Seismic activity and deformation process of the overriding plate in the northeastern Japan subduction zone

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Abstract

We estimated the temperature distribution within the crust of the northeastern Japan arc from P wave velocity perturbations obtained by travel time tomography. By comparing the estimated temperature distribution with the focal depth distribution of shallow, precisely relocated microearthquakes, we found that the brittle to ductile or stick-slip to stable-sliding transition occurs at the ~400°C isotherm and that the transition depth has considerable lateral variations. The brittle seismogenic zone, the upper portion of the crust, becomes locally thin in the P wave low-velocity areas, where the temperature is estimated to be relatively high. Concentration of shallow microearthquakes, high topography and relatively large contractile deformation of the crust are also observed in these low-velocity areas. Active faults are not distributed in the low-velocity areas but lie just along the edge of those areas or outside them. All these observations suggest that earthquake occurrence and deformation within the crust is governed, to a considerable degree, by the thermal regime of this volcanic arc, which is characterized by a horizontally inhomogeneous distribution of temperature. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

Northeastern (NE) Japan overlies a typical subduction zone, where the Pacific plate subducts beneath the land area at a rate of ~10 cm/year. Many studies based on seismic observations have revealed the detailed seismic structure of this volcanic arc.

In the present study, we have estimated the thermal structure of the crust in the NE Japan arc from the tomographic images and compared it with the shallow seismicity, crustal deformation and some other observations. Then we constructed a qualitative model to explain the observed pattern of the shallow seismic activity within the overriding continental plate in this subduction zone. Horiuchi et al. (1982a,b) and Zhao et al. (1990, 1992a) estimated the spatial distributions of the...
depths to the Conrad and Moho discontinuities beneath the land area using shallow earthquake data. Their results show that both discontinuities are deep in the middle of the land area and become shallow toward the coasts of the Japan Sea and the Pacific Ocean. Pn velocity is as low as about 7.5 km/s beneath the land area, but it changes abruptly across the coasts to 8.0–8.2 km/s beneath the Japan Sea and the Pacific Ocean.

Tomographic inversions of P and S wave arrival times from local earthquakes in NE Japan revealed seismic low-velocity bodies in the crust and mantle wedge beneath active volcanoes (Hasemi et al., 1984; Obara et al., 1986). Tomographic images of the crust and upper mantle beneath NE Japan subsequently obtained by Zhao et al. (1992b, 1994) and Zhao and Hasegawa (1993) updated the work of Hasemi et al. (1984) and Obara et al. (1986) by improving the resolution. The subducted Pacific plate is imaged as a high-velocity zone with a thickness of 80–90 km, in good agreement with the results from the reflected waves at the bottom of the slab (Umino and Hasegawa, 1993; Hasegawa et al., 1994). Low-velocity zones are distributed continuously from the crust to the mantle wedge beneath active volcanoes. They are inclined toward the backarc side in the mantle wedge and are nearly parallel to the dip of the subducted Pacific plate.

The three-dimensional seismic attenuation structure of the crust and upper mantle in NE Japan has also been estimated from observations of seismic wave attenuation (Umino and Hasegawa, 1984; Hashida and Shimazaki, 1984, 1987; Matsumoto and Hasegawa, 1989; Matsuzawa et al., 1989). Recently, Tsumura et al. (1996, 2000 — this issue) estimated P wave attenuation structure, source parameters and site response spectra simultaneously by inverting the spectral data of many earthquakes, improving on previous studies of the attenuation structure in NE Japan. Their results clearly show the existence of the inclined high-Q Pacific plate and low-Q bodies in the crust and in the mantle wedge. The spatial distribution of the low-Q bodies roughly coincides with that of the low-velocity bodies imaged by the travel time tomography (Zhao et al., 1992b, 1994; Zhao and Hasegawa, 1993).

2. Shallow seismic activity beneath the land area

Fig. 1 shows the epicenter distribution of shallow (shallower than 40 km) microearthquakes located by the Tohoku University seismic network in the NE Japan subduction zone. Many shallow microearthquakes occur along or in the vicinity of the boundary between the subducted Pacific plate and the overriding continental plate (the main thrust zone) beneath the Pacific Ocean, as do the large destructive earthquakes that have repeatedly inflicted damage to this district. Highly concentrated clusters of events beneath the Japan Sea are the aftershocks of the 1983 Nihonkai Chubu Earthquake (M7.7) and of the 1993 Hokkaido Nansei-oki Earthquake (M7.8). These events, and others occurring beneath the Japan Sea, are considered to be the activity along or in the vicinity of a new plate boundary currently being formed along the eastern edge of the Japan Sea (Nakamura, 1983; Kobayashi, 1983).

In addition to the seismic activity along or in the vicinity of the plate boundaries beneath the Pacific Ocean and the Japan Sea, shallow earthquakes also occur within the overriding continental plate beneath the land area. As can be seen from Fig. 1, these intraplate microearthquakes are not distributed homogeneously in space but concentrate to some particular areas. The activity level of these events is relatively high along the volcanic front or the central mountainous range (Ou Backbone Range), which runs nearly parallel to the axis of the Japan trench.

In order to see the focal depth distribution of these events in detail, we relocated the shallow events that occurred in the middle of the land area during the period of August 1991 to June 1996 by using the same method of source-region station corrections as that of Hasegawa and Yamamoto (1994). The station coverage of the seismic network of Tohoku University is relatively dense there, and thus precise determination of hypocenters, especially focal depths, can be expected from the relocation with source-region station corrections. Hypocenters thus relocated are added to the data set of Hasegawa and Yamamoto (1994), which results in a large data set containing the events from November 1988 to June 1996. Fig. 2
shows a N–S vertical cross-section of these relocated events along the volcanic front in the middle of NE Japan. Deep, low-frequency microearthquakes, which have been found in a depth range of 22–47 km beneath active volcanoes and have anomalously low predominant frequencies of both P and S waves (1.0–5.5 Hz for P waves and 1.5–4.5 Hz for S waves; Hasegawa and Yamamoto, 1994), are also shown in Fig. 2.

We can see in Fig. 2 that the depth limit of
shallow normal events is about 15 km or less, and it changes remarkably with location. The cut-off depth for the normal shallow seismicity is as deep as 15 km in the focal areas of the 1896 Rikuu Earthquake (M7.2) and the 1970 Akita-ken Nantobu Earthquake (M6.2), but it becomes locally as shallow as 10 km beneath active volcanoes.

3. Estimation of temperature distribution in the crust

The regional variation of the cut-off depth for the shallow inland events and its close correlation with the distribution of active volcanoes lead us to conjecture that the cut-off depth is prescribed by the crustal temperature. In order to confirm...
this, we tried to estimate the temperature distribution of the crust from the P wave velocity distribution obtained by travel time tomography.

Estimation of temperature distribution in the crust is made following the method of Sato and coworkers (Sato et al. 1989, 1997; Sato, 1995). We simply assume that the P wave velocity perturbation at each location is caused by the temperature perturbation there, although other factors such as rock types or the content of volatiles also cause the differences in seismic velocity. Then we estimate the temperature perturbation from the P wave velocity perturbation, by using P wave velocity data determined at high pressure and temperature in the laboratory (Hughes and Cross, 1951; Hughes and Maurette, 1956, 1957; Lin, 1977; Kern, 1978; Christensen, 1979).

We assume that the temperature \( T(\varphi, \lambda, h) \) at a certain location \((\varphi, \lambda, h)\) can be expressed as:

\[
T(\varphi, \lambda, h) = T_0(h) + \Delta T(\varphi, \lambda, h),
\]

where \( T_0(h) \) is the mean or standard temperature at depth \( h \), and \( \Delta T(\varphi, \lambda, h) \) is the temperature perturbation at the location \((\varphi, \lambda, h)\); \( \varphi \) and \( \lambda \) are latitude and longitude respectively. \( \Delta T(\varphi, \lambda, h) \) is given by:

\[
\Delta T(\varphi, \lambda, h) = \Delta V(\varphi, \lambda, h)(\partial V/\partial T)_h,
\]

where \( \Delta V(\varphi, \lambda, h) \) is the velocity perturbation, and \((\partial V/\partial T)_h \) is the temperature derivative of the P wave velocity at depth \( h \). The temperature derivative of the P wave velocity at each depth (or pressure) can be estimated from laboratory measurement data (Sato et al., 1989, 1997; Sato, 1995).

In the present study, we adopted the thermal structure model by Furukawa and Uyeda (1986) based on heat flow data as the standard, i.e. the mean temperature \( T_0(h) \) at depth \( h \). Since we are focusing our discussion on the temperature in the upper crust, the temperature derivative of the P wave velocity for granite compiled by Sato et al. (1997) is used for the present estimation. The estimated temperature derivative has uncertainties of several percent (Sato et al., 1997). The P wave velocity perturbations for the crust by Zhao et al. (1992b) are used as \( \Delta V(\varphi, \lambda, h) \). Then we estimate the temperature distribution in the crust from Eqs. (1) and (2).

4. Relationship between P wave velocity structure, active faults, crustal deformation and other observations

The estimated distribution of temperature is shown in Fig. 2 on the NS vertical cross-section of shallow seismicity by three isothermal contour lines of 300, 400 and 500°C. As clearly seen from Fig. 2, the estimated temperature in the crust changes laterally; the isotherms become locally shallow beneath the active volcanoes (solid triangles). The cut-off depth of the shallow seismicity is in good agreement with the 400°C isotherm. This provides evidence that the depth limit of the shallow seismicity in this region is prescribed mainly by the temperature distribution of the crust, which varies significantly not only in the vertical direction but also in the horizontal direction.

The cut-off depth for the shallow seismicity, sharply delimited and closely correlated with the horizontally varying temperature distribution, can be interpreted as the zone of the brittle to ductile transition or the stick-slip to stable-sliding transition due to increasing temperature with depth (e.g. Brace and Byerlee, 1970; Kobayashi, 1976; Meissner and Strehlau, 1982; Sibson, 1982; Tse and Rice, 1986; Shimamoto, 1991). The relatively shallow depth of the transition zone at \(~ 15 \) km is considered to be due to a steep geothermal gradient in this volcanic arc that is formed by magmatic underplating and intrusion (e.g. Shimamoto, 1991). The upper 15 km of the crust forms a brittle seismogenic zone; lower temperatures allow brittle failure or stick-slip in rock rather than aseismic deformation. The lower portion of the crust and the mantle wedge below it are governed by creep or plastic flow (i.e. aseismic deformation).
locally elevated, if the horizontally compressional stress field due to the plate convergence is acting in the direction of E–W to ESE–WNW in this subduction zone. To understand this process better, we further compared the P wave velocity structure obtained by travel time tomography with other observations. Here we simply assume that the P wave velocity perturbation at each location reflects the temperature perturbation there.

Fig. 3 shows the distribution of P wave velocity perturbations at 40 km depth in the uppermost mantle beneath NE Japan determined by travel time tomography (Zhao et al., 1992b). Among the tomographic images of all depth levels, the image at 40 km depth (Fig. 3) has the best spatial resolution because this depth slice was most densely covered by both the horizontally propagating Pn waves from the crustal events and the vertically traveling rays from the intermediate-depth earthquakes in the slab. The tomographic images of the crust (0–35 km) were mainly determined by the direct waves (Pg) and refracted waves within the crust (P') from the shallow events, in addition to the up-going rays from the slab events. For a seismic wave from a crustal event, the first arrival is the Pg or P' wave when the epicentral distance is shorter than about 140 km, and it is the Pn wave when the epicentral distance is longer than 140 km. Because the spacing between stations of the Tohoku seismic network is 40–50 km, most of the arrival times from the shallow events are those of Pn waves, and there are few Pg and P' waves (Zhao et al., 1992b). For this reason, the tomographic image of the Pn velocity (Fig. 3) has much higher resolution, except for the central part of the land area where clusters of stations are located. Therefore, we prefer to use the Pn velocity image to represent the P wave velocity structure of the crust in NE Japan. In other words, we assume that P wave velocity perturbations in the crust are closely correlated with Pn velocity perturbations, because we expect that the horizontal variations of temperature in the crust are formed mainly by upward movement of mantle material and heat into the crust. In fact, the crust right above the low Pn velocity areas generally has low P wave velocities in regions of relatively dense station coverage (Zhao et al., 1992b).

Active faults (Active Fault Research Group, 1991) are shown by red lines in Fig. 3; active volcanoes are also shown by the red triangles. We can see in this figure that active faults are not distributed within P wave low-velocity areas (red or yellow colors) but just along the edge or outside of them. Moreover, the length of each fault system seems to be governed by the distribution of the low-velocity areas, i.e. the ends of a fault system are often cut by the low-velocity areas or the length of a fault system corresponds to the areal extent of a low-velocity area.

Fig. 4 shows the topography of NE Japan together with active faults. Comparison with Fig. 3 shows that high topography areas approximately coincide with P wave low-velocity areas except for the Kitakami Mountain Range located in the forearc side (38.8–40.3°N, 141.2–142°E). Active faults are traces of the past large inland earthquakes that have occurred there repeatedly, and are sites of future large earthquakes. Hence, the good correlation between the spatial distributions of low-velocity areas, active faults and mountain ranges suggests that horizontal variations of temperature in the crust play an important role for the generation of shallow earthquakes, and for the deformation process of the crust within the overriding continental plate.

The information on crustal deformation for recent years in NE Japan can be obtained from the result of triangulation measurements made by the Geographical Survey Institute of Japan. Fig. 5 shows a crustal strain map for the last about 100 year period modified from the Geographical Survey Institute of Japan (1994). Coseismic strains associated with large inland earthquakes that
Fig. 4. Surface topography and active faults in NE Japan. The topography is shown in meters by the color scale on the top. Active faults and active volcanoes are shown by red lines and red triangles respectively.
Fig. 5. Map showing the crustal strain for the last ~100 year period (Geographical Survey Institute of Japan, 1994) and epicenters of shallow microearthquakes (depth ≤40 km). Elements of the triangulation net with contractions larger than $10^{-5}$, less than $10^{-7}$, and with extensions in the plate convergence direction are shown by red, yellow and blue triangles respectively.
occurred during the period are subtracted and not included in this map. Elements of the triangulation net with contractions larger than $10^{-5}$ in the direction of the plate convergence are shown by red triangles. We can see from Fig. 5 that most of the microearthquakes (dots) occurred in areas with large contractions in the plate convergence direction rather than in areas with small contractions or extensions. Comparison with Fig. 3 also shows that the P wave low-velocity areas roughly correspond to the areas of large contractions.

5. Discussion

Based on the observations described in the previous sections, we propose a qualitative model of earthquake generation and crustal deformation within the overriding plate for this subduction zone, as schematically shown in Fig. 6. Under the horizontally compressional stress field caused by the plate convergence, stress concentration will take place in or around the areas where the brittle seismogenic zone is locally thin and so its mechanical strength is comparatively weak.

The correlation of abundant shallow microearthquakes, high topography, large contraction, and low P wave velocities (i.e. the areas with thin brittle seismogenic zone) suggests that relatively large deformation of the crust is in progress there, and that it is accompanied by seismic activity and land elevation. The total seismic moment release by earthquakes is too small to cause the crustal deformation of about $10^{-7}$/year estimated from triangulation measurements (Wesnousky et al., 1982), which suggests that a significant portion of the deformation is caused by aseismic or plastic deformation, even in the upper brittle seismogenic zone. The numerous microearthquakes within the seismogenic zone in those areas are presumed to be those accompanying such aseismic deformation. The aseismic deformation would cause land elevation, which may, at least partially, be the mechanism of the mountain building seen in Fig. 4. An exception for this is the Kitakami Mountain Range in the forearc side, where seismic activity and active faults are extremely scarce, suggesting that a different mechanism is working there. Huang et al. (1997) conducted a dynamic modeling using a two-dimensional finite-element method and inferred that the Kitakami Mountain Range is elevated by the strong plate coupling along the main thrust zone of the plate boundary just east of it beneath the Pacific Ocean.

Active faults are not distributed within the low-velocity areas but along the edge of or outside them (Fig. 3). This indicates that large earth-
quakes associated with surface traces of faults do not occur in the brittle seismogenic zone of the low-velocity areas, although microearthquakes occur very frequently there. Perhaps, owing to the high temperature, those areas become too weak in their mechanical strength to generate large earthquakes, and so the stress is released mainly by aseismic deformation. On the contrary, in the surrounding areas the materials are strong enough to sustain large stress concentrations so that the large earthquakes occur that produce faults that reach the Earth’s surface.

In NE Japan, distinct S wave reflectors have been detected in the midcrust below the brittle seismogenic zone at five locations (Mizoue et al., 1982; Horiuchi et al., 1988; Iwase et al., 1989; Hasegawa et al., 1991; Hori and Hasegawa, 1991; Matsumoto and Hasegawa, 1996, Matsumoto and Hasegawa, 1997). Their locations are shown in Fig. 3, together with those of low-frequency microearthquakes. All these midcrustal S wave reflectors are located in or around the P wave low-velocity areas beneath active volcanoes (Hasegawa et al., 1991). They generate reflected S waves with anomalously large amplitudes, which can be explained by a large velocity contrast across a discontinuity underlain by low-rigidity material, similar to the 'bright spot' detected beneath the Rio Grande Rift near Socorro, New Mexico (Sanford et al., 1973; Brown et al., 1980; Brocher, 1981; Ake and Sanford, 1988).

Matsumoto and Hasegawa (1996) used observed spectral ratios of reflected to direct S waves at stations near Nikko-Shirane volcano, the southern end of NE Japan, and found that the midcrustal S wave reflector body has an areal extent of 15 km x 15 km, becomes shallow toward the volcano, and has a thickness of only about 100 m and a very low rigidity. This indicates that the reflector is a thin magma body and perhaps acts as the temporary storage site of magma at midcrustal levels.

Recently, similar S wave reflectors have been detected in the midcrust below the brittle seismogenic zone at five additional locations in NE Japan (squares in Fig. 3), all having very large reflection coefficients like the previous ones (Kono et al., 1997; Umino, 1997). However, the locations of these reflectors are slightly removed from any active volcanoes, compared with the reflectors previously found. They are located below the focal areas of the 1896 Rikuu (M7.2), 1970 Akita-ken Nantobu (M6.2) and 1962 Miyagi-ken Hokubu (M6.5) earthquakes. This leads to a suspicion that the material within these reflector bodies is high-temperature magma, similar to the case of the reflector body beneath the Nikko-Shirane volcano. Yet, anomalously large reflection coefficients, especially for S waves, require that the interior of the reflector bodies is filled with fluid-like material. An alternative candidate for it is water in the state of a super-critical fluid.

Although the newly detected reflector bodies are not close to any active volcanoes, they are still located in or around the Pn low-velocity areas in the same manner as the previous ones. This suggests that the material forming the reflector bodies, whether it is magma or super-critical fluid, is supplied mainly from the upper mantle below, and that the reflectors are not uniformly distributed in space at midcrustal levels. It seems that they are more prevalent in or around the low-velocity areas where the base of the brittle seismogenic zone is locally elevated. We infer that, under the horizontally compressional stress field in this subduction zone, the contractile deformation in the lower crust below the seismogenic zone occurs more easily in those low-velocity areas. The fluid- or magma-filled reflectors, together with a lower strength and a locally thin brittle seismogenic zone above them, will allow more contractile deformation in the low-velocity areas, resulting in crustal shortening, land elevation, and mountain ranges, as schematically shown in Fig. 6. Magmatic underplating or intrusion may have contributed not only to the crustal weakening but also to the formation of the topographic highs.

Exceptionally deep, low-frequency microearthquakes have been detected in the lowermost crust and the uppermost mantle beneath this vol-
canic arc, as mentioned above. They occur well below the base of the brittle seismogenic zone and at depths where rheological properties of rocks are in the ductile regime. Hasegawa and Yamamoto (1994) pointed out that these anomalous events may be generated by the magmatic activity of mantle diapirs in the mantle wedge. Similarly to the S wave reflectors, the low-frequency events are also located in or around the P wave low-velocity areas. Actually, these low-frequency events have been found at levels of the lowermost crust or the uppermost mantle right under many of the midcrustal reflectors detected to date, suggesting that the main source supplying the materials, such as super-critical fluid or magma into the midcrustal reflector bodies, originates in the mantle diapirs in the uppermost mantle.

Fig. 6 is a presently inferred across-arc vertical cross-section schematically showing the seismic activity and the crustal deformation in this arc based on the observations as described above. The P wave low-velocity and low-Q zones in the uppermost mantle, which are inclined toward the backarc side, perhaps correspond to ascending flow of subduction-induced convection in the mantle wedge (Hasegawa et al., 1991). The migration of mass in the ascending flow causes upwelling of hot mantle material, and consequently produces low-velocity, low-Q zones, which are the manifestation of mantle diapirs. We suppose that materials further rising from the mantle diapirs, such as magma or slab-derived water, make their appearance as deep, low-frequency microearthquakes at levels of the lowermost crust or the uppermost mantle, and as distinct S wave reflector bodies (bright reflective layers) at midcrustal levels. Their upward migration raises the temperature and reduces the seismic wave velocity of crustal materials around them, causing the brittle seismogenic zone above them to become locally thinner and weaker.

Subject to the horizontally compressional stress field in the plate convergence direction, the contractile deformation will take place mainly in the low-velocity, low-Q areas because of the thinner brittle seismogenic zone, and the weaker upper and lower crust due to the higher temperature there. The deformation proceeds partly in small earthquakes, but probably mainly in plastic deforma-

mation, causing the crustal shortening, upheaval and mountain building there. In those areas, sufficient stress cannot be stored to generate large earthquakes; they occur within the seismogenic zone along the edge of the low-velocity areas or outside them.

We have assumed simply that low velocities are caused mainly by higher temperature. As mentioned above, other factors also cause low velocities. The presence of volatiles, such as water, reduces the velocity and the effective normal stress, which might elevate the stick-slip to stable-sliding transition depth (e.g. Scholz, 1990). If this is the case, low-velocity areas also have a thin seismogenic zone. Heat flow data is critical to discriminate, but unfortunately the number and coverage of measurement points of heat flow is very limited. Therefore, we cannot compare the presently inferred temperature variation with heat flow data for the whole area shown in Fig. 2. Although limited, heat flow measurements were made at several points in the northern part of the area. The measurements show locally high heat flow values amounting to 160–200 mW/m² in the northern volcanic area in Fig. 2 with locally elevated isotherms [see fig. 3 in Nagao and Uyeda (1995)]. This, in part, seems to support the present assumption that the seismic velocity perturbation in the crust is mainly governed by the temperature perturbation.

Actually, seismic low velocities obtained by travel time tomography must be caused by a coupled effect of the many causes mentioned above, although the main cause seems to be a higher temperature in the present case. High-resolution seismic tomography studies for Vp/Vs and anelasticity structure would provide important information for discriminating the effect of volatiles from that of temperature anomaly.

6. Concluding remarks

We have shown that seismic velocity perturbations reflect temperature perturbations within the crust. A close correlation between the seismic velocity structure, seismic activity, active faults and other observations is also found, which sug-
suggests that earthquake occurrence and deformation processes in the crust are governed, to a considerable degree, by the horizontally inhomogeneous distribution of temperature in the crust.

Imaging the seismic velocity structure of the crust with a much higher spatial resolution, as well as much denser measurements of heat flow and GPS observations, will provide important information for a deeper understanding of the processes of earthquake occurrence and aseismic deformation of the crust within the overriding plate. An intensified seismic observation has been carried out by the Japanese university groups since October 1997 in an area of $60 \text{ km} \times 100 \text{ km}$ in NE Japan, including the Ou Backbone Range where seismic velocity is low and contractile deformation is considered to be large. In that area, 50 temporary seismic stations and seven GPS stations have been deployed and the data from each of the stations have been collected continuously by using a satellite communication system together with other stationary stations. Seismic refraction and reflection profiling experiments are also made across the arc by the Research Group for Explosion Seismology and other groups in Japan. By analyzing the new data collected by this joint seismic observation, we may gain a much better understanding of the processes actually occurring in the NE Japan arc.

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References

Ake, J.P., Sanford, A.R., 1988. New evidence for the existence and internal structure of a thin layer of magma at mid-


Zhao, D., Horiuchi, S., Hasegawa, A., 1990. 3-D seismic velocity structure of the crust and the uppermost mantle in the northeastern Japan arc. Tectonophysics 181, 135–149.