High strain rate zone in central Honshu resulting from the viscosity heterogeneities in the crust and mantle

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Abstract

A linear zone with high strain rates along the backarc side of central Honshu (the so-called Niigata–Kobe Tectonic Zone) has been revealed by the dense GPS array (GEONET) operated by the Geographical Survey Institute of Japan. In order to explore the origin of this zone, we examine rheological heterogeneities in the crust or mantle above a subducting slab, which possibly contribute to the high strain rates. We calculate velocity profiles at the surface of the upper plate due to loading and unloading of the subducting Philippine Sea plate using a 2-dimensional finite element method. From the numerical experiments, we find that the high strain rates can be reproduced by viscosity heterogeneities either in the crust or the mantle. If viscosity heterogeneity in the crust is an origin, the conditions to be satisfied are (i) an effective elastic thickness of 5 km for short-term loading ($10^2$–$10^3$ yr), (ii) a low viscosity ($<10^{19}$ Pa s) in the lower crust beneath the zone, with a normal viscosity ($>10^{21}$ Pa s) in the ambient lower crust, and (iii) a uniform viscosity of $10^{19}$ Pa s in the upper mantle. On the other hand, if viscosity heterogeneity in the mantle is an origin, the conditions to be satisfied are (i) an effective elastic thickness of ~30 km for short-term loading, (ii) a low viscosity (~$10^{18}$ Pa s) at least down to a depth of 60 km in the zone and to a depth of 10 km further seaward below the Moho depth, and (iii) a viscosity of $10^{20}$ Pa s in the ambient mantle. The crust heterogeneity model contradicts the 15-km cut-off depth of intraplate seismicity and the ~20-km effective elastic thickness previously inferred from gravity and topography in central Honshu. The fact that Quaternary active faults and intraplate earthquakes are not particularly concentrated in the zone also favours the mantle viscosity heterogeneity model. Thus, we prefer the mantle heterogeneity model. We suggest that upward movement of water dehydrated from the subducting Philippine Sea and Pacific plates, partial melting of the mantle above the Pacific plate, and serpentinization in the wedge mantle above the Philippine Sea plate are possible origins of the low viscous
upper mantle. Possible low viscosities in the upper mantle wedge presented in this study seem to provide important constraints for subduction processes of oceanic plates and strain accumulation processes in the upper plate.

1. Introduction

The dense Global Positioning System (GPS) array developed by the Geographical Survey Institute of Japan (GEONET) has revealed the presence of a high strain rate zone northwest of central Honshu [1]. This high strain rate zone, ~500 km long in the NE–SW direction and ~100 km wide (Fig. 1), suffers contraction in the WNW–ESE direction (~10^{-7}/yr), which is a few times larger than in the surrounding regions [1]. Because this zone has been called the Niigata–Kobe Tectonic Zone, we use NKTZ to denote it, even if it might not be a Quaternary tectonic feature as shown below. Fig. 1 also shows the tectonic elements where NKTZ is present. Beneath central Honshu, the Philippine Sea plate is subducting in the northwest direction along the Nankai and Sagami Troughs. Further to the east, the Pacific plate is subducting beneath the northern Honshu (the Okhotsk plate) and the Philippine Sea plate along the Japan and Bonin Trenches, respectively.

Several previous studies have attempted to explore the origin of NKTZ. Hashimoto and Jackson [2] detected the high strain rate zone from analyses of traditional geodetic data, such as triangulation, prior to the deployment of GPS. On the basis of the block-fault model proposed by Matsu’ura et al. [3], they interpreted the high strain rate zone by means of slip deficit on the faults at block boundaries. Since continuous GPS data became available, a sharp transition in velocity vectors (e.g., with respect to Eurasia) was noticed along NKTZ after removing the interplate locking effects at subduction zones, and some studies [4,5] proposed NKTZ to be a plate boundary, such as a boundary between the Eurasian and Okhotsk (North American) plates.

However, Quaternary crustal deformation in the Japanese islands does not evidently indicate a feature like a plate boundary [6–11]. For example, active faults are not particularly concentrated in this zone (see Fig. 1 of [6] and Fig. 3 of [8]). Historical intraplate earthquakes also do not show concentration in this zone (see Fig. 1 of [6] and Fig. 4 of [8]). Iio et al. [12] regarded NKTZ as an intraplate deformation zone, and proposed a qualitative model for the origin of NKTZ. In their model, the lower crust beneath NKTZ has a much lower viscosity than that in the ambient lower crust; they believe that this is caused by dehydration from the subducting Pacific plate. They also noted that the observed GPS velocity vectors are fairly uniform seaward of NKTZ, and explained this by a decoupling zone in the wedge mantle above the Philippine Sea plate, in addition to the weakened lower crust below NKTZ.

Hyodo and Hirahara [13] quantitatively investigated the model proposed by Iio et al. [12] using a 3-dimensional viscoelastic finite element model, in which the elastic crust having a thickness of 30 km is floating over the viscoelastic asthenospheric mantle, and the upper plate is loaded by interseismic locking and unlocking of the Pacific plate at the southernmost Japan Trench. They showed that the high strain rates in the zone can be realized by either of the two models with locally weakened crust: (1) the crust beneath NKTZ has an elastic thickness of 15 km and 50% reduction in rigidity and (2) the crust beneath NKTZ has an elastic thickness of 5 km. They preferred the former model, because the 5-km elastic thickness of the crust is too small compared with the 15-km cut-off depth of intraplate seismicity in this region [14,15]. The fairly uniform GPS velocities seaward of NKTZ are obtained in their model by introducing a viscoelastic mantle with a viscosity of ~10^{19} Pa s and a long loading interval between interplate earthquakes (~1000 yr). Note also that Wang et al. [16] showed that the viscoelastic mantle can produce more uniform deformation of the forearc in the Cascadia subduction zone than the elastic mantle.

Although the models of Hyodo and Hirahara [13] explain the high strain rates nicely, we note that there
remain some difficulties. The refraction studies in this region do not indicate any such large reduction in rigidity beneath NKTZ [17] that does not favour the preferred model. Furthermore, the thin elastic thickness beneath NKTZ in both models would enhance the concentration of permanent surface deformation during the Quaternary, and might not be favourable if we take into account the fact that Quaternary active faults and intraplate seismicity are not particularly concentrated in NKTZ.

Although NKTZ seems to be important for revealing strain accumulation processes in the crust of subduction zones, its origin has not, therefore, been elucidated yet. In order to examine the origins of such remarkable features of surface tectonic deformation, rheological structures in the whole Earth should be taken into account. Rheological heterogeneities in the wedge mantle in a subduction zone, not only in the crust, might also be a possible factor contributing to surface crustal movements. Many petrologic and numerical studies have discussed the possible existence of mantle wedges weakened by the water dehydrated from the subducting slabs [18–21]. For example, from dynamic modelling of the Tonga–
Kermadec subduction zone, Billen and Gurnis [19] suggested that the viscosity in the wedge mantle is possibly more than 10 times smaller than in the surrounding mantle. In addition, James et al. [22] estimated a mantle viscosity value ranging from $5 \times 10^{18}$ to $5 \times 10^{19}$ in the Cascadia subduction zone from the post-glacial rebound data. If there is a low-viscosity zone in the mantle wedge beneath a limited region, it would affect surface crustal deformation due to loading and unloading by a subducting plate, and might realize high strain rates in a narrow zone such as NKTZ.

In this study, we quantitatively evaluate the effects of viscosity heterogeneities, not only in the crust but also in the upper mantle on the interseismic deformation field associated with subduction using a 2-dimensional finite element model. Savage [23] proposed a kinematic model to represent crustal deformation in a subduction zone. In this model, interseismic locking is described by virtual normal faulting (back-slip or slip-deficit) on the main thrust zone superposed on a steady-state aseismic slip. The steady-state aseismic slip is assumed to produce no observable deformation. The accumulated back-slip is released completely at the time of an earthquake at the thrust zone and the cycle repeats. Based on this modelling of loading and unloading, we calculate the surface velocity profiles of the upper plate in a 2-dimensional space, and compare them with the observed GPS velocities to examine what kind of rheological structures in the crust and mantle are consistent with the observed data.

Hyodo and Hirahara [13] assumed 50% locking at the thrust in the southernmost Japan Trench in their finite element model. The 50% coupling and the repeat time of 1000 yr assumed by them produces a coseismic slip of 40 m, which seems to be too large noting that the seismic coupling in the southern part of the Japan Trench is generally weak [24,25] and seismic slip of large earthquakes amounts to a few meters at most.

In this study, we assume that the high strain rate zone is produced by subduction of the Philippine Sea plate along the Suruga and Sagami Troughs (see Fig. 1). Obviously this seems to be an oversimplification in such an area located west of an arc–arc junction, and loading by the Pacific plate might also affect crustal deformation in central–southwest Japan, as shown by Hyodo and Hirahara [13]. In particular, the deformation in the northern part of NKTZ might be affected by the Pacific plate subduction. The area also might be affected by the interaction between the Eurasian and Okhotsk (North American) plates (see Fig. 1). However, the aim of this study is to make clear the importance of the viscosity heterogeneities in the mantle wedge above a subducting slab and not to reproduce exactly the observed surface deformation. We, therefore, neglect the above complexities. We await future studies for more complete treatments in a 3-dimensional scheme.

2. Model descriptions

A vertical section from the Tokai district to the Japan Sea directed N70°W along the long side of the rectangle in Fig. 1 is constructed using a finite element model (see Fig. 2). The surface velocities relative to the Eurasian plate at the GPS stations (1997 to 1999, GEONET) in the rectangle area projected on this section shown in Iio et al. [12] are used for the comparison with the numerical calculations. The origin of the distance corresponds to line O–O' in Fig. 1. Although the direction of the section is perpendicular to the Japan Trench, it is also perpendicular to the Suruga Trough (Fig. 1). Because the direction of the Philippine Sea plate subduction differs only by 17° from that of the Pacific plate subduction (Fig. 1), the comparison seems eligible. The size of the whole model is 200 km in depth and 5400 km in horizontal distance, and the total numbers of elements and nodes are 6180 and 6417, respectively. We employ the finite element code TEKTON [26] to solve the mechanical equilibrium equation under the following conditions; at the left and right boundaries, horizontal displacements are zero, vertical displacements at the bottom are zero, and the surface is stress-free. We assume that the thickness of the subducting Philippine Sea plate is 80 km [27] and the dip angle changes from 20° to 30° at a depth of 80 km.

The succession of back-slip and following sudden forward slip at the thrust zone due to the earthquake cycles is incorporated into the model using the split node technique [28]. In this method, the fault displacement or the displacement increment is converted into equivalent nodal force, but the sum of
nodal forces applied to the system is zero. The backslip rate is estimated from the convergence velocity of the subducting Philippine Sea plate and a depth-dependent coupling ratio (i.e., a ratio of the backslip to the convergence velocity).

Comparing the vertical surface movements of the peninsulas of SW Japan during the interseismic and coseismic periods with those calculated from dislocation models in the Nankai subduction zone, Hyndman et al. [29] indicated that the extent of coupling of the main thrust zone is controlled by the temperature on the plate interface. The coupling ratio is mostly zero from the surface down to a depth corresponding to 150 °C, and it is almost full from there down to a depth corresponding to 350 °C, with a transition zone extended to the depth at 450 °C. Following this, we estimate the depth-dependent coupling ratio on the thrust zone from the thermal structure beneath the Tokai district calculated by Seno and Yamasaki [30]. In their thermal structure, temperatures of 150 °C, 350 °C and 450 °C are located at depths of about 5 km, 30 km and 45 km, respectively. We then assume that the coupling ratio is full at depths of between 5 and 35 km and zero at depths shallower than 5 km and deeper than 35 km. This is consistent with the estimation of the interplate coupling in the Suruga Trough based on the inversion analysis of the vertical GPS data [31], in which the interface is fully coupled at depths of between 6 and 30 km.

We assume that the convergence velocity of the Philippine Sea plate is 40 mm/yr [32]. Some studies [5,6,33] propose that there is a microplate east of the Suruga Trough (Izu microplate), which might reduce the convergence velocity to half of the above value. However, such a large reduction is not seen in the GPS velocities around the Suruga Trough (see Fig. 3 of Mazzotti et al. [34]). We also confirmed that such a small convergence velocity cannot explain the observed GPS velocities in NKTZ, even if viscosity structures are modified. Accordingly, backslip with a rate of 40 mm/yr is assigned along the plate boundary at depths of between 5 and 35 km. Along the Nankai–Suruga Troughs, great earthquakes have periodically occurred every 100–200 yr, associated with subduction of the Philippine Sea plate [35]. Therefore, we assume the repeat time of the earthquakes, i.e., the
interseismic loading period, to be 150 yr. After 10 cycles of loading and unloading, we obtain steady-state solutions. For a comparison with the observation, we use the result 140 yr after the previous earthquake in the tenth cycle, which roughly corresponds to the observation time of GPS since the previous large interplate earthquake in 1855 in the Suruga Trough.

Fig. 2 also shows the rheological structure used in this study. The thickness of the continental crust is 30 km [36]. The subducting Philippine Sea plate is assumed to be purely elastic. The elastic layer on the continent has a thickness $T_e$, and is underlain by the Maxwell viscoelastic lower crust and mantle. The values of elastic properties of each layer adopted in this study are listed in Table 1. We construct two models with viscosity heterogeneities either in the crust or in the mantle. The viscosities of the standard normal lower crust and mantle are denoted by $\eta_{lc}$ and $\eta_{lm}$, respectively. In the model with viscosity heterogeneities in the crust, a low-viscosity zone in the ductile lower crust is located beneath NKTZ at horizontal distances of between $-290$ and $-190$ km west of the Suruga Trough, whose viscosity is denoted by $\eta_{wlc}$.

In the model with viscosity heterogeneities in the wedge mantle (see Fig. 2), a 30-km-thick upper zone with a viscosity of $\eta_{wm1}$ above a 40-km-thick lower zone with a viscosity of $\eta_{wm4}$ are located in the wedge mantle beneath NKTZ. The wedge mantle seaward of NKTZ is also divided into two layers: a 10-km-thick upper zone with a viscosity of $\eta_{wm2}$ above a 20-km-thick lower zone with a viscosity of $\eta_{wm3}$. We obtain viscosity values of these zones and an elastic thickness that explain the observed GPS velocities, and discuss them in the later section.

### Table 1

<table>
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<th>Young’s modulus (Pa)</th>
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<td>$3.3 \times 10^{10}$</td>
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<tr>
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<td>$1.91 \times 10^{11}$</td>
<td>0.258</td>
</tr>
<tr>
<td>Sea plate</td>
<td></td>
<td></td>
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</tbody>
</table>

All values are the same as in Suito and Hirahara [49].

### 3. Numerical results

#### 3.1. Influence of viscosity heterogeneities in the ductile lower crust

Rheological heterogeneities only in the crust are taken into account in the following models. Fig. 3 shows a comparison of the observed surface velocities (small circles) with calculated ones (lines) as a function of distance from the trench. The observed velocities at distances between $-40$ and $30$ km are not used for comparison because these include stations in Izu Peninsula on the Philippine Sea plate (see Fig. 1).

Figs. 3(a) and (b) show the predicted surface velocity profiles as a function of $\eta_{lc}$ for the models with $T_e=15$ and 5 km, respectively. Adopted $\eta_{wlc}$ is $10^{18}$ Pa s, and the whole upper mantle has a uniform viscosity of $10^{19}$ Pa s. As can be seen in these figures, the models with higher $\eta_{lc}$ result in larger velocities, and the behaviour is insensitive to $T_e$. The predictions of the models with $T_e=15$ km are not consistent with the observed velocity profile. However, when $T_e=5$ km, the surface velocities obtained by the models with $\eta_{lc}$ more than $10^{21}$ Pa s are consistent with the observations; i.e., a high strain rate zone with a steeper velocity gradient than in the surrounding regions is reproduced. In contrast, the velocities predicted by the models with $\eta_{lc}$ less than $10^{20}$ Pa s are significantly smaller than the observations.

Fig. 3(c) shows the effects of $\eta_{wlc}$, $\eta_{lc}$ and $\eta_{lm}$ on the predicted surface velocity profile. Even though the viscosity $\eta_{wlc}$ is $10^{19}$ Pa s, the predictions of the models with $\eta_{lc}$ more than $10^{21}$ Pa s are consistent with the observations. However, the models with $\eta_{wlc}$ more than $10^{20}$ Pa s have a flatter velocity profile than that observed even if a significant viscosity contrast exists between $\eta_{wlc}$ and $\eta_{lc}$. In addition, the model with $\eta_{lm}=10^{20}$ Pa s predicts significantly smaller surface velocities than the observations.

#### 3.2. Influence of viscosity heterogeneities in the mantle

The viscosity heterogeneities in the wedge mantle contributing to the surface velocity profile are examined here. Fig. 4 shows a comparison between the observed and calculated surface velocities as a function of the distance from the trough axis. Fig. 4(a)
Fig. 3. Comparison of surface velocity profiles predicted by the models with viscosity heterogeneity in the lower crust with the observed velocities. Observed eastward velocities in the rectangle depicted in Fig. 1 are plotted by the open circles. Origin of the distance is the solid line O-O in Fig. 1. Dependence of the surface velocity profile on the viscosity of the lower crust $\eta_{lc}$ obtained by the model with (a) $T_e=15$ km and (b) $T_e=5$ km. (c) Dependence of the surface velocity profile on the viscosity of the weakened lower crust $\eta_{wlc}$ and upper mantle $\eta_m$ obtained by the model with $T_e=5$ km, where the viscosity structures are indicated in the figure.
Fig. 4. Comparison of surface velocity profiles predicted by the model with viscosity heterogeneity in the upper mantle with the observed velocities. Thickness of elastic layer $T_e$ is 30 km. (a) Dependence on the viscosity $\eta_m$ of the upper mantle with the localized low-viscosity zone in the wedge mantle ($\eta_{wm1}=10^{18}$ Pa s) beneath NKTZ. Surface velocity profiles without the localized weak zone are also plotted by the dashed lines. (b) Dependence on the viscosity $\eta_{wm2}$ ($=\eta_{wm3}$) of seaward wedge mantle with the localized low-viscosity zone in the wedge mantle ($\eta_{wm1}=10^{18}$ Pa s) beneath NKTZ. (c) Dependence on the viscosity $\eta_{wm2}$ of the seaward wedge mantle with the localized low viscosity zone in the wedge mantle ($\eta_{wm1}=10^{18}$ Pa s) beneath NKTZ.
shows the predicted surface velocities, where $T_e$ is 30 km, $\eta_{wm1}$ is $10^{18}$ Pa s, and the viscosity of the other wedge mantle ($\eta_{wm2-4}$) are equal to $\eta_m$. The high-velocity gradient in NKTZ is not realized in these models. For the model with $\eta_m=10^{18}$ Pa s, the predicted surface velocities are almost constant (about 3 cm/yr) except near the trough. When $\eta_m=10^{19}$ Pa s, the predicted surface velocities at distances between $-400$ and $-230$ km are significantly higher than the observations, but the predictions almost match the observations at distances between $-230$ and $-40$ km. On the contrary, the predicted surface velocities obtained by the models with $\eta_m$ larger than $10^{20}$ Pa s are apparently smaller than the observations. The surface velocities for the models with $\eta_m=10^{19}$ and $10^{20}$ Pa s excluding the localized low-viscosity zone are depicted by the dashed curves, indicating that the localized low-viscosity zone increases the surface velocity over a wider area than its width.

Fig. 4(b) shows the predicted surface velocity profile under the condition that the whole seaward wedge mantle is uniformly weakened ($\eta_{wm3} = \eta_{wm2}$). $T_e$ is 30 km, $\eta_m$ and $\eta_{wm1}$ are $10^{20}$ and $10^{18}$ Pa s, respectively, and $\eta_{wm4}$ is equal to $\eta_m$. As can be seen in the figure, the surface velocity profile is dependent on the viscosity of the seaward wedge mantle. The predicted surface velocities are significantly smaller than the observations, when $\eta_{wm3}$ ($=\eta_{wm2}$) is more
than $10^{19}$ Pa s. Although the surface velocity profiles obtained by the models with $\eta_{wm3}$ ($=\eta_{wm2}$) less than $5\times10^{18}$ seem to be closer to the observations, the feature of the high strain rate zone is not so clearly realized.

Fig. 4(c) shows the predicted surface velocity profile, in which the dependence on the viscosity of the uppermost seaward wedge mantle ($\eta_{wm2}$) is examined. As can be seen in the figure, in the model with $\eta_{wm2}=10^{18}$ Pa s, the high strain rate zone is modestly realized. When $\eta_{wm2}$ is more than $5\times10^{18}$ Pa s, the surface velocities in NKTZ become smaller than the observations. Although we do not show the sensitivity to $\eta_{wm3}$, it is noted that as $\eta_{wm3}$ increases the velocity gradient decreases at distances between $-150$ and $-100$ km, and becomes steeper at distances between $-260$ and $-170$ km, which conforms closer to the observations. In addition, from several trial calculations, we confirm that the surface velocity profile is almost insensitive to $\eta_{wm4}$ (the viscosity of the lowest part beneath NKTZ).

Surface velocities might be dependent on the elastic layer thickness $T_e$. Fig. 5(a) and (b) show the surface velocity profiles as a function of $\eta_{lc}$ for models with $T_e=5$ and 15 km, respectively. The ductile lower crust has a uniform viscosity $\eta_{lc}$. Mantle viscosities are such that $\eta_{wm1}=\eta_{wm2}=10^{18}$ Pa s, $\eta_{wm3}=10^{19}$ Pa s, and $\eta_{wm4}=\eta_m=10^{20}$ Pa s. For the model with $T_e=5$ km, if $\eta_{lc}$ is less than $10^{20}$ Pa s, the predicted velocities are significantly smaller than the observations. On the contrary, if $\eta_{lc}$ is more than $10^{21}$ Pa s, the predicted velocities can mostly explain the observations. Such model behaviour is regardless of $T_e$ (see also Fig. 5(b)). Because the relaxation time for a viscosity of more than $10^{21}$ Pa s ($\tau \geq 500$ yr) is longer than the loading period (~150 yr), this implies that $T_e$ for short-term loading should be ~30 km to explain the observed velocity profile.

4. Discussion

4.1. Rheological heterogeneity in the crust for explaining NKTZ

If the observed velocity profile is explained by viscosity heterogeneity only in the crust, $T_e$ must be ~5 km, and there should be a low-viscosity lower crust beneath NKTZ. This is simply because a uniform thickness for $T_e$ larger than 15 km raises the lower crust viscosity beneath NKTZ, which fails to explain the high strain rates. The result that $\eta_{lc}$ is larger than $10^{21}$ Pa s in the normal lower crust outside NKTZ implies that $T_e$ is ~30 km outside NKTZ for short-term loading. This result is thus similar to the latter model obtained by Hyodo and Hirahara [13].

The 5-km thickness of $T_e$ beneath NKTZ is not consistent with the 15-km cut-off depth of intraplate seismicity in NKTZ [14,15] because the effective elastic thickness is slightly greater to the seismogenic thickness. Furthermore, the effective elastic thickness in central Honshu obtained by the correlation between gravity and topography is about 20 km [37]. Noting that the time scale of topographic loading is longer than $10^2$ to $10^3$ yr, this thickness contradicts the 5-km-thick $T_e$ beneath NKTZ. The thin $T_e$ is neither favoured if the Quaternary deformation and intraplate seismicity are not particularly concentrated in NKTZ, as discussed before.

4.2. Rheological heterogeneity in the mantle for explaining NKTZ

In the models with viscosity heterogeneities only in the mantle, a low-viscosity zone localized only beneath NKTZ does not realize the observed high strain rates. However, if a low-viscosity (~$10^{18}$ Pa s) layer of about 10-km thickness also exists at the uppermost part in the seaward wedge mantle, the predicted surface velocity profile almost matches the observation (Fig. 4(c)). The viscosity in the lower part of the seaward wedge mantle ($\eta_{wm3}$) is preferred to be larger than $5\times10^{18}$ Pa s to obtain a high strain rate zone. In addition, we show that the viscosity in the wedge mantle at depths >60 km ($\eta_{wm4}$) does not affect the surface velocity profile, which means that it is enough for the wedge mantle to be weakened at least to a depth of 60 km to realize a high strain rate zone at the surface. The necessity of a low-viscosity zone at the uppermost part in the seaward wedge mantle was qualitatively suggested by Iio et al. [12] as mentioned before, although a weakened lower crust beneath NKTZ was assumed in their model.

Hybrid models with viscosity heterogeneities both in the crust and the upper mantle are possible. We have conducted calculations for such models, but do
not show the results because the origin of NKTZ becomes obscured.

4.3. Viscosity of the non-weakened lower crust

The rheology of the ductile lower crust may be important to constrain models for surface deformation. For both the lower crust and mantle heterogeneity models, the viscosity of the non-weakened lower crust should be larger than $10^{21}$ Pa s. In spite of its importance, the viscosity of the lower crust has not been yet known well. Nakada et al. [38] investigated the convective coupling between the ductile lower crust and the upper mantle induced by the upwelling of hot mantle. From the time scale of the sedimentary basin formation in northwest Kyushu, Japan, they estimated the lower crustal viscosity to be $10^{19}$–$10^{20}$ Pa s. A similar lower crustal viscosity has been obtained beneath the Basin and Range [39,40]. However, these low values were obtained in very hot regions. It is well known that the temperature beneath the Basin and Rage is abnormally high. The geotherm beneath northwest Kyushu was estimated using a pyroxene geothermometer, and the temperature just below the Moho possibly reaches almost 1000 °C [38]. It is not, therefore, surprising that the lower crust in such hot regions has very low viscosities, because the viscosity of rocks is highly sensitive to temperature. On the contrary, Masek et al. [41] constrained the viscosity of the lower crust to be less than $10^{22}$ Pa s to explain the rift flank uplifts in Tibet, and Allemand et al. [42] estimated it to be about $10^{21}$ Pa s from the flow induced by the sediment loading in the Paris basin. The crust beneath these areas is not so hot [43], and these viscosity values are consistent with that estimated in this study.

The viscosity in the crust and/or mantle can also be estimated from the ductile strength profile for an assumed strain rate using the rheological parameters derived from the laboratory experiments. We show the temperature and viscosity distributions in the crust and mantle estimated using such a method in Fig. 6. The temperature profile is calculated by solving a 1-dimensional steady-state equation for heat conduction with a surface heat flow of 65 mW/m², which is the average value observed in the Tokai district [44,45]. Depicted viscosities in each layer are based on dislocation creep with rheological parameters of wet quartzite [46] for the upper crust, anorthite [47] for the lower crust, and wet olivine [48] for the mantle, where the thicknesses of the upper and lower crust are both 15 km. As can be seen in the figure, whereas the viscosity in the crust is dependent on the assumed strain rate, it is mostly larger than $10^{21}$ Pa s, which is consistent with the value estimated in this study. Thus, it is inferred that the viscosity of the lower crust in central Honshu is more than $10^{21}$ Pa s, and the lower crust is mostly elastic for short-term loading.

4.4. Origin of the weakened mantle

In our preferred model, we estimated a viscosity of $10^{18}$ Pa s in the uppermost part of the mantle wedge beneath NKTZ and seaward of it. This value is even less than the value of $10^{19}$–$10^{20}$ Pa s for the ambient normal mantle. The latter value is similar to $10^{19}$ Pa s obtained from the postseismic deformation associated with the 1896 Riku-u earthquake, N. Honshu, by Suito and Hirahara [49] and Thatcher et al. [50], and
from the sedimentary basin formation in northwest Kyushu by Nakada et al. [38]. On the other hand, the upper mantle viscosity inferred from the post-glacial rebound is $10^{20} - 10^{21}$ Pa s [51,52], which is by one order larger. This might be because the estimation of viscosity from the post-glacial rebound is usually done in a tectonically less-active region. In fact, James et al. [22] obtained $\eta_m$ of $10^{18} - 10^{19}$ Pa s in the Cascadia subduction zone even from the post-glacial rebound.

The low viscosities of the upper mantle in subduction zones may be attributed to partial melting and/or presence of dehydrated water. As has been discussed in a number of studies, the dehydrated water from the subducting slab would induce partial melting of the wedge mantle [18,53], thus reducing the viscosity of mantle rocks [54]. The viscosity of rocks also significantly decreases with an increase in the water content [21,48]. A low-viscosity zone in the mantle wedge may appear due to an upward movement of water dehydrated from the serpentine within or above the subducting plate. However, in the study region where the Pacific plate is subducting, this effect would be overwhelmed by the effect of partial melting discussed above.

The lower viscosity in the uppermost zone seaward of NKTZ may be partly caused by serpentinitization of peridotite in the forearc mantle wedge. Serpentinitized peridotite can be characterized by low seismic velocity and high Poisson’s ratio [55–60]. Kamiya and Kobayashi [55,56] detected a high Poisson’s ratio beneath the Kanto and Tokai regions (Fig. 7), and interpreted its origin to be serpentinitization in the wedge mantle at depths of between 30 and 45 km above the subducting Philippine Sea plate. Experimental results indicated that the frictional strength of serpentinitized peridotite is significantly low [61–65]. In addition to its low frictional strength, the serpentinitized peridotite might have a lower viscosity than the normal peridotite [66–69], although sufficient experiments have not been conducted.

Another special reason for the low viscosity, pertinent to the southwestern part of NKTZ where no volcanoes exist, is subduction and dehydration of the serpentinitized mantle within the subducting Philippine Sea plate [70,71]. Although the dehydration

![Fig. 7. Tectonic elements in central Honshu–Izu islands region (modified from [73]). The Philippine Sea plate is subducting along the Sagami and Nankai Troughs with velocities of 27 mm/yr and 40 mm/yr, respectively [32,72]. Izu Peninsula is colliding with central Honshu. The contours show the upper surface of the intraslab seismicity [74]. The fault planes are those of the Taisho and Genroku Kanto earthquakes [75], the Tokai earthquake [76]. The possibly serpentinitized wedge mantle revealed by the seismic tomography [55,56] is shaded in green. The red triangles indicate active volcanoes. NKTZ is indicated by the bold shaded line.](image-url)
loci in the oceanic crust of the subducting young Philippine Sea plate are generally shallower than 50 km, the down-dip edge of the serpentine stable area within or above the slab reaches a depth of 80 km beneath southwest Japan south of NKTZ, and possible dehydration of the serpentinized mantle within the subducting Philippine Sea plate [70, 71] might contribute to weakening the mantle wedge beneath the non-volcanic southwestern part of NKTZ.

5. Conclusions

In this study, we explore rheological heterogeneities in the crust or mantle beneath central Honshu, Japan, to explain the origin of a high strain rate zone (Niigata–Kobe Tectonic Zone, NKTZ) detected by the dense GPS network. We investigate interseismic surface deformation of the upper plate associated with the subduction of the Philippine Sea plate beneath central Honshu using a 2-dimensional finite element model, with various viscosity structures in the crust or the mantle.

In order to explain NKTZ by viscosity heterogeneities in the crust, the following condition is required: elastic layer thickness $T_e$ is 5 km and lower crust beneath NKTZ has a viscosity less than $10^{19}$ Pa s, the ambient normal lower crust has a viscosity larger than $10^{21}$ Pa s, and the mantle below has a uniform viscosity of $10^{19}$ Pa s. On the other hand, in order to explain NKTZ by viscosity heterogeneities in the mantle, viscosities of the uppermost mantle beneath and seaward of NKTZ must decrease to $\sim 10^{18}$ Pa s with a normal mantle viscosity of $10^{20}$ Pa s, and effective elastic thickness $T_e$ should be $\sim 30$ km.

We prefer the mantle heterogeneity model for explaining NKTZ because the crust heterogeneity model with an elastic thickness of 5 km is inconsistent with the 15-km cut-off depth of intraplate seismicity in NKTZ and the 20-km effective elastic thickness inferred from the gravity and topography data in central Honshu.

In our preferred model, a regional difference in the Maxwell relaxation time between weakened and non-weakened mantle is responsible for NKTZ. This implies that NKTZ is a feature having a relaxation time of $10^2$–$10^3$ yr, and is not directly related to Quaternary crustal deformation, which has a longer time scale. We suggest that the reduction in viscosity of the uppermost mantle wedge might be caused by serpentinization above the subducting Philippine Sea slab, presence of water and generation of melts by dehydration from the subducting Pacific plate, and addition of water into the mantle wedge beneath southwest Japan due to dehydration of the Philippine Sea plate. This study provides important constraints for subduction processes of oceanic plates and strain accumulation processes in the upper plate.

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References

35. M. Ando, Source mechanisms and tectonic significance of historical earthquakes along the Nankai Trough, Japan, Tectonophysics 27 (1975) 119–140.


