Crustal Deformation Cycle and Interplate Coupling in Shikoku, Southwest Japan

四国における地殻変動サイクルとプレート間カップリング

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ABSTRACT

Geodetic data for the last 100 years in Shikoku, southwest Japan, are compiled in order to reveal the crustal deformation associated with the earthquake cycle at the Nankai trough, where the Philippine Sea plate is subducted beneath the Eurasian plate. Significant characteristics of the crustal deformation in Shikoku for the last earthquake cycle including the 1946 Nankaido earthquake (M8.1) are 1) interseismic subsidence at 5 mm/year and coseismic rebound at Muroto Promontory, the southeastern end of Shikoku Island, 2) coseismic subsidence and its postseismic recovery at the northern and western coast of Tosa Bay and 3) postseismic subsidence along the northern coast. The first and second are cyclic deformation with a small secular movement, but the third is an irreversible movement as far as the last earthquake cycle is concerned.

Crustal deformation data for various stages of the earthquake cycle are inverted to estimate interplate coupling distribution by using a back-slip inversion technique with ABIC (Akaike's Baysien Information Criterion). Strongly coupled areas are confined to the shallower part less than 30 km in depth on the plate boundary throughout the whole cycle. But a postseismic slip might occur below 30 km in depth just after the 1946 Nankaido earthquake. It is one of the possible sources of the postseismic subsidence along the northern coast. The maximum back-slip rate is estimated as 5 to 6 cm/year in the NW direction. This value is well coincide with the geodetically observed relative plate motion at the Nankai trough.

From the estimated back-slip rate distribution and the assumed relative plate motion, the history of the relative displacement on the plate boundary is reproduced. The progress of relative displacement is significantly different between the shallower

part (< 30 km) and the deeper part (> 30 km). At the shallower part, the main part of the relative displacement is a coseismic slip for the shallower part. At the deeper part, on the contrary, the relative displacement takes the forms of a postseismic after slip and an interseismic steady sliding. The total amount of slip for the last 100 years is about 6 m over the whole plate boundary. Therefore, the present state of strain accumulation on the plate boundary is considered to be similar to that of 100 years ago. This means that it will take several tens of years for the next Nankaido earthquake to occur. Thinking of the relative displacement on the plate boundary for the whole earthquake cycle, the postseismic after slip at the deeper part plays an important role. Without the after slip, the relative displacement at the deeper part never accumulates enough to complete the whole cycle during about 100 years.

For the history of the relative displacement on the plate boundary, there have been only some numerical simulation results based on theoretical models. This study reveals what actually occurs on the plate boundary surface based on the inversion analysis of the crustal deformation data. Quasi-real-time precise monitoring of the relative displacement on the plate boundary surface, together with the progress in physics of earthquake generation, will realize the detection of precursory phenomena of large earthquakes in the future.

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Chapter 1. Introduction

Plate motion is the most essential source of various crustal activities, such as earthquakes, volcanic eruptions, crustal deformation, topography and so on. Apart from a driving force system of plate motion, we can interpret various crustal activities as a result of relative plate motion. Especially, among such various crustal activities, major crustal deformations are concentrated in the plate boundary region. Among various types of plate boundaries, convergent boundaries accompanying plate subduction attract much seismological and tectonophysical interest. Occurrence of large earthquakes, high seismicity, existence of volcanoes, island arc formation and many other interesting features characterize plate subduction zones.

It is well-known that there is a periodicity about occurrence of great earthquakes at many subduction zones. This periodic occurrence of great earthquakes is called an 'earthquake cycle'. An earthquake cycle can be roughly divided into the following four stages, preseismic stage in which seismogenic strain is accumulated in a source area, coseismic stage which contains the occurrence of an earthquake with instantaneous stress release, postseismic stage in which viscoelastic stress relaxation in the asthenosphere takes place, and interseismic stage in which stress accumulation starts preparing for the next event. In terms of the stress-strain state, the earthquake cycle is a repetition of accumulation and release of elastic stress-strain. The stress-strain state of the earth is reflected on the deformation of the earth's crust. In accordance with the progress of an earthquake cycle, the crustal deformation is expected to show periodic characteristics.

In this study, characteristics of crustal deformation during nearly one earthquake cycle at a typical low-angle subduction zone are described on the basis of geodetic data. The subduction zone treated here is the Nankai trough lying along southwest Japan (Figure 1). Geodetic data collected through routine surveys by the Geographical Survey Institute of Japan (GSI) are mainly processed. We interpret the crustal deformation in terms of interplate coupling, which contains very important information about processes of crustal strain accumulation and earthquake generation.

We use the term of interplate coupling to represent the intensity of plate interaction, which is related with seismogenic stress-strain accumulation on plate boundaries. In the case of a transcurrent plate boundary or a plate subduction zone, the intensity of interplate coupling is related with the shear strain energy accumulated at the plate boundary. Macroscopically, the intensity of interplate coupling is defined as a ratio of the seismic energy release rate to the total shear energy accumulation rate estimated from relative plate motion. There are distinct differences of coupling strength among various subduction zones [Kanamori, 1977]. The differences appear in seismic activity, the type of great earthquakes, the pattern of crustal deformation and the surface topography. Generally speaking, interplate coupling is considered to be strong at low angle subduction zones, such as Chile and Alaska, and weak at high angle subduction zones like Mariana. It is an evidence of the regional variation of interplate coupling that mega-thrust earthquakes occur only in low angle subduction zones.

If we look into a single subduction zone in more detail, lateral- or depth-dependent inhomogeneity of the coupling strength will be found. Plate boundary configuration, relative plate motion, rheological structure, thermal structure, chemical composition, and other factors may be related to such variations in coupling strength.

The Nankai trough is one of typical subduction zones (Figure 1), where the Philippine Sea plate is being subducted under the Eurasian plate along the Nankai trough. Relative plate motion was used to be estimated from , for example, sea-floor spreading rates, transform fault azimuths and earthquake slip vectors [Minster and Jordan, 1978], but nowadays it is directly observed by using space geodesy techniques such as GPS and VLBI. The relative plate motion between the Philippine Sea plate and the Eurasian plate is estimated to be 4 ~ 7 cm/year in NW-SE direction. [Matsuzaka et al., 1991; Seno et al., 1993; Tsuji, 1995]. The Philippine Sea plate must be consumed by subduction at the Nankai trough as much as this rate. This gives a kinematic boundary condition to the coupling problem.

Configuration of the subducted Philippine Sea plate in southwest Japan, which might have a significant effect on the interplate coupling, has been determined from the hypocenter distribution of microearthquakes [Mizoue et al. 1983; Okano et al., 1985; Miura et al. 1991]. The subduction angle at the Nankai trough is about 10 ~ 20 degree. The plate boundary slope become steeper at the eastern and western ends of Shikoku, that is around the Kii strait and the Bungo strait. The subducted slab can be followed as far as the northern coast of Shikoku, where the depth of plate boundary surface is about 50 km. Although we cannot find a clear continuous seismicity any more, Nakanishi [1980] has suggested the existence of a subducted slab under the southern part of the Chugoku district from the analysis of ScS phase.

The Nankai trough is one of major candidates which cause a great earthquake as large as magnitude 8 in and around Japan. The latest event there was a series of two earthquakes with an interval of only 2 years, the Tonankai earthquake (M7.9) in 1944 and the Nankaido earthquake (M8.1) in 1946. The source areas of these two earthquakes estimated from tsunami and geodetic data extend side by side for about 400 km long along the Nankai trough and about 100 km wide [Ishibashi, 1981; Ando, 1982; Yabuki and Matsu'ura, 1992]. According to the study of historical literature, such seismic events of magnitude 8 or so have occurred repeatedly along the Nankai trough [Ando, 1975; Utsu, 1984]. Table 1 shows a series of seismic events along the Nankai trough. The whole area broke at once in some events like the Hoei earthquake in 1707, and two events occurred successively with a very short interval in other cases. Therefore the whole area along the Nankai trough can be considered as a single seismogenic zone. The average recurrence interval is about 120 years for recent five events (Table 1). The Nankai trough is very suitable for this kind of research, because the seismic recurrence cycle can be investigated through reliable historical literature, and because the crustal deformation dataset is well-suited through repeated geodetic surveys.

The Shikoku Island, southwest Japan, is located on the land side of the Nankai trough (Figure 1). The Median Tectonic Line (MTL), which is one of the major tectonic line in the Japan island-arc, runs across the central part of the Shikoku Island in the E-W direction and divides geologically it into the northern and southern parts. The axis of the Nankai trough lies 100 to 150 km southeast from the southern coast of Shikoku. The southern part of the island is considered to be formed as a result of the accretion of oceanic sediments associated with the subduction of the Philippine Sea plate [Taira and Nakamura, 1986].

Geodetic surveys in Japan was started in the late 19th century by the Military Land Survey of Japan. They established an extensive and dense network consisting of thousands of control points and benchmarks throughout Japan, and determined their position and heights by triangulation and leveling. The density and uniformity of the network was the highest one at that time and perhaps even now in the world. After the completion of the survey for the whole land, surveys were repeated occasionally in order to reconstruct the geodetic network which was distorted by big disastrous seismic events like the 1891 Nobi (M8.0), the 1923 Kanto (M7.9), the 1927 Tango (M7.3) earthquakes or the big volcanic eruptions of Sakurajima in 1914. After the World War II, the survey work was succeeded by the present Geographical Survey Institute and re-surveys of the network were conducted several times. After 1970's, re-surveys of the nationwide geodetic network have become an important part of the Japanese earthquake prediction program, because understanding crustal strain accumulation is considered to be the first step in earthquake prediction.

All these surveys have been done based on the same criteria so that we can treat their data as a uniform dataset. Thus we have accumulated a precious dataset of crustal deformation of Japanese islands during the last 100 years, which is comparable to the seismic recurrence interval at the Nankai trough. We are able to reproduce the crustal deformations for the last one earthquake cycle now. This reproduction is very important for understanding physical subduction processes.

There have been several studies which dealt with crustal deformations in Shikoku [Thatcher, 1984; Miyashita, 1989; Minematsu, 1992; Sato and Matsu'ura, 1992; Fukahata et al., 1994]. Compared with these studies, the presen study deals with a comprehensive dataset including both vertical and horizontal data, covering the latest period. These data are collected from a database system of the Geographical Survey Institute and sometimes from original databooks.

The shallow angle subduction under Shikoku provides us for another favorable condition for this research. Although the trough axis lies about 150 km southeast off Shikoku, the depth to the plate boundary surface in the land area is from 20 to 40 km [Mizoue et al., 1983]. This situation is not so advantageous as in the Tokai area, central Japan, where the depth to plate boundary surface is around 10 km at its minimum. The depth range from 20 to 40 km corresponds to the main source region of the 1946 Nankaido earthquake [Yabuki and Matsu'ura, 1992], though the shallow part of the source region, which is under the sea, may not be included. The interplate coupling in this region is reflected on the crustal deformations in Shikoku. As the computational method for crustal deformation fields due to a dislocation source in an elastic half-space has been already established, we can estimate fault slip distributions from observed crustal deformations by using an inversion technique developed by Yabuki and Matsu'ura [1992]. This inversion technique, named 'back-slip inversion', is employed in order to estimate the distribution of coupling intensity on the plate boundary surface. The present problem is a kind of geophysical inverse problems, in which the number of free parameters for model description is larger than the number of observed data. Then, to solve the problem, we incorporate prior information into observed data in a inversion procedure and make a very flexible model, called a Bayesian model. ABIC (Akaike's Bayesian Information Criterion) is used for the selection of an optimum parametric model from the Bayesian model.

The study of interplate coupling is very important in various meaning. At first, overall description of a seismic recurrence cycle becomes possible. Especially, it is very important that we can understand how the stress-strain accumulation proceeds during the earthquake cycle. Then, together with results of laboratory experiments about rock mechanics and theoretical studies about earthquake generation, we can understand the whole process of earthquake cycle, which might lead us to realize a prediction of large earthquakes. For example, Tse and Rice [1986] and Stuart [1988] have developed an earthquake instability model based on the friction law of the plate boundary surface, which was obtained through laboratory experiments. According to their results, steady slip proceeds in the deeper part of the plate boundary, while shallower part slips episodically at the time of seismic event. And one of the important results of those simulations is the possibility of a precursory slip on the plate boundary before large seismic events. We can expect to detect such a precursory slip through precise measurement of crustal deformation. This study will be the first step to realize it.

Processing of geodetic survey data and crustal deformation characteristics in Shikoku are described in Chapter 2. The concept and the method to estimate interplate coupling distributions are described in Chapter 3. Estimated interplate coupling distributions are discussed in Chapter 4. Finally, Chapter 5 is assigned to related discussion.

Chapter 2. Crustal Deformation in Shikoku

2.1. Data

Data of geodetic observation such as leveling, triangulation, trilateraion, and tidal observation, are used to describe the crustal deformation field in Shikoku for the past 100 years. Geodetic surveys have been done by the Geographical Survey Institute of Japan and its antecedent, the Military Land Survey of Japan. These data are kept in the Geographical Survey Institute as both digital data and written form now. The first part of this study is to compile these data to reproduce crustal deformation field not only spatially bur also temporally. Each dataset and its processing method are explained at first, and the characteristics of the crustal deformation fields for individual stages are discussed.

2.1.1. Leveling Survey

In Japan, the Military Land Survey, which was the antecedent of the Geographical Survey Institute, started the nationwide leveling survey in the late 19th century of the Meiji era. In Shikoku the first leveling survey was conducted from 1886 to 1896. Several re-surveys have been conducted since then as summarized in Table 2. We have results of eight overall surveys in Shikoku during about 100 years. The average interval of re-surveys is around 15 years, but the recent surveys have been conducted more frequently than before. In these surveys, the first two surveys were before the 1946 Nankaido earthquake, and the third one was conducted just after the event as a part of recovery survey. Five surveys have been conducted afterwards.

The leveling routes in Shikoku are shown in Figure 2. There is a big loop nearly along the coastline of the island and two routes across the central mountain area in mainly N-S direction, which divide the big loop into three smaller loops. All these routes have been surveyed repeatedly in each survey. The small loop which passes through Tosa-shimizu, the southwestern end of the island, was established in 1955. There are also several supplementary routes which connect tidal stations with leveling routes.

Leveling survey has been conducted basically in the same style and the uniform regulations for observation errors have been kept through all the times. Survey is conducted using a leveling instrument with a tripod and two leveling rods. One measurement gives us a differential height of a subsection whose length is less than 100 m. As benchmarks are established every 2 km in usual, the differential height between two benchmarks is obtained as a sum of differential heights of subsections. A relative height is measured down to the order of 0.1 mm and permitted closure error of a round-trip survey is $2.5\text{mm}\times S^{1/2}$ (S is a surveyed distance in km). This regulation has been kept from the beginning of the leveling survey so that the precision of datasets is very homogeneous.

All the data of these leveling routes are summarized by the Geographical Survey Institute as "Collection of the results of precise leveling along primary benchmarks" which is published annually, and a database system of the leveling data, which is called LAGSAS [Geographical Survey Institute, 1987], is available in the Geographical Survey Institute. The leveling dataset consists of a series of relative heights between adjacent benchmarks along each leveling route. Date of the survey and the distance between benchmarks are attached to the relative height data.

2.1.2. Triangulation and trilateration survey

Horizontal crustal deformations are obtained by surveying a triangulation network. A triangulation network in Shikoku is shown in Figure 3 and the history of the observation is listed in Table 2. The triangulation network in Shikoku was established also in the late 19th century. There is hierarchy of three networks different in spacing, which are called the first, the second and the third order triangulation network. An average spacing for the first, the second and the third order triangulation network is about 20 km, 8 km and 5 km, respectively. Re-surveys have been conducted for the first order network and a part of the second order network. The first re-survey in Shikoku was conducted just after the 1946 Nankaido Earthquake in order to re-organize the heavily distorted network. Recent re-surveys were done from 1980 to 1984 and from 1989 to 1990. The recent two re-surveys were different from the former one in their purpose. These re-surveys were done following the Earthquake Prediction Program of the Japanese government. As the coastal area of the Bungo strait is designated as the area of specified observation for earthquake prediction, the network of recent re-surveys in this area is denser than rest of the island.

Change from triangulation survey to trilateration survey was an important turning point in the history of geodetic survey. Of all four surveys in Shikoku, the former two surveys were done by angular measurement and the latter two were done by distance measurement. A remarkable progress in measuring technology enabled us to use a laser distance meter which can measure a distance up to 50 or 60 km with the precision of around 1 ppm.

Now there has come another drastic change in the method of measuring horizontal crustal deformation. Global Positioning System (GPS) has almost taken the place of optical distance measurement. GPS is originally a navigation system using satellites. The differential positioning technique enable us to realize the precision better than 10^{-7}, which is one order better than the optical method. In Shikoku, Tabei et al. [1994] detected northwestward displacement vectors through repeated GPS observations. In Japan, also, the Geographical Survey Institute installed a nationwide permanent GPS network in 1994 [Abe and Tsuji, 1994]. More than 200 GPS receivers are working everyday now in 1995. This system realize precise real time monitoring of the crustal deformation all over Japan. In addition to high precision of the measurement, GPS has a merit that the station coordinates are estimated in a geocentric reference frame. In other words, we have a reference point at nearly infinite distance. We can utilize this network for various purposes such as estimating a crustal deformation field in Japan, investigation of large earthquakes and detecting earthquake precursors.

2.1.3. Tidal observation

Tidal observations have a big importance in treating vertical crustal deformation in Shikoku because we have no absolute reference point about vertical deformations else. That is due to the following two factors. At first, the leveling routes in Shikoku are isolated from the other routes of Japan, separated by the sea. Another factor is that all the island is influenced by the crustal deformation cycle associated with repetition of large earthquakes at the Nankai trough. Tidal records can be a reference of the absolute vertical displacements. Tidal stations used in this study are shown in Figure 2. Only the Kure station is maintained by GSI, and other 7 stations belong to Japan Meteorological Agency (JMA).

Figure 4 shows the annual mean sea level changes of tidal stations in Shikoku. Annual mean sea level data are mainly taken from Coastal Movements Data Center [1991], and supplemented by Kawasumi [1956] for the data before 1950. Being more specific, the data before 1950 at Komatsushima are borrowed from that of Toyomasu tidal station which is about 15 km southeast of Komatsushima. On the other hand, as Tosa-shimizu station was replaced in 1951, no change between 1950 and 1951 annual means is assumed. We can clearly see the large coseismic change at 1946 Nankaido earthquake at Kochi, Tosa-shimizu and Uwajima, while Takamatsu and Komatsushima show significant postseismic changes without large coseismic changes. Kochi also shows a postseismic sea level fall.

Raw tidal records contain various effects such as oceanic tide, oceanic current, atmospheric pressure, crustal deformation and so on. In order to extract absolute vertical crustal deformation from tidal records, Kato and Tsumura [1979] elaborated a method to process tidal records and analyzed the vertical movements along Japanese coast. Ozawa et al. [1994] used the method of Kato and Tsumura to analyze additional datasets from 1951 to 1993. Figure 5 is a processed monthly mean sea level of tidal station in Shikoku by Ozawa et al. [1994]. In the method of Kato and Tsumura, an original monthly tidal record is corrected for seasonal variations, an atmospheric pressure effect and localities. A secular variation in the sea level and the effect of real crustal deformations are contained in the final result inseparably. However, tidal stations at Hosojima, Tonoura, Wajima and Oshoro have tidal records for more than 80 years and all of them don't show any clear sea level change larger than 10 cm (Figure 6) [Coastal Movement Data Center, 1991]. Because these four stations are thought to be stable from tectonic point of view, secular sea level change can be neglected reasonably. Then the results of tidal analysis (Figure 5) can be treated as vertical crustal deformation data.

2.2. Processing of Data

In order to describe a crustal deformation cycle in Shikoku, the whole period is divided into five stages, which are 1) preseismic, 2) coseismic, 3) postseismic, 4) interseismic-1 and 5) interseismic-2. Used datasets for each stage are shown in Table 3. As the horizontal crustal deformation observation were not so frequent as vertical ones, only two stages, the coseismic and the interseismic-2 stages, have horizontal deformation dataset. Crustal deformation datasets of the other three stages consist of vertical deformation alone.

The lengths of five stages in Table 3 are different one another. The shortest one is postseismic-2 stage (about 10 years), and the longest one is preseismic stage (about 35 years). Therefore the usage of crustal deformation rate is more suitable for comparison between different stages than crustal deformation itself. Practical data arrangement process is mentioned below.

2.2.1 Vertical deformation processing

In the leveling survey, a direct observable is a relative height between two adjacent benchmarks dh_{ij} (i and j are benchmark indices). If we have two datasets of leveling observed in the same area in different times, we have a number of equations,

$$dh_{i,j}(t_2) - dh_{i,j}(t_1) = (V_i - V_j) \times (t_2 - t_1) \quad . \tag{1}$$

Here, V_i and V_j are the vertical displacement rates of benchmarks i and j. We have a set of similar equations for all the combinations of adjacent benchmarks.

Simultaneous equations (1) cannot determine the absolute vertical displacement rates of benchmarks but only relative ones so that we must introduce some reference points. A result of tidal analysis is used in this research for this purpose. Absolute vertical displacement rate is estimated for tidal stations at Takamatsu, Komatsushima, Muroto, Kochi, Kure, Tosa-shimizu and Uwajima. All these tidal stations are very close (~ 2 km) to leveling routes and the leveling results between each tidal station and its connection benchmark show no significant change for all the period (Table 4) [Coastal Movement Data Center, 1991]. So the vertical displacement rates of the connection benchmarks are assumed as that of tidal stations estimated by Ozawa et al. [1994]. The assumed vertical displacement rates of tidal stations for each stage are shown in Table 5. Equations (1) are simultaneously solved with these assumptions. This procedure is schematically shown in Figure 7 for the simplest case. As is obvious from Table 5, there are no tidal records available for preseismic stage and coseismic stage. Thus the northern end of the Shikoku Island, around Takamatsu, is assumed to be stable during these stages. Because a preseismic subsidence rate at Muroto promontry becomes almost the same as that of interseismic stages and coseismic deformation at Takamatsu was only 7 cm [Takamatsu Meteorological Observatory, 1950], this assumption is reasonable. Vertical displacement rates of benchmarks for each stage are listed in Table 6.

Estimated vertical deformation rate contains errors caused by various factors. The closure error of each survey is shown in Table 7. The closure error is a kind of indices representing the consistency of a series of surveys. According to Table 7, the first survey and the third survey have extraordinarily large closure errors. These large errors are ascribed to deformations during each survey. For example, it takes at least several months to complete a leveling survey throughout Shikoku and crustal deformations can proceed during the period. Then a later measurement has a possibility to be inconsistent with an earlier one. There surely existed a rapid postseismic rebound during the period of the third leveling survey just after the 1946 event [Geographical Survey Institute, 1972]. The data of this survey contain results from observations in various stages of the postseismic rebound, which may be an origin of large closure errors. Also, as the first survey took about ten years to complete surveying all over Shikoku, significant crustal deformation might proceed during that period. Except for these two surveys, the maximum closure error for the biggest loop is 59.5 mm, and 2.6 mm at its minimum. These closure errors are distributed over whole leveling routes through a network adjustment. As an average of closure errors for 6 surveys except for the first and the third ones is 22.9 mm, when the discrepancy between two relative heights for clockwise and counterclockwise half-loop is considered, the precision of relative vertical displacement rate during 10 years is expected to be around 1 mm/year ($22.9 \div 2 \div 10$ year = 1.15 mm/year). This value is thought to be a representative value of estimation error of vertical displacement rates.

2.2.2 Horizontal deformation processing

As we have four datasets of horizontal measurement, three pairs of measurements are made to estimate horizontal deformation field.

Horizontal deformations during the interseismic-2 stage are obtained by two trilateration surveys using the laser distance measurement technique. Scaling errors are considered to be very small because the measurement method was the same for both surveys and distances between trilateration points are measured directly. The data are compiled in the form of baseline length change between two surveys. In the processing like this, as all data are independent one another, a single erroneous data doesn't affect neighboring network at all. Horizontal displacement vectors, $\boldsymbol{u}_1$ and $\boldsymbol{u}_2$, at control points 1 and 2 are connected with the baseline length change $dl_{1,2}$ as

$$(\boldsymbol{u}_1 - \boldsymbol{u}_2) \cdot \boldsymbol{e}_{1,2} = dl_{1,2} \quad (2)$$

where $\boldsymbol{e}_{1,2}$ in (2) is a direction cosine between the control points 1 and 2.

On the other hand, the first triangulation survey in the late 1800's and recovery survey just after the 1946 event were done by the triangulation method. Although the precision of the triangulation is inferior to that of the trilateraion by laser ranging, closure errors of both triangulation surveys are represented fairly well by Gaussian law [Geographical Survey Institute, 1952]. In this study, displacement vectors by the Geographical Survey Institute [1952] are used for coseismic stage data. The latter survey was done just after the 1946 event, but the former one was about 50 years before that. Thus we should mind that the coseismic deformation here contains the preseismic deformation for about 50 years, too. It is impossible to know the amount of preseismic displacement exactly, but analogizing from the vertical displacement of the same stage, the amount of preseismic deformation seems to be at least 30 % of the total displacement.

We have another horizontal displacement dataset for 1949/52 ~ 1980/84. The old observation was triangulation but the other one was trilateration. Thus there might exist systematic errors. Moreover, crustal deformations in this stage contain two different deformation phases, postseismic and interseismic-1, inseparably. This dataset is not used for the analysis to estimate the distribution of interplate coupling in the later section because of these difficulties.

2.3. Characteristics of Crustal Deformation in Shikoku

Table 2 shows the division of the whole period into the 5 stages associated with the earthquake cycle at subduction zones. Characteristics of the crustal deformation in each stage is stated below.

2.3.1. Preseismic stage

Figure 8(a) shows the vertical displacement rate distribution of the preseismic stage (1886/96 ~ 1929/37). Only vertical deformation data exist for this stage. The displacement pattern shows a rapid subsidence rate of about 5 mm/year at Muroto Promontory, the southeastern end of Shikoku. Meanwhile the rest of the island upheaved. The uplift rate is bigger in the central and the western part. The eastern and the northern coastal area is stable within the estimation error. Although there is no reliable reference point in this dataset, a subsidence rate of 5 mm/year at Muroto

Promontory is very similar to the interseismic-2 stage as is stated later. As both stages are considered to be typically interseismic from the position in an earthquake cycle, this coincidence supports the assumption that the northern Shikoku was stable in the preseismic stage.

Comparing with the later interseismic stages, any clear differences cannot be found about preseismic stage. As far as this dataset is concerned, we cannot find any imminent precursors to the 1946 event. The latter survey of this stage was done about 9 to 17 years before the 1946 event. So more distinct precursory deformations might appear after this period. In that sense, this stage should have been called 'interseismic'.

2.3.2. Coseismic stage

Figure 8(b) shows the vertical displacement distribution during the coseismic stage (1929/37 ~ 1947), and Figure 9(a) is the horizontal displacement obtained from two surveys in 1890 ~ 1901 and 1949 ~ 1952. Nankaido earthquake of December 21, 1946 occurred in this stage. Since coseismic deformation is considered to be much bigger than preseismic one in its magnitude, it will be justified that this dataset is treated as coseismic one, though this stage also contains a rather long preseismic period. As the previous event occurred in 1854 [Usami, 1987], horizontal deformation data are obtained from the comparison between the survey just after (within 6 years) the event and the survey which was conducted about a half of interseismic period before. Therefore about a half of the crustal strain, which was accumulated after the previous event, might be included in the dataset. We may underestimate the coseismic deformation because the interseismic strain accumulation is roughly in the opposite sense against the coseismic strain release, that is, the coseismically released strain was NW-SE extension (Figure 10(a)) while the interseismically acumulated strain was NW-SE contraction (Figure 10(b), 10(c)). If the total amount of the interseismic strain accumulation is nearly equal to the coseismic strain release, we must multiply the coseismic deformation by 1.5 in this case. The effect of postseismic rebound is also contained in this dataset. In spite of those possibility, we will treat this horizontal dataset as horizontal coseismic deformation data, because there is no information about the actual horizontal deformation before the 1946 event. The crustal deformation field of recent period has an after effect of the 1946 event. As the 1854 event and the 1946 event are not the same, the recent deformation field is not the same as before 1946. So the uncertainty in the preseismic deformation field before 1946 is big. Therefore it is better to treat this dataset as the coseismic deformation than introducing uncertain assumptions. But we must remember being careful about the analysis of this dataset and its interpretation.

Okuda [1950] summarized and discussed the vertical coseismic crustal deformation and the Geographical Survey Institute [1952] summarized horizontal one. Raw dataset of leveling and horizontal network result for this stage are derived from these papers. The network-adjusted leveling data and triangulation data are treated as the coseismic deformation itself, not as the crustal deformation rate, because it is obvious that the major part of this deformation took place at the moment of the earthquake occurrence.

Vertical deformation pattern shows nearly 1 m uplift at Muroto Promontory, and the amount of uplift decreases northward rapidly to 50 cm subsidence in the northern coast area of Tosa Bay. This pattern is roughly in an opposite sense against the preseismic deformation. In both the northern and the western part of the island, vertical deformation is as small as several centimeters.

According to the network adjusting calculation by the Geographical Survey Institute [1952], horizontal displacements show an elastic rebound pattern against a northwestward motion of the Philippine Sea plate in southeast Shikoku (Figure 9(a)). The biggest displacement is at the control point near Muroto Promontory, which is 1.8 m in the ESE direction. The magnitude of horizontal displacement vectors decreases in northern and western part, but still as large as 1m around Tosa Bay. However, horizontal displacement varies much in southwestern Shikoku. Although the number of control points is quite a few, displacement vectors are directed in N or NW. There are some possible interpretations of these vectors. At first, we should be careful whether they are artificial results due to contaminated data or inappropriate data processing or not. But it is unlikely that all the observations contain large errors only in the southwest Shikoku. Distribution of horizontal strain (Figure 10(a)), which is obtained by the network adjustment of observed angle changes, shows that strain field in southwest Shikoku is rather uniform, though they are different from other parts of Shikoku. Thus it is concluded that these apparently strange displacement vectors show real crustal deformations.

We can interpret this displacement pattern in some ways. The first explanation is that the source area of the 1946 Nankaido Earthquake is confined at the western coast of Tosa Bay and no crustal strain was relaxed in southwest Shikoku. Suppose the strain accumulation had proceeded similarly under the whole Shikoku, as southwest

Shikoku was not the source area of 1946 event, seismogenic crustal strain remained there even after the earthquake. These displacement vectors contain not only the coseismic displacement but also the preseismic one, and these vectors may show preseismic deformations mainly. One reason of this inference is that the vertical displacement was also very small in southwest Shikoku, compared with the southeastern and the central part. Another possibility is that the crustal deformation in southwest Shikoku was due to faulting different from the main Nankaido thrust. There were no major seismic events reported in the coseismic period except for 1946 Nankaido, but, as Kato [1983] proposed, there might be a coseismic faulting of spray fault system around Ino Promontory, just northeast of Nakamura. This hypothesis is supported by tidal data at Tosa-shimizu which shows coseismic uplift of 40 cm at the 1946 event (Figure 11). This uplift can be explained qualitatively by simple extension of the Ino-misaki fault, which is a high angle reverse fault proposed by Kato [1983], to the southwest. The coseismic uplift at Tosa-shimizu cannot be explained by the first interpretation of strange vectors. It is most likely that effects of two factors stated above are responsible for the crustal deformation in southwest Shikoku together.

2.3.3. Postseismic stage

Figure 8(c) is the vertical displacement rate distribution of the postseismic stage (1947 ~ 1968/71). This period has only a vertical dataset. The pattern of vertical displacement for this period has a strong contrast against the coseismic one. Two remarkable characteristics are found from Figure 8(c). The first characteristic crustal deformation is a belt of rapid upheaval trending along the Nankai trough, which is situated just north of the coseismic uplift area. The maximum uplift rate is about 15 mm/year at the western coast of Tosa Bay, but it can be larger in the early postseismic stage as deduced from the tidal record at Kochi (Figure 4). As the uplift area looks closely related to the coseismic deformation, this rapid uplift is considered to be caused by a viscoelastic stress relaxation in the asthenosphere or some other after effects of the 1946 event. Instantaneous crustal deformations at a seismic event are relaxed to some extent by viscous flow in the asthenosphere in order to be gravitationally stabilized. Matsu'ura et al. [1981], Sato and Matsu'ura [1988] and Matsu'ura and Sato [1989] developed a method for calculating a viscoelastic response of the earth due to faulting and we can estimate a postseismic relaxation pattern by assuming a fault model of a seismic event. As the coseismic crustal deformation data discussed in 2.3.2 are inverted to obtain coseismic slip distribution on the plate boundary in the later section, the postseismic deformations are estimated based on this fault model.

The other characteristic of this period is a rapid subsidence of the northern coastal area of Shikoku. This subsidence is clear also from the tidal observations (Figure 4). Takamatsu and Komatsushima show rapid subsidence as fast as 5 cm/year from 1947 to 1950. The subsidence rate decreased after 1950 but lasted until late 1960's. The total amount of subsidence reached around 30 cm. This postseismic subsidence is larger than the coseismic ones, which is a very rare case. It was reported that this abnormal subsidence became a social problem, because the salt farms along the coastal area was damaged by a spring tide of a typhoon, intensified by the land subsidence [Ogasawara, 1948, Takamatsu Meteorological Observatory, 1950]. The abnormal subsidence was observed also at the tidal station at Shimotsu, the western coast of Kii Peninsula and at Sumoto, the Awaji Island (Figure 12). The subsidence area extended very widely along the Seto inland sea. It is unclear whether this postseismic subsidence in northern Shikoku is a common feature for every postseismic period in the earthquake cycle at the Nankai trough, because we have only one example. More researches for historical literature will be necessary to make out this point.

Looking at the annual mean sea level change (Figure 4), the postseismic subsidence at Takamatsu and Komatsushima changed its trend around 1950 apparently. The subsidence rate looks around 5 cm/year before 1950, and it decreased to less than 1 cm/year after 1950. On the other hand, the change in postseismic uplift rate at Kochi before and after 1950 is much smaller than that of Takamatsu and Komatsushima. These facts suggest that at least two different mechanisms are responsible for the postseismic crustal deformations. Candidates for the possible mechanism causing such deformations are discussed in later sections.

2.3.4. Interseismic-1 stage

Figure 8(d) is the vertical displacement rate distribution of the interseismic-1 stage (1968/71 ~ 1981/82). Although there is only a vertical dataset and the distribution of data points is rather sparse, some characteristics of the deformation field can be drawn from Figure 8(d). After big changes in coseismic and postseismic stage, crustal deformations calmed down in this stage. There still remains upheaval area around Tosa Bay, which can be confirmed by the tidal observation (Figure 4). Tidal record at Kochi shows an uplift of about 10 cm in this period, which corresponds to the uplift rate of 7 ~ 9 mm/year. Tidal record at Kure shows a similar change as Kochi, though the record started from 1973. Comparing with the deformation pattern in the postseismic stage, the center of uplift moves from the western coast to the northern coast of Tosa Bay. On the other hand, uplift in the eastern part of Shikoku had ceased almost completely. As an interpretation of this deformation pattern, it is considered that there exists a clear difference in time constant of postseismic deformation between eastern Shikoku and western Shikoku. The postseismic deformation in the central or western Shikoku has a longer relaxation time. It might reflect difference in the crustal structure, because the postseismic deformation pattern is sensitive to the lithosphere thickness as will be shown later in the present study, and the isodepth contour of the Philippine Sea plate has a big bend under the Kii strait, so the effective lithosphere thickness might be different in this region from the surrounding region. Or it may reflect the difference in slip distribution on the plate boundary surface including temporal sense. When we try to interpret postseismic deformation field in terms of the slip distribution on the plate boundary, we must think of not only its spatial distribution but also its temporal change. As for the uplift at the northern coast of Tosa Bay, we can find two phases of uplift from tidal analysis at Kochi (Figure 4). The first phase was clear in 1950's but calmed down after middle 1960's. Then, the second phase of uplift starts from middle 1970's. It can be found out at Kure tidal station as well. Rapid uplift at the northern coast of Tosa Bay corresponds to this second phase. The second phase uplift may be ascribed to healing of plate boundary surface, aseismic slip event or some other unknown phenomena. When the healing, that is the recovery of the interplate coupling strength, occurs around the source region which ruptured at the 1946 event, the area of back-slip will extend and the magnitude of the back-slip vector will increase. On the contrary, the aseismic slip will appear as the fore-slip distribution. But there are no data available for verifying these possibilities.

Overall features of crustal deformations in northern and eastern Shikoku are the typical ones of an interseismic period of Shikoku, but they look like postseismic ones in central and western Shikoku. A localized subsidence is observed around Muroto Promontory up to 5 ~ 6 mm/year.

2.3.5. Interseismic-2 stage

Figure 8(e) shows the vertical displacement rate of the Interseismic-2 stage (1981/82 ~ 1990/94). The baseline length change rate distribution obtained from two trilateration surveys in 1980 ~ 1984 and 1989 ~ 1990 is shown in Figure 9(b). Characteristics of the crustal deformation in an interseismic stage became more evident than the previous period. On the other hand, the difference among interseismic and preseismic stages is not clear, because the temporal resolution of the geodetic dataset is poor and the preseismic stage in the present study contains a fairly long period which should be called interseismic.

The horizontal displacement field shows contraction in the NW-SE direction, but this trend is not so clear as other periods. That is because the duration of this stage is only less than 10 years and it is too short to accumulate enough strain to be observed by the laser ranging technique. The accumulation rate of horizontal strain is of the order of 10^{-7}, and the repeatability of laser ranging is of the order of 10^{-6} for the baselines consisting of the Shikoku network. So the accumulated strain is likely to be comparable to a measurable limit. This causes the bad S/N ratio in this stage.

On the other hand, the vertical displacement rate is rather small as a whole. Some exceptions exist around Muroto Promontory and northeastern Shikoku. A localized subsidence around Muroto Promontory is a common feature for the preseismic and interseismic stages. The subsidence rate is about 5 mm/year there. Uplift in the northern coast of Tosa Bay which was remarkable in the previous stage, has stopped. It seems that various postseismic effects such as a viscoelastic stress relaxation, an after slip, and healing has disappeared almost completely in this period, and the deformation pattern has become a typical one for an interseismic period. One thing which cannot be made out now is the subsidence in the northeastern part of the island. It looks as if the subsidence become larger northeastward. It is difficult to interpret this subsidence in terms of interplate coupling, because the depth to the plate boundary is more than 40 km and the subducted Philippine Sea slab is too hot to cause interplate coupling at this depth. But it is interesting that the crustal deformation seems to correspond to surface topography, that is, the subsiding area is adjacent to the Sea of Harima and the subsidence rate increases seaward.

2.4. Crustal Deformation Cycle in Shikoku

As a summary of this chapter, vertical displacement and displacement rate data are listed in Table 6, and a list of triangulation points and horizontal crustal strains is given in Table 8. General features of the crustal deformation cycle are described as follows.

At first, as to the horizontal crustal strain (Figure 10), compression strain axes lie in the NNW-SSE or NW-SE direction in the central and east Shikoku for the postseismic (Figure 10(b)) and the interseismic (Figure 10(c)) periods. The magnitude of the compressive strain decrease northward. These compression axes slightly rotate counterclockwise in the western part of Shikoku, where the compressive strains are not so significant as the eastern part. But as the NNW-SSE compression had been accumulated before 1946 and have not been released yet, there remains NNW-SSE compression in the southwestern part of Shikoku. According to Okano et al. [1980], focal mechanisms in Shikoku show E-W compression for shallow (< 23 km) events and N-S compression for deeper (>23 km) events. These shallow and deep events correspond to earthquakes in the upper crust and inter- or intra-plate earthquakes around the plate boundary surface, respectively. Then we face a paradox that the NNW-SSE compression is prevailing for crustal strains while the E-W compression is obtained from seismic events. As one possibility, this paradox may be solved as follows. The NNW-SSE compression is a result of plate convergence at the Nankai trough. On the other hand, the E-W compression is a characteristic of the stress field in southwest Japan [Tsukahara and Kobayashi, 1991]. Magnitude of the NNW-SSE compression varies in accordance with the progress of earthquake cycle ,and the NNW-SSE compression has been presumably smaller than the value of the long term average for the whole cycle. Since the E-W compression stress is relatively stable for a long time, it is prevailing in the postseismic stage when a cyclic component is small. If it is true, as the earthquake cycle proceeds, the NNW-SSE compression will become a major focal mechanism.

About the vertical displacements, Figure 13 shows contour maps of vertical displacement or vertical displacement rate for all stages and characteristics of vertical dispacement through the earthquake cycle become clear. We can identify a clear cyclic movement around Muroto Promontory. Interseismic subsidence rate at the southeastern tip of Shikoku is always about 5 mm/y, and it is rather constant throughout the interseismic period. If the typical value of coseismic uplift is 1 m and the earthquake recurrence time is 120 years, we get 40 cm for one cycle offset. Some part of this offset will be relaxed by viscoelastic response, but it is considered that a considerable uplift accumulate through the repetition of earthquake cycles. It is consistent with the theoretical results of Matsu'ura and Sato [1989], in which the steady uplift rate of Muroto is estimated as 2 ~ 3 mm/yr.

Similar vertical movements but in opposite sense are observed in the northern and the western coast area of Tosa Bay. The land upheaves during the interseismic period and subsides at the time of the earthquake. However, the interseismic uplift is not a steady one like subsidence at Muroto Promontory but shows an exponential curve, suggesting a viscoelastic effect.

On the other hand, in the northern part of Shikoku, crustal deformations were rather small in the preseismic and the coseismic periods while large subsidence occurred and continued after the 1946 event. The postseismic subsidence is continuing even now. As we have only one example, we cannot conclude that this subsidence is a common feature through repeated earthquake cycles. But maybe we can say that the deformation cycle in Shikoku is not perfectly cyclic and a non-cyclic part of the crustal deformation is related to the deformations with a very long time scale as pointed by Fukahata et al. [1994].

Chapter 3. Back-slip Inversion

In order to estimate the strength distribution of interplate coupling from the crustal deformation data discussed in the previous section, we use a 'back-slip inversion' technique. The concept of back-slip and details of inversion analysis are described in this chapter.

3.1. Concept of Back-slip

It is widely accepted that the strength of plate interaction, or interplate coupling, varies with depth along a plate boundary surface [Shimamoto, 1990; Yoshioka et al, 1993]. A subducted oceanic plate and an overlying continental plate stick each other at an intermediate depth, but at shallower or deeper part of the plate boundary they are decoupled so as to move freely in accordance with the relative plate motion. The situation of the interplate coupling in an interseismic stage is illustrated in Figure 14(a). As a result of steady slip at shallower and deeper parts, tectonic strain accumulates in the strongly coupled region. In order to describe such a situation, the plate interaction at subduction zones can be expressed as a conceptual superposition of two components [Savage, 1983]. One is a steady plate subduction, which is described as uniform slip distribution along the whole plate boundary surface (Figure 14(b)). This part is called a 'steady' term hereafter. The steady term is prescribed by the macroscopic relative plate motion, which can be directly measured by space technique such as GPS, VLBI and SLR. Another is a perturbation of slip motion from the steady part, caused by coupling or decoupling on a part of the plate boundary (Figure 14(c)). It is described as a distribution of fault slip localized to the coupled region. The slip distribution will vary in its extension and magnitude in accordance with the progress of earthquake cycle. This part is called a 'coupling' term hereafter. When we describe an interaction on a coupled plate boundary, the coupling term becomes a backward slip so as to cancel out the steady term on the coupled region. That is because the coupling term is called a 'back-slip'. A magnitude of a back-slip vector indicates strength of interplate coupling. So if the estimated back-slip vector cancel out the relative plate motion exactly, we can interpret that the two plates are perfectly coupled there. Figure 14 shows a schematic diagram of dividing the total plate interaction into the two parts.

The effect of the steady term appears in a very long time sacale on topography such as marine terraces [Sato and Matsu'ura, 1988]. Although the dislocation source is distributed on the infinitely long plate boundary in Figure 14(b), we can neglect the effect of the dislocation sources within and below a viscoelastic layer according to Sato and Matsu'ura [1988]. In back-slip inversion, the effect of the steady part is neglected because the crustal deformation field produced by the steady term has a rather long wavelength and its magnitude is considered to be much smaller than that produced by the coupling term. According to the theoretical calculation for Shikoku by Sato and Matsu'ura [1992], vertical deformation rate of the steady term, which corresponds to the average uplift rate over one earthquake cycle, is less than 1 mm/year except for the southeastern tip of Shikoku. Thus the steady deformation is comparable to the noise level for the most part of Shikoku. At the southeastern tip of Shikoku, the steady motion is uplift and the typical interseismic motion is subsidence. So it is likely that the coupling term is underestimated around there. Of course, more accurate analysis including viscoelastic effects is preferable. However, the crustal deformation data are not sufficient for resolving the temporal variation necessary for distinguishing viscoelastic effects. Therefore the back-slip model is adopted for the present analysis.

We get information about interplate coupling from observed geodetic data as a back-slip distribution on the plate boundary. There are three possibilities in the polarity of the coupling term (Figure 15). During strain accumulation in an interseismic stage, no relative motion occur across the completely coupled plate boundary surface. In this case, the coupling term works so as to cancel the steady term, that is, the coupling term is the same as the steady term in its magnitude, and opposite in its direction. We call this situation as 'back-slip'. When the magnitude of the estimated coupling term is zero, it means no plate interaction there, that is, the relative displacement proceeds at the same rate of relative plate motion, and neither accumulation nor release of crustal strain occurs. This situation is called 'no slip' hereafter. If the relative displacement rate on the plate boundary exceeds the rate of relative plate motion, there occurs strain release on the plate boundary. Then, we obtain a coupling term in the same direction as the steady motion. This situation is called as 'fore-slip'. In the present study, the results of inversion analysis are represented as back-slip distributions or back-slip rate distributions, but it should be noted that they contain all the three cases.

With the concept of back-slip, we can treat the inversion problem in a simple and unified way. The observed crustal displacement data are rearranged in the form of the crustal deformation rates, and then the deformation rates are inverted to back-slip rate distributions on the plate boundary. This inversion procedure is very similar to that of estimating coseismic fault slip distributions, developed by Yabuki and Matsu'ura

[1992]. But note that the meaning of back-slip (rate) is different from that of coseismic slip. We need to add a steady term to the back-slip distribution in order to reproduce an actual slip distribution.

3.2. Inversion Procedure

Processing of the back-slip inversion is almost the same as that of the coseismic slip estimation from a crustal deformation dataset. The difference between them is that the input data are not crustal displacements but displacement rate and that the effect of steady term must be added before interpreting the inverted result. The outline of the Yabuki and Matsu'ura [1992]'s method is followed to formulate the back-slip inversion procedure used here.

3.2.1. Surface deformation due to back-slip

We can regard the back-slip inversion problem as a geodetic inversion problem for estimating a slip distribution on a fault whose geometry is already known. The geometry of the plate boundary surface can be resolved from a hypocenter distribution of microearthquakes. We consider a discontinuity in tangential displacement across a fault surface S buried in a homogeneous, isotropic, elastic half-space. The representation theorem for a static displacement field for such a case has been theoretically solved [Maruyama 1964; Manshinha and Smylie, 1971]. The representation theorem is expressed in a very simple form using the moment tensor [Backus and Malcahy, 1976a,b] and the Green's tensor as

$$u_i(x)=\sum_{p=1}^{3}\sum_{q=1}^{3}\int_S G_{ip,q}(x,\xi)m_{pq}(\xi)dS(\xi) \qquad (i=1,2,3), \tag{3}$$

with

$$m_{pq}(\xi)=\mu[n_p(\xi)\Delta u_q(\xi)+n_q(\xi)\Delta u_p(\xi)] \qquad (p,q=1,2,3), \tag{4}$$

where $G_{ip,q}$ is the derivative of Green's tensor $G_{ip}(x,\xi)$ with respect to ξ_q, $m_{pq}(\xi)$ is the moment tensor density distribution, $n_j(\xi)$ is a normal vector of the boundary surface at ξ, and $\Delta u_j(\xi)$ is ordinary a relative displacement vector across the boundary surface at ξ but a back-slip vector in this case. Analytical expression of $G_{ip,q}(x,\xi)$, which were first obtained by Maruyama [1964] and rewritten in a compact form by Yabuki and Matsu'ura [1992], are given in Appendix A. Since the slip vector $\Delta u_j(\xi)$ has only tangential components, $\Delta u_j(\xi)$ must satisfy the next condition,

$$\sum_{j=1}^{3} n_j(\xi)\Delta u_j(\xi)=0 \qquad \text{for all } \xi \text{ on S.} \tag{5}$$

This implies that only two components of the slip vector $\Delta u_j(\xi)$ are independent variables. So the surface displacement field can be described in the next form,

$$u_i(x)=\sum_{j=1}^{2}\int_S H_{ij}(x,\xi)\Delta u_j(\xi)dS(\xi) \tag{6}$$

with

$$H_{i1}=\mu\left\{2n_1(G_{i1,1}+G_{i3,3})+n_2(G_{i1,2}+G_{i2,1})+\frac{n_1n_2}{n_3}(G_{i2,3}+G_{i3,2})+\frac{n_1{}^2+n_3{}^2}{n_3}(G_{i1,3}+G_{i3,1})\right\}$$
$$H_{i2}=\mu\left\{2n_2(G_{i2,2}+G_{i3,3})+n_1(G_{i1,2}+G_{i2,1})+\frac{n_1n_2}{n_3}(G_{i1,3}+G_{i3,1})+\frac{n_2{}^2+n_3{}^2}{n_3}(G_{i2,3}+G_{i3,2})\right\}$$
$$\tag{7}$$

3.2.2. Parametric expansion of back-slip distribution

The back-slip distribution is represented by a linear combination of basis functions Φ_{kl}, defined on the coordinate system referring to the flat surface of the earth, as

$$\Delta u_j(\xi_1,\xi_2)=\sum_{k=1}^{K}\sum_{l=1}^{L}a_{jkl}\Phi_{kl}(\xi_1,\xi_2)\quad j=1,2. \tag{8}$$

where a_{jkl} are expansion coefficients. Substituting (8) into (6), we obtain the following expression of the surface displacement field.

$$\begin{aligned}u_i(x)&=\sum_{j=1}^{2}\int_S H_{ij}(x,\xi)\Delta u_j(\xi)dS(\xi)\\&=\sum_{j=1}^{2}\sum_{k=1}^{K}\sum_{l=1}^{L}F_{ijkl}(x)a_{jkl}\end{aligned} \tag{9}$$

with

$$F_{ijkl}(x)=\int_S G_{ij}(x,\xi)\Phi_{kl}(\xi_1,\xi_2)dS\,. \tag{10}$$

Then, we have a finite number (2×K×L) of expansion coefficients a_{jkl}, which are the model parameters of this analysis. As the basis functions Φ_{kl}, we take bicubic B-spline functions defined as follows.

$$\Phi_{kl}(\xi_1,\xi_2)=f(\xi_1;\xi_k,d)f(\xi_2;\xi_l,d) \tag{11}$$

$$f(s;s_0,d)=\begin{cases}\dfrac{2}{3}-\dfrac{(s-s_0)^2}{2d^3}\{2d-|s-s_0|\} & (0\le|s-s_0|\le d)\\[2ex] \dfrac{1}{6d^3}(2d-|s-s_0|)^3 & (d\le|s-s_0|\le 2d)\,.\\[2ex] 0 & (|s-s_0|\ge 2d)\end{cases} \tag{12}$$

Here, (ξ_k, ξ_l)denote the grid point, and the spatial resolution of this representation is controlled by the grid interval d. An important property of bicubic B-splines is that the second derivative of the function is continuous everywhere. The linear combination of B-splines has the same characteristics. As the strength of rocks is finite, internal stress of the earth must take a finite value. Then the slip distribution becomes continuous and smooth to some extent. The linear combination of B-splines satisfies this physical constrain by nature. Figure 16 shows a graphical image of a bicubic B-spline function. It takes non-zero value in the area of $4d \times 4d$. Superposition of B-splines is done with a shift d.

3.2.3. Observation equations

Suppose the crustal deformation (rate) is observed at N_p points. Then we can construct a set of observation equations.

$$u_i(x_p) = \sum_{j=1}^{2}\sum_{k=1}^{K}\sum_{l=1}^{L} F_{ijkl}(x_p)a_{jkl} + e_{ip} \qquad (p = 1 \sim N_p) \tag{13}$$

e_{ip} is an error vector of observed data which includes modeling errors as well. The observation equations can be written in a matrix form as

$$\boldsymbol{d} = \boldsymbol{Ha} + \boldsymbol{e}, \qquad \boldsymbol{e} \sim N(\boldsymbol{0}, \sigma^2 \boldsymbol{E}). \tag{14}$$

Here, the error vector $\boldsymbol{e}$ is assumed to be Gaussian. From observation, only relative magnitude is assumed to be known for each component of the error vector and it is reflected in the matrix $\boldsymbol{E}$. An absolute magnitude of the error vector is controlled by an unknown scale factor σ^2, which is estimated through the inversion procedure.

The observation equation (14) is equivalently described by a likelihood function which takes the form of probability density.

$$p(d|a;\sigma^2) = (2\pi\sigma^2)^{-N/2}\|E\|^{-1/2}\exp\{-\frac{1}{2\sigma^2}(d - Ha)^t E^{-1}(d - Ha)\} \quad (15)$$

where N is the number of data, $\|E\|$ is the determinant of E, and the superscript t means the transposition of the vector.

3.2.4. Prior distribution and a Bayesian model

As is mentioned before, finiteness of shear strength results in finiteness of shear stress. Therefore the back-slip distribution must be smooth in some degree. Let's define a quantity which represents roughness of the back-slip distribution as follows.

$$r = \sum_{j=1}^{2}\int_S \frac{1}{n_3}\left[\left(\frac{1}{h_1^2}\frac{\partial^2 \Delta u_j}{\partial \xi_1{}^2}\right)^2 + 2\left(\frac{1}{h_1 h_2}\frac{\partial^2 \Delta u_j}{\partial \xi_1 \partial \xi_2}\right)^2 + \left(\frac{1}{h_2^2}\frac{\partial^2 \Delta u_j}{\partial \xi_2{}^2}\right)^2\right] d\xi_1 d\xi_2\,, \quad (16)$$

with

$$h_i = \frac{\partial}{\partial \xi_i} f(\xi_1, \xi_2) \ (i=1,2). \quad (17)$$

Here, $\xi_3 = f(\xi_1, \xi_2)$ gives a configuration of the plate boundary surface. Substituting the parametric expansion (8) of the back-slip distribution into the definition (16), we obtain the following expression of roughness;

$$r = \sum_{j=1}^{2}\sum_{k=1}^{K}\sum_{l=1}^{L}\sum_{k'=1}^{K}\sum_{l'=1}^{L} a_{jkl} B_{klk'l'} a_{jk'l'} \quad (18)$$

with

$$
\begin{aligned}
B_{klk'l'} &= \frac{1}{\bar{n}_3 \bar{h}_1^{\,4}} \int \frac{d^2 f_k(\xi_1)}{d\xi_1^{\,2}} \frac{d^2 f_{k'}(\xi_1)}{d\xi_1^{\,2}} d\xi_1 \int f_l(\xi_2) f_{l'}(\xi_2) d\xi_2 \\
&+ \frac{2}{\bar{n}_3 \bar{h}_1^{\,2} \bar{h}_2^{\,2}} \int \frac{df_k(\xi_1)}{d\xi_1} \frac{df_{k'}(\xi_1)}{d\xi_1} d\xi_1 \int \frac{df_l(\xi_2)}{d\xi_2} \frac{df_{l'}(\xi_2)}{d\xi_2} d\xi_2 \, . \\
&+ \frac{1}{\bar{n}_3 \bar{h}_2^{\,4}} \int f_k(\xi_1) f_{k'}(\xi_1) d\xi_1 \int \frac{d^2 f_l(\xi_2)}{d\xi_2^{\,2}} \frac{d^2 f_{l'}(\xi_2)}{d\xi_2^{\,2}} d\xi_2
\end{aligned} \tag{19}
$$

As each integrant takes non-zero value only in a limited portion on the boundary surface, we can consider that h_1, h_2 and n_3 take constant values $\bar{h}_1$, $\bar{h}_2$ and $\bar{n}_3$. The equation (18) can be rewritten in a simple quadratic form of the parameter vector $\boldsymbol{a}$ as,

$$r = \boldsymbol{a}^t \boldsymbol{B} \boldsymbol{a} \, . \tag{20}$$

The prior information on the smoothness of a back-slip distribution is represented in a form of probability density function

$$q(\boldsymbol{a}; \mu^2) = (2\pi\mu^2)^{-P/2} \|\Lambda_P\|^2 \exp\left(-\frac{1}{2\mu^2} \boldsymbol{a}^t \boldsymbol{B} \boldsymbol{a}\right) . \tag{21}$$

Here, P is the rank of the matrix $\boldsymbol{B}$, $\|\Lambda_P\|$ represents the product of non-zero eigenvalues of the matrix $\boldsymbol{B}$, and μ^2 is an unknown parameter which controls the strength of the prior constraints on the parameter vector $\boldsymbol{a}$.

We can combine this prior information with the observation equation based on Bayes' theorem. Likelihood of the Bayesian model is given as

$$
\begin{aligned}
l(\boldsymbol{a}; \sigma^2, \mu^2 | \boldsymbol{d}) &= c p(\boldsymbol{d} | \boldsymbol{a}; \sigma^2) \cdot q(\boldsymbol{a}; \mu^2) \\
&= c (2\pi\sigma^2)^{-(N+P)/2} (\alpha^2)^{P/2} \|\boldsymbol{E}\|^{-1/2} \|\Lambda_P\|^{1/2} \exp\left\{-\frac{s(\boldsymbol{a})}{2s^2}\right\} ,
\end{aligned} \tag{22}
$$

with

$$s(\boldsymbol{a}) = (\boldsymbol{d} - \boldsymbol{H}\boldsymbol{a})^t \boldsymbol{E}^{-1} (\boldsymbol{d} - \boldsymbol{H}\boldsymbol{a}) + \alpha^2 \boldsymbol{a}^t \boldsymbol{B} \boldsymbol{a} \, . \tag{23}$$

A new parameter α^2 ($=\sigma^2/\mu^2$) was introduced here.

The likelihood of the model parameter vector $\boldsymbol{a}$ also contains two unknown parameters σ^2 and α^2, which control respectively, the magnitude of data errors and the relative weight of prior information to observed data. These parameters are called hyperparameters, because they are distinguished from ordinary model parameters in the point that they define a parametric model itself.

The problem to be solved here is, first, to find optimum hyperparameters σ^2 and α^2, and then, to get the best estimates of model parameters $\boldsymbol{a}$ for the given values of σ^2 and α^2.

3.2.5. ABIC and optimum hyperparameters

The first thing to do in solving the problem is to find the optimum values of the hyperparameters σ^2 and α^2. Determination of hyperparameters means to select an optimum parametric model from the infinite number of parametric models described by a Bayesian model. For this purpose we can use a Bayesian Information Criterion (ABIC) proposed by Akaike [1977, 1980]. The ABIC minimization criterion is based on the principle of maximizing an information entropy. Practically, ABIC is defined as

$$ABIC(\sigma^2,\alpha^2) = -2\log L(\sigma^2,\alpha^2) + C, \tag{24}$$

where

$$L(\sigma^2,\alpha^2) = \int l(\boldsymbol{a};\sigma^2,\alpha^2|\boldsymbol{d})d\boldsymbol{a} \tag{25}$$

is called the marginal likelihood of σ^2 and α^2. The values of σ^2 and α^2 which are determined so as to minimize ABIC are the best estimates of hyperparameters. From (22) and (25), a concrete form of the marginal likelihood is obtained through the integration with respect to $\boldsymbol{a}$ as

$$L(\sigma^2,\alpha^2)=c(2\pi\sigma^2)^{-(N+P-M)/2}(\alpha^2)^{P/2}\|E\|^{-1/2}\|\Lambda_P\|^{1/2} \\ \times\left\|H^T E^{-1}H+\alpha^2 B\right\|^{-1/2}\exp\left\{-\frac{1}{2\sigma^2}s(\boldsymbol{a}^*)\right\}, \quad (26)$$

where

$$\boldsymbol{a}^*=(H^t E^{-1}H+\alpha^2 B)^{-1}H^t E^{-1}\boldsymbol{d}. \quad (27)$$

Here, $\boldsymbol{a}^*$ is a solution vector of $\delta s(\boldsymbol{a})=0$ for a prescribed value of α^2. From the necessity condition for the minimum of ABIC, the following conditions must hold true.

$$\partial L/\partial\sigma^2=0 \quad (28)$$

$$\partial L/\partial\alpha^2=0 \quad (29)$$

From (28), the next relation is analytically obtained.

$$\sigma^2=s(\boldsymbol{a}^*)/(N+P-M) \quad (30)$$

Substituting the relation (30) into the expression (25) of the marginal likelihood function, we can express ABIC is expressed as a function of α^2,

$$ABIC(\alpha^2)=(N+P-M)\log s(\boldsymbol{a}^*)-P\log\alpha^2+\log\left\|H^t E^{-1}H+\alpha^2 B\right\|+C \quad (31)$$

Search for the minimum of ABIC is conducted by trial and error changing α^2. As ABIC is usually a rather smooth function of α^2 (Figure 17), the minimum of ABIC is easily found. After the optimum α^2 is found, (30) yields the best estimation of σ^2.

3.2.6. Best estimation of model parameters

Once hyperparameters are fixed to their optimum value $\hat{\sigma}^2$ and $\hat{\alpha}^2$, the

Bayesian model gives a likelihood function of model parameters. Then the best estimation of model parameters can be obtained based on the maximum likelihood principle. Maximization of $l(\boldsymbol{a};\hat{\sigma}^2,\hat{\alpha}^2|\boldsymbol{d})$ is realized by minimizing $s(\boldsymbol{a})$. Then the best estimation of the model parameters is obtained by solving an equation $\delta s(\boldsymbol{a}) = 0$, which is written in a more concrete form as

$$H'E^{-1}(d - Ha) + \hat{\alpha}^2 Ba = 0\,. \tag{32}$$

The solution of this equation is given by

$$a^* = (H'E^{-1}H + \hat{\alpha}^2 B)^{-1} H'E^{-1}d\,. \tag{33}$$

Then, $s(\boldsymbol{a})$ can be rewritten as

$$s(a) = s(a^*) + (a - a^*)'(H'E^{-1}H + \hat{\alpha}^2 B)(a - a^*)\,. \tag{34}$$

From (22) and (34) we can see that the posterior distribution of model parameters $\boldsymbol{a}$ is Gaussian and so the covariance matrix is given as

$$C = \hat{\sigma}^2 (H'E^{-1}H + \hat{\alpha}^2 B)^{-1}. \tag{35}$$

Chapter 4. Coupling Distribution along the Nankai Trough

The crustal deformation data of Shikoku are analyzed with the back-slip inversion technique in order to estimate the strength distribution of interplate coupling between the Philippine Sea plate and the Eurasian plate along the Nankai trough. Analyzing various datasets for different stages of the earthquake cycle in the same way, similarities and differences between various stages of the earthquake cycle are discussed. Summarizing the results of these analysis, we can reproduce the history of relative displacement on the plate boundary surface.

4.1. Model Parameters and Conditions

A rectangle region of 255 km×176 km is taken on the earth's surface. The projection of this region onto the plate boundary surface defines the source region for all the datasets (Figure 18). The rectangle region is taken so that the longer side of the rectangle become parallel to the strike (N65°E) of the Nankai trough. The rectangle is delineated so as to correspond to the plate boundary depth of 10 km to 40 km beneath Shikoku, which is thought to include the most part of strong interplate coupling for this subduction zone. As only the plate boundary along Shikoku is treated in the present study, the whole source region of the 1946 Nankaido earthquake cannot be covered. Crustal deformation in Kii Peninsula must be analyzed for that purpose, but the effect of 1944 Tonankai earthquake cannot be neglected in that case. The rectangle is also delineated so as not to be too far from data points, because the resolution of the inverted back-slip distribution will become worse and worse as a source become apart from data points.

Configuration of the plate boundary surface is determined by the hypocenter distribution of microearthquakes. Owing to the efforts of improving seismic observation network, the hypocenter distribution in and around Japan is resolved very well. Hypocenter distributions from Mizoue et al. [1983], Okano et al. [1985] and Miura et al. [1991] are referred for the plate boundary depth and make a dataset of iso-depth contour's trajectory on the plate boundary surface for every 10 km. The plate boundary surface is described by a linear combination of bicubic B-splines to satisfy the dataset. The observation equations to determine the depth of the plate boundary surface are

$$h(x_n) = \sum_{k=1}^{K_S} \sum_{l=1}^{L_S} b_{kl} \Phi_{kl}(x_n) + e_n \quad (n = 1, N_S) . \tag{36}$$

Here, $h(x_n)$ is a plate boundary depth at x_n, $\Phi_{kl}(x_n)$ are the bicubic B-splines defined in (11) and (12), and e_n is a random error containing both a reading error and a modeling error. The depth of the plate boundary surface is determined for the rectangle region of 360 km × 300 km, covering the model source region. The total number of B-splines used for the description of the plate boundary surface is 21 × 16 = 336. Parameters b_{kl} are estimated based on the ABIC minimization principle with a prior constraint that the plate boundary surface is smooth. Figure 18 shows the iso-depth contours of the plate boundary surface expressed by a linear combination of bicubic B-splines. The slope of the plate boundary along the Nankai trough has a low dip angle of about 10 degree under Shikoku and it becomes steeper under the Kii strait. Referring to the microearthquake distribution, overall depth resolution of the plate boundary surface is as good as within several kilometers from the width of hypocenter distribution except for the southwestern part of Shikoku. In the southwestern part of Shikoku we have only a sparse hypocenter distribution because of the shortage of observation points and maybe because of low seismicity. Also in the sea area, the plate boundary surface is poorly resolved. Observations using ocean bottom seismometers are necessary to improve the configuration of the plate boundary surface.

The back-slip distribution, which represents the strength of interplate coupling on the plate boundary surface, is expressed by the linear combination of bicubic B-splines as stated in the Chapter 3. For simplicity of calculation, first, we take grid points on the flat earth's surface. Then the grid points and the corresponding small rectangles are projected onto the curved plate boundary surface. As only a tangential slip is treated, there are two independent components of the back-slip vector. We take a strike-slip component and a dip-slip component as the two independent components. It should be noted that the direction of strike-slip or dip-slip differs from subfault to subfault, because the plate boundary is not a plane but a curved surface. So a unit strike-slip vector and a unit dip-slip vector must be calculated for each subfault.

The grid intervals are taken to be 15 km in the direction parallel to the trough axis and 14.5 km in the perpendicular direction. Then the model source region is divided into $17 \times 12 = 204$ sub-regions. $20 \times 15 = 300$ bicubic splines are necessary to describe the distribution of each slip component over the whole region. Since the strike-slip and dip-slip components are independent with each other, the total number of model parameters is 600. The number of parameters is fairly big relative to the number of data. Then it is necessary to introduce prior constraints on the parameters. The ABIC minimum criterion can lead us to the optimum balance between the information from observed data and the prior information.

Considering that the interseismic strain accumulation is due to the relative plate motion, the direction of back-slip should lie within a narrow range around the direction of the relative motion between the Philippine Sea and the Eurasian plates. This prior information can be incorporated into the inversion analysis by adding an inequality condition of

$$\theta_P - \delta\theta \le \tan^{-1}(u_D / u_S) + \phi \le \theta_P + \delta\theta \tag{37}$$

for all the points on the plate boundary surface. Here, θ_P denotes the direction of relative plate motion, u_D and u_S are the magnitude of the dip slip and the strike slip respectively, and ϕ is the strike at the source region. $\delta\theta$ prescribes the width of acceptable range and is set as 45° in this analysis. Lawson and Hanson [1974] developed algorithms named NNLS (non-negative least squares) and LDP (least distance programming) for solving least squares problems with linear inequality constraints (LSI). These algorithms are adopted in order to incorporate the above linear inequality constraint. As the relative motion between the Philippine Sea and the Eurasian plates, we assumed 5 cm/year in N50°W, according to Seno [1977] and Seno et al. [1993].

4.2. Back- Slip Distribution during an Earthquake Cycle

The datasets of crustal deformation rates are analyzed to estimate the back-slip rate distributions on the plate boundary surface for the preseismic, postseismic, interseismic-1 and interseismic-2 stages. As for the coseismic stage, since instantaneous deformation at the seismic event is considered to be big in comparison with steady deformation during the period, the deformation field can be treated as a coseismic one. Thus, for the coseismic period, the fault slip distribution on the plate boundary surface is estimated. The results of inversion analysis are shown in Figures 19(a)~(e). Discussion about the result for each stage follows below.

4.2.1. Preseismic stage

Figure 19(a) shows the inverted back-slip rate distribution for the preseismic stage (1886/96 ~ 1929/37). The back-slip rate takes the maximum value at the center of Tosa Bay where the depth of the plate boundary surface is about 20 km. The maximum back-slip rate is about 6 cm/year in N40°W direction. Considering the estimation error (~ 1cm/year) and the relative plate motion (5 cm/year in N50°W, according to Seno et al. [1993]), the Philippine Sea and the Eurasian plates have been coupled almost completely under Tosa Bay. The back-slip rate decreases to 2 cm/year at the depth of 30 km. The back-slip rate also decreases in the eastern part, around the Kii strait, though the resolution is poor. Geodetic data along the eastern coast of Kii Peninsula will be necessary to improve the resolution in this area. The directions of most back-slip vectors lie within N30°~50°W, which is consistent with the relative plate motion at the Nankai trough. As is mentioned later, the back-slip distribution for the preseismic stage looks quite similar to that for the interseismic stages. Because the interval of geodetic surveys is too long and because the deformations just before the 1946 event are not included, there aren't any significant difference in the back-slip distribution between the preseismic stage and the interseismic stages. In that sense, we cannot find any short-term precursor of the 1946 event from the back-sip distribution.

4.2.2. Coseismic stage

Figure 19(b) is not a back-slip distribution, but a forward slip distribution for the coseismic stage (1929/37~1947, vertical; 1891/1901~1949/52, horizontal) including the occurrence of Nankaido event in 1946. It should be noted that this coseismic slip distribution is likely to be underestimated because the crustal deformation data include the preseismic and the postseismic deformations, as mentioned in 2.3.2. There are many studies on the source model of the earthquake based on geodetic data and/or tsunami data [Fitch and Scholz, 1972; Ando, 1975; Ando, 1982; Yabuki and Matsu'ura, 1992; Satake 1993]. Though the used datasets in the present study are almost the same as previous studies, there exist significant differences in some cases. The estimated coseismic slip distribution in the present study (Figure 19(b)) is similar to that of Yabuki and Matsu'ura's [1992] in the position of the maximum slip (just west off Muroto), the maximum amount of slip (6~7m), and the general pattern of slip distribution under Shikoku, although Yabuki and Matsu'ura analyzed Kii Peninsula and Shikoku simultaneously. On the other hand, some previous studies like Fitch and Scholz [1972] gave significantly different results, because of both lack of flexibility in parametric models and inappropriate assumptions. For example, the dip angle of 30 degree, which is assumed in Fitch and Scholz, is too high for the plate boundary beneath Shikoku.

The main part of the coseismic slip is confined to the area on the plate boundary surface shallower than 30 km. This slip magnitude takes the maximum value of 6.6 m at west off Muroto. If the relative velocity between the Philippine Sea and the Eurasian plates is assumed to be 5 cm/year, the fault offset of 6.6 m corresponds to a recurrence time of 132 years, which almost agrees with the average recurrence interval (~ 120y) estimated from historical records. However, there are some factors increasing ambiguity of this estimation. One factor is a no-uniform slip distribution on the plate boundary. If the slip distribution is averaged for the whole source area, the coseismic slip will become much smaller. Another factor is that the present inverted results gives only a minimum estimate of coseismic slip distribution, because the analyzed data contains an effect of strain accumulation process as mentioned before. Moreover, the slip distribution has very poor resolution for the sea area. Comparing the present result with that of Satake [1993], who has analyzed tsunami data as well, the estimated source region in the present study is confined to a rather small area near the land of Shikoku. That is simply because submarine crustal deformation in offshore is sensitive to tsunami data, but not to geodetic data on land. The inversion scheme used here tends to assign null values to parameters which are not resolved well by observations. As for the offshore slip distribution, Satake [1993] seems to propose more reliable results. On the other hand, we should also think of a spray fault system for the shallower part, which was not incorporated even in the analysis of Satake. Unfortunately, we don't have data for resolving a motion of such a complicated fault system at least for the 1946 event.

The slip vectors are directed around N110°E, a little rotated counterclockwise from the value expected from the relative plate motion. The discrepancy from the relative plate motion may be ascribed to the errors contained in the geodetic dataset. On the other hand, recent GPS observation results shows that horizontal displacement vectors at Mutoro Promontory relative to Ikeda, Tokushima Prefecture is in the WNW (N70°W) direction [Tabei et al., 1994], which is consistent with the estimated coseismic slip vectors. Then, it is possible that actual relative plate motion should be a little rotated counterclockwise from the assumed direction.

4.2.3. Postseismic stage

Figure 19(c) is the postseismic (1947~1968/71) back-slip distribution. An inequality constraint on the slip direction is not used in the inversion analysis of this stage, because we can expect both back-slip (stress accumulation) and fore-slip (after slip) for this period. The estimated back-slip distribution shows a relatively big fore-slip (negative back-slip) distribution up to 12 cm/year, which suggests a postseismic slip on the plate boundary. The maximum fore-slip is obtained at southeast off Muroto though the estimation error is also as large as 10 cm/year there. As is stated repeatedly, the offshore back-slip distribution is not so reliable. However, we can notice some interesting points in this distribution. The first point is a small back-slip rate in the main area of the coseismic rupture (Figure 19(b)). It is considered that the coseismic rupture area can neither release nor store tectonic strain in this period, because there remains no stress to be released and the coupling strength has not been recovered just afetr the earthquake. On the other hand, the second point to be noticed is that a significant fore-slip occurs at the deeper extension of the coseismic rupture area whose depth is about 30 km. The significantly large fore-slip area extends downward along the plate boundary. As is mentioned before, the northern coast of Shikoku subsided very rapidly from 1947 to 1950. It is very plausible that this subsidence was caused by an after slip at the deeper part of the plate boundary. If the main process of the after slip finished until 1950, the actual fore-slip rate will become much larger.

By the way, it is expected that a large part of the postseismic deformation is caused by viscoelastic stress relaxation in the asthenosphere, which is not incorporated in the present analysis. Then, in order to estimate the viscoelastic effect after the 1946 event, a model calculation is conducted by using the source parameters of the 1946 event estimated here. The calculation of viscoelastic response follows Matsu'ura and Sato [1989], which are slightly modified version of Matsu'ura et al. [1981]. The mathematical formulation of viscoelastic deformation due to a dislocation source is summarized in Appendix B.

The structural model used for the viscoelastic calculation is shown in Table 9, where the thickness of the lithosphere is set to 35 km based on several seismological studies using surface waves [Kanamori and Abe, 1968; Yoshii et al., 1974; Seekins and Teng, 1977]. A pattern of viscoelastic response is very sensitive to the lithosphere thickness H. In the present case, as the coseismic rupture of a brittle regime extended deeper than 30 km, the elastic lithosphere must be thicker than 30 km. Figure 20(a) shows the calculated viscoelastic deformation rate during 20 years after the 1946 event for the present source model in the case of $H = 35$ km. For reference, Figures 20(b) and 20(c) show the results in the cases of $H = 40$ km and 50 km, respectively. The characteristic wavelength of the calculated deformation pattern becomes larger as the lithosphere becomes thicker. The viscoelastic effect can be regarded as the effect of imaginary source located at the mirror point of the actual source in the lithosphere across the boundary between the lithosphere and the asthenosphere. Then, the thicker the lithosphere is, the deeper the imaginary source is located, which cause a longer characteristic wavelength of the deformation pattern. The deformation pattern of Figure 20(a) is intuitively an unexpected one, in which the vertical motion occurs in the same sense as coseismic one in a wide area. However, since the viscoelastic deformation pattern is always complex, such a case sometimes occurs as demonstrated in Matsu'ura et al. [1981].

Subtracting the viscoelastic response in Figure 20(a) from the observed data in Figure 8(c), and we get the distribution of corrected vertical deformation rates as shown in Figure 21. Comparing Figure 21 with Figure 8(c), we can see that the postseismic uplift in south Shikoku is intensified and the uplift center moves to northern coast of Tosa Bay from the surroundings. On the other hand, there still remains rapid subsidence along the northern and eastern coast of Shikoku.

The corrected vertical deformation rates (Figure 21) are inverted in order to estimate the back-slip distribution. The obtained result (Figure 22) is not so different from Figure 19(c), but a larger fore-slip just below Shikoku and a back-slip below Tosa Bay denoting a strain accumulation appear as new features. The largest fore-slip rate is about 12 cm/year and it corresponds to the fault offset of 2.4 m for 20 years. Temporal variations in tidal records at Takamatsu and Komatsushima suggest that this after slip was concentrated between 1947 and 1950. So the most part of the after slip occurred within 5 years after the event. Both Figure 19(c) and Figure 22 show large fore-slip distribution below the Kii strait. In spite of poor resolution in this area, together with the fact that coseismic rupture didn't occur below the Kii strait [Yabuki and Matsu'ura, 1993] and there was an activation of seismicity after the 1946 Nankaido earthquake, this large fore-slip suggests that the stress is relaxed aseismically below the Kii strait

4.2.4. Interseismic-1 stage

The inverted back-slip distribution (Figure 19(d)) shows a typical strain accumulation pattern again in the interseismic-1 stage (1968/71 ~ 1981/83). The maximum back-slip rate is about 6.5 cm/year at the center of Tosa Bay. The back-slip rate decreases downward along the plate boundary and takes a small value under eastern Shikoku along the Kii strait. These features are very similar to those in the preseismic stage. As discussed in 2.3.4., however, note that the second phase of two-step uplift, which is possibly mean the occurrence of the healing or the aseismic slip on the plate boundary surface, is observed at the northern coast of Tosa Bay, which cannot be resolved by the dataset now used.

The direction of back-slip is N30°W, which is different from the relative plate motion at the Nankai trough by 20 degree. If the back-slip vector contains an estimation error of 35 % of its magnitude, the direction can change up to 20 degree ($\tan 20° \cong 0.35$). It is possible that this discrepancy means the degree of estimation error in the analysis. The strongly coupled region is shallower than 30 km, and the back-slip rate takes the maximum value at the depth of 20 km.

4.2.5. Interseismic-2 stage

The back-slip distribution in the most recent period is shown in Figure 19(e).

The maximum back-slip rate is 4.8 cm/year at southwest off Muroto, where the plate boundary depth is about 17 km. The magnitude is a little smaller and the location shifts eastward slightly comparing with the interseismic-1 stage. But, considering estimation errors, the difference between these two stages is not significant. Conjecturing from the crustal deformation patterns, the interseismic-2 stage differs from the interseismic-1 stage in the dissipation of postseismic uplift motion along Tosa Bay, which cannot be described well by the back-slip model. The depth extension of the strongly coupled area is very similar to the interseismic-1. No significant signals, which might be a precursor to the next seismic event, are found. The interplate coupling is very weak on the plate boundary deeper than 30 km. This depth-dependent characteristic in coupling strength seems to be common during the whole earthquake cycle at the Nankai trough except for the postseismic stage.

The back-slip distribution in this stage is the only one based on both vertical and horizontal data. Since the line strain data are incorporated, it is expected that the direction of the back-slip is more trustworthy. It is interesting that the back-slip vectors rotate counterclockwise around Muroto. The back-slip vectors are directed to N30°W in the eastern part and to N70°W in the western part. This direction change may be an effect of large slab pull at the Bungo strait, where the Philippine Sea plate is subducted with a higher angle.

4.3. Spatio-temporal Variation in Actual Slip on the Plate Boundary

From the results in Chapter 4.2., we can obtain the cross section views of actual slip or slip rate distributions on the plate boundary. The back-slip rate is interpreted as a peturbation from a steady slip along the plate boundary, corresponding to the relative plate motion. Therefore the actual slip rate on the plate boundary surface is given by the sum of the back-slip rate and the steady slip rate. It should be noted that the back-slip has the opposite sign to the steady slip. The cross section views of the actual slip (rate) distributions along the lines A-a, B-b, C-c, and D-d in Figure 18 are shown in Figures 23(a)~(d), respectively. The coseismic slip distribution is given at the top of each figure, and the back-slip rate distribution of the other stages are given below in order of the progress of the earthquake cycle.

In the present study, the relative motion of the Philippine Sea plate to the Eurasian plate is taken to be 5.0 cm/year in the N50°W direction [Seno et al., 1993]. The corresponding steady slip rate is denoted by a horizontal dashed line in each plot. It is clear that the deeper limit of the strongly coupled region (with actual slip rates) is around 30 km in depth in the interseismic and preseismic stages. This is a common characteristic for 4 plots in Figure 23.

If the coupling strength of the plate boundary changes like Figure 23, we can expect that the nucleation process of a large earthquake occurs around the depth of 30 km as a result of the relative displacement change along the plate boundary surface. It is consistent with the simulation result of Stuart [1988]. So it is considered that earthquakes at subduction zones would start from a deeper part of plate boundaries. However, about the 1946 Nankaido earthquake, seismic rupture is thought to start from a shallower part, just south of Kii Peninsula [Kanamori, 1972]. Although the origin of the initial rupture at the shallow part under Kii Peninsula is not clear, we can understand the seismic rupture under Shikoku at the 1946 Nankaido earthquake was induced by the rupture under Kii Peninsula. It is considered that the plate boundary under Shikoku had accumulated enough seismigenic energy at that time, but the accumulated energy was not enough for generating a dynamic rupture by itself. When we deal with the earthquake cycle along subduction zones, such a chain reaction is very important and it can affect the recurrence interval of earthquakes.

Then, summing up the relative displacements for all stages, we can obtain the cross section views of the actual slip history on the plate boundary surface. The ccumulative plots of the actual slip along A-a, B-b, C-c, and D-d during the past 100 years are shown in Figures 24(a) ~ (d), respectively. The relative displacements for the preseismic, the postseismic, and the interseismic stages are obtained by multiplying the magnitude of the slip rate in Figures 23(a) ~ (d) by the corresponding stages. The coseismic displacement is simply added to the profiles. In Figures 24(a) ~ (d), the slip distribution at the end of the 19th century, when the geodetic surveys began, is taken as a reference state. Each curve denotes the ccumulative relative displacement from the reference state, and the difference between the adjacent curves denotes the amount of relative displacement during the corresponding stage of the earthquake cycle.

The steady slip caused by the relative plate motion must be consumed everywhere on the plate boundary surface on a long term average. In this sense, the whole plate boundary should slip by the same amount after the completion of one earthquake cycle. Since the resolution of the inversion analysis is not so good, this condition isn't necessarily satisfied over the whole plate boundary. Neverthless all four profiles in Figure 24 show rather flat curves under the land area after the interseismic-2 stage, which means that the almost whole cycle has finisfed. From Figure 24(c), which is a profile C-c across Muroto, we can see that the preceding aseismic slip at depths (PRE) amounted to 2m, and then the 1946 Nankaido earthquake brought about the coseismic slip of 6 m at an intermediate depth (CO). The postseismic slip occurred around the depth of 30 km (POST) and it amounted to 3 m at its maximum, while only a small slip occurred in the intermediate depth. In the two interseismic stages, the steady aseismic slip takes place at the rate of relative plate motion only at the deeper part just like the preseismic stage. For the whole earthquake cycle, the whole plate boundary slips about 6 m, which corresponds to the total amount of the relative plate motion during the recurrence time of 120 years.

In the western part, the profile A-a shows a rather large postseismic slip at depths in comparison with the coseismic slip. The shallower part of the profile A-a, which is poorly resolved because of the lack of data, shows a remarkable delay of the relative displacement on the plate boundary surface. From the analysis of the 1946 event, it is probably true that the plate boundary of this portion didn't slip so much. Thus, although it is only a speculation, the area off west Shikoku might have a potentiality to generate a large earthquake. The profiles B-b can be interpreted in the same way as the profile C-c. However, the easternmost profile D-d is a little problematic. The source area of the 1946 event slips as much as 2 m in the postseismic period, and the total amount of slip there is 7 m. Because this area is poorly resolved, we can ascribe it to the estimation error. Nevertheless, this large postseismic slip might have some relation to the activation of seismicity in the Kii strait after the 1946 event [Utsu, 1957].

There have been some studies which simulated the whole process of earthquake cycles at transform faults or subduction zones with a fault constitutive law. Tse and Rice [1986] and Stuart [1988] are examples of those studies. In those studies, relative displacement history on the plate boundary surface was simulated by using a rate- and state-dependent friction law. For example, Figure 7 of Tse and Rice and Figure 5 of Stuart are similar to Figures 24(a) ~ (d) of the present study. The difference is that the plots of the present study are obtained from the inversion analysis of actually observed data. Because of insufficient data, the results of this inversion analysis contain considerable errors and the temporal resolution is poor. Nevertheless, the present study has great significance because the observation of crustal deformations is improved day by day now. For example, a GPS continuous monitoring network can provide us for the precise location of stations everyday. So, it is possible to draw a daily profile of relative displacement like Figure 24. We may be able to detect some precursory phenomena expected by model calculation prior to the occurrence of an earthquake.

Chapter 5. Discussion

Depth extension of interplate coupling

The depth extension of the strongly coupled area along the Nankai tough is bounded by the depth of 30 km through the interseismic and preseismic stages. The source region of the 1946 Nankaido earthquake is bounded similarly. Therefore, it is concluded that the strain accumulation on the plate boundary surface under Shikoku is limited to the shallow (< 30 km) part and that the deeper part moves more smoothly without accumulation of seismogenic strain.

It is considered that such a depth dependency of interplate coupling is due to the change of material property, that is, brittle-ductile transition. Thermal structure of the earth's interior, increasing confining pressure, existence of H_2O, and some other factors may contribute to this change. Shimamoto [1990] proposed the hypothesis about the depth-dependent interplate coupling for the plate boundary in the Tohoku district. According to the Shimamoto's hypothesis, the interplate coupling in the shallower part of the plate boundary is weakened by the enormous amount of H_2O provided by metamorphism and dehydration of rocks. On the other hand, the deeper part of the plate boundary is decoupled by the ductile property due to high temperature. The results of the present study suggest that the temperature on the plate boundary become high enough for rocks to be ductile at the depth of 30 km. Dragert et al. [1993] and Hyndman et al. [1995] have analyzed the crustal deformation data to estimate a thermal structure of the plate boundary at the Cascadian subduction zone and the Nankai trough. They concluded that the interplate coupling is weakened at the temperature of 350°C and vanishes at 450°C for both subduction zones, and that the plate boundary depth of this transition is around 30 km for the Nankai trough. It is a matter of course that the present result agrees with Hyndman et al.'s well, because both results are based on the same data basically. The important thing is that they got the same temperature 350°C for different subduction zones. Thus we can deduce that the temperature of 350°C is a common limit for every subduction zone to cause interplate coupling. According to Shimamoto[1990], the seismogenic zone in the Tohoku district extends from about 30 to 60 km in depth. The extension of a strongly coupled region is significantly different between Shikoku and Tohoku. Then it is maybe concluded that this difference is due to different thermal structures, that is, the different ages of the Philippine Sea and the Pacific plates. The Pacific plate, which is being subducted under Tohoku, was formed at the Eastern Pacific rise and has moved to the Japan trench. So the Pacific plate is not only older than the Philippine Sea plate but also more cooled. Then, the isotherm of 350°C can reach the deeper part on the plate boundary in the Tohoku district. The Philippine Sea plate is a young (15 ~ 30 Ma) plate and relatively hot, so that it is easily heated.

As for the shallow part (< 15 km), the crustal deformation data cannot resolve the strength of interplate coupling well. According to the joint inversion of geodetic data and tsunami data for the 1946 event by Satake [1993], the coseismic slip extended to the shallower part of the plate boundary, that is, nearly to the surface. However, it is not clear how shallow the limit of the plate coupling is, because the shallow decoupled plate boundary would easily slip passively regardless of the accumulated strain. It is likely that the effect of such a fault slip is reflected on the Satake's result.

Temporal variations in the slip rate distribution

There is very little information about the temporal variations in the slip rate distribution. It is due to poor temporal resolution of the geodetic data in the past. It is easy to distinguish the coseismic and postseismic stages from other stages, but no significant difference is found between the interseismic and preseismic stages. In particular, data of the interseismic periods with a short duration (~ 10 years) has a rather bad S/N ratio because of small signals, though there have been some improvements in surveying technology.

On the other hand, the geodetic data of the coseismic and postseismic stages contain effects from several different phases. As is stated in 2.3.2. and 2.3.3., coseismic horizontal vectors contain preseismic deformation as well, and postseismic deformations have some exponentially decaying phases which are different each other in their time constants. Some studies like Miyashita [1987] took a epoch-reduction method for such cases but there is no significant difference between their results and the present study.

Utilization of space technology like GPS and VLBI has improved the accuracy of geodetic observation very much and quasi-real-time monitoring of crustal deformations by GPS has been in operation recently in Japan [Abe and Tsuji, 1994]. This drastic change will enable us to catch even small signals from a part of plate boundary surface and we can resolve temporal variations in the slip-rate distribution much better.

Comparison with the Suruga trough

Yoshioka et al. [1993, 1994] analyzed the recent crustal deformation data to estimate the spatial distribution of interplate coupling in the Tokai and Kanto districts. Their main result for the Tokai district is that a strongly coupled region is extending form 10 to 30 km in depth and the maximum back-slip rate is 4.0 cm/year in the average direction of N30°W. The result of Shikoku agrees with that of Tokai in the depth extension of a coupled region, but the interseismic back-slip rate in Shikoku is 1.2 ~ 1.5 times larger than that of Tokai. Considering the error level, significance of the difference in the back-slip rate is disputable. But it is likely that the back-slip rate differs from Shikoku to Tokai because of the difference in the relative plate motion. According to Seno et al. [1993], the magnitude of the relative plate motion vectors increases from Tokai to Shikoku along the Suruga-Nankai trough. The Philippine Sea plate is being subducted from the Suruga-Nankai trough, while the northern tip of the plate, Izu Peninsula, is colliding with the Japan island arc, and the plate subduction has not been confirmed there from seismicity [Ishida, 1992]. Therefore, the motion of the Philippine Sea plate is considered to be decelerated or locked at the northern tip, and the relative plate motion is decreased at the adjacent Tokai area. Shikoku is situated about 300 km west of the Tokai area so that the relative plate motion in Shikoku is not influenced by the collision of Izu Peninsula. Moreover, the Nankai trough is contiguous to the Ryukyu trench which have a higher angle subduction. So we can expect a larger slab pull force at the western part of the Philippine Sea plate. The faster subduction rate and the larger back-slip rate in Shikoku can be attributed to this drag force.

The direction is not well defined in the present analysis because of the lack of horizontal deformation data. However, the back-slip vectors are directed from N30°W to N70°W in most cases, though the direction constraint allows the range from N5°W to N95°W. The estimated direction is roughly coincident with the relative plate motion between the Eurasian plate and the Philippine Sea plate by Seno et al. [1993]. Yoshioka et al. [1993] mentioned the discrepancy in direction between the relative plate motion and the estimated back-slip is related to the motion of supposed North American plate, which should be called the Okhotzk plate according to Seno et al. [1994] now. Although the result of the present study is not different from Yoshioka et al. [1993]'s significantly in the back-slip direction, the plate subduction at the Nankai trough is considered to be more representative motion of the Philippine Sea plate relative to the Eurasian plate than that at the Suruga trough, because it is not affected by the collision of Izu Peninsula so much. Scattering of the direction and magnitude of back-slip vectors will represent the precision of the analysis.

Back-slip rate, coupling strength and seismic recurrence cycle

The estimated magnitude of back-slip rates for Shikoku is 5 ~ 6 cm/year and the maximum coseismic slip is 6.6 m. The area of the maximum coseismic slip roughly coincides with the area of the maximum back-slip rate. If we simply divide the maximum coseismic slip by the maximum back-slip rate, we get the recurrence interval of 110 ~ 130 years. This value seems to be a fairly good estimation compared with the recurrence interval obtained from historical records.

On the other hand, the seismic moment M_0 or the seismic moment rate $\dot{M}_0$ can be calculated for each stage as follows;

$$M_0 = \mu \int_S \Delta u(\xi) dS(\xi), \qquad \dot{M}_0 = \mu \int_S \Delta \dot{u}(\xi) dS(\xi) \tag{38}$$

where $\Delta u(\xi)$ is the back-slip distribution on the plate boundary surface and μ is the rigidity of the earth's crust that is assumed to be 5×10^{10} N•m^{-2} here. The estimated moments and moment rates are summarized in Table 10. The seismic moment of the 1946 Nankaido earthquake, that is M_0 for coseismic stage of the present study, is 2.3×10^{21} N•m and it corresponds to M_w = 8.2 from an empirical relation between a seismic moment M_0 and a moment magnitude M_w [Kanamori, 1977],

$$\log M_0 = 1.5 M_w + 9.1. \tag{39}$$

Ando [1982] divided the whole source region of this earthquake into two parts and obtained $M_0 = 3.2\times10^{21}$ N•m for the western part near Shikoku. So the results of the present study is thought to be consistent with the Ando's model in this point.

The preseismic or the interseismic moment rate, which is thought to represent a strain accumulation rate along this part of the plate boundary, are estimated to be 4.6 ~ 6.2×10^{19} N•m. If we assume the recurrence interval as 120 years, the ratio of seismic moment to total tectonic moment becomes 30 ~ 40 %. This estimation contains the unresolved shallow part and the coseismic moment release could be larger because of the underestimation of the coseismic crustal deformation as discussed in 2.3.2. On the other hand, we can investigate the seismic coupling problem from the relative displacement profiles in Figure 24. Though the estimation of the back-slip distribution is contaminated due to poor resolution, the general characteristics are very consistent each other. The area corresponding to the coseismic slip forms less than a half of the whole area. Instead, there is a considerable strain release at the postseismic stage, which is comparable to the coseismic one. The rate of seismic coupling is almost 100 % at the intermediate depth (20~25 km) and it reduces to 20~30 % at the depth of 30 km. As is obvious from this example, the after slip at the deeper part of the plate boundary plays an important role associated with the interplate coupling along the Nankai trough. As a fairly large part of the strain release takes the form of this after slip, the seismic recurrence interval will be much longer without the after slip. We might misunderstand the state of interplate coupling if we neglect the effect of this after slip.

All the plate boundary surface is considered to slip as large as 6 m during the last 100 years including the occurrence of 1946 Nankaido earthquake. Therefore, the present state of the plate boundary surface is the same as that of about 100 years ago Therefore, the next great earthquake at the Nankai trough is expected to occur 40 ~ 50 years after from now.

Viscoelastic Modeling

Significant postseismic crustal deformations were observed after the 1946 Nankaido earthquake [Thatcher, 1984]. There have been several 2-dimensional calculations for this postseismic recovery motion [Thatcher and Rundle, 1984; Miyashita, 1987; Sato and Matsu'ura, 1992]. On the other hand, though many source models are proposed for the 1946 Nankaido earthquake, there has never been 3-dimensional viscoelastic calculation which test the consistency of those models with the postseismic deformations due to the viscoelastic relaxation in the asthenosphere are investigated for the observed postseismic deformations. In the present study, the source models by Ando [1982], Satake [1993] and the present study are investigated. The viscoelastic calculation scheme used here is that of Matsu'ura and Sato [1989], which is a slightly modified version of Matsu'ura et al [1981], is described in Appendix B. The thickness of the lithosphere is assumed to be 35 km [Kanamori and Abe, 1968; Yoshii et al., 1974; Seekins and Teng, 1977]. About the viscosity of the asthenosphere, Matsu'ura and Iwasaki [1983] obtained 5×10^{18} Pa·s from the analysis of an after effect of the 1923 Kanto earthquake. Used structural model of the earth is shown in Table 9.

As Rundle [1982] showed, a crustal deformation pattern is very sensitive to the extension of coseismic rupture within the lithosphere. If the upper half of the lithosphere is ruptured, a main feature of the postseismic deformation is subsidence. On the other hand, if the lower half of the lithosphere is ruptured, uplift around the fault is a main feature. Then close look into the postseismic deformation pattern can determine the depth extension of seismic rupture.

Principal characteristics of the postseismic stage are the exponential subsidence at Muroto revealed by leveling surveys [Geographical Survey Institute, 1972] and the exponential uplift in the northern coast of Tosa Bay revealed by tidal observation. Looking at the postseismic deformation pattern expected from three source models, all of them cannot explain the actual situation well (Figure 25(a) ~ (c)). We can explain this discrepancy by introducing an after slip effect as we showed in the present study. The necessity of the after slip effect in Shikoku was pointed out by Sato and Matsu'ura [1992] based on the forward modelling of vertical deformation pattern.

Although the after slip in the deeper part is thought to be a principal origin of this discrepancy, there are other candidates. Insufficient knowledge about the earth's rheological structure may cause the discrepancies between the calculations and the observations. Another candidate is an inappropriate source model of the 1946 event. Geodetic data are contaminated with various errors as mentioned before and estimation processes are also responsible. Including the present study, all the source models should be checked out whether they are consistent with both the coseismic crustal deformations and postseismic ones. All the source models of the 1946 Nankaido earthquake have not been checked in that way so far. Therefore, constructing such a complete source model will become our future work.

Chapter 6. Conclusion

As concluding remarks, the main results of the present study are summarized as follows:

1. Based on the geodetic data in Shikoku for the nearly one earthquake cycle at the Nankai trough, crustal deformation patterns for various stages in the earthquake cycle are described. The crustal deformation patterns are roughly classified into three types, that is coseismic, postseismic and interseismic. There are no clear differences between the interseismic and preseismic stages. Seismogenic strain is accumulated in the interseismic stage and it is instantaneously released coseismically. Deformation patterns of these two stages are reasonably interpreted by elastic modeling. On the other hand, a viscoelastic effect isn't negligible in the postseismic stage. But even if the viscoelastic response is corrected, there surely remains considerable crustal deformations throughout Shikoku, which is considered to be caused by the after slip at the deeper part following 1946 Nankaido earthquake. This subsidence looks irreversible, though the main part of deformations in south Shikoku is rather cyclic.

2. The obtained crustal deformation patterns are inverted by back-slip inversion technique with ABIC minimization principle in order to estimate interplate coupling on the plate boundary. The representative value for an interseismic back-slip is 5 ~ 6 cm/year in the NW direction, which is consistent with the relative plate motion between the Philippine Sea plate and the Eurasian plate. As far as the data analysis of this study is concerned, no significant temporal changes in the coupling strength cannot resolved among the interseismic stages. On the other hand, the spatial variations in the coupling

strength is clear. The area of the strong coupling is confined to the shallow (<30 km) part. This limit depth is thought to correspond to the depth where brittle-ductile transition begin as a result of high temperature.

3. The history of the actual slip on the plate boundary surface is reproduced based on the back-slip inversion results. Plots of the cumulative relative displacement on the plate boundary surface shows that an after slip forms a considerable part of the total displacement, and that the after slip plays an important role in determining the recurrence interval of large earthquakes at the Nankai trough. The present state of the plate boundary surface is thought to be almost the same as that of 100 years ago when geodetic surveys were started in Japan. Therefore, it will take 40 to 50 years for the next great earthquake at the Nankai trough to occur. As the extensive and intensive GPS network for continuous observation comes into operation recently, we can improve the resolution of the result drastically and quasi real-time monitoring of the interplate coupling on the plate boundary will become possible in the near future, which might have a possibility to detect precursors to the next great earthquake at the Nankai trough.

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Appendix A

According to Yabuki and Matsu'ura [1992], surface deformation field of an isotropic, homogeneous, elastic half space due to a moment tensor density $m_{pq}(\xi)$ on a fault plane S is expressed as

$$u_i(\boldsymbol{x}) = \int_S G_{ip,q}(\boldsymbol{x},\xi) m_{pq}(\xi) dS(\xi) \quad (i=1,2,3) \tag{A1}$$

where $G_{ip,q}$ is a spatial derivative of Green's tensor G_{ip} with respect to ξ_q. Analytical expressions for $G_{ip,q}$ are given as follows:

$$G_{11,1} = \frac{1}{4\pi\mu}\frac{X}{R^3}\left[F_1\frac{X^2}{R^2} + \frac{3\gamma}{(z+1)^2} - 1\right], \tag{A2}$$

$$G_{12,2} = \frac{1}{4\pi\mu}\frac{X}{R^3}\left[F_1\frac{Y^2}{R^2} + \frac{\gamma}{(z+1)^2} - 1\right], \tag{A3}$$

$$G_{13,3} = \frac{1}{4\pi\mu}\frac{X}{R^3}\left[3z^2 + \gamma - 1\right], \tag{A4}$$

$$G_{11,2} + G_{12,1} = \frac{1}{2\pi\mu}\frac{Y}{R^3}\left[F_1\frac{X^2}{R^2} + \frac{\gamma}{(z+1)^2}\right], \tag{A5}$$

$$G_{11,3} + G_{13,1} = \frac{3}{2\pi\mu}\frac{X^2}{R^4}z, \tag{A6}$$

$$G_{12,3} + G_{13,2} = \frac{3}{2\pi\mu}\frac{XY}{R^4}z, \tag{A7}$$

$$G_{21,1} = \frac{1}{4\pi\mu}\frac{Y}{R^3}\left[F_1\frac{X^2}{R^2} + \frac{\gamma}{(z+1)^2} - 1\right], \tag{A8}$$

$$G_{22,2} = \frac{1}{4\pi\mu}\frac{Y}{R^3}\left[F_1\frac{Y^2}{R^2} + \frac{3\gamma}{(z+1)^2} - 1\right], \tag{A9}$$

$$G_{23,3} = \frac{1}{4\pi\mu}\frac{Y}{R^3}\left[3z^2 + \gamma - 1\right], \tag{A10}$$

$$G_{21,2} + G_{22,1} = \frac{1}{2\pi\mu}\frac{X}{R^3}\left[F_1\frac{Y^2}{R^2} + \frac{\gamma}{(z+1)^2}\right], \tag{A11}$$

$$G_{21,3} + G_{23,1} = \frac{3}{2\pi\mu}\frac{XY}{R^4}z, \tag{A12}$$

$$G_{22,3} + G_{23,2} = \frac{3}{2\pi\mu}\frac{Y^2}{R^4}z, \tag{A13}$$

$$G_{31,1} = \frac{1}{4\pi\mu}\frac{1}{R^2}\left[F_2\frac{X^2}{R^2} - z + \frac{\gamma}{z+1}\right], \tag{A14}$$

$$G_{32,2} = \frac{1}{4\pi\mu}\frac{1}{R^2}\left[F_2\frac{Y^2}{R^2} - z + \frac{\gamma}{z+1}\right], \tag{A15}$$

$$G_{33,3} = \frac{1}{4\pi\mu}\frac{1}{R^2}\left[3z^3 + (\gamma - 1)z\right], \tag{A16}$$

$$G_{31,2} + G_{32,1} = \frac{1}{2\pi\mu}\frac{XY}{R^4}F_2, \tag{A17}$$

$$G_{31,3} + G_{33,1} = \frac{3}{2\pi\mu}\frac{X}{R^3}z^2, \tag{A18}$$

$$G_{32,3} + G_{33,2} = \frac{3}{2\pi\mu}\frac{Y}{R^3}z^2, \tag{A19}$$

where

$$X = x_1 - \xi_1, \quad Y = x_2 - \xi_2, \quad Z = -\xi_3, \\ R = \sqrt{X^2 + Y^2 + Z^2} \tag{A20}$$

and

$$z = Z / R, \\ F_1 = 3 - \gamma\frac{z+3}{(z+1)^3} \quad F_2 = 3z - \gamma\frac{z+2}{(z+1)^2} \tag{A21}$$

with

$$\gamma = \frac{\mu}{\lambda + \mu}. \tag{A22}$$

Appendix B

Matsu'ura et al [1981] developed a theoretical calculation method for viscoelastic response of a layered half-space due to a buried fault in a elastic layer. Matsu'ura and Sato [1988] and Sato and Matsu'ura [1989] incorporated a gravitational effect and also obtained a description in case that the source exists in a viscoelastic layer. Their results are summarized here.

Suppose a point dislocation with slip time function $H(t)$, is located at (0, 0, h) in a gravitating layered medium, Laplace transform of the vertical component of the surface displacement at (r, ϕ, 0) in a cylindrical coordinate system is obtained from elastic solution in accordance with the correspondence theorem by replacing the slip time function $H(t)$ by $1/s$, and the Lamé constants λ and μ by

$$\hat{\lambda} = \frac{\lambda \tau s + \kappa}{\tau s + 1} \tag{B1}$$

$$\hat{\mu} = \frac{\mu \tau s}{\tau s + 1} \tag{B2}$$

where $\kappa=(\lambda+2\mu/3)/(\lambda+2\mu)$ is the bulk modulus, $\tau=\nu/\mu$ is the Maxwell relaxation time, and s is the Laplace transform variable. The expression of vertical displacement in the s domain is given by

$$\hat{u}_z(r,\phi,s,h) = -\frac{1}{4\pi}\int_0^\infty \boldsymbol{J}'(r,\phi,\xi)[\boldsymbol{K}(h,\xi)\boldsymbol{W}(h,\xi) + \boldsymbol{K}'(h,\xi)\boldsymbol{W}'(h,\xi)]\xi d\xi \tag{B3}$$

where

$$J = \begin{bmatrix} \frac{1}{4}\sin 2\delta J_0(\xi r) \\ -\cos 2\delta \sin \phi J_1(\xi r) \\ \frac{1}{4}\sin 2\delta \cos 2\phi J_2(\xi r) \end{bmatrix} \tag{B4}$$

$$K = ce^{-d\xi} \begin{bmatrix} 2b & -4a & -a & -b \\ a & -b & 0 & 0 \\ 2b & 0 & a & b \end{bmatrix} \tag{B5}$$

$$K' = ce^{-d\xi} \begin{bmatrix} b + 3h'\xi a & 4a + 3h'\xi b & 4a + 3h'\xi b & b + 3h'\xi a \\ h'\xi b & b + h'\xi a & b + h'\xi a & h'\xi b \\ -b + h'\xi a & h'\xi b & h'\xi b & -b + h'\xi a \end{bmatrix} \tag{B6}$$

with

$$a = (1 + e^{-2h'\xi}), \quad b = (1 - e^{-2h'\xi}) \tag{B7}$$

and

$$c = \frac{4}{\prod_{j=1}^{m} (1 + e^{-2H_j\xi})}. \tag{B8}$$

Here, a point source exists in the m-th layer, $J_n(\xi r)$ is the Bessel functions of order n, and h' is the distance from the source to upper boundary of the m-th layer. $\boldsymbol{W}$ and $\boldsymbol{W}'$ are the s-dependent vectors whose elements are given by

$$W_j = \frac{1}{s} \frac{(P_{11} + P_{21})(Q_{3j} + Q_{4j}) - (P_{31} + P_{41})(Q_{1j} + Q_{2j})}{(P_{11} + P_{21})(P_{32} + P_{42}) - (P_{12} + P_{22})(P_{31} + P_{41})}, \tag{B9}$$

and if m-th layer is elastic,

$$W_j' = \frac{1}{s} \alpha_m W_j, \tag{B10}$$

else, if m-th layer is viscoelastic,

$$W_j' = \frac{1}{s}[\alpha_m + (1-\alpha_m)\frac{\kappa}{\tau s + \kappa}]W_j \tag{B11}$$

where

$$P = E_{n+1} \cdot D_n \cdot F_n \cdots D_1 \cdot F_1 \tag{B12}$$

$$Q = E_{n+1} \cdot D_n \cdot F_n \cdots D_{m+1} \cdot F_{m+1} \cdot D_m \tag{B13}$$

with

$$E_{n+1} = \begin{bmatrix} 1 & 0 & 0 & 0 \\ 0 & \hat{\alpha}_{n+1} - 1 & 2 - \hat{\alpha}_{n+1} & 0 \\ -1 & 0 & 0 & 1 \\ 0 & 1 & -1 & 0 \end{bmatrix}, \tag{B14}$$

$$D_j = \begin{bmatrix} 1 & 0 & 0 & 0 \\ 0 & 1 & 0 & 0 \\ 0 & 0 & \hat{\delta}_j & 0 \\ 0 & 0 & 0 & \hat{\delta}_j \end{bmatrix}, \tag{B15}$$

$$F_j = \begin{bmatrix} 1 + \hat{\alpha}_j \beta_j T_j & \hat{\alpha}_j (T_j - \beta_j) - T_j & \hat{\alpha}_j (\beta_j - T_j) + 2T_j & -\hat{\alpha}_j \beta_j T_j \\ \hat{\alpha}_j (\beta_j + T_j) - T_j & 1 - \hat{\alpha}_j \beta_j T_j & \hat{\alpha}_j \beta_j T_j & 2T_j - \hat{\alpha}_j (\beta_j + T_j) \\ \hat{\alpha}_j (\beta_j + T_j) & -\hat{\alpha}_j \beta_j T_j & 1 + \hat{\alpha}_j \beta_j T_j & T_j - \hat{\alpha}_j (\beta_j + T_j) \\ \hat{\alpha}_j \beta_j T_j & \hat{\alpha}_j (T_j - \beta_j) & \hat{\alpha}_j (\beta_j - T_j) + T_j & 1 - \hat{\alpha}_j \beta_j T_j \end{bmatrix}$$

, (B16)

and

$$\hat{\alpha}_j = \frac{\hat{\lambda}_j + \hat{\mu}_j}{\hat{\lambda}_j + 2\hat{\mu}_j}, \quad \beta_j = \xi H_j, \quad \delta_j = \frac{\hat{\mu}_j}{\hat{\mu}_{j+1}}, \quad T_j = \tanh \beta_j. \tag{B17}$$

Figure Captions

Figure 1 Tectonic setting and submarine topography of southwest Japan. The Philippine Sea plate (PHS) and the Eurasian plate (EUR) are bounded by the Suruga-Nankai trough. Areas enclosed with dashed lines are the tsunami source areas of 1944 Tonankai and 1946 Nankaido earthquakes. MTL denote the Median Tectonic Line.

Figure 2 Leveling routes and tidal stations in Shikoku.

Figure 3 Location map of triangulation points in Shikoku. Double squares , squares with cross and solid squares denote the first, the supplementary first and the second order triangulation points, respectively.

Figure 4 Annual mean sea level change of tidal stations in Shikoku. Upward change corresponds to the land subsidence.

Figure 5 Vertical crustal deformation of tidal stations in Shikoku estimated from monthly tidal record by Ozawa et al. [1994] using a method of Kato and Tsumura [1979]. Upward change corresponds to the land uplift.

Figure 6 Annual mean sea level at Hosojima, Tonoura, Wajima and Oshoro.

Figure 7 Schematic diagram of solving vertical displacement of benchmarks from leveling and tidal record. Squares numbered 1, 2, 3 are benchmarks and a triangle T is tidal station. $h_{i,j}$ is a differential height between benchmarks i and j, V_T is a vertical displacement rate of a tidal station estimated from tidal analysis. V_i is a vertical displacement rate of benchmark i.

Figure 8 Vertical crustal deformations of each earthquake cycle stage in

Shikoku. Circles denote subsidence and crosses denote uplift. (a) Preseismic (1886/96 ~ 1929/37) vertical displacement rate, (b) coseismic (1929/37 ~ 1947) vertical displacement, (c) postseismic (1947 ~ 1968/71) vertical displacement rate, (d) interseismic-1 (1968/71 ~ 1981/82) vertical displacement rate, (e) interseismic-2 (1981/82 ~ 1990/94) vertical displacement rate.

Figure 9 Horizontal deformations of each earthquake cycle stage in Shikoku. (a) Horizontal displacement vectors during 1890/1901 ~ 1949/52 (preseismic, coseismic), (b) baseline length change rate during 1980/84 ~ 1989/90 (interseismic).

Figure 10 Horizontal crustal strain rate distribution. (a) 1890/1901 ~ 1949/52 (preseismic, coseismic), (b) 1949/52 ~ 1980/84 (postseismic, interseismic), (c) 1980/84 ~ 1989/90 (interseismic).

Figure 11 Differential daily mean sea level between Hosojima and Tosashimizu tidal stations before and after the 1946 Nankaido earthquake (1946/12/21). Upward change corresponds to the uplift of Tosashimizu. Tosashimizu is supposed to be uplifted by 30 cm.

Figure 12 Annual mean sea level at Shimotsu and Sumoto tidal stations before and after the 1946 Nankaido earthquake. Upward displacement means land subsidence.

Figure 13 Contour map of the vertical displacement (rate) in Shikoku for each stage in the earthquake cycle. Contour intervals and the units are shown in each figure. Solid lines denote positive values (uplift), and the broken lines denote negative values (subsidence). (a) Preseismic vertical displacement rate. (b) Coseismic vertical displacement. (c) Postseismic vertical displacement rate.

(d) Interseismic-1 vertical displacement rate. (e) Interseismic-2 vertical displacement rate.

Figure 14 Schematic diagram of back-slip model. Plate interaction at subduction zone. The locked effect at an intermediate depth (a) is described by superposition of uniform steady sliding on the whole plate boundary (b) and back-slip at the locked portion (c).

Figure 15 Interpretation of back-slip model in 3 cases. (a) Back-slip denotes interplate coupling. (b) Back-slip denotes strain release (c) No slip denotes no coupling.

Figure 16 Graphical view of a bicubic B-spline function.

Figure 17 An example of ABIC minimum search. ABIC is minimized at α^2=0.067.

Figure 18 Iso-depth contour of modeled plate boundary surface. A dashed rectangle is a model source region in back-slip inversion. Map is plotted after counterclockwise rotation by 25 degree. Lines A-a, B-b, C-c and D-d are the lines on which cross section views are plotted.

Figure 19 Back-slip distribution for each stage of an earthquake cycle estimated by back-slip inversion. The maximum amplitude of back-slip vector and the contour interval in each plot is shown above each figure. Shaded area is the area where estimation error (E) of back-slip vector is larger than a threshold (E_0). (a) preseismic back-slip rate; E_0 = 1 cm/year, (b) coseismic slip; E_0 = 2 m , (c) postseismic back-slip; E_0 = 6 cm/year, (d) interseismic-1 back-slip; E_0 = 1 cm/year, (e) interseismic-2 back-slip; E_0 = 1 cm/year.

Figure 20 Viscoelastic response of the 1946 Nankaido earthquake during 20 years after the shock. Figure 19(b) is assumed as the coseismic slip. (a) lithosphere thickness $H = 35$ km. (b) $H = 40$ km. (c) $H = 50$ km.

Figure 21 Viscoelastically corrected vertical displacement rate of the postseismic stage.

Figure 22 Postseismic back-slip distribution estimated from Figure 21. Shaded area is the area where estimation error (E) of back-slip vector is larger than a threshold $E_0 = 10$ cm/year.

Figure 23 Cross section profile of estimated relative displacement (coseismic) and relative displacement rate (others) on the plate boundary for each stage of earthquake cycle. Coseismic, postseismic, interseismic-1, interseismic-2. Relative plate motion of 5 cm/year in N50°W is assumed (dashed line). Plate boundary depth distribution is plotted at the bottom. (a) Profiles for A-a. (b) Profiles for B-b. (c) Profiles for C-c. (d) Profiles for D-d.

Figure 24 Cross section profile of accumulated relative displacement on the plate boundary surface. Plate boundary depth distribution is plotted at the bottom. (a) Profile for A-a. (b) Profile for B-b. (c) Profile for C-c. (d) Profile for D-d.

Figure 25 Vertical deformation distribution due to viscoelastic response of the 1946 Nankaido earthquake during 20 years after the shock. Lithosphere thickness $H = 35$ km. Assumed source model is (a) Ando [1982], (b) Satake [1993] and (c) this study.