Spatiotemporal distribution of interplate coupling in southwest Japan from inversion of geodetic data

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Received 20 December 2002; revised 15 November 2003; accepted 9 December 2003; published 26 February 2004.

[1] We determine the spatiotemporal distribution of slip (or slip deficit) on the subduction interface of the Nankai trough over an entire earthquake cycle using geodetic data (including leveling, triangulation, and trilateration, sea level, and GPS surveys) obtained during the past 100 years in southwest Japan. We develop a new inversion method that accounts for long-term crustal deformation, coseismic (earthquake) displacements, and stress relaxation of the viscoelastic asthenosphere. From this analysis we obtain a model that shows postseismic afterslip on the deeper part of the plate interface following the 1946 Nankaido earthquake. Significant afterslip is found beneath central Shikoku that totals about 0.8 m. The slip deficit rate during the interseismic period is 5–6 cm/year in the N50°/C176°W–N60°/C176°W direction, which is consistent with the relative plate motion between the Philippine and Amurian plates. A fully locked region is found in the shallower portion (<30 km), and the slip deficit rate has a maximum at a depth of about 20 km. The plate interface deeper than 30 km is slowly slipping. The amount of slip deficit reaches about 3.3 m off Shikoku and about 2 m off the Kii peninsula 50 years after the 1946 earthquake. This suggests that fault healing and full interplate coupling has occurred by this time along the Nankai trough. (INDEX TERMS: 1206 Geodesy and Gravity: Crustal movements—interplate (8155); 7209 Seismology: Earthquake dynamics and mechanics; 7260 Seismology: Theory and modeling; 8150 Tectonophysics: Plate boundary—general (3040); KEYWORDS: interplate coupling, Geodetic data, spatiotemporal distribution, viscoelastic asthenosphere, stress relaxation, afterslip


1. Introduction

[2] Southwest Japan is one of the most studied subduction zones in the world. The Philippine plate is subducting in the N55°W direction by 6.3–6.8 cm/year beneath the Amurian plate along the Nankai trough off Shikoku [Miyazaki and Heki, 2001]. It is well known that great interplate earthquakes have occurred repeatedly with intervals about 90–150 years in southwest Japan along the plate boundaries [e.g., Shimazaki and Nakata, 1980; Thatcher, 1984]. Periodic crustal deformation is often observed in association with great earthquake cycles and many models have been used to explain them. For instance, Savage and Prescott [1978], Thatcher [1983], and Cohen and Kamer [1984] proposed earthquake cycle models for transform faults like the San Andreas Fault, and for convergent plate boundaries, Savage [1983] proposed a simple earthquake cycle model that is widely used. The periodic crustal deformation in southwest Japan is explained by Thatcher and Rundle [1979] as fault motion in an elastic lithosphere and viscoelastic asthenosphere. Also, Sato and Matsu'ura [1988] and Matsu'ura and Sato [1989] modeled the stable subduction of a plate at a convergence boundary. However, these studies are based on general models, and there have not been many region-specific studies that estimate the accumulation of tectonic stress from inversion of observed data using viscoelastic models. Although forward modeling of viscoelastic deformation histories have been compared with real data in Thatcher [1984], Miyashita [1987], and Wang et al. [1994, 2001].

[3] The most recent events were the 1944 Tonankai (M7.9) and the 1946 Nankaido (M8.0) earthquakes. There have been many studies of the mechanisms of these two earthquakes. The work of Kanamori [1972] is based on teleseismic data, and the studies of Fitch and Scholz [1971], Ando [1975], Yoshioka et al. [1989], Yabuki and Matsu'ura [1992], and Sagiya and Thatcher [1999] are based on geodetic data. Also, Aida [1981], Ando [1982], Satake [1993], and Tanioka and Satake [2001] use tsunami data to analyze these earthquakes. Although the derived fault parameters should not be directly compared because of...
different sources of information, these studies found that coseismic slip is widely distributed in the region from the Kii Peninsula to Shikoku along the Nankai trough and the slip on the west side of the Kii channel is larger than on the east side.

[4] Previous studies have attempted to constrain interplate coupling by inverting geodetic data [Yoshioka et al., 1994; Ito et al., 1999; Sagiya, 1999; Nishimura et al., 2000; Ito et al., 2000], but the displacement fields in these studies are calculated using dislocation theory in a semi-infinite homogeneous elastic medium. Furthermore, these studies failed to clarify the relationship of the obtained fault slip with the earthquake cycle and are not very useful for the discussion of temporal variations. To construct a model covering the entire earthquake cycle, it is important to incorporate a viscoelastic asthenosphere. The shallow part of the Earth can be modeled with an elastic lithosphere and a viscoelastic asthenosphere. The observed crustal deformation near plate boundaries is due not only to the elastic deformation from earthquakes and interplate coupling, but also to viscoelastic relaxation of the asthenosphere. In order to study the strain accumulation process during the earthquake cycle, it is important to derive spatiotemporal changes in slip or slip deficit on a plate interface. This can be done by an inversion analysis of long-term crustal deformation data that incorporates viscoelastic effect.

[5] It is appropriate to select southwest Japan as the study area, since it has one of the longest histories of geodetic data. Other studies have also attempted to obtain temporal distribution of accumulated slip on the plate interface between the subducting Philippine Sea and the overriding plates, by applying an inversion technique to crustal deformation data (leveling, triangulation and trilateration, sea level, and GPS surveys) of about 100 years (see Figure 1).

2. Formulation of the Inverse Problem

[7] This section explains the formulation of the inverse problem. We attempt to estimate spatiotemporal slip distribution on a plate interface by using the discrete spatiotemporal crustal deformation data on the surface. We assume a medium which consists of an elastic lithosphere and a Maxwell viscoelastic asthenosphere, so that our model contains the effect of viscoelastic relaxation of the asthenosphere.

2.1. Viscoelastic Response to Slip

[8] If the source is given at point \((x, t)\) on the plate interface at time \(t\), we obtain the displacement \(W(x, t)\) at a point on the surface \(x = (x, y)\) at time \(t\) from the integration over time as

\[
W_i(x, t) = \sum_{j=1}^{2} \int_{-\infty}^{t} \int_{-\infty}^{\infty} G_{ij}(x, t; \xi, \tau) u_i(\xi, \tau) d\xi d\tau
\]

Equation (1) represents the viscoelastic response at a surface point at time \(t\) when a unit slip is given on the plate interface \(\Sigma\) at time \(\tau\). We attempt to obtain the spatiotemporal distribution of accumulated slip on the plate interface by using the discrete spatiotemporal crustal deformation data on the surface. We assume a medium which consists of an elastic lithosphere and a Maxwell viscoelastic asthenosphere, so that our model contains the effect of viscoelastic relaxation of the asthenosphere.

Figure 1. Tectonic setting of southwest Japan. The Philippine plate (PH) is subducting beneath the Amurian plate (AM) at the Nankai trough. Squares denote tidal stations (corresponding to Table 2).

[9] In this study, we develop three-dimensional viscoelastic response functions to model deformation in southwest Japan. We attempt to obtain the spatiotemporal distribution of accumulated slip on the plate interface between the subducting Philippine Sea and the overriding plates, by applying an inversion technique to crustal deformation data (leveling, triangulation and trilateration, sea level, and GPS surveys) of about 100 years (see Figure 1).
the plate interface. A more detailed expression for \( G_i^j(x, t; \xi, \tau) \) is shown by Matsu'ura et al. [1981] and Iwasaki and Matsu'ura [1982]. Because the viscoelastic response depends on the history of slip, the lower limit for the integration for time should be \( -\infty \). However, it is sufficient to replace this with the effective relaxation time \( \tau_e \) of the whole system because there is no "memory" of older slip. When the asthenospheric viscosity is assumed to be \( 5 \times 10^{18} \) Pa \( \times \) s, the effective relaxation time \( \tau_e \) of the whole system is about 50 years. This viscosity is estimated from crustal deformation following the 1923 Kanto earthquake [Matsu'ura and Iwasaki, 1983].

### 2.2. Observation Equation

[9] In order to estimate the spatiotemporal distribution of slip rate on a plate interface, the slip rate is expressed as a superposition of basis functions \( T_m(\tau) \) and \( X_\beta(\xi) \) concerning time and space, respectively. The spatial basis functions are bicubic B-spline functions, while the temporal basis functions are the delta function for coseismic slip and B-spline functions for interseismic slip. Details of these basic functions will be given in Section 3.6. Using the spatiotemporal distribution of slip on the plate interface represented by B-spline functions, surface deformation \( d_i \) is expressed as

\[
d_i = \sum_{j=1}^{K} \sum_{k=1}^{L} \sum_{m=1}^{M} H_{ijklm} a_{ijklm} + e_i, \quad (i = 1, \ldots, N),
\]

with

\[
H_{ijklm} = \int_{-\infty}^{t} \int \sum G_i^j(x, t - \tau; \xi, 0) X_\beta(\xi) T_m(\tau) d\xi d\tau.
\]

Here, \( a_{ijklm} \) and \( e_i \) are the coefficients of the superposed basis functions (model parameters), and the random error, respectively. Our purpose is to find the model parameters as the best solution. \( K \) and \( L \) are numbers of B-spline functions in the length and width direction of the model fault, respectively. \( M \) is the number of temporal basis functions. Here, we consider the likelihood function of \( a_{ijklm} \). The observation equation (2) contains errors \( (e) \), which consist of measurement errors and modeling errors. For simplicity we assume the observed data are mutually independent and the errors obey \( N(0, \sigma^2 E) \). Then, we have the probability density function (pdf) of the model parameter \( a \) as

\[
p(d|a; \sigma) = \frac{1}{(2\pi\sigma^2)^{N/2}} \exp \left[ -\frac{1}{2\sigma^2} \left( d - Ha \right)^T E^{-1} \left( d - Ha \right) \right].
\]

where \( ||E|| \) and \( \sigma^2 \) denote the absolute value of the determinant of \( E \) and an unknown scale factor for the variance of \( e_i \), respectively. The information supplied by the observed data is extracted through this likelihood function.

### 2.3. Prior Constraints

[10] We assume the slip rate perturbation \( \Psi u(\xi, \tau) \) on the plate interface is smooth in space and time except for rapid coseismic changes, and we incorporate this assumption as a priori information into our inversion scheme. Specifically, we add a constraint that the spatial Laplacian is nearly zero at each time step:

\[
0 = \Delta \Psi u(\xi, \tau) + e_1,
\]

where \( \Delta \) and \( e_1 \) are the Laplacian operator and Gaussian error, respectively. It is necessary to set boundary conditions in order to obtain the Laplacian at all spatial locations. Decoupling is dominant along shallower and deeper portions of plate interface because of high pore pressure due to the existence of water and low viscosity due to high temperature [Hyndman and Wang, 1993], respectively. Thus we assume the slip rates at the shallow and deep edges of the plate interface are zero. The slip distribution across the eastern boundary is assumed to be continuous along strike, because there should be coupling into the Tokai area. However, the boundary condition on the Kyushu side is assumed to be zero, because decoupling on the plate boundary off Kyushu is suggested by Ito et al. [1999].

[11] The coseismic slip rate is much larger than slip rate during the interseismic period, so that a hyperparameter that controls smoothness for the spatial slip distribution cannot control the huge amounts of slip in the coseismic time step. We separate the hyperparameter \( \alpha_C \) that control smoothness of the spatial slip distribution into two hyperparameters \( (\alpha_C^C, \alpha_C^L) \) that control the smoothness of slip distribution in the coseismic and interseismic time periods respectively.

[12] The second constraint is that changes with time are small in order to represent a smooth temporal change of slip rate at each point:

\[
0 = \Psi u(\xi, \tau - 1) - \Psi u(\xi, \tau) + e_2,
\]

where \( e_2 \) is the Gaussian error. In this study, because of the cyclic condition is introduced, the slip distribution at all time steps can be differentiated except during the coseismic stage.

[13] The third constraint is that the direction of the slip rate is nearly equal to that of the subducting plate (\( \psi \)):\[
0 = \text{CS1} - \text{CS2} + e_3,
\]

\[
\text{CS1} = \sin(\psi_d) \Psi u_1(\xi, \tau) + \cos(\psi_d) \Psi u_2(\xi, \tau),
\]

\[
\text{CS2} = \sin(\psi_d) \Psi u_2(\xi, \tau) + \cos(\psi_d) \Psi u_2(\xi, \tau),
\]

where \( e_3 \) is again the Gaussian error.

[14] Geodetic data have little resolving power of slip distribution on the fault plane outside the network. In subduction zones a large portion of plate interface is usually located under the ocean, but geodetic data are available only on land. Therefore we must add a new hyperparameter that controls the coseismic slip distribution in the shallowest portions of the plate interface. Here we use the slip distribution derived from tsunami and geodetic data by Tanioka and Satake [2001] as a priori information. Prior information is expressed as follows:

\[
0 = \Psi u(\xi_1, 1, \tau_{\text{co}}) - \Psi u_{a\text{pri}}(\xi_1, 1) + e_4,
\]

where \( \Psi u(\xi_1, 1, \tau_{\text{co}}) \) and \( \Psi u_{a\text{pri}}(\xi_1, 1) \) denote the slip distribution along the shallowest portions of the plate.
interface in the coseismic step and a priori slip distribution, respectively.

We can represent the above four constraints of equations (4)–(7) in a vector form as follows, using model parameters $a$:

$$p_a + e_i = 0. \quad (8)$$

where $i = 1C, 1I, 2–4$.

We may represent the above constraints in the form of the pdf of the model parameters $a$ with five hyperparameters $(\alpha_1, \alpha_2, \alpha_3, \alpha_4)$. The constraint equations (8) are not independent. Therefore, according to Fukahata and Mats'ura [2001], the pdf can be written as follows:

$$p(a; \sigma^2) = (2\pi\sigma^2)^{-1/2} \times \prod_{i=1C}^4 \sigma_i^2 p_i P_i \times \exp \left[ -\frac{1}{2\sigma^2} \sum_{i=1C}^4 \sigma_i^2 \parallel P_i a \parallel \right]. \quad (9)$$

where $i = 1C, 1I, 2–4$. \( \alpha_i^2 \) are hyperparameters that control the roughness of spatial, temporal, directions of the slip distribution and a priori slip distribution, respectively.

### 2.4. Bayesian Modeling and ABIC

We estimate the minimum ABIC by the quasi-Newton method ([Davidon-Fletcher-Powell (DFP) method] using a linear minimum technique [Davidon, 1959].

### 3. Data and Model Setting

#### 3.1. Leveling Surveys

In southwest Japan, the Military Land Survey conducted the first leveling surveys from 1886 to 1899, and the surveys have been repeated 7 times since then by its successor, the Geographical Survey Institute (GSI). In total, we have data from 8 leveling surveys in Shikoku, the Kii peninsula and the central Kyushu region, over about 120 years. The average interval between resurveys is about 15 years, but the recent surveys have been conducted more frequently. We use all leveling data from 1886 until 2001 in the inversion analysis.

We use changes in height between consecutive surveys compiled by GSI [Kunimi et al., 2001]. Height differences are measured down to the order of 0.1 mm. The leveling surveys were carried out along roads on which bench marks are located with an interval of approximately 2 km. We select data points with an average spacing of approximately 3 km. The total number of bench marks and leveling data are 523 and 2177, respectively.

The leveling data give the changes in height difference between neighboring bench marks $(x_{i-1}, y_{i-1})$ and $(x_i, y_i)$ measured at time $t_j$ and $t_{j-1}$. If leveling data are written as $d_i(x_i, y_i, x_{i-1}, y_{i-1}; t_j, t_{j-1})$, the equation relating the vertical displacement calculated by equation (1) and the observation data is written as

$$d_i(x_i, y_i, x_{i-1}, y_{i-1}; t_j, t_{j-1}) = W_3(x_i, y_i, t_j) - W_3(x_{i-1}, y_{i-1}, t_j) - W_3(x_i, y_i, t_{j-1}) + W_3(x_{i-1}, y_{i-1}, t_{j-1}), \quad (11)$$

where $W_3(x_i, y_i, t_j)$ is the calculated vertical displacement at a point $(x_i, y_i)$ at time $t_j$. The threshold for closure for a round trip survey was $1.5 \times \sqrt{2S}$ (mm) until 1965 and $2.5 \times \sqrt{S}$ (mm) after 1965. Because the observation error is smaller than the threshold for a round trip survey, we usually use $\sigma = 1.5 \times \sqrt{2S}$ (mm) until 1965 and $\sigma = 2.5 \times \sqrt{S}$ (mm) after 1965 as the error. Figures 2a–2d show the observed spatiotemporal vertical displacement field from these data.

#### 3.2. Triangulation and Trilateration Surveys

The triangulation network in southwest Japan was also established in the late nineteenth century. Control points have been occupied six times and the surveys are called Meiji (1888–1901), Tango–EQ (1928), Showa 1st (1948–1964), Showa 2nd (1968–1972), Precise 1st (1973–1985) and Precise 2nd (1977–1987). The average spacing for the first-order triangulation network is about 50 km.

We use triangulation and trilateration data for which adjustments were made by Hashimoto [1990] and Ishikawa and Hashimoto [1999]. Hashimoto [1990] obtained distance change rates between the first-order control points using the weighted least squares method. The total number of control points and data are 57 and 544, respectively.
Figure 2. Vertical crustal deformation at each time period in southwest Japan from leveling data. Circles and crosses denote uplift and subsidence, respectively. (left) Observed deformation. (right) Differences between the observed displacement and the calculated displacement from the inverted slip distribution. These vertical displacement fields are height differences between each benchmark, calculated from equation (11). The total number of benchmark marks and leveling data are 523 and 2177, respectively. (a, e) Preseismic (<1946) vertical displacement rate. (b, f) Coseismic (1946) vertical displacement rate. (c, g) Postseismic (1946–1971) vertical displacement rate. (d, h) Interseismic (1971–1982) vertical displacement rate.
[23] The change in distance between control points \((x_{1i}, y_{1i})\) and \((x_{2i}, y_{2i})\) at time \(t_j\) and at time \(t_{j-1}\) are used in this inversion analysis. For the data \(d(x_{1i}, y_{1i}, x_{2i}, y_{2i}, t_j, t_{j-1})\), the equation that relates the horizontal displacement calculated by equation (1) and the observation data is written as

\[
\begin{align*}
\Delta R_{ij} &= d(x_{i1}, y_{11}, x_{i2}, y_{i2}, t_j, t_{j-1}) \\
&= \frac{S_X_{ij}}{W_1(x_{1i}, y_{1i}, t_j)} + \frac{S_{Y ij}}{W_2(x_{1i}, y_{1i}, t_j)} \\
&\quad - \frac{S_X_{ij}}{W_1(x_{2i}, y_{2i}, t_j)} - \frac{S_{Y ij}}{W_2(x_{2i}, y_{2i}, t_j)} \\
&\quad - \frac{S_X_{ij}}{W_1(x_{1i}, y_{1i-1}, t_{j-1})} - \frac{S_{Y ij}}{W_2(x_{1i}, y_{1i-1}, t_{j-1})} \\
&\quad + \frac{S_X_{ij}}{W_1(x_{2i}, y_{2i-1}, t_{j-1})} + \frac{S_{Y ij}}{W_2(x_{2i}, y_{2i-1}, t_{j-1}),}
\end{align*}
\]

where \(W_1(x, y, t, t)\) and \(W_2(x, y, t, t)\) are the calculated horizontal displacements in the \(x\) and \(y\) directions, respectively, at a point \((x, y)\) at time \(t_j\).

[24] The errors in the triangulation and trilateration data change with time mainly because the surveys carried out before laser distance meters became available have much larger errors (scale error). The errors of the older surveys are assumed to be \(1 \times 10^{-5}\), while the errors of surveys using laser distance meters are \(2 \times 10^{-6}\) [Hashimoto, 1990]. The errors we used for the inversion analysis are shown in Table 1. Figures 3a–3d show the observed spatiotemporal horizontal displacement field for the triangulation and trilateration data.

### 3.3. Sea Level Observations

[25] Sea level observations in southwest Japan started in the mid 20th century and are still being continued. The locations of tidal stations used in this study are shown in Figure 1. Raw tidal records contain various effects such as ocean tides, ocean currents, atmospheric pressure, and crustal deformation. In order to extract the vertical crustal deformation information from tidal records, we incorporate an improved analysis method based on Kato and Tsumura [1979]. We can remove annual or seasonal variations, such as atmospheric pressure and ocean tides, using the yearly mean sea level records (Figure 4). Hence the yearly mean sea level record can be decomposed into ocean current effects and the crustal deformation. We divide southwest Japan into three regions [Kato and Tsumura, 1979], according to the characteristics of the ocean current effects, as shown in Table 2. Within each region the mean sea levels at the stations fluctuate similarly, although there are also many variations which can be traced throughout all regions. The effects of the ocean currents are expressed as

\[
d_{\text{region}} = H_{\text{region}} + S_{\text{region}} + \epsilon_{\text{region}},
\]

where \(d_{\text{region}}\) and \(S_{\text{region}}\) are the tidal records in a region and the common components due to ocean effects in a region, respectively. We use observation equation (13) for sea level instead of equation (2) and simultaneously estimate the common components \(S_{\text{region}}\). The eustatic seal level change is included in the common components \(S_{\text{region}}\). The total number of tidal stations, data and unknown parameters \(S_{\text{region}}\) are 18, 845 and 167, respectively.

### 3.4. GPS Observations

[26] Recently, the Geographical Survey Institute of Japan completed the installation of a continuous GPS observation network in Japan, which enables us to investigate the recent crustal movements. The GPS Earth Observation Network (GEONET) currently consists of about 1200 permanent sites. We use the results of Hatanaka et al. [2001] who reanalyzed phase center variations of all types of GEONET sites and recalculated all site coordinates.

[27] The GPS coordinates are obtained in the ITRF97. We converted the GPS data into displacements relative to the Amurian (AM) plate using the Euler vector between the Amurian and Philippine plates estimated by Miyazaki and Heki [2001].

[28] We estimated the annual rate of change of the GPS coordinates by a least square method. The modeled time series can be expressed as a linear combination of a constant\((A_1)\), a linear trend \((A_2t)\), and annual changes:

\[
d(t) = A_1 + A_2 t + A_3 \sin(2\pi t/365.25) + A_4 \cos(2\pi t/365.25) \\
+ A_5 \sin[2(2\pi t/365.25)] + A_6 \cos[2(2\pi t/365.25)] + e(t),
\]

where \(t\), \(A\), and \(e(t)\) are time (in days), observed displacements or estimated amplitudes of each term, and residuals, respectively.

[29] The observation period of GPS data used in this study is from April 1996 to December 2001. We use horizontal displacements \((A_2 \times \text{length of observation period})\) at each GPS station in our inversion analysis. Miyazaki and Heki [2001] showed evidence of interarc collision in Japan. Hence we remove data from the GPS stations in the Kinki region where the collision effect may be large.

[30] The vertical component of the GPS data derived from routine analysis is also based on ITRF97. We estimated a correction value from an average vertical displacement velocity in a reference region, which is considered unaffected by the subducting Philippine plate. We obtained –0.347 cm/year as the correction value from 26 GPS
stations, along the Japan Sea coast in the Chugoku region, which are located sufficiently far from the plate boundary. We calculated displacement rates by subtracting this correction from the vertical displacement rate at each GPS station in the study area. We use the vertical displacements (vertical displacement rate × length of observation period) in our inversion analysis. The total number of GPS station and data are 143 and 429, respectively.

[31] The uncertainties of the GPS displacements (one standard deviation of displacement rate × the length of observation period) for each time period in southwest Japan from triangulation and trilateration surveys. (left) Differences between the observed horizontal displacements and the calculated displacements from the inverted slip distribution. These deformation fields are the distances between each control point, calculated from equation (12). The total number of control points and data are 57 and 544, respectively. (a, e) Coseismic (1946) horizontal crustal deformation field. (b, f) Postseismic (1858–1970) horizontal crustal deformation rate. (c, g) Interseismic (1970–1979) horizontal crustal deformation rate. (d, h) Interseismic (1979–1986) horizontal crustal deformation rate.

Figure 3. (right) Horizontal crustal deformation fields for each time period in southwest Japan from triangulation and trilateration surveys. (left) Differences between the observed horizontal displacements and the calculated displacements from the inverted slip distribution. These deformation fields are the distances between each control point, calculated from equation (12). The total number of control points and data are 57 and 544, respectively. (a, e) Coseismic (1946) horizontal crustal deformation field. (b, f) Postseismic (1858–1970) horizontal crustal deformation rate. (c, g) Interseismic (1970–1979) horizontal crustal deformation rate. (d, h) Interseismic (1979–1986) horizontal crustal deformation rate.
observation period) are based on a least squares fit to original time series. Figure 5 (black arrows) shows the observed horizontal and vertical displacement field from the GPS data.

3.5. Structure Model

To calculate viscoelastic response functions, it is necessary to assume an Earth structure. We tested three different Earth structure models. MODEL1 has an elastic layer over an elastic half-space. MODEL2 has two layers over a half-space. The two layers represent the elastic lithosphere and the viscoelastic asthenosphere, and the half-space is the elastic mesosphere. The thickness of surface layer is 30 km, and the viscosity of the asthenosphere is assumed to be $10^{18}$ Pa · s, as estimated from crustal deformation following the 1923 Kanto earthquake [Matsu’ura and Iwasaki, 1983]. MODEL3 has a layer over a half-space. The layer and half-space represent the lithosphere and viscoelastic asthenosphere (viscosity is assumed to be $10^{18}$ Pa · s), respectively. The details of these models are described in Table 3. The shape of the fault in the models follows the real geometry of the subducted Philippine Sea slab as shown in Figure 6.

3.6. Model Setting

The average recurrence time of great earthquakes is about 120 years, but we adopt a recurrence time of 92 years in our model, because the most recent interval between the 1854 Ansei earthquake and 1946 Nankaido earthquake is 92 years. The 1944 Tonankai earthquake occurred outside of the model area. We set the model source period from 1947 to 2039 and assume great earthquakes occur in 1947 and 2039. The time intervals of the observed data (leveling, triangulation and trilateration, sea level and GPS) are from

Figure 4. Yearly mean sea level changes (land uplift taken as positive) at each station. (a) Located in region A. (b) Located in region B. (c) Located in region C. See Table 2.
1883 to 2001, from 1893 to 1986, from 1941 to 2002, and from 1996 to 2001, respectively.

If the system’s response to slip on the plate interface is completely elastic, we need to model slip history only for the observation period. However, we need to consider the slip history before the first observations because of the viscoelastic effects in our model. In order to deduce the slip history from geodetic data during the 92 year period, we introduce the cyclic assumption that the same slip history repeats in each model source period (Figure 7). In this way, geodetic data before 1946 can be treated like those before 2038 and constrain the behavior near the end of the cycle. With this constraint, we have crustal deformation data that covers the whole modeling period (Figure 8). Since we incorporate viscoelasticity into our model, viscoelastic deformation due to any events before 1947 should also be included. However, we think that the 92 years of the model source period is long enough for these past effects to die out, since the effective relaxation time $\tau_e$ of the entire system is about 50 years, as described in the preceding section. We do not assume the cyclic boundary condition that the total forward slip and slip deficit must sum to a particular value.

To represent the temporal distribution of the slip rate, we distribute 21 B-Spline functions of $1/476$ every four years from 1947 to 2039, and delta functions at 1947 and 2039 to

### Table 2. List of Tide Gauge Stations

<table>
<thead>
<tr>
<th>No.</th>
<th>Latitude, deg</th>
<th>Longitude, deg</th>
<th>Station Name</th>
<th>Observation Period</th>
<th>Region</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>34.4853</td>
<td>135.8244</td>
<td>Toba</td>
<td>1950–2002</td>
<td>A</td>
</tr>
<tr>
<td>2</td>
<td>34.0767</td>
<td>136.2075</td>
<td>Owase</td>
<td>1954–2002</td>
<td>A</td>
</tr>
<tr>
<td>3</td>
<td>33.5881</td>
<td>135.8961</td>
<td>Uragami</td>
<td>1950–2002</td>
<td>A</td>
</tr>
<tr>
<td>4</td>
<td>33.4758</td>
<td>135.7733</td>
<td>Kushimoto</td>
<td>1950–2002</td>
<td>B</td>
</tr>
<tr>
<td>5</td>
<td>33.6833</td>
<td>135.3758</td>
<td>Shirahama</td>
<td>1967–2002</td>
<td>B</td>
</tr>
<tr>
<td>6</td>
<td>34.2219</td>
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<td>Hosijima</td>
<td>1941–2002</td>
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*aStation numbers correspond to the locations in Figure 1.*

---

**Figure 5.** Displacement rates at each GPS station calculated from the inverted slip distribution (white arrows) and the observed displacement rates (black arrows). Observed horizontal and vertical displacement rate fields are relative to the Amurian plate and the north part of the Chugoku district, respectively (black arrows). Contour lines denote vertical displacement rates, which are calculated from the inverted slip distribution. Units are in cm/year.
represent the coseismic slip. Figure 7d shows the temporally variable distribution of slip with the cyclic condition. Hence basis functions in the time domain are written as follows:

$$T_1(t) = \delta(t),$$
$$T_{23}(t) = \delta(t - 92),$$
$$T_l(t) = \begin{cases} 
(\tau - \tau_l + 4) & (\tau_l - 4 \leq \tau \leq \tau_l) \\
(4 + \tau_l - \tau) & (\tau_l \leq \tau \leq \tau_l + 4) \\
0 & (\tau \leq \tau_l - 4, \tau \leq \tau_l + 4) 
\end{cases},$$

with

$$\tau_l = 4(l - 1) \ (l = 2, \ldots, 22).$$

The three-dimensional upper surface of the subducting Philippine plate is obtained from the spatial distribution of microearthquakes [Nakamura et al., 1997]. The strike of the model source region is almost parallel to the Nankai trough. We divide the 230 × 500 km spatial model source region into 6 × 12 subsections and distribute 9 × 15 bicubic B-splines so that they homogeneously cover the whole region (Figure 6). The distribution of each slip rate component (strike slip and thrust) in the model source region is expressed by the superposition of the 9 × 15 bicubic B-splines with various amplitudes. The boundary condition for model source region is assumed to be semi-fixed, eliminating the bicubic B-spline functions which express slip distributions that are not consistent with the constraints described in section 2.3. Therefore stress concentrations, which appear near the edge of the model source regions for conventional uniform slip models, are considerably suppressed. Finally the total number of model parameters is \(3168 (\approx 6 \times 12 \times 22 \times 2)\).

[36] Our problem is to determine the optimal values of the 3335 (≈ 3168 + 167) model parameters and the five hyperparameters from 2177 leveling, 544 triangulation and trilateration, 845 tidal gauge and 429 GPS data.

4. A Numerical Experiment

[37] In the preceding section we developed a new method for inversion of geodetic data based on ABIC. Now we examine the validity of this method using a numerical experiment. The synthetic data are inverted to recover the

![Figure 6](image-url)  
**Figure 6.** Model fault region in this study. The isodepth distribution of the upper surface of the subducted Philippine plate is based on Nakamura et al. [1997]. Units are in km.

![Figure 7](image-url)  
**Figure 7.** Schematic of the decomposition of slip history with B-Spline functions. The boundary condition of time is the cyclic condition. (a) Accumulated slip history at a point on the plate interface. (b) Temporally variable slip rate on the plate interface. (c) Amplitude of discrete B-spline functions of 1°. Discrete B-spline functions are used to express the continuous slip rate. The inversion estimates these amplitudes of the spline functions. (d) Distribution in time domain of B-spline functions of 1° and a delta function used to express the continuous slip distribution.
slip model that was used to generate these data. The validity
of our inversion method will be checked by comparing the
inverted slip distribution with the input model slip. In order
to deduce the slip history from geodetic data during the
period spanning the occurrence of a great earthquake, we
introduce the cyclic assumption that the same slip history
repeats in each model source period. We adopt MODEL2
which has the viscoelastic asthenosphere as the structure
model. Details of the model setting and the structure model
are described in the sections 3.6 and 3.5, respectively.

The slip distribution used to generate synthetic data
is shown in Figure 9a. On the plate interface we continu-
ously distribute slip vectors with different magnitudes and
direction (Figures 9a–9c). For this slip (deficit) distribution,
we compute vertical surface displacements at 523 bench
marks, horizontal surface displacements at 57 control
points, vertical surface displacements at 18 tide gauge
stations, and horizontal and vertical surface displacements
at 143 GPS stations. The temporal distribution of the
synthetic data is based on the real observation data set
(see sections 3.1–3.4). We add random noise with zero
mean and standard deviation corresponding to the real
observation data, to these theoretical surface displacements.

We search numerically for hyperparameters which
minimize the ABIC defined in equation (10) by the quasi-
Newton method using the linear minimum method. Since
we use model slip distribution for coseismic slip, the hyper-
parameter \(a_4\) is not included in this inversion. Figures 9d–9f
show the slip distribution obtained by inverting the synthetic
data (Table 4). The coseismic slip distribution and spatio-
temporal slip distribution during the interseismic period are
reproduced very well. Hence the effectiveness of this inver-
sion analysis is confirmed by this test.

To demonstrate the validity of ABIC, we show the
inverted slip distributions for inappropriate choices of the
values of \(a\). We test two cases in which the values of \(a_{ij}^2\)
and \(a_2^2\) are much greater than the optimal values. Figures 10a
and 10b show a case in which the value of \(a_{ij}^2\) is much
greater than its optimal value. Larger values of \(a_{ij}^2\) imply
that the prior constraint on the roughness of the spatial slip
distribution are stronger. In this case the subpeak existing in
the model slip distribution disappears, because the prior
constraint is too strong. On the other hand, Figures 10c and
10d show a case in which the value of \(a_2^2\) is much greater
than its optimal value. Larger values of \(a_2^2\) imply that the
prior constraint on the roughness of temporal slip distribu-
tion are stronger. In this case there are few temporal slip
changes.

5. Results of Inversion
5.1. Results With Elastic Response Function

Before using the viscoelastic response function, we
invert the data using an elastic Earth model (MODEL1). We
use the slip distribution derived from tsunami and geodetic
data by Tanioka and Satake [2001] as a priori information
(see Table 5). The inversion result is shown in Figure 11.
This figure shows the perturbation term of the slip rate on
the plate interface. The inverted spatiotemporal slip distri-
bution has the following features.

1. Two areas of large coseismic slip are separated by
the low-slip zone extending along the Kii channel (see
Figure 11a).

2. An area of large slip deficit (over 3 cm/year) during
the interseismic period reaches to about 30–40 km in depth
in the northeastern part of the source model (see Figures
11d–11f).

3. There are unphysical, very large slip deficits (over
9 cm/year) during the 30 years after the occurrence of the
main shock (see Figure 11d).

4. The spatial pattern of coseismic slip distribution
agrees well with the spatial pattern of slip deficit distribu-
tion during the interseismic period (see Figures 11a and
11d–11f).

The pattern of coseismic slip estimated in this
inversion is consistent with other studies. The maximum
slip estimated by Yabuki and Matsuzawa [1992] was about
6.5 m, and by Ando [1975] about 6 m, while the estimate
of this study is about 12 m. This difference is mainly
because in our analysis the observation period for the
coseismic stage spans over 10 years and may include some
postseismic deformation. The inverted magnitude of coseismic slip is larger than the geodetic inversion result (11 m) by Sagiya and Thatcher [1999], and the magnitude of slip estimated from the recurrence times of earthquakes. The locations of the two peaks obtained here are almost the same as that found by Yabuki and Matsu’ura [1992]. The area of large slip deficit at 30 to 40 km depth during the interseismic period, is consistent with other studies [e.g., Ito et al., 1999; Mazzotti et al., 2000]. The very large slip deficit following the main shock seems inconsistent with the plate rate, since it is much larger than the convergence rate (6.3–6.8 cm/year) at the Nankai Trough. The last feature corresponds to a general interpretation of large coseismic slip in the region where strain is accumulated.

As described in the preceding section, for better modeling of the geodetic data, we need to consider the viscoelastic asthenosphere. Because the effective relaxation time \( t_e \) of the whole system is about 50 years, the observed crustal deformation during the last 100 years includes response not only to a great earthquake and interplate

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<th>Prior Constraint</th>
<th>( \alpha_1^C )</th>
<th>( \alpha_1^I )</th>
<th>( \alpha_2^C )</th>
<th>( \alpha_2^I )</th>
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<td>0.0832</td>
<td>6.2180</td>
<td>0.7607</td>
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*ABIC is calculated from equation (10). \( C' \) of equation (10) is assumed to be zero.*