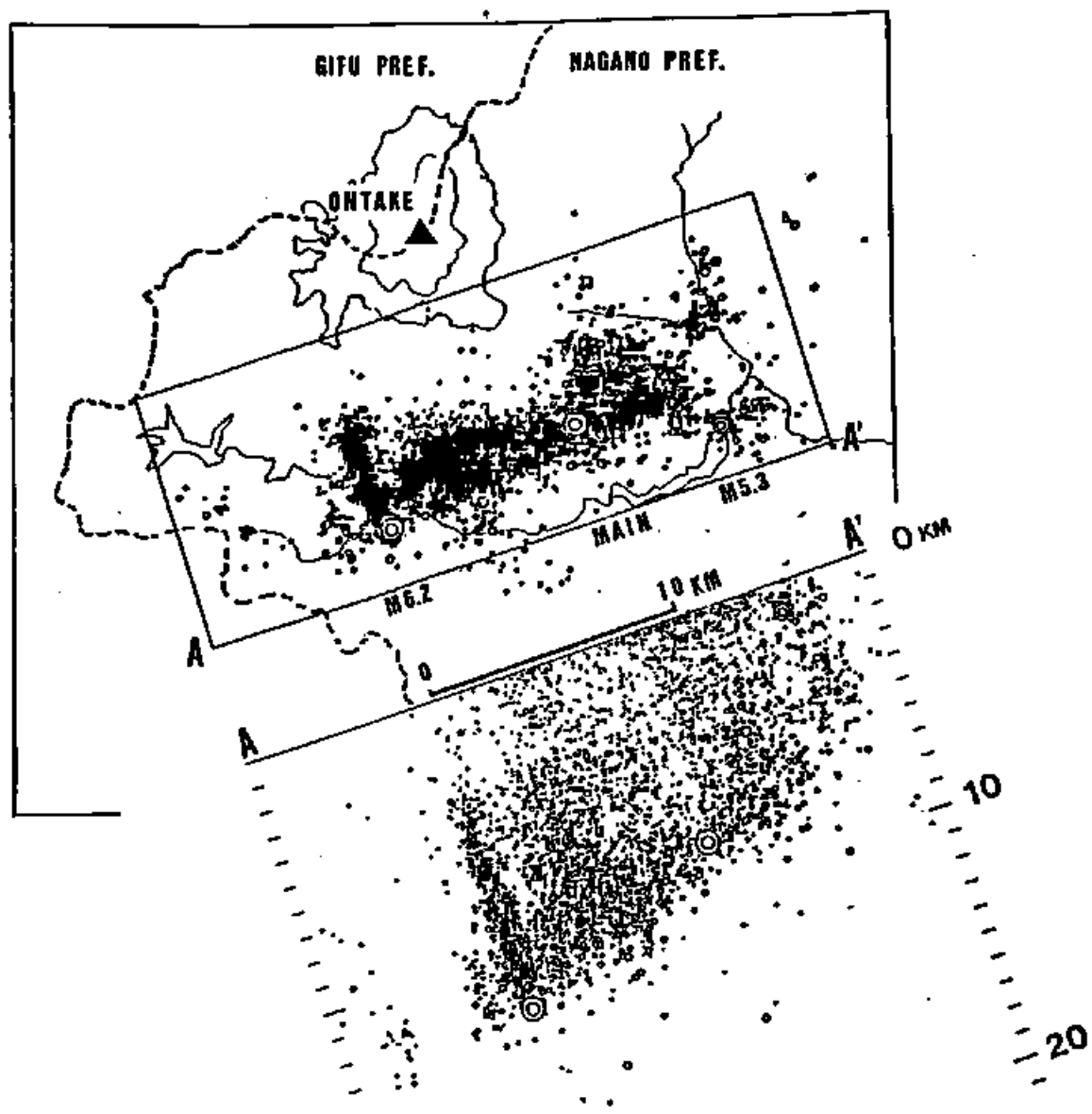


Fig. 13.

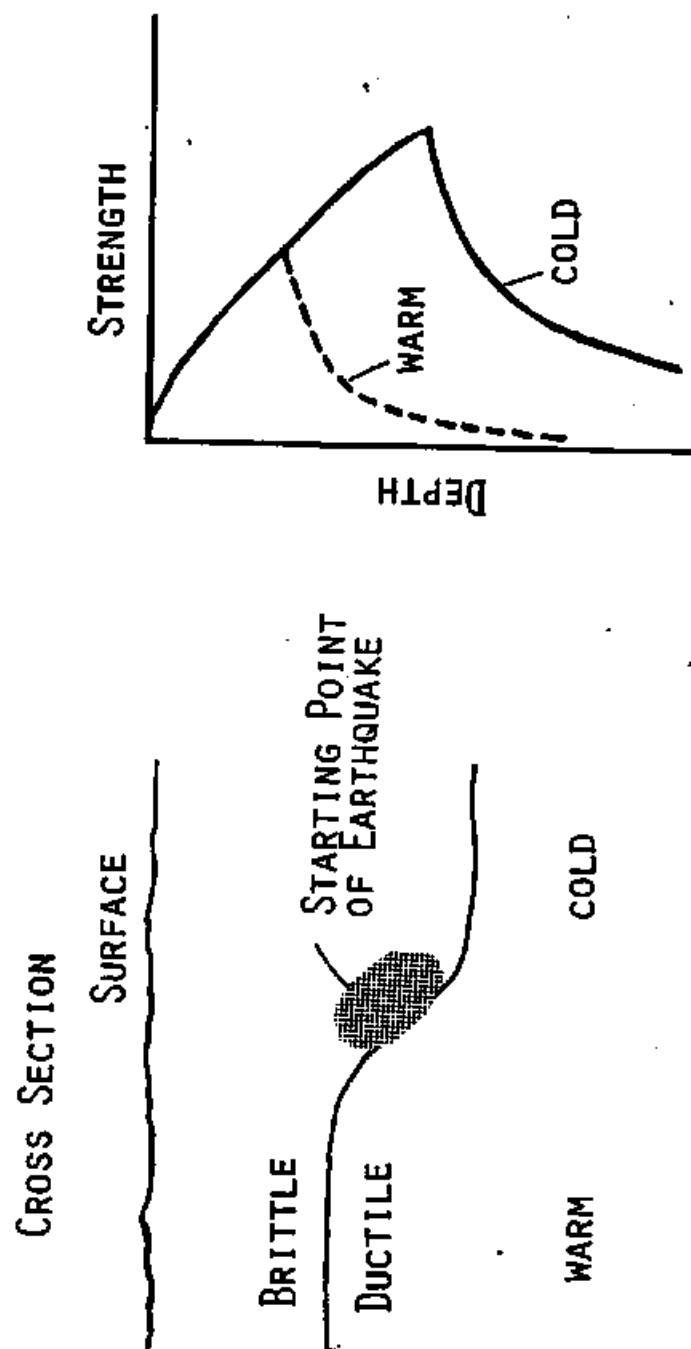


Fig. 14.

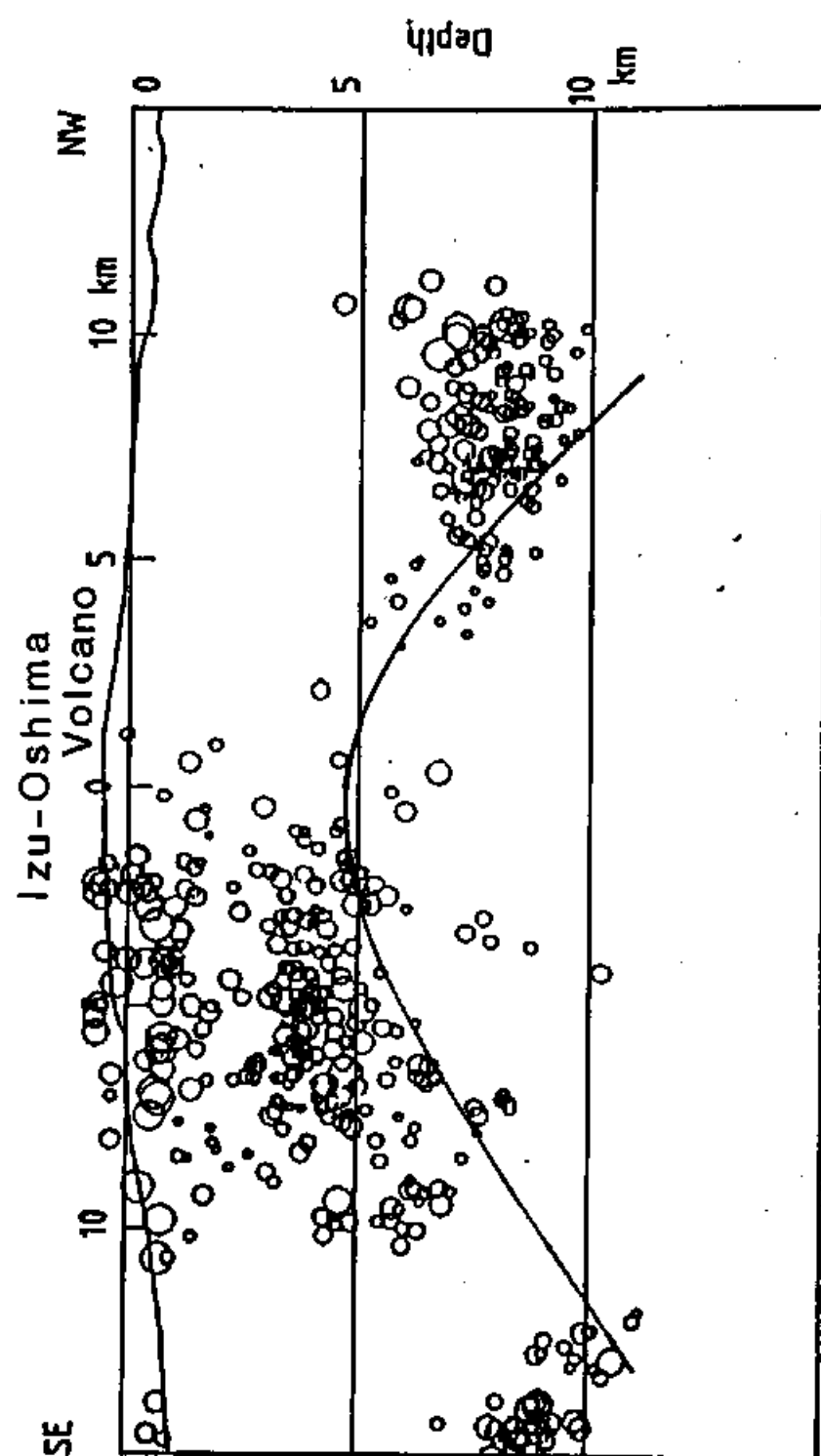


Fig. 15.

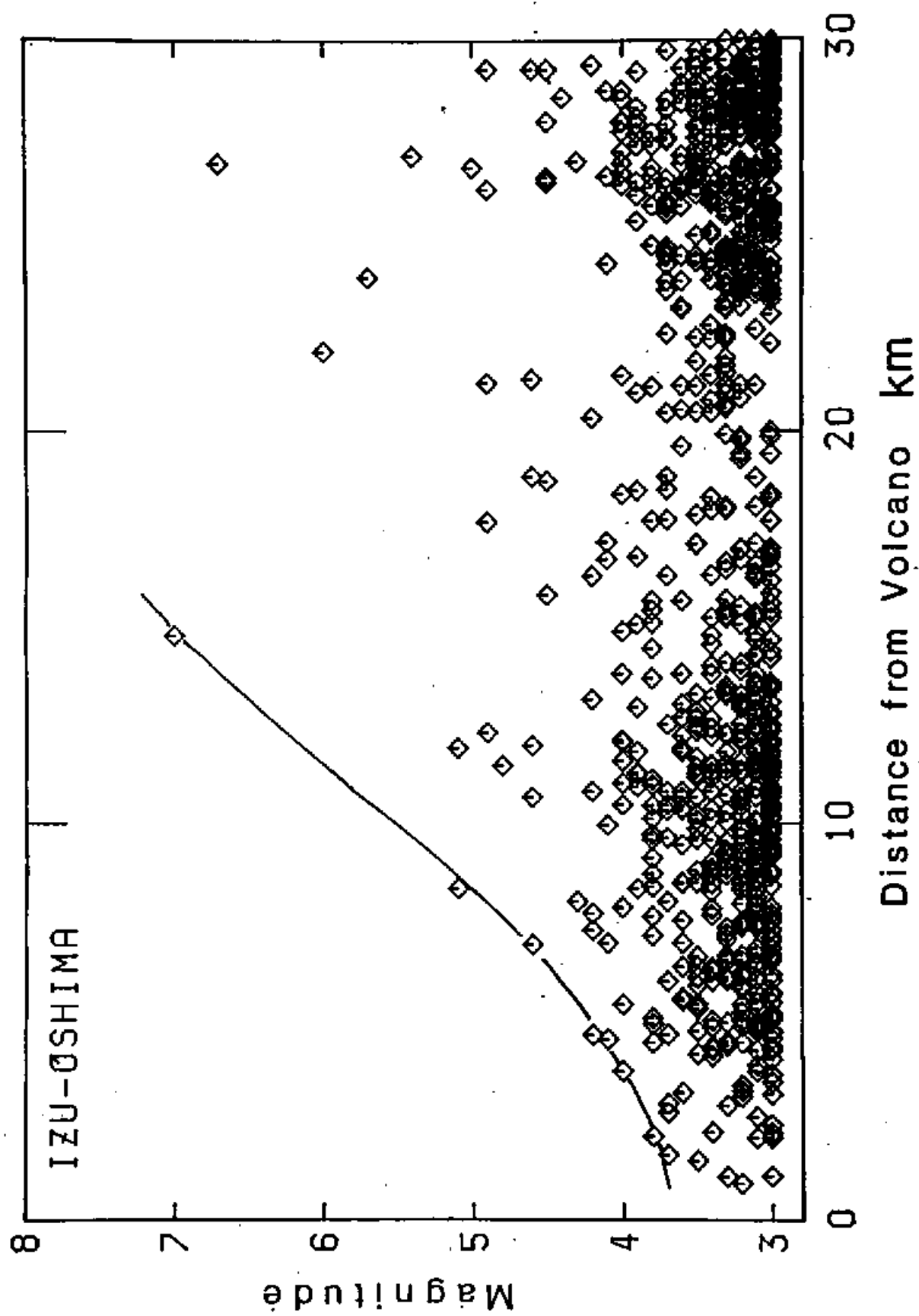


Fig.16.

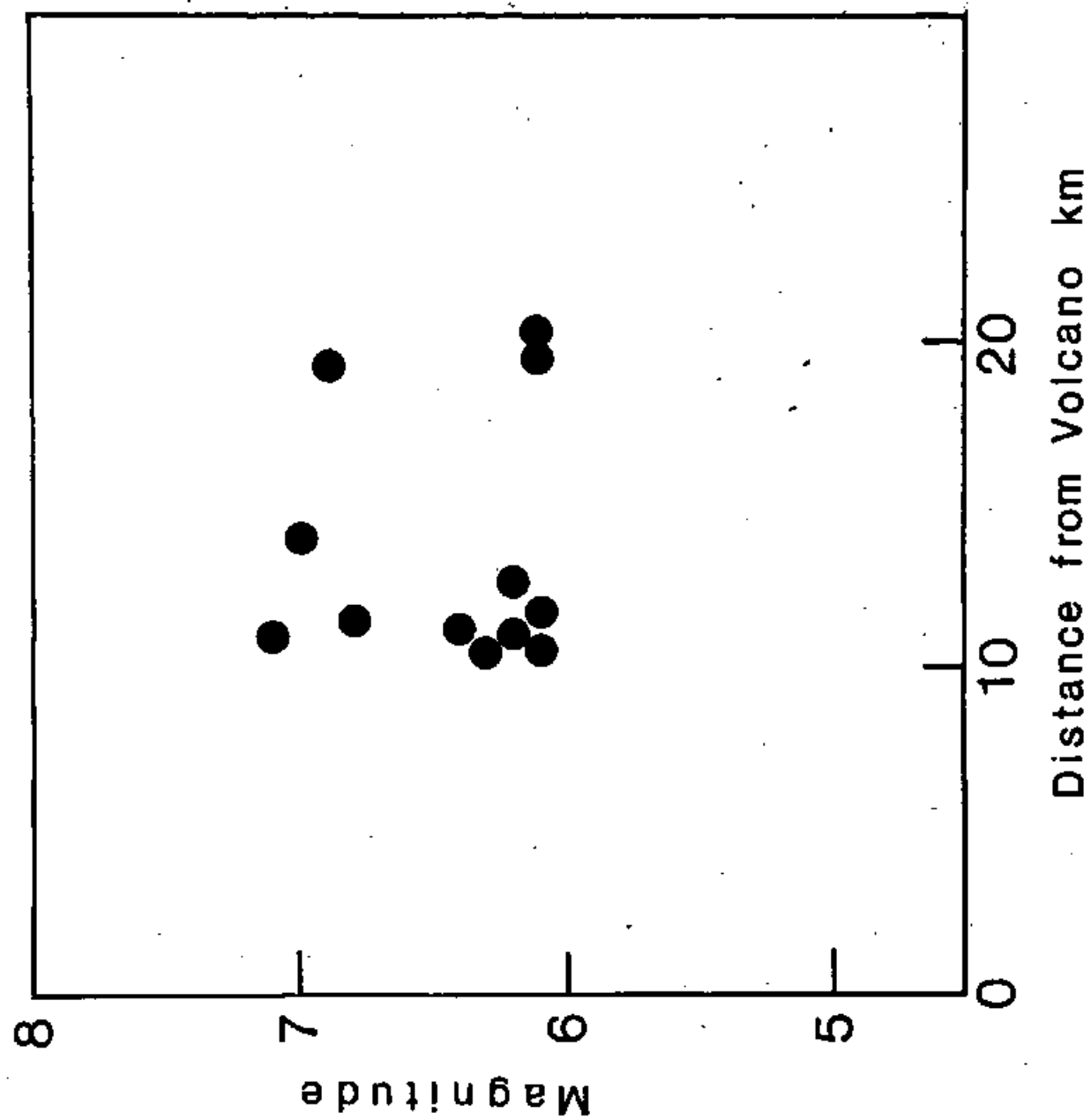


Fig. 17.