Vertical deformation of the Japanese islands, 1996–1999

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[1] The interseismic vertical deformation field of the Japanese islands is derived from the Global Positioning System (GPS) data between 1996 and 1999. Although vertical components of the GPS velocities have rarely been used as important information because of their higher noise level, information on the vertical deformation field will enable us to separate rigid plate motions from deformation due to interplate coupling because horizontal displacements contain both effects, whereas vertical displacements contain only the interplate coupling effect. The results show that (1) the precision of the vertical velocity is generally better than 2 mm/yr, (2) the existence of clear uplift peaks of about 8 mm/yr in the Shikoku and Tokai regions is consistent with strong coupling with a gradual brittle-plastic transition zone at the Nankai and the Suruga trough, respectively, (3) clear uplift peaks do not seem to exist in the Tohoku region because the Japan trench plate boundary is too far oceanward from the GPS sites, (4) significant uplift is obtained in the western part of Hokkaido, which may be associated with the long-term postseismic deformation due to the 1983 Japan Sea and the 1993 Hokkaido-nansei-oki (M = 7.6) earthquakes, and (5) uplift in northeastern Japan along the Japan Sea coast is apparently associated with distributed compression due to incipient subduction.

INDEX TERMS: 1206 Geodesy and Gravity: Crustal movements—interplate (8155); 1229 Geodesy and Gravity: Reference systems; 1243 Geodesy and Gravity: Space geodetic surveys; 1244 Geodesy and Gravity: Standards and absolute measurements; 8124 Tectonophysics: Earth’s interior—composition and state (1212); KEYWORDS: vertical deformation, interseismic deformation, brittle-plastic transition zone, interplate coupling


1. Introduction

[2] The Japanese islands are located among the convergence of four plates including the Eurasian (EUR), the North American (NA) (or the Okhotsk; OKH), the Pacific (PAC), and the Philippine Sea (PHS) plate (Figure 1). PAC subducts beneath NA (or OKH) at the Kuril-Japan trench and beneath PHS along the Izu-Bonin trench. PHS subducts beneath EUR at the Suruga-Nankai trough and beneath NA (or OKH) along the Sagami trough.

[3] Some studies have suggested the presence of microplates around the Japanese islands; for example, Zonenshain and Savostin [1981] proposed the existence of OKH on which northeastern Japan would lie, and Savostin et al. [1983] and Cook et al. [1986] proposed the existence of the Amurian plate (AMR; Figure 1) on which southwestern Japan would lie. However, the motion and even the existence of these plates are still ambiguous because crustal deformation in the Japanese islands is complicated by cyclic interplate coupling.

[4] Recent development of the continuous Global Positioning System (GPS) network in the Japanese islands allows us to see with high spatial resolution how the Japanese islands deform in the interseismic period. For example, Kato et al. [1998] and Sagiya et al. [2000] estimated the horizontal strain rate field from the interseismic deformation field of the Japanese islands, pointing out that the Japanese islands are generally under compression due to plate convergence. There have been several studies to invert for the interplate coupling from the horizontal displacement field [e.g., Ito et al., 2000; Miyazaki and Heki, 2001]. Employing only the horizontal velocity components as data, however, does not permit the separation of rigid plate motions and interplate coupling strain because horizontal displacements include both contributions. Thus rigid plate motions could be mapped into the interplate coupling, leading to a misunderstanding of its spatial distribution. To avoid this problem, Mazzotti et al. [2000] employed volumetric strain data obtained by spatial differentiation of GPS horizontal velocities to invert for the interplate coupling, based on the idea that rigid plate motions do not produce volumetric strains. However, the spatial resolution of interplate coupling is then poor because the spatial differentiation of displacement data results in the introduction of additional noise to the data to be inverted.
absolute vertical velocity field is obtained from raw GPS data.

2.1. Principal Component Analysis

The PCA, also known as the empirical orthogonal function analysis, has been commonly used to detect common mode signals in tidal data [e.g., Savage and Thatcher, 1992; Savage, 1995]. It resolves data, which are simultaneously observed, into a superposition of several common modes, each of which are orthogonal, and it can detect common modes at various spatial scales.

A displacement vector for the jth site, \( \mathbf{r}' \), is decomposed by

\[
\mathbf{r}' = \sum_{i=1}^{\infty} \alpha'_i \mathbf{u}_i
\]

where \( \mathbf{u}_i \) represents the ith mode of orthogonal functions derived by the PCA and \( \alpha'_i \) represents a coefficient for the ith mode at the jth site. Note that

\[
\delta_{ij} = \begin{cases} 
1 & \text{if } i = j \\
0 & \text{if } i \neq j 
\end{cases}
\]

where \( \delta_{ij} \) is the Kronecker’s delta. See, for example, Menke [1989, section 10.4] for more details.

2.2. Data Selection

We used continuous GPS data obtained by the Geographical Survey Institute (GSI) to obtain interseismic velocities. GSI gives one solution coordinate per day for each GPS site. We selected GPS data analyzed by GSI between 21 March 1996 (\( t = 1996.22 \)) and 31 December 1999 (\( t = 2000.00 \)), for the PCA even when there are some missing observations. Although the data in 2000 and 2001 are available, we ignored the data after \( t = 2000.00 \) because in 2000 the deformation field was perturbed by volcanic activity at Usu volcano [Tobita et al., 2001] and in the northern Izu islands [Nishimura et al., 2001], and by coseismic deformation due to the Tottori earthquake (\( M_w = 6.6 \)).

GPS sites in which any large (more than a meter) jumps occurred mainly due to antenna change were removed from the analysis. In this sequence of data selection, 502 out of 998 stations are left for the analysis. Distribution of these stations are shown in Figure 2, indicating that they are distributed all over Japan.

Even these selected GPS sites have some missing observations. Because all the observations should be done simultaneously for the PCA to work, the missing observations are interpolated by cubic splines.

2.3. Result

Figure 3 depicts the eight lowest principal component modes, and Figure 4 plots the 10 highest eigenvalues normalized by the highest one. Mode 1 of Figure 3 corresponds to the linear interseismic deformation, and mode 2 corresponds to deformation with an annual cycle, causes of which are still controversial [Murakami and Miyazaki, 2001; Heki, 2001]. Higher modes are neglected in this study because the eigenvalue associated with the mode 3 signal is less than 20% of that of the mode 1 signal (Figure 4).
Here we extracted only mode 1 as interseismic deformation. Thus the displacement of the $j$th site is described by
\[ r_j = a_j u + e_j \]
where $u$ equals $u_1$ in equation (1) and $a_j$ is chosen so that the residual $e_j$ is minimized. Figure 5 shows a sample of the comparison between the raw data and the extracted time series for two nearby stations, demonstrating that the raw data are well denoised by applying the PCA.

3. Obtaining Absolute Vertical Velocities

Although the procedure described so far allows us to obtain a velocity field, it does not reflect the absolute vertical velocity field because the GPS data we used are processed to give an a priori velocity relative to Tsukuba (the star in Figure 6). Figure 5 (right) shows that the GPS data we used gives an a priori subsidence of $\sim 6$ mm/yr of Tsukuba, while the absolute vertical velocity in Tsukuba is
expected to be near zero because Tsukuba is located far from the plate boundary.

Thus the GPS vertical velocity field needs to be combined with tide gage data to obtain absolute vertical velocities because they represent absolute vertical velocities if corrected for global sea level change. This section describes how the obtained GPS velocity field is combined with the tide gage data to obtain the absolute vertical velocity field.

### 3.1. Time Series Analysis of Tide Gage Data

First we need to obtain vertical velocities at each tide gage. Vertical velocities of 18 tide gages, operated by GSI, are calculated from the yearly averaged tide gage data between 1974 and 1999 (Figure 6). We used PCA, just as with the GPS data to removed common modes. Figure 7 shows the eight lowest modes of principal components and Figure 8 illustrates the 10 highest eigenvalues normalized by the highest one. Figure 9 shows examples of raw tide gage data after an eustatic correction of 1.8 mm/yr [Douglass, 1991] and the time series in which the mode 1 signal is extracted. The results shown in Table 1 appear satisfactory because vertical velocities at relatively stable sites such as WAJ, NEZ, KAS, and KAR (Figure 6) are within ±1 mm/yr.

### 3.2. Combining GPS Velocities With Tide Gage Velocities

This section describes how the GPS velocity field is combined with the tide gage velocity field to obtain absolute vertical velocities. Because velocities obtained from tide gage data represent absolute velocities, the reference frame of GPS velocities must be adjusted to that of the tide gage velocities. First, we chose a GPS site which is collocated with, or if not collocated, nearest to each tide gage. Then the selected GPS vertical velocities are adjusted to the tide gage velocities by

\[
dV = \frac{1}{N} \sum_{i=1}^{N} (V_i^t - V_i^g)
\]

where \(dV\) is the adjustment, subscripts \(t\) and \(g\) represent tide gage and GPS, respectively, superscript \(i\) represents \(i\)th data, \(V\) is velocity of tide gages and GPS stations, and \(N\) is the number of tide gages.

### 4. Error Analysis

Estimating error in a solution is important to assess its reliability. In this case, we need to consider the total error as a superposition of errors in (1) velocities derived from the raw GPS data when the PCA is applied, (2) velocities of tide gages when the PCA is applied, and (3) translation parameters. This section discusses how the error in the solution is obtained.

#### 4.1. Errors in Velocities Obtained From Raw GPS Data

The GPS velocity of the \(i\)th station is calculated from the raw data by

\[
v_i^G = \alpha_i / s_G
\]

where \(\alpha_i\) is the coefficient in equations (1) and (3) for the \(i\)th site and \(s_G\) is the velocity of mode 1 obtained by the PCA (Figure 3). Because both \(\alpha_i\) and \(s_G\) have errors, let us first explore each separately.
The variance associated with \( \sigma_{j}^{2} \), \( \sigma_{s}^{2} \), is given from equation (3) by
\[
\sigma_{j}^{2} = \frac{\mathbf{u}^{T} \mathbf{u}^{-1} \mathbf{u}^{T} \sigma_{j}^{2}}{C_{0}/C_{1}^{2}}
\]
where \( \sigma_{j}^{2} \) is the variance associated with individual data for the \( j \)th site derived by taking standard deviation of the raw data after removing the linear trend.

The variance associated with \( s_{G} \), \( \sigma_{sG}^{2} \) is represented by a sum of white noise and random walk components as [Zhang et al., 1997]
\[
\sigma_{sG}^{2} = \frac{\phi_{wh}^{2}}{T^{2}} \frac{12(N_{e} - 1)}{N_{e}^{2}} + \frac{\phi_{rw}^{2}}{T}
\]
where \( \phi_{wh} \) and \( \phi_{rw} \) are magnitudes of white noise and random walk in the time series of the mode 1 signal (Figure 3), respectively. \( T \) is the span of observations, and \( N_{e} \) is the number of observations. Another noise model with a combination of white noise and flicker noise, whose noise level is inversely proportional to frequency, has been proposed more

Figure 7. Eight lowest modes of principal components for tide gage data.

Figure 8. Eigenvalues of 10 lowest modes normalized by the eigenvalue of the mode 1 (with the highest eigenvalue) for tide gage data.

Figure 9. An example of time series of tide gages at WAJ, ONI, KAR, and KUR, whose locations are shown in Figure 6. Gray dots denote raw data and black lines show denoised time series.
recently [Mao et al., 1999; Nikolaidis, 2002], but we employed a model with random walk because of the simplicity of calculation and because the choice of the noise type does not strongly affect the result.

[22] Solving the nonlinear least squares problem reveals that white noise of 1.6 mm and random walk noise of 2.9 mm/√yr best represent the noise property. Substituting these values, with $T = 3.78$ years, and $N = 1381$, into equation (7) yields $\sigma_{sv} = 1.49$ mm/yr. Note that the random walk component dominates the amplitude of the noise because $N$ in equation (7) is large in this case.

[23] Because both $\alpha_i$ and $\sigma_s$ in equation (5) are assumed to have Gaussian noise, the error in velocity $v_G^i$ is not represented by simple Gaussian noise. Hence the variance in GPS velocities derived from the raw data is obtained by a numerical simulation which calculates $v_G^i = (\alpha_i^j + \Delta\alpha_i^j)(\sigma_G + \Delta\sigma_G)$ many, say 2000, times, where $\Delta\alpha_i^j$ and $\Delta\sigma_G$ are perturbations to $\alpha_i^j$ and $\sigma_G$, respectively, according to the estimated variance of them.

4.2. Errors in Tide Gage Velocities

[24] The velocity of the $j$th tide gage, $v_T^j$, is calculated by

$$v_T^j = \beta^j s_T$$

where $s_T$ and $\beta^j$ depict the velocity of mode 1 obtained by the PCA and the associated coefficient for the $j$th tide gage, respectively. As was done for the error in the raw GPS velocities, we first computed the error of both $\beta^j$ and $s_T$ separately, and then estimated the variance of $v_T^j$ by a numerical calculation. Note that we did not consider the random walk component of the mode 1 signal for the tide gage time series because the noise spectrum of the mode 1 signal does not have a significant random walk component. We only considered the white noise, which equals 28.4 mm, for the analysis.

4.3. Total Error

[25] The total error of the velocity for the $j$th site, $\sigma_{total}^j$, is given by

$$\sigma_{total}^j = \sigma_{sv}^j + \sigma_{dv}$$

where $\sigma_{sv}^j$ and $\sigma_{dv}$ are variances associated with the GPS velocities from the raw data and adjustment of the reference frame. From equation (4), $\sigma_{dv}$ is calculated by

$$\sigma_{dv} = \frac{1}{N} \sum_{i=1}^{N} (\sigma_{sv}^j + \sigma_{dv}^i)$$

where $\sigma_{dv}^i$ is the variance associated with the tide gage velocity for the $j$th site. Combining equations (9) and (10) yields

$$\sigma_{total}^j = \sigma_{sv}^j + \frac{1}{N} \sum_{i=1}^{N} (\sigma_{sv}^j + \sigma_{dv}^i)$$

[26] The total error of the vertical velocity at the $i$th site, $\sigma_{total}^i$, are described by equation (11).

5. Results

[27] Figure 11 shows the distribution of vertical velocities in the Japanese islands overlayed by depth contours of subducting plates compiled from the distribution of micro-earthquakes by Ishida [1992] and Sagiya and Thatcher [1999] for the Philippine Sea plate and Hasegawa et al. [1985] for the Pacific plate.

[28] The result shows a clear uplift pattern up to 8 mm/yr from Shikoku to Tokai parallel to the Nankai trough and clear subsidence pattern near the Suruga trough, indicating the overriding Eurasian plate is down dragged by the subducting Philippine Sea plate. No clear uplift pattern is seen along the Pacific coast of Kyushu, indicating weak or no interplate coupling between the Eurasian and Philippine Sea plate there, as expected from lack of M8 earthquakes in that area [e.g., Shiono et al., 1980]. On the contrary, the deformation pattern in northeast Japan indicates no such clear evidence for interplate coupling along the Kuril-Japan trench. This is probably because GPS sites are too far from the trench to resolve the vertical component of interplate coupling.

[29] Clear uplift up to 10 mm/yr and smaller (~6 mm/yr) uplift are seen in a small region of western Hokkaido. This

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Table 1. Tide Gage Locations With Their Velocities and Uncertainties After the Eustatic Correction

<table>
<thead>
<tr>
<th>Site</th>
<th>Latitude, °N</th>
<th>Longitude, °E</th>
<th>Velocity, mm/yr</th>
<th>SD, mm/yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABU</td>
<td>35.13</td>
<td>139.62</td>
<td>1.05</td>
<td>0.53</td>
</tr>
<tr>
<td>AKU</td>
<td>31.72</td>
<td>130.28</td>
<td>2.05</td>
<td>0.68</td>
</tr>
<tr>
<td>ASA</td>
<td>40.84</td>
<td>140.83</td>
<td>0.78</td>
<td>0.45</td>
</tr>
<tr>
<td>HOS</td>
<td>32.02</td>
<td>131.47</td>
<td>2.76</td>
<td>0.77</td>
</tr>
<tr>
<td>KAI</td>
<td>34.27</td>
<td>135.07</td>
<td>5.57</td>
<td>1.01</td>
</tr>
<tr>
<td>KAR</td>
<td>33.48</td>
<td>129.85</td>
<td>1.11</td>
<td>0.50</td>
</tr>
<tr>
<td>KAS</td>
<td>37.40</td>
<td>138.57</td>
<td>0.47</td>
<td>0.63</td>
</tr>
<tr>
<td>KAT</td>
<td>35.17</td>
<td>140.27</td>
<td>1.04</td>
<td>0.45</td>
</tr>
<tr>
<td>KUR</td>
<td>33.41</td>
<td>133.40</td>
<td>5.82</td>
<td>1.19</td>
</tr>
<tr>
<td>MIK</td>
<td>36.23</td>
<td>136.17</td>
<td>−1.29</td>
<td>0.56</td>
</tr>
<tr>
<td>NEZ</td>
<td>38.23</td>
<td>139.51</td>
<td>0.18</td>
<td>0.53</td>
</tr>
<tr>
<td>OGI</td>
<td>38.06</td>
<td>138.47</td>
<td>2.50</td>
<td>0.61</td>
</tr>
<tr>
<td>OII</td>
<td>39.97</td>
<td>139.78</td>
<td>3.67</td>
<td>0.61</td>
</tr>
<tr>
<td>ONI</td>
<td>34.82</td>
<td>136.87</td>
<td>8.50</td>
<td>0.88</td>
</tr>
<tr>
<td>OSH</td>
<td>43.29</td>
<td>140.60</td>
<td>1.10</td>
<td>0.46</td>
</tr>
<tr>
<td>SOM</td>
<td>38.03</td>
<td>140.84</td>
<td>−4.17</td>
<td>0.53</td>
</tr>
<tr>
<td>TGO</td>
<td>35.59</td>
<td>134.33</td>
<td>0.00</td>
<td>0.46</td>
</tr>
<tr>
<td>WAJ</td>
<td>37.38</td>
<td>136.89</td>
<td>0.58</td>
<td>0.46</td>
</tr>
</tbody>
</table>
uplift may indicate the long-term postseismic deformation associated with the 1993 Hokkaido-nansei-oki ($M = 7.6$) earthquake. Uplift of $\sim 6$ mm/yr in northwestern Honshu adjacent to the Japan Sea may be associated with the convergent plate boundary just offshore and represented by earthquakes such as the 1983 Japan Sea earthquake ($M = 7.6$). It is known that related deformation, associated with thrust faulting such as the 1896 Riku-u earthquake [Thatcher et al., 1980] extends onshore in this region.

6. Analysis of Subduction Coupling

Here we discuss the deformation field in southwest and northeast Japan separately because our result shows two clear uplift patterns in southwest and northeast Japan. We also discuss transient deformation during the observation period.

6.1. Southwest Japan

Our result shows an uplift pattern similar to those of previous studies [Savage and Thatcher, 1992; Savage, 1995] obtained from tide gages for southwest Japan. Both previous studies and our result have clear uplift peaks with

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**Figure 11.** Distribution of vertical velocities with lines for the seismic coupling analysis (Figure 13), in which lines A, B, C correspond to Nankaido, Tonankai, and Tokai regions, respectively, and depth contours of subducting Philippine Sea Plate (solid) and Pacific Plate (dotted). Contour interval is 10 km for the Philippine Sea plate and 50 km for the Pacific plate. Note that the scale of vertical velocity is limited between $-4$ mm/yr and 6 mm/yr to get more resolution, although there are some areas such as southwestern Hokkaido, where vertical velocities exceed 6 mm/yr.

**Figure 12.** Vertical deformation normalized by the amount of dislocation as a function of horizontal distance of the locked portion obtained by Savage [1983], who assumed elastic, homogeneous, and isotropic half-space. Solid and broken lines are for a fault with dip angles of 10° and 30°, respectively.
similar locations, indicating that our results are reliable and the state of plate coupling has not changed much for several decades. However, our results show almost no vertical deformation along the Japan Sea, while Savage and Thatcher [1992] show uplift of about ~3 mm/yr. It seems reasonable to think that the vertical uplift is nearly zero along the Japan Sea in southwestern Japan because the area is far from the Nankai trough and from the nascent plate boundary along the Japan Sea coast of northeast Japan. The uplift in their study may be due to an overestimate of the eustatic correction for sea level changes [Douglas, 1991] or higher noise level in the tide gage data.

We modeled the depth distribution of seismic coupling using two-dimensional dislocation models [Savage, 1983] (Figure 12) for three areas: the focal region of the 1946 Nankaido earthquake, that of the 1944 Tonankai earthquake, and the Tokai seismic gap (lines A, B, C of Figure 11, respectively). Two models, with and without a brittle-plastic transition zone (BPTZ), are examined for each area. The model with no BPTZ has a sharp cutoff of seismic coupling from full to no coupling at a certain depth, which is to be determined by solving a nonlinear inverse problem, while a model with BPTZ has the seismic coupling decreasing gradually with depth and the upper and lower bound of BPTZ are to be calculated by solving a nonlinear inverse problem as well.

Figure 13 shows the spatial distribution of seismic coupling with the comparison between the observed and calculated deformation for both models and all three regions, indicating that the full coupling region extends to about 30 km for the model without BPTZ and BPTZ extends between about 20 and 40 km for the model with BPTZ for all three regions, consistent with a thermal study [Hyndman et al., 1995] that 30 km corresponds to BPTZ (300°C–500°C).
The results indicate that the model with a BPTZ fits the observation better than the model without a BPTZ. Is BPTZ really required? We use a statistical test to address this problem. Table 2 shows the result of the F test [Abramowitz and Stegun, 1965], which tests whether increasing the number of model parameters significantly improves the fit to the data. It indicates that BPTZs are required at the 90% confidence level for the Tokai area and 95% confidence for other areas. The gradual decrease of the vertical deformation away from the trench cannot be explained without a gradual BPTZ.

There are three basic models of interseismic strain accumulation; the viscoelastic coupling model [Nur and Mavko, 1974; Savage and Prescott, 1978; Thatcher, 1983], in which a strong brittle layer is underlain by a weak viscoelastic layer, the strong plastosphere model [e.g., Bourne et al., 1998], in which shear in a strong plastosphere drives slip on the faults in the weak schizosphere, and the deep slip model [Savage and Prescott, 1978], in which the fault extends through a ductile shear zone. For strike-slip faults, such as the San Andreas, geodetic data [Gilbert et al., 1994] and seismic reflection studies [Henstock et al., 1997; Parsons, 1998] favor the deep slip model. Savage [1995] pointed out that a viscoelastic coupling model is inconsistent with the tide gage data for the Nankai trough region. In the case of subduction, the strong plastosphere model can be ruled out because the loading is driven by the downgoing plate rather than the lower crust.

Our results confirm that interseismic strain accumulation is governed by the deep slip model. The above three models have been difficult to distinguish for vertical faults like the San Andreas fault [Mavko, 1981], but can be distinguished with vertical motions for a dipping fault like the Nankai trough because some GPS sites are just above the fault and thus provide higher resolution of slip at depth. The deep slip model, which our data favor, is consistent with the geological interpretation of mylonite belts extending into the plastic layer of the crust [Scholz, 1988, 2002].

In Figure 13d, the trench-normal velocities with respect to stable Eurasia, using the new reference frame of M. G. Kogan et al. (manuscript in preparation, 2003), are compared with those predicted by our fit to the vertical data with and without a BPTZ. The horizontal velocities were calculated with the formulation of Freund and Barnett [1976], with a correction from Rani and Singh [1992]. As can be seen, the fit is quite reasonable both with and without a BPTZ. This shows that horizontal data are not very sensitive to the interplate coupling.

Although a clear uplift pattern exists between Shikoku and Tokai, no areas with significant vertical velocities exist in Kyushu area. This result reflects the low coupling between the Philippine Sea plate and the Eurasian plate in this area. This has already been inferred from the absence of historical large (M ≈ 8) earthquakes [e.g., Shiono et al., 1980] and that horizontal velocities in this area with respect to the stable Eurasian craton are toward the trench rather than away from the trench [Kato et al., 1998; Sagiya et al., 2000].

### 6.2. Northeast Japan

As in southwest Japan, vertical deformation in northeast Japan have previously been studied for decades with conventional geodetic techniques. For example, Kato [1979] and El-Fiky et al. [1997] obtained vertical velocities in northeast Japan from leveling and tide gage data. Their results are not, however, consistent with each other because the deformation in northeast Japan is heavily contaminated by large earthquakes such as the 1968 Tokachi-oki earthquake (M = 7.9), the 1978 Miyagi-oki earthquake (M = 7.4), and the 1994 Sanriku-oki earthquake (M = 7.5) along the Kuril-Kurile trench and the 1983 Japan Sea earthquake (M = 7.7) and the 1993 Hokkaido-nansei-oki earthquake (M = 7.6) in the Japan Sea, whose source regions are shown in Figure 1. Such contamination is inherent if we try to obtain vertical deformation by conventional geodetic techniques because they require longer periods of observations than GPS data to get reliable results, hence the probability that the vertical deformation is disturbed by coseismic deformations is higher.

Our results show no significant vertical deformations along the Pacific coast in northeast Japan, consistent with previous studies on interplate coupling using horizontal GPS data and data from conventional geodetic techniques, which shows no interseismic coupling below 100 km depth [El-Fiky and Kato, 1999; Ito et al., 2000]. The significant uplift observed along the Japan Sea coast therefore cannot be due to interplate coupling along the Japan-Kuril trench.

What causes this anomalous uplift along the Japan Sea in northeast Japan? It is most straightforward to speculate that the uplift is caused by afterslip or viscoelastic rebound due to large earthquakes such as the 1983 Japan Sea (M = 7.6) earthquake and the 1993 Hokkaido-nansei-oki earthquake (M = 7.5). Recently, Ito and Yoshioka [2000] tried to explain the anomalous deformation by viscoelastic rebound due to these large earthquakes, but the observed data did not fit the model well. A simple afterslip model assuming an isotropic, elastic, and homogeneous medium cannot explain the deformation field very well either (J. Fukuda, personal communication, 2001). This is a region of incipient subduction [e.g., Tamaki and Honza, 1985] and the deformation is not localized along a subduction interface as in mature subduction zones, so it may be hard to model the deformation with simple rectangular faults. Large thrust

<table>
<thead>
<tr>
<th>Nankaido</th>
<th>Tonankai</th>
<th>Tokai</th>
</tr>
</thead>
<tbody>
<tr>
<td>(D)</td>
<td>29</td>
<td>32</td>
</tr>
<tr>
<td>(D_{bptz}) km</td>
<td>16</td>
<td>20</td>
</tr>
<tr>
<td>(D_{bottom}) km</td>
<td>40</td>
<td>42</td>
</tr>
<tr>
<td>(\chi_1^2)</td>
<td>33.23</td>
<td>26.29</td>
</tr>
<tr>
<td>(\chi_2^2)</td>
<td>12.00</td>
<td>18.39</td>
</tr>
<tr>
<td>(F)</td>
<td>2.77</td>
<td>1.43</td>
</tr>
<tr>
<td>(F(0.05))</td>
<td>1.54</td>
<td>1.51</td>
</tr>
<tr>
<td>(F(0.1))</td>
<td>1.41</td>
<td>1.39</td>
</tr>
<tr>
<td>Number of data</td>
<td>59</td>
<td>65</td>
</tr>
</tbody>
</table>

\(\chi_2^2\) represents the square of weighted sum of residuals calculated by \(\chi_2 = [1/(N-M)] \sum_{i=1}^{N} (e_{i}^2/\sigma_{i}^2)\) where \(N\) and \(M\) are numbers of observations and model parameters, respectively, \(e_{i}\) is the residual between the observed and the calculated data for the \(i\)th data, and \(\sigma_{i}\) is the variance for the \(i\)th data, for the model with abrupt and gradual BPTZ, respectively. \(F\) is the static calculated by \(F = \chi_2^2/\chi_2^2\). \(F(0.05)\) and \(F(0.1)\) are \(F\) values for the 95% and 90% confidence. If the \(F\) static is above these values, the model is considered to be significantly improved with the confidence of the specified value.

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Table 2. Results of the Seismic Coupling Analyses
earthquakes such as the 1896 Riku-u earthquake ($M = 7.5$) [Thatcher et al., 1980] have occurred on land in this area, leading to the interpretation that it is a region of distributed compression.

6.3. Transient Deformation

[42] During the period of our analysis, local transient deformation episodes were observed at some GPS sites. Figure 14 shows three-component time series for GPS sites affected by the transient deformation. Figure 14 (left) shows time series of station 950473 (location shown in Figure 6) during a silent earthquake. Hirose et al. [1999] showed that the moment release during the slow earthquake is equivalent to a $M/C\geq247$ earthquake, despite the lack of any $M>5$ earthquakes. Note that this is in the region of Kyushu where our results indicated that the plate boundary is decoupled. Figure 14 (middle) shows time series of station 940095 (location shown in Figure 6) during two $M>6$ earthquakes and their afterslip. Ozawa et al. [2001] modeled the deformation to conclude that the moment release lasted more than 2 years. Figure 14 (right) shows time series of station 93062 (location shown in Figure 6) during a seismic swarm. Aoki et al. [1999] showed that the deformation was due to the intrusion of 0.35-m-thick dike. Figure 14 indicates that transient deformation is not as clearly observed for the vertical component as the horizontal components for all events. Specially, a slow earthquake at the Bungo channel (Figure 14, left) did not seem to affect the estimate of interseismic vertical velocity. On the other hand, two other events seem to have affected the estimates significantly. Despite these disturbances, deformation in most of the area in our data set were not contaminated by transients. Thus our estimate for vertical velocities (Figure 11) is considered to be reliable.

7. Conclusion

[43] We derived the distribution of interseismic vertical deformation field in the Japanese islands from continuous GPS data between 1996 and 1999 by employing principal
component analysis to remove spatially correlated noise components, and the results show that the precision of the vertical velocity is generally better than 1 mm/yr.

[44] A clear uplift peak of ~8 mm/yr are seen from Shikoku to Tokai, while no uplift pattern is observed in the Kyushu area. This indicates strong coupling between the Philippine Sea plate and the Eurasian plate at the Suruga-Nankai trough and weak coupling off Kyushu. The existence of a brittle-plastic transition zone between 20 and 40 km depth in the former area indicates that interseismic strain accumulation is explained well by the deep slip model, in which the fault extends to depth as a ductile shear zone. This is consistent with the geological interpretation of mylonite belts extending into the plastic layer of the crust.

[45] Another clear uplift pattern is observed along the Japan Sea in the northeastern Japan, which may be associated with the long-term postseismic deformation due to the 1983 Hokkaido-nansei-oki (M = 7.6) earthquake and with distributed compression associated with incipient subduction at the eastern rim of the Japan Sea. On the other hand, clear uplift peaks do not seem to exist in the Tohoku region because the plate boundary between the Pacific plate and the North American (Okhotsk) plate (Japan trench) is too far seaward from the GPS sites.

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