Temporal change of interplate coupling in northeastern Japan during 1995–2002 estimated from continuous GPS observations

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SUMMARY

Temporal change of deformation in northeastern Japan is clarified by continuous Global Positioning System (GPS) observations from 1995 April to 2002 March. The observed GPS velocity is approximately parallel to the direction of plate convergence on east and west plate boundaries of northeastern Japan and shows post-seismic transient deformation around source regions of the $M_w$ 7.8 1993 Hokkaido-Nansei-Oki and the $M_w$ 7.7 1994 Sanriku-Haruka-Oki earthquakes. We interpret the source of the observed deformation as contemporary interplate coupling on the east subducting boundary to the Pacific Plate and the west collision boundary to the Amurian Plate. Using elastic dislocation theory, we inverted horizontal and vertical velocities of 212 GPS stations to estimate interplate coupling on both boundaries. The estimated coupling during 1995–2002 is spatially heterogeneous, however, it is temporally almost constant except for the region around the 1993 and 1994 earthquakes. After-slip of the 1994 earthquake occurred over the coseismic rupture area and its downdip extension on the plate boundary for 0.3–1.3 yr after the earthquake. After-slip continued only in the downdip extension for later periods and decayed with time. Weak coupling was recovered in the eastern part of the coseismic rupture area 3.3 yr after the earthquake. Interplate coupling on the Pacific Plate was strong in two regions, Miyagi-Oki and Tokachi-Oki. The west plate boundary is tightly coupled except for the source areas of three large earthquakes that occurred in 1964, 1983 and 1993. The apparent decoupling of the source areas of these earthquakes implies long-term post-seismic deformation as a result of viscoelastic relaxation in the subseismogenic lithosphere.

Key words: crustal deformation, fault slip, Global Positioning System (GPS), northeastern Japan, subduction zone.

1 INTRODUCTION

The northeastern Japan island arc is a subduction zone in which seismicity is very high. One of the densest networks of geophysical observations, including seismometers and Global Positioning System (GPS), in the world is continuously monitoring seismicity and crustal deformation there. The exact boundaries of tectonic plates around northeastern Japan are still being debated, however, it is widely accepted that northeastern Japan belongs to the North American (or Okhotsk) Plate (Fig. 1). The Pacific Plate is subducting westwards beneath northeastern Japan at a rate of 8–9 cm yr$^{-1}$ along the Japan trench (DeMets et al. 1990; DeMets et al. 1994; Seno et al. 1996). The Amurian (or Eurasian) Plate is converging eastwards on the west side of Japan at a rate of 1–2 cm yr$^{-1}$ (Seno et al. 1996; Heki et al. 1999). For the interseismic period of interplate earthquakes on both plate boundaries, the northeastern Japan arc is compressed by surrounding plates from both the east and the west.

Recently, many studies have used GPS measurements to estimate the contemporary interplate coupling, or degree of plate boundary locking, on subduction interfaces in the world (e.g. Sagiya 1999;
McCaffrey et al. 2000; Darby & Beavan 2001). Most studies assume a steady rate of deformation at GPS stations during measurement periods because of limitations of GPS campaign data. However, transient deformations associated with slow earthquakes and aseismic slip have been detected by continuous GPS observations in subduction zones (e.g. Ozawa et al. 2001; Dragert et al. 2001). Also, in northeastern Japan, coseismic and post-seismic transient deformation associated with the $M_w$ 7.7 1994 Sanriku-Haruka-Oki earthquake was observed in GPS data (Heki et al. 1997; Nishimura et al. 2000). Continuous GPS observations in northeastern Japan are helping to clarify the temporal deformation pattern in the subduction zone during post-seismic and interseismic periods of the earthquake cycle.

Some studies have estimated the distribution of interplate coupling and after-slip of the 1994 Sanriku-Haruka-Oki earthquake on the subducting Pacific Plate (Ito et al. 2000; Nishimura et al. 2000; Mazzotti et al. 2000). Nishimura et al. (2000) concluded that the region of after-slip expanded southwards and westwards on the plate boundary from the coseismic rupture region. However, they studied for only a year after the earthquake temporal changes of distribution of after-slip and coupling distribution. Mazzotti et al. (2000) estimated the annual instantaneous coupling ratio on the Pacific Plate from 1995 to 1997, however, they did not estimate after-slip distribution of the 1994 earthquake. No studies considered the deformation caused by the plate boundary on the west side of northeastern Japan. A number of large earthquakes have occurred on both the east and west plate boundaries of northeastern Japan in the recent past. The $M_w$ 7.8 1993 Hokkaido-Nansei-Oki earthquake occurred southwest offshore of Hokkaido and generated devastating tsunamis that caused over 200 deaths. The $M_w$ 7.7 1994 Sanriku-Haruka-Oki earthquake occurred on the subducting plate boundary east of northern Tohoku and caused severe loss of lives and extensive damage to buildings. Earthquakes of ca $M_w$ 7.5 have occurred roughly every 40 yr on the plate boundary in the Miyagi-Oki region, east of central Tohoku (latitude is 38–38.5° N). The last $M_w$ 7.4 event occurred in 1978 (Seno et al. 1980). The Headquarters for Earthquake Research Promotion of the Japanese government estimated the probability of a $M_w$ 7.5 earthquake before 2020 to be 80 per cent in the Miyagi-Oki region (report is available at http://www.jishin.go.jp/main/index-e.html in Japanese). The locked part of the subduction interface leads to the accumulation of elastic strain that will be ruptured by subsequent earthquakes. Therefore, estimation of interplate coupling...
and its temporal change is important for assessing seismic potential in Japan. In this paper, we present the annual velocity at continuous GPS stations in northeastern Japan for each year from 1995 April to 2002 March. Then, we estimate the distribution of interplate coupling and after-slip on the two plate boundaries east and west of northeastern Japan. We discuss the temporal change of the estimated coupling and after-slip distribution following the 1994 Sanriku-Haruka-Oki earthquake and the downdip limit of interplate coupling and interplate earthquakes on the subducting Pacific Plate. We use moment magnitude denoted by \( M_w \) as magnitude of earthquake, if it is available. Otherwise, we use body and surface wave magnitude determined by the Japan Meteorological Agency and denoted as \( M \).

## 2 GPS Data and Velocity Estimation

### 2.1 GPS data and baseline analysis

We use the data of the nationwide continuous GPS network, named GPS Earth Observation NETwork (GEONET), operated by the Geographical Survey Institute of Japan (GSI) (Miyazaki et al. 1998; Sagiya et al. 2000). GEONET in northeastern Japan commenced in 1994 October. In the study area, the number of stations was 33 as of 1994 October. GSI installed 97, 54, and 16 stations in 1996, 1997 and 1998, respectively (Fig. 1). The GPS phase data of GEONET after the end of 1996 March are processed by Bernese 4.2 software (Astronomical Institute 1996) to estimate daily coordinates on a routine basis (Hatanaka et al. 2003). The reference frame of the estimated coordinates is International Terrestrial Reference Frame 1997 (ITRF97).

Tohoku University and Hirosaki University started regional continuous GPS observations in 1994, independently. We use data of eight GPS stations of Tohoku University and four GPS stations of Hirosaki University from 1995 April to 1996 March. These GPS stations are shown as open squares in Fig. 1(b). Because GPS baseline analyses are processed by each institute independently and the analysis procedures differ, it is difficult to combine coordinates of GPS stations estimated by their routine analyses. Therefore, we simultaneously processed GPS data of GSI, Tohoku University, and Hirosaki University before 1996 March with GIPSY 2.6.1 software using the precise point positioning technique (Zumberge et al. 1997). We resolved ambiguities by the double-differencing of solutions after precise point positioning and estimated daily coordinates of GPS stations referred to ITRF97. Because the strategy is different between the GSI’s routine analysis with Bernese 4.2 software and our analysis with GIPSY 2.6.1 software, we found offsets in estimated coordinates. Daily coordinates from 1996 March 25 to 31 were estimated with both software. We corrected the offset by subtracting the difference of average coordinates between the Bernese 4.2 and GIPSY 2.6.1 solutions for this period. Fig. 2 shows three [N–S, E–W, up–down] components of daily solutions in the selected baselines whose end station is 940049. The locations of baseline GPS stations are indicated in Fig. 1(b).

### 2.2 Estimation of site velocity

We fitted a piecewise straight line to each component of relative coordinates to estimate annual horizontal and vertical site velocities relative to station 940049 from 1995 April to 2002 March. Each line segment is one year long from April to March of the next year because additional GPS stations commenced observations in April. We used bi-weighted robust estimation (e.g. Nakagawa & Koyanagi 1982) to eliminate effects of outlier data. Fitting of piecewise lines makes the variation of the estimated annual velocities smaller than when we fit individual lines for yearly GPS time-series. In addition, we modelled steps of relative coordinates in time-series by using step functions. The steps are associated with earthquakes, volcanic activity, and GPS antenna and receiver replacements. We found significant displacement associated with an earthquake swarm off Matsumae in 1995 October, four M 5-class earthquakes in Onikoube in 1996 August (Nakamura et al. 1999), a swarm and a M 6.1 earthquake near Iwate volcano in 1998 (Miura et al. 2000; Nishimura et al. 2001a) and the eruption of Usu Volcano in 2000 March (Nakada 2001) (Fig. 1b). There were artificial steps caused by replacements of GPS antennas and receivers at 16 stations in 2001 November and December. The fitted lines including steps are shown in grey in Fig. 2. We tried to remove abrupt displacements resulting from seismic and volcanic events in the time-series, however, it is difficult to remove transient deformation continuing for more than a month, such as Iwate and Usu volcanoes activities.

The standard deviation of the velocities assuming a white-noise model ranges from 0.11 to 0.92 mm yr\(^{-1}\) for horizontal components and from 0.54 to 3.59 mm yr\(^{-1}\) for vertical components. Studies using GPS time-series (e.g. Langbein & Johnson 1997; Zhang et al. 1997) show that uncertainties based on a white-noise model are underestimated by 2–6 times as a result of long-term correlation in GPS time-series, compared with time-correlated-noise models. Following Langbein & Johnson (1997), we assumed the uncertainties as \( \sqrt{R W^2 + WN^2} \), where RW is a random-walk contribution and WN is a white-noise contribution estimated above. We assumed uncertainties of random-walk noise as 1 mm/\(\sqrt{\text{yr}}\) for horizontal data and 3 mm/\(\sqrt{\text{yr}}\) for vertical data. The annual horizontal velocities and uncertainties are shown in Fig. 3.

### 2.3 Main features of annual velocities

The estimated horizontal velocity field shown in Fig. 3 is relative to the 940049 station on the coast of the sea of Japan. We did not translate and rotate this velocity field relative to any stable rigid plate, for example, one of the three rigid plates to which northeastern Japan may belong, the North American Plate, the Okhotsk Plate and an independent micro-plate (e.g. Seno et al. 1996; Heki et al. 1999). We neglect a component of the rigid rotation that is contained by the observed relative velocity because the rigid rotation of the North American Plate (Argus & Gordon 1991; Sella et al. 2002) suggests that the relative velocity with respect to 940049 is less than 1 mm yr\(^{-1}\) at all stations used in this study. We chose this approach because our concern is the regional velocity field in northeastern Japan and its temporal change.

The most distinctive feature of the estimated velocities in Fig. 3 is westward movement on the Pacific ocean side. The direction of westward vectors is approximately parallel to the relative plate motion between the Pacific Plate and the overriding continental plate based on plate models such as NUVEL-1 (DeMets et al. 1990) and REVEL (Sella et al. 2002), suggesting interplate coupling between them. Though large westward vectors were observed in southern Hokkaido and southern Tohoku in the period 1995–2002, eastward vectors were observed in northern Tohoku (39.5°–41.5°N) from 1995 April to 1997 March (Figs 3a and b). These transient velocity vectors in northern Tohoku represent post-seismic deformation from the 1994 Sanriku-Haruka-Oki earthquake (Heki et al. 1997; Nishimura et al. 2000). In northern Kanto (36.5°–37°N), westward...
vectors are smaller than those in southern Tohoku in the period 1995–2002, especially from 2000 April to 2001 March. The deceleration of the deformation may be associated with the intense volcanic activity in the northern Izu islands from 2000 June to August, (Nishimura et al. 2001b). On west side of Japan, we found eastward vectors in southernmost Tohoku, northern Tohoku and central Hokkaido during the period 1995–2002. These vectors also suggest interplate coupling on the plate boundary west of northeastern
Japan. However, the direction of vectors in southwestern Hokkaido (42°–43°N, 139°–141°E) gradually changed from west to east with time. A radial pattern around Iwate volcano from 1998 April to 1999 March (Fig. 3d) suggests inflation of a magma chamber (Nishimura et al. 2001a). A more complex pattern around Usu volcano from 2000 April to 2002 March (Figs 3f and g) implies magma movements. These patterns could not be removed by the step-function procedure.

2.4 Vertical deformation

Fig. 4 shows vertical velocities from 1999 April to 2000 March. The observed velocities at half of the stations are significantly larger than

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Figure 3. Horizontal annual velocity fields relative to 940049 indicated as Ref. (a) 1995 April to 1996 March. (b) 1996 April to 1997 March. (c) 1997 April to 1998 March. (d) 1998 April to 1999 March. (e) 1999 April to 2000 March. (f) 2000 April to 2001 March. (g) 2001 April to 2002 March.
the observation uncertainties. It is difficult to discern a regional character because of spatial incoherence even over short distances. We suspect that the vertical velocities are heavily contaminated by local disturbances such as monument instability and subsidence by water pumping. Nevertheless, we notice a distinctive uplift of southwestern Hokkaido. This uplift is supported by levelling surveys in 1993 and 1998 (Geographical Survey Institute 1999). A relative subsidence in the Kitakami basin (38.5–40°N, 141°E) in central Tohoku is observed, which is consistent with subsidence noted by El-Fiky et al. (1997) from levelling data for 1978–1995.

3 COUPLING DISTRIBUTION ON THE PLATE BOUNDARIES

3.1 Inversion method

We interpret the main source of the observed velocity in northeastern Japan as contemporary coupling between the Pacific Plate subducting at the Japan trench and the overriding plate. In addition, high strain rates observed by GEONET along the west coast of Japan (Sagiya et al. 2000) are attributed to plate convergence there (Nakamura 1983; Kobayashi 1983; Seno et al. 1996). The convergence rate at the plate boundary along the west side of northeastern Japan is about one-fifth of that at the Japan trench. However, the convergence may cause comparable displacements at the GPS stations because the plate boundary in the sea of Japan is much closer to the land than the Japan trench. We, therefore, allow coupling on both plate boundaries.

We use a geodetic inversion method developed by Yabuki & Matsu’ura (1992). Their method combines deformation data with a priori information about the smoothness of fault-slip distribution using Bayes’ theorem to construct a flexible Bayesian model with hyperparameters. The hyperparameters control the weight between the observed data and a priori information. The numerical value of the hyperparameters is objectively determined by minimizing Akaike’s Bayesian information criterion (ABIC) proposed by Akaike (1980). We assume a homogeneous elastic half-space to calculate surface displacements induced by fault slip (Okada 1985).

We use the concept of back-slip proposed by Savage (1983) to calculate surface deformation resulting from locking on the subducting plate boundary. The strain state of full locking on a part of the subducting plate boundary is mathematically equivalent to that resulting from superposition of the nominal normal slip (back-slip) opposite to the plate convergence rate on the locked part and uniform sliding over the whole plate boundary. Here, we neglect the deformation resulting from the uniform sliding and estimate the magnitude of slip deficit with respect to plate relative motion by inversion of the observed deformation. In an actual case, the uniform sliding should produce some long-wavelength deformation because the asthenosphere is approximately viscoelastic and the subducting plate does not travel to infinite depth. However, Matsu’ura & Sato (1989) showed that the rate of long-wavelength deformation resulting from the uniform sliding is much smaller than that resulting from the interplate coupling. Therefore, we expect that the back-slip model is a sufficiently good approximation to explain the observed velocities. We estimated the slip distribution not in the whole region of the plate boundary, however, in the region shallower than ca 120 km.

The shape and location of the plate boundary between the subducting Pacific Plate and northeastern Japan are constrained by hypocentre distribution of large earthquakes and microearthquakes (Nishizawa et al. 1992; Umino et al. 1995; Hino et al. 1996; Igarashi et al. 2001). The surface formed by the hypocentres dips at 5° or less near the Japan Trench and dips at 30° to the west of 143°E in the region east of central Tohoku. Fig. 5 shows a smoothed surface passing through the hypocentres. The curved surface of the plate boundary
plates along the Japan trench is about 65 cm yr\(^{-1}\) of relative plate motion between the North American and Pacific plates. GPS observations are 73 ± 15 cm yr\(^{-1}\) (Ito et al. 2000) and ca 60 cm yr\(^{-1}\) (Nishimura et al. 2000) east of Tohoku, and 71–73 cm yr\(^{-1}\) east of Tohoku and Hokkaido (Mazzotti et al. 2000). The direction of back-slip should be nearly opposite to the relative plate motion if a single fault accommodates the plate motion. Though many seismological studies show that the slip direction of interplate earthquakes has small-scale variation, the slip direction of interplate earthquakes and after-slip to release the stress accumulated by interplate coupling should be approximately in the same direction as the relative plate motion. Therefore, we fixed the slip direction to 70° NW and estimated only the slip magnitude. The direction is determined by trial and error testing to optimize for small inversion residuals. We allow negative slip magnitude that means sliding is faster than steady plate motion. After-slip and slow earthquakes are two examples.

The western plate boundary is widely accepted as a convergent plate boundary. The Eurasia Plate was once considered to be subducting beneath northeastern Japan (Nakamura 1983; Kobayashi 1983). However, three large thrust-type earthquakes that recently occurred along the plate boundary show complexity beyond a simple subduction configuration. The \(M = 7.5\) 1964 Niigata earthquake has a high-angle, west-dipping fault (Matsushashi et al. 1987). On the other hand, the \(M = 7.8\) 1983 Japan sea earthquake has a shallow-angle, east-dipping fault (Kosuga et al. 1986). The 1993 Hokkaido-Nansei-Oki earthquake has both east-dipping and west-dipping segments (Tanioka et al. 1995). Ohtake (1998) proposed that two plates were colliding on the boundary.

Opening and closing vertical tensile faults are used as an approximation of the locked thrust and normal faults to calculate interseismic deformation (e.g. McClusky et al. 2001). Shimazaki & Zhao (2000) explained the GPS velocities in central Japan by a combination of rigid block motion and opening of a tensile fault on the block boundary. They interpreted the opening rate of the tensile fault as the plate convergence rate on the collision boundary, which is analogous to back-slip on subducting faults (Savage 1983) and strike-slip faults (Matsu’ura et al. 1986). Here, we assume a simple vertical tensile fault whose strike is 8° NE (Fig. 5) because we do not have...
enough information to assign a realistic fault geometry. The fault length and width are chosen to be 780 km and 30 km, respectively. We divided the tensile fault into 26 segments of 30 km length and estimated the opening or closing rate of each segment as part of the inverse solution.

### 3.2 Inversion results

Fig. 6 shows the estimated slip distribution on the western and eastern plate boundaries for seven years. Red regions on the subducting model fault mean the region where back-slip, representing interplate...
coupling, is estimated. No slip means stable sliding at a rate of rigid plate motion, that is, no coupling. Blue regions indicate slip faster than the rigid plate motion, which we call forward-slip. The real slip rate on the plate boundary is the sum of the estimated slip rate and the rigid plate motion. For the western plate boundary, red bars east of the fault indicate the opening rate of the virtual tensile fault. Blue bars west of the fault represent the closing rate of the virtual tensile fault. Opening of the virtual tensile fault is interpreted as the pushing of northeastern Japan at the plate boundary by the Amurian Plate.

The errors of the estimated slip depend heavily on the fault location for all inversion periods from 1995 to 2002. The errors of the estimated slip on the subducting plate boundary are typically ca 4 cm yr\(^{-1}\) at the northern and southern edges of the model fault, 1–2 cm yr\(^{-1}\) in the offshore region and ca 1 cm yr\(^{-1}\) under land. Slip errors on the virtual tensile fault are ca 4 cm yr\(^{-1}\) in the northernmost segment and 0.5–2 cm yr\(^{-1}\) in the remaining part. We show three figures associated with the inversion for the period from 1999 April to 2000 March in Fig. 7. Fig. 7(a) shows a comparison between the observed and calculated horizontal velocities. It demonstrates good
agreement. Fig. 7(b) shows the residual horizontal velocities on a scale four times larger than in Fig. 7(a). The directions of residual velocities appear to be almost random. We think that most GPS stations having large residual velocities suffer from local non-tectonic effects. One common characteristic for such stations is large seasonal variations in their time-series. Monument instability from freezing of water in soils, local deformation from groundwater pumping and interception of radio waves by vegetation were reported as the cause of anomalous seasonal variations of GEONET (e.g. Abe et al. 1999). Fig. 7(c) shows calculated vertical velocities. Agreement between observed (Fig. 4) and calculated velocities is very poor because of less weighting and large noise of vertical data. We calculated the mean misfit of the inversion for each data period (Table 1). The misfit is not so different for each data period. The mean misfits for horizontal data are approximately 1.5 mm yr\(^{-1}\), which is a similar value to the observed uncertainties. On the other hand, the misfits for vertical data are approximately 5 mm yr\(^{-1}\) and are larger than the assumed data uncertainties. The poor agreement (Figs 4 and 7c) and large misfits (Table 1) for vertical data imply that vertical velocities estimated for one year do not necessarily show tectonic deformation.

4 DISCUSSION

4.1 Interplate coupling on the subducting plate boundary

The most distinctive temporal change of slip distribution on the subducting Pacific Plate is found at 39.7–42.0\(^{\circ}\)N in Fig. 6. For the period from 1995 April to 1996 March (Fig. 6a), forward-slip is found from the Japan trench to the western edge of the model fault. After that period, the region of forward-slip is limited to the deep western part of the model fault and the slip rate decays with time except for the period from 2001 April to 2002 March (Fig. 6g). The forward-slip is interpreted as after-slip of the 1994 Sanriku-Haruka-Oki earthquake (Heki et al. 1997; Nishimura et al. 2000). The period from 1995 April to 1996 March corresponds to 3 to 14 months after the earthquake and the forward-slip region includes
the coseismic rupture region defined by aftershocks, however, it is much wider. In particular, the forward-slip region extends beyond the downdip (westward) end of the coseismic rupture region by 100 km. The maximum forward-slip is at the western edge of the coseismic rupture. The moment of forward-slip at $39^\circ -42^\circ$ N is $1.5 \times 10^{19}$ N m ($M_w 7.4$) for one year. The forward-slip region in later periods (Figs 6b–f) occupies the western and the northern areas of the coseismic rupture, whose depth is from 40 to 80 km.

There are two regions of local maximum forward-slip at 40.5° N, 142° E and 41.5° N, 143° E in Figs 6(b)–(e). Fig. 8 shows the comparison of estimated slip from 1999 April to 2000 March (Fig. 6e) with the fault segments of seven large ($M > 7.2$) earthquakes after 1950 (Aida 1978; Seno et al. 1980; Kosuga et al. 1986; Matsuhashi et al. 1987; Nagai et al. 2001). The first region is at the downdip extension of the coseismic fault rupture, which is clearly shown in Fig. 8. Large interplate earthquakes in subduction zones such as the 1946 Nanakaido (Thatcher & Rundle 1984), the 1960 Alaska (Cohen 1998), the 1964 Chile (Barrientos et al. 1992) and the 1978 Miyagi-Oki earthquakes (Ueda et al. 2001) were accompanied by downdip post-seismic slip. The maximum forward-slip rate is less than 1.9 cm yr$^{-1}$ for the period from 2000 April to 2001 March (Fig. 6f) and is nearly equal to the estimation error. It suggests that the after-slip became insignificant in 2000, that is, 5 yr after the main shock.

The distribution of forward-slip from 2001 April to 2002 March (Fig. 6g) is different from that for the previous periods (Figs 6b–f). The moment of forward-slip at $39^\circ -42^\circ$ N is $4.3 \times 10^{19}$ N m ($M_w 7.0$) and is three times as large as that for the year following 2000 April. Therefore, it is difficult to consider the forward-slip as continuing after-slip of the 1994 earthquake. We speculate that this forward-slip is after-slip of another earthquake or is a slow earthquake. A $M 6.4$ low-angle thrust-type earthquake whose epicentre is denoted as a star in Fig. 6(g) occurred in the centre of the forward-slip region on 2001 August 14. This is one of three large ($M > 6$) earthquakes that occurred on the subducting plate boundary during the study period. The other two earthquakes occurred in the forward-slip region in 1995 December. It is difficult to detect GPS displacement change associated with the $M 6.4$ earthquake in the time-series of GPS baselines because of noisy daily solutions and seasonal variations. The two Sanriku-Oki earthquakes, one in 1989 ($M 6.9$) and the other in 1992 ($M 7.2$), occurred south of the source region of the 1994 Sanriku-Haruka-Oki earthquake and were followed by prominent strain changes measured by extensometers. Miura et al. (1993) and Kawasaki et al. (1995) suggested that post-seismic deformation was caused by post-seismic after-slip and that the total moment following the earthquakes may be 2 to 16 times larger than the seismic moment of the earthquakes. These observations may imply that it is not rare for the moment release of after-slip to be much larger than that of the earthquake.

Next, we discuss the distribution of back-slip where the hanging-wall plate is dragged by the subducting plate as a result of strong interplate coupling. We found two regions where large back-slip is estimated for all time periods. One has a maximum slip in the Miyagi-Oki region at 38° N, 142.5° E, where the occurrence of an $M 7.5$ earthquake is predicted. The maximum back-slip rate there varies from 8.3 to 10.2 cm yr$^{-1}$ over all periods. Taking the error of the estimated slip rate into consideration, the slip rate is approximately the same as the plate convergence rate of 8.3 cm yr$^{-1}$ between the North America and Pacific plates at the Japan trench predicted by NUVEL-1A (DeMets et al. 1994). The comparable rates imply that the fault between the two plates is completely locked. The contour line of 8 cm yr$^{-1}$ of back-slip rate, which represents nearly full coupling, is located ca 80 km east of the Pacific coast where the plate boundary megathrust is ca 40 km deep. Coupling becomes weaker to the west. The 2 cm yr$^{-1}$ contour line locates beneath land where the interface depth is 60–90 km. The back-slip almost vanishes at a depth of 80–100 km.

The other region has a maximum back-slip rate in Tokachi-Oki at 42.5° N, 144° E. The maximum rate ranges from 12 to 19 cm yr$^{-1}$ at the northern edge of the model fault and it is unrealistically larger than the convergent rate predicted by NUVEL-1A. There are large uncertainties of the estimated slip rate, reaching 5 cm yr$^{-1}$. The slip rate at the southern and the northern edges tends to be larger than expected because we do not consider the interplate coupling outside of model faults. We, therefore, conclude that this maximum back-slip rate is not real and that the interplate coupling is very strong there. The strong coupling regions in Miyagi-Oki and Tokachi-Oki include the source areas of the $M 8.2$ 1954 Tokachi-Oki and the $M 7.4$ 1978 Miyagi-Oki earthquakes. These regions, thus, have high potential for future large earthquakes.

In addition, we point out that the third region of local maximum of back-slip is in the Sanriku-Oki region at 40° N, 143.5° E. Though forward-slip is found in that region for 1995 April to 1996 March

Figure 8. Comparison of source regions of large earthquakes ($M > 7.2$) after 1950 and inversion result for 1999 April to 2000 March. Areas with broken lines are the source regions of the large earthquakes whose names are indicated. The areas of the 1968 Tokachi-Oki and the 1994 Sanriku-Haruka-Oki earthquakes are high slip regions where moment release density is more than $3 \times 10^{16}$ N m cm$^{-3}$ (Nagai et al. 2001). Areas of the other earthquakes are shown by rectangles of their coseismic dislocation models (Aida 1978; Seno et al. 1980; Kosuga et al. 1986; Matsuhashi et al. 1987; Tanioka et al. 1995). Thin dotted line denotes the western limit of the zone where low-angle thrust-type earthquakes occur on the plate boundary (Igarashi et al. 2001). Solid contour lines of the inversion slip rate are for 1999 April to 2000 March, with 2 cm yr$^{-1}$ contour interval (same as in Fig. 6e). Shaded area indicates the region of forward-slip with magnitude $>$2 cm yr$^{-1}$. The length of bars in the sea of Japan indicates the collision rate at the plate boundary.
(Fig. 6a), it changed to back-slip after 1998 April. The change is interpreted to be healing of the fault zone that ruptured during the 1994 Sanriku-Haruka-Oki earthquake. Recent seismological studies (Nakayama & Takeo 1997; Nagai et al. 2001) show that the coseismic slip of the 1994 earthquake was concentrated in a small region, called an asperity, at the centre part of the aftershock region. Furthermore, this asperity overlapped one of two asperities identified for the $M_w$ 8.2 1968 Tokachi-Oki earthquake. Nagai et al. (2001) compared average slip in the asperity of the 1994 earthquake to interseismic slip deficit calculated by the plate convergence rate and concluded that the asperity on the plate boundary was fully locked for the interseismic period between the 1968 and 1994 earthquakes. Therefore, we expect the immediate recovery of interplate coupling after the 1994 earthquake in the asperity if the present earthquake cycle is like the previous one. The estimated maximum back-slip rate is $3 \pm 5$ cm yr$^{-1}$ in the asperity for 1998 April to 2002 March, i.e. ca 50 per cent coupling. However, 100 per cent coupling of the asperity is also permitted because our geodetic inversion does not have enough resolving power to detect this small asperity of size $50 \times 20$ km$^2$ (Nagai et al. 2001).

In the region beneath and offshore of southernmost Tohoku and northern Kanto (south of 37.5°N), the estimated back-slip rate is less than 7 cm yr$^{-1}$ and suggests interplate coupling that is weaker than that of the northern (Miyagi-Oki) region (Fig. 6). The seismic coupling coefficient, which is defined by the ratio of the rate of coseismic slip to the rate of relative plate motion, is 19 per cent and less than 3 per cent, in Miyagi-Oki and in the region east of northern Kanto, respectively (Pacheco et al. 1993). The seismic coupling coefficient is much smaller than the interplate coupling estimated by geodetic observations, which means that most of the accumulated strain is released by aseismic slip such as after-slip and slow aseismic slip (e.g. Heki et al. 1997; Mazzotti et al. 2000). Both coupling coefficients offshore of northern Kanto, which are smaller than those in Miyagi-Oki, imply positive correlation between them. Deep coupling at the southern edge of the model fault for a period from 1996 April to 1999 March (Figs 6b–d) is not significant because of the large estimated uncertainty there.

Miura et al. (2001) found small (ca 8 mm) transient deformation observed only at one GPS station on an offshore natural-gas platform for a few months starting in 2001 February. They suggested that an aseismic slip with $M_w$ 6.6 occurred at 37–37.3°N on the subducting Pacific Plate. Figs 6(f) and (g) suggest that interplate coupling in the region east of northern Kanto, including the fault of the aseismic slip suggested by Miura et al. (2001), was weaker from 2000 April to 2002 March. However, an intense volcanic activity in the northern Izu islands occurred from 2000 June to August. Nishimura et al. (2001b) attribute several mm of the northeastward motion in northern Kanto to the volcanic activity. Thus, it is difficult to judge if apparent coupling weakening is related to aseismic slow slip. Our land-based GPS network may be unable to detect aseismic slip events associated with weak subduction coupling and episodic creep of the kind occurring on the creeping section of the San Andreas fault (Scholz 1990).

Mazzotti et al. (2000) estimated interplate coupling on the subducting Pacific Plate by using GEONET data. They suggested that the subducting plate boundary in northeastern Japan was completely locked from the Japan trench to a depth of 55–70 km except for the region in and around the rupture area of the 1994 Sanriku-Haruka-Oki earthquake and that large scale heterogeneity of interplate coupling does not exist in the study area. Our results differ. The 100 per cent coupling is limited to depths between 20 and 40 km with spatially heterogeneous coupling across and along the Japan trench. The main difference between the data sets is that Mazzotti et al. (2000) used GPS velocities with respect to the Eurasia Plate, whereas we used relative velocities with respect to a GPS station in northeastern Japan. GPS measurements (Heki et al. 1999) and slip vectors of large earthquakes near the west coast of Japan (Seno et al. 1996) support significant westward movement of northeastern Japan with respect to the Eurasia Plate at a rate of 1–2 cm yr$^{-1}$. Referring velocities to the Eurasia plate may, thus, lead to overestimates of interplate coupling. We, therefore, believe that the spatially heterogeneous coupling estimated in this study is more strongly established by our analysis. Lateral variation of interplate coupling is supported by spatial variation of the observed westward velocities along the east coast of northeastern Japan, which is approximately parallel to the plate boundary (Fig. 3).

Fig. 6 shows similar pattern of interplate coupling except for the region around the 1994 Sanriku-Haruka-Oki earthquake and edges of the model fault for each of the seven years. We suspect that the coupling at the deep extension of after-slip beneath central Tohoku in Fig. 6(d) may be an artefact because of deformation caused by the earthquake swarm activity of Iwate volcano. Our result suggests that interplate coupling on the subducting plate boundary is spatially heterogeneous and changes temporally in the area where the recent large earthquake occurred. It also suggests that the intensity of the coupling is almost constant in most areas, at least over the timescale of our measurements.

4.2 Interplate coupling on the boundary in the east rim of the sea of Japan

We found a very similar pattern of slip distribution on the boundary along the western coast of northeastern Japan during all seven years, except for the segment at 42–43°N where a negative rate is estimated. The opening of the virtual tensile fault can be interpreted as the colliding and pushing of northeastern Japan by the Amurian Plate. We interpret that the negative rate of the virtual fault represents thrust creeping in shallow depth. We found four red regions of a high collision rate (>2 cm yr$^{-1}$) representing a tightly coupled plate boundary in Figs 6(b)–(g). The maximum rate is over 20 cm yr$^{-1}$ at the northern edge. The rate at 38.5–40.2°N reaches 7 cm yr$^{-1}$. These rates are unrealistically large because the plate convergent rate on the boundary is expected to be less than 2 cm yr$^{-1}$ according to rigid plate models (Heki et al. 1999; Seno et al. 1996). We think the high rate is caused by our simplified plate geometry and that the actual plate boundary may be closer to land at the segment where the high rate is estimated. The section of the virtual tensile fault where the estimated rate is negative at 42–43°N corresponds to the source area of the 1993 Hokkaido-Nansei-Oki earthquake. The slip rate decayed with time and almost vanished in 2002. It can be interpreted as after-slip of thrust faulting, which is consistent with coseismic slip of the 1993 earthquake.

As shown in Fig. 8, the segment where the collision rate of the boundary is small or negative is restricted to source regions of three large earthquakes, the $M 7.5$ 1964 Niigata, the $M 7.8$ 1983 Japan sea and the 1993 Hokkaido-Nansei-Oki earthquakes. It suggests that the fault strength in the megathrusts of these large earthquakes has not yet returned to interseismic values. Weak strength of these faults contrasts with quick recovery of fault strength after the 1994 Sanriku-Haruka-Oki earthquake. Another possible explanation for negative or small positive collision rate is long-term post-seismic deformation as a result of viscoelastic relaxation. The upper-mantle viscosity beneath northeastern Japan is estimated to be $0.7–1.3 \times 10^{19}$ Pa s (Thatcher et al. 1980; Suito & Hirahara 1999). Theoretical
post-seismic deformation for the three earthquakes resulting from viscoelastic relaxation can roughly cancel out the deformation resulting from plate collision and continues at more than 1 mm yr$^{-1}$ for several decades in the epicentral area (Nishimura 2000). In the case of the 1994 Sanriku-Haruka-Oki earthquake, viscoelastic relaxation is not very important for the first few years after the main shock because theoretical post-seismic deformation rate is at most 2 mm yr$^{-1}$ at the GPS stations and much smaller than the observed deformation rate. However, it is generally difficult to distinguish the deformation resulting from viscoelastic relaxation from that resulting from fault slip (Savage & Prescott 1978). Viscoelastic relaxation is probably the biggest contribution to systematic modelling errors in our inversion results.

4.3 Downdip limit of interplate coupling beneath northeastern Japan

Our inversion results show 100 per cent coupling above a depth of 40 km and 0 per cent coupling at a depth of 80–100 km in Miyagi-Oki, one of the most tightly coupled regions. The interval between 40 km and 80–100 km is a transition region where interplate coupling decreases from 100 to 0 per cent with depth. This downdip limit is deeper than the 60–70 km depths estimated from previous studies using GPS measurements (Ito et al. 2000; Mazzotti et al. 2000; Nishimura et al. 2000). We found that the maximum depth of interplate coupling was ca 50–70 km when our model did not have a plate boundary along the western coast of northeast Japan. Including the western plate boundary makes data fitting significantly better if a larger number of parameters is taken into consideration. Therefore, we believe our result is more accurate. We attribute the deeper extent of interplate coupling compared to other subduction zones (Sagiya 1999; McCaffrey et al. 2000; Darby & Beavan 2001) to the old age and cold temperature of the subducting Pacific slab. Hyndman & Wang (1993) concluded that interplate coupling is controlled by temperature on subduction interfaces and that coupling is weaker above 350°C and vanishes at 450°C. Numerical simulation for a thermal model of northeastern Japan (Wang & Suyehiro 1999; Peacock & Wang 1999; Peacock 2001) shows that the depth on the subduction interface at 350°C and 450°C is 60–70 km and 70–100 km, respectively. The partial coupling to a depth of 80–100 km estimated in this study is, thus, supported by the thermal model of northeastern Japan.

From a seismological viewpoint, the downdip limit of interplate coupling may correspond to the maximum depth where low-angle thrust-type earthquakes occur on the plate interface. Igarashi et al. (2001) studied microearthquakes in the northeastern Japan subduction zone and found the downdip limit is 50–70 km depth (Fig. 8) and shallower than our result. Numerical simulation of fault slip during earthquake cycles in subduction zones using prescribed friction laws (e.g. Stuart 1988; Kato & Hirasawa 1997) shows that during the interseismic period faults lock in both the velocity-weakening region and the adjacent and deeper velocity-strengthening region of the friction law. Simulation suggests that after-slip following the main shock occurs mainly in the velocity-strengthening region, which is the downdip extension of the coseismic rupture region. Therefore, we think that the maximum depth of interplate earthquakes coincides with the downdip limit of the velocity-weakening region and that the maximum depth of interplate coupling for interseismic period can be deeper than that of interplate earthquakes. The 1978 Miyagi-Oki earthquake had after-slip on both the coseismic rupture area and on a downdip extension at a depth of 40–100 km (Ueda et al. 2001).

Our results show that the after-slip of the 1994 Sanriku-Haruka-Oki earthquake continued for approximately 5 yr at the depth of ca 40–80 km. Both after-slip episodes occurred at a depth where the plate boundary is partly locked in Miyagi-Oki. The strain accumulated at this depth may be released mainly by after-slip. The observed after-slip distributions are consistent with slip behaviour at the downdip extension of the velocity-weakening region expected by numerical simulation using the friction law.

5 CONCLUSION

We estimated annual velocities at continuous GPS stations in northeastern Japan from 1995 April to 2002 March and detected temporal changes in horizontal velocity including post-seismic deformation of the 1993 Hokkaido-Nansei-Oki and 1994 Sanriku-Haruka-Oki earthquakes. Assuming elastic dislocation theory, we investigated the annual spatial distribution of interplate coupling and temporal changes on plate boundaries on both eastern and western sides of northeastern Japan through inversion of horizontal and vertical GPS velocities. The main results are:

(i) The after-slip area of the 1994 Sanriku-Haruka-Oki earthquake expanded into the area adjacent to the coseismic rupture area, especially its downdip extension. Approximately 1.3 yr after the main shock, after-slip occurred mainly on the downdip extension of the coseismic rupture area at a depth of ca 40–80 km. The after-slip almost ceased ca 5 yr after the main shock.

(ii) Interplate coupling on the subducting plate boundary is spatially heterogeneous along the Japan trench, however, it does not change significantly over seven years except for the region around the 1994 Sanriku-Haruka-Oki earthquake where the fault changed from completely decoupled to partially coupled with time.

(iii) There are two regions, Miyagi-Oki and Tokachi-Oki, where the subduction interface was tightly coupled over several years. These regions would seem to have a large potential for future earthquakes.

(iv) Interplate coupling in Miyagi-Oki was ca 100 per cent at depth range 20–40 km and 0 per cent at depth range 80–100 km, decreasing with depth. The downdip limit of interplate coupling is deeper than that of interplate thrust microearthquakes.

(v) The western plate boundary was tightly coupled except for the source areas of recent large earthquakes, the 1993 Hokkaido-Nansei-Oki, the 1983 Japan Sea and the 1964 Niigata earthquakes, in the period 1995–2002. Weak coupling of the source areas of large earthquakes suggests an unhealed fault zone after the earthquakes or the effect of long-term post-seismic deformation as a result of viscoelastic relaxation of the subseismogenic lithosphere.

ADDENDUM

A M 8.0 thrust-type earthquake occurred southeast of Hokkaido on 2003 September 26. This earthquake occurred on the part of plate boundary called Tokachi-Oki region where we noted a large potential for great earthquakes in this study. It is probably a recurrence event of the 1952 Tokachi-Oki earthquake (Miura et al. 2004). Considering the occurrence of this event, there is little possibility of the large earthquake in Tokachi-Oki region in the next few decades.

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REFERENCES


APPENDIX: RESOLVING POWER OF SLIP DISTRIBUTION

To examine the resolving power of the inversion method used in this study, we performed numerical experiments by using synthetic data. The configuration of the model faults is the same as those described in Section 3. We checked the resolving power with two kinds of GPS station distributions. One distribution is stations as they existed in 1995 April to 1996 March (Fig. 3a), the other is for 1998 April to 1999 March (Fig. 3d). The number of GPS stations for each period is 47 and 196, respectively. We assigned slip distributions with checkerboard patterns of 120 km² meshes (Fig. A1) and 60 km² meshes (Fig. A2). We then calculated theoretical displacements at GPS stations and added Gaussian random noise whose standard deviations are 1 mm for horizontal data and 3 mm for vertical data. The inversion procedure and conditions are just the same as for actual data. The inverted results of the synthetic data are shown in Figs A1 and A2.

For 120 km² meshes (Fig. A1), the checkerboard pattern on the subducting model fault is fairly well reproduced in the west–two-thirds part of the fault with both GPS station distributions. Because we used a small roughness of slip distribution as a priori information in the inversion, the inverted slip is smoother than the original slip. The magnitude of the estimated slip is larger than the original in the centre of each constant slip square and smaller outward. The reproduction of the virtual tensile fault representing the colliding plate boundary in the sea of Japan is good in both station distributions. In the case of the 60 km² mesh (Fig. A2), the checkerboard pattern on the subducting model fault is fairly well reproduced in the west–two-thirds part of the fault with the 1998–1999 station distribution. The inverted slip with the 1995–1996 station distribution is too smooth to recognize the assigned pattern except in the region.
along the coastline. The inverted slip pattern of the virtual tensile fault under the sea of Japan is in good agreement with the original for both station distributions. We found that large slip was estimated at the northern and southern edge of the model fault where we assigned no slip. This result suggests that the slip near the northern and southern edge is not significant. Slip more than 150 km east from the coastline is poorly resolved with all mesh sizes and station distributions. The numerical test suggests the geodetic data have very limited resolving power for offshore fault slip on the subduction boundary, as also noted by Sagiya & Thatcher (1999).

In summary, the spatial resolution on the subducting fault is about 60 km and 120 km with the GPS stations from 1998 to 1999 and 1995 to 1996, respectively. However, slip located ca 150 km east of the coastline is poorly resolved by both station distributions. Our inversion has a resolving power of at least 60 km wide patches for the virtual tensile fault under the sea of Japan.