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# Stress–strain 'paradox', plate coupling, and forearc seismicity at the Cascadia and Nankai subduction zones

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**Abstract**

In the continental forearcs of the Cascadia and Nankai subduction zones, geodetic strain measurements indicate crustal contraction in the direction of plate convergence (nearly 'margin-normal'), but earthquake focal mechanisms and other stress indicators show maximum compression along strike ('margin-parallel'). It is evident that the geodetic strain signals reflect the temporal fluctuations of elastic stress associated with subduction earthquake cycles, not the absolute stresses. The absolute margin-normal stress is not only much less than the margin-parallel stress but also no greater than the lithostatic value, indicating a very weak subduction thrust fault. Weak subduction faults are also consistent with geothermal data which require very low frictional heating along the faults. Great subduction earthquakes occur at very low shear stresses and cause small perturbations to the forearc stress regime. Because these small elastic stress perturbations are relatively fast, they give large strain rates that are detected by geodetic measurements. The present nearly margin-normal contraction in both places is due to the locking of the subduction fault and the consequent increase in elastic stress in the direction of plate convergence. With the margin-parallel compression being dominant, the small increase in margin-normal stress may change the forearc seismicity in Cascadia from a mixture of thrust and strike–slip types into mainly thrust before the next great subduction earthquake. In Nankai, the change should be from strike–slip to quiescence, as observed in the Nankai forearc prior to the 1944–1946 great subduction earthquakes. 

*Keywords:* earthquake; stress–strain; subduction zone; plate coupling

## 1. Introduction

At the Cascadia subduction zone in western North America, the Juan de Fuca plate subducts beneath the North America plate at a convergence rate of ca. 45 mm year$^{-1}$ (Fig. 1). At the Nankai subduction zone in southwest Japan, the Philippine Sea plate subducts beneath the Eurasia plate [or the Amurian microplate (Heki et al., 1998)] at a similar rate (Fig. 2). In the continental forearc areas of both subduction zones, active crustal seismicity and deformation take place. However, it has been an outstanding problem in both places that the direction of the maximum compressive stress ($\sigma_1$) and the direction of contemporary crustal contraction are nearly perpendicular to each other. The maximum compression as shown by earthquake focal mechanisms and other stress indicators is margin-parallel, but crustal shortening as determined by geodetic measurements is nearly margin-normal, in the direction of plate convergence.

The stress and strain 'paradox' for the Cascadia

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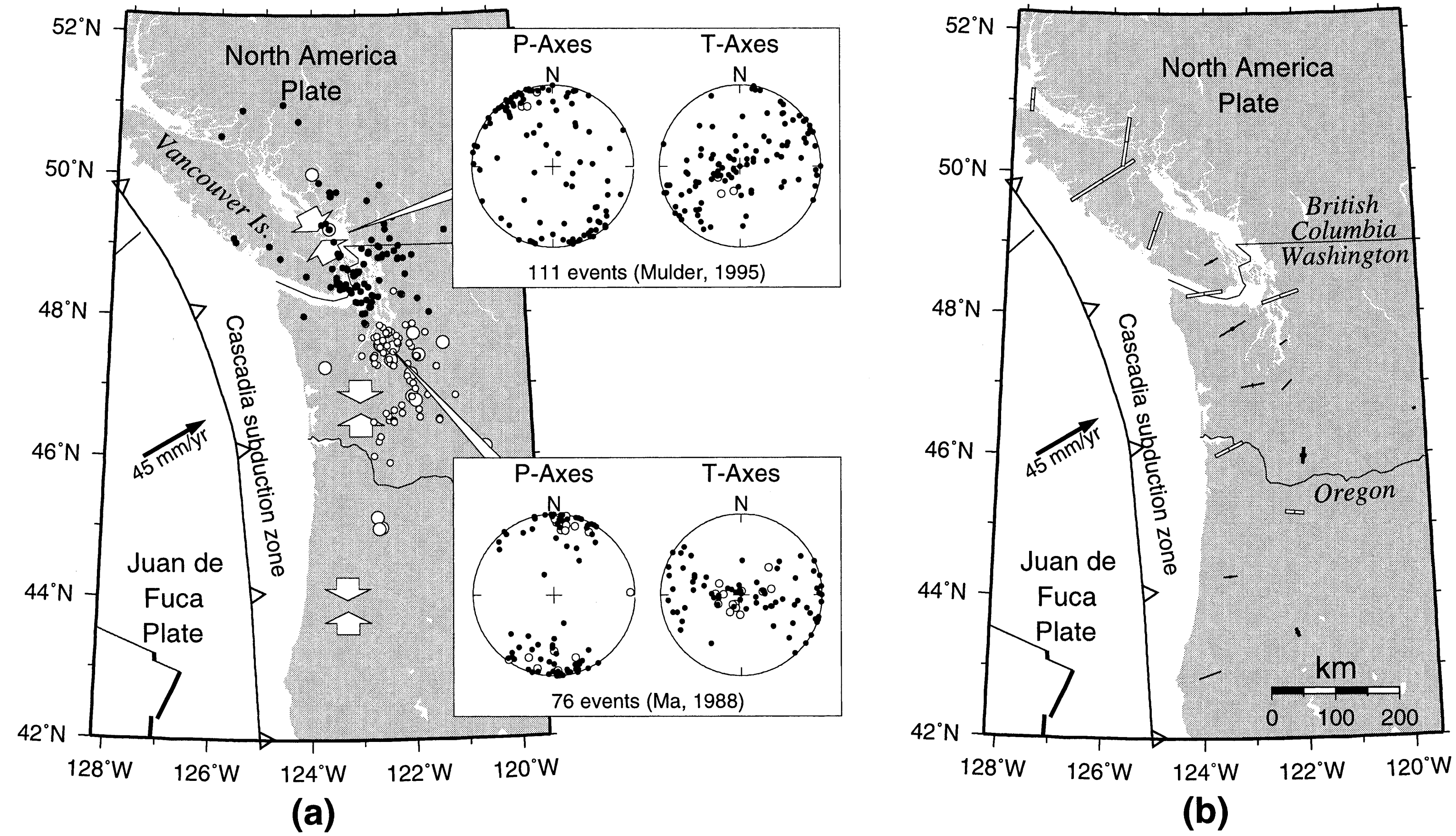


Fig. 1. (a) A summary of the stress regime of the Cascadia forearc. Small solid and open circles on map are continental earthquakes ($M<3$) used by Mulder (1995) and Ma (1988), respectively, to determine focal mechanism solutions shown in insets. Large open circles on map and in insets are events ($M \geqslant 4$) since 1993 for which moment tensor solutions have been provided by the Oregon State University (http://quakes.oce.orst.edu, 1998). Large arrow pairs indicate the orientation of maximum compressive stress from various stress indicators (see text for details). (b) A summary of geodetic strain rate measurements in the Cascadia forearc. A thin solid bar indicates contraction, and a thick solid bar indicates extension. A hollow bar indicates maximum contraction where only shear strain rates were determined, assuming uniaxial contraction. The largest strain rate shown is $2.3 \times 10^{-7}$ year$^{-1}$.

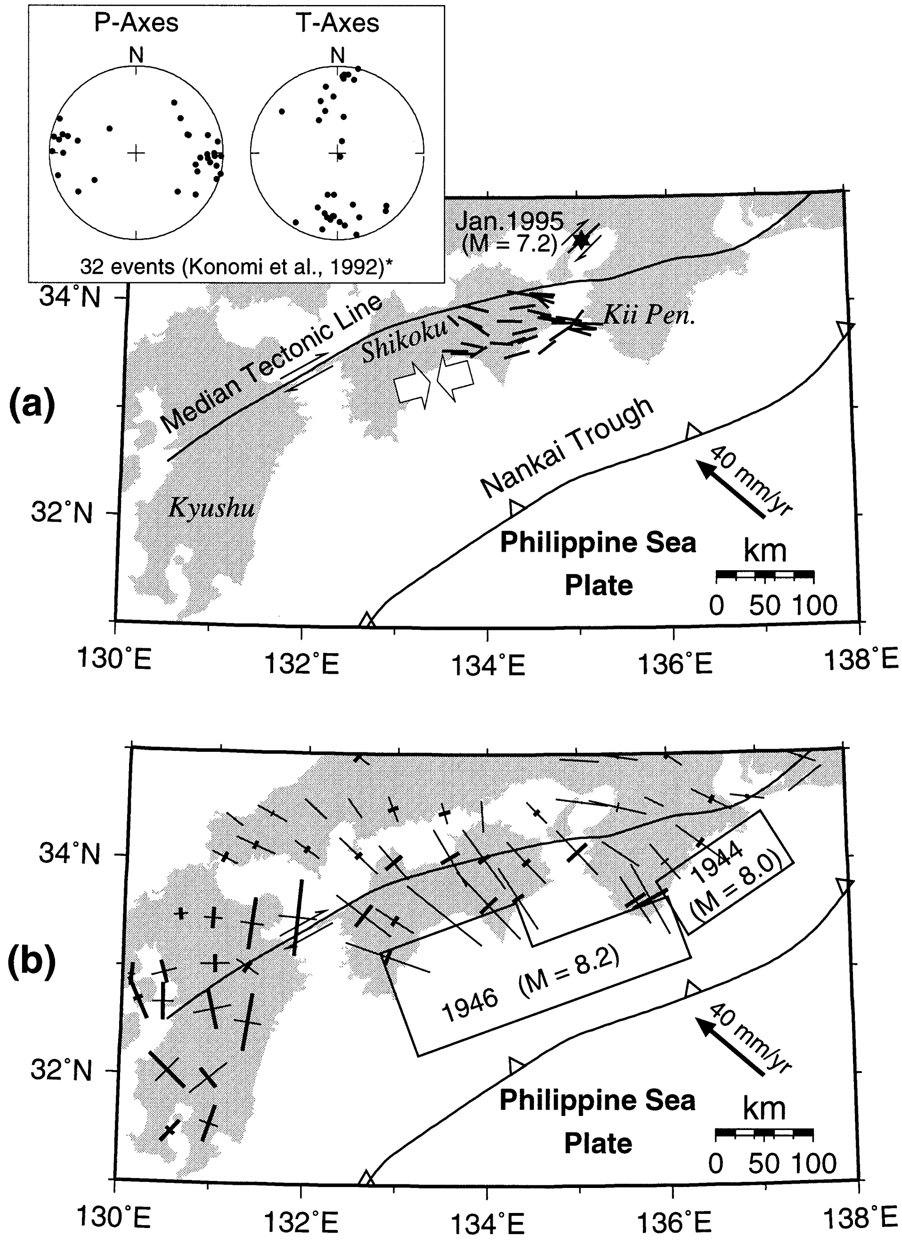


Fig. 2. (a) A summary of the stress regime of the Nankai forearc. Bars on map indicate locations and *P*-axis orientations of continental earthquakes ($3 \leq M < 4$) analysed by Konomi et al. (1992) and Konomi (personal communication, 1999). The composite focal mechanism solutions for these events are shown in the inset. The large arrow pair indicates the direction of the maximum compressive stress from various stress indicators (see text for details). (b) Strain rate tensors derived from site velocities of the continuous GPS network GEONET over a 1-year period in 1997–1998. A thin bar represents contraction. The largest strain rate shown is $2.8 \times 10^{-7}$ year$^{-1}$.

subduction zone has been discussed by Sbar (1983) and Wang et al. (1995a,b). Sbar (1983) considered the difference between the seismic and geodetic data to reflect crustal strain at very different time scales. Wang et al. (1995a,b) further pointed out that the stress indicators constrain the absolute crustal stress field, but geodetic measurements reflect the temporal stress changes associated with subduction earthquake cycles. For the Nankai subduction zone, the same 'paradox' has also long been noticed. This article is a synthesis of the observations and theoretical aspects of this problem for both places. I summarize the evidence for the stress and strain regimes, review the mechanics behind the 'paradox', and explore their implications for plate coupling and earthquake hazards. The paper deals with the qualitative aspects of the fundamental concepts. Previous and on-going mechanical and thermal modelling that provides the quantitative basis for these concepts are given in various references.

## 2. Stress and strain rate observations

### *2.1. Forearc stress indicators*

#### *2.1.1. Focal mechanisms of continental earthquakes*

Earthquake focal mechanisms provide the most important constraints for the directions and relative magnitudes of the principal stresses in the forearc. The fault plane solution for a single earthquake is an indicator of the directions of the principal components of the incremental stress (tensor) induced by coseismic fault slip. These directions may be different from those of the regional stress, since slip may occur on a less favourably oriented weak fault. This is especially true for large events on major faults, of which the subduction earthquakes are typical examples. However, for many small earthquakes that occur on randomly oriented faults, the average directions of the *P*- and *T*-axes represent the directions of maximum and minimum compressive regional stresses, respectively (Sbar, 1982).

Crosson (1972) reported focal mechanism solutions from the Puget Sound area, Washington State, that indicated N–S compression ($\sigma_1$) of the continental crust. Zoback and Zoback (1980, 1991) and Sbar (1982) compiled larger data sets that yielded the same pattern. Fig. 1a shows the locations and focal mechanisms of 111 continental forearc earthquakes that occurred during 1984–1991 in British Columbia (Mulder, 1995; Wang et al., 1995b) and 76 events that occurred during 1982–1985 in Washington (Ma, 1988; Ma et al., 1996). The average *P*-axis direction determined by Ma (1988) in Washington is N–S, and that by Mulder (1995) in British Columbia is NW–SE, both parallel with the strike of the subduction zone. Also included in Fig. 1a are 17 larger ($M \geq 4$) events that occurred during 1994–1997 in these areas with moment tensor solutions provided by the Oregon State University (http://quakes.oce.orst.edu, 1998), and the moment magnitude 5.5, 1993 Scott Mills, Oregon, earthquake (Nabelek and Xia, 1995). The focal mechanisms of these more recent events are consistent with those reported by Ma (1988) and Mulder (1995) in their respective areas. In some studies at Cascadia, the reported average *P*-axes are oblique to the margin, such as for a few large earthquakes that occurred in northernmost Cascadia (Rogers, 1979) and in a small area near Mount St. Helens (Weaver and Smith, 1983). These *P*-axes may be complicated by local structures (Ma et al., 1996), but even so, they are still more consistent with margin-parallel compression than with margin-normal compression.

The maximum compressive stress in the continental forearc of the Nankai subduction zone had once been thought to be in the direction of plate convergence (e.g. Nakamura and Uyeda, 1980). However, as more earthquake data became available, a pattern similar to Cascadia emerged. As summarized by Tsukahara and Kobayashi (1991), the maximum compressive stress is nearly margin-parallel south of the Median Tectonic Line (MTL) in the Shikoku area and is at an angle with strike farther inland. Although our interest is mainly in the Shikoku area where maximum geodetic contraction is observed, it is worth pointing out that north of the MTL, the maximum compressive stress is still ca. 40° more oblique to normal than the direction of plate convergence, obviously not the simple result of a push by the subducting

Philippine Sea plate. The *P*-axes of many small crustal earthquakes in Shikoku were reported by Okano et al. (1980) and Kimura and Okano (1992) to be margin-parallel. *T*-axes for the same events were not published, but these earthquakes have been reported to be mainly of the strike–slip type (Tsukahara and Kobayashi, 1991). Ichikawa (1971) and Shiono (1977) also reported margin-parallel *P*-axes. Konomi et al. (1992) presented focal mechanism solutions for 20 continental crustal events (depth < 23 km) in eastern Shikoku and vicinity. The solutions were later refined, with more events added (Konomi 1999, personal communication). The locations and the horizontal projection of the *P*-axes of these events are shown in Fig. 2a. They are mostly strike–slip events under margin-parallel compression, consistent with previous findings. The 1995 Hyogoken-Nanbu earthquake (known as the Kobe earthquake) (Fig. 2a) was also a strike–slip event under E–W compression (Ishikawa, 1995).

#### 2.1.2. *In situ stress measurements*

In situ measurements constrain the near surface state of stress and provide corroborative information for the stresses determined for the larger area and depth range from focal mechanisms. Borehole breakouts data from a number of wells in western Oregon (Werner et al., 1991) and one well in western Washington (Magee and Zoback, 1992) indicated N–S compression (Fig. 1a). There is only one available in situ stress measurement in Shikoku. It was conducted using the method of stress relief (Tanaka, 1987) and yielded a nearly E–W maximum compression in the horizontal plane. These in situ measurements are in accord with margin-parallel compression.

#### 2.1.3. *Crustal anisotropy from shear-wave splitting studies*

Differential speeds of shear waves polarized in different directions when travelling in the crust are commonly attributed to crustal anisotropy due to the alignment of fluid-filled fractures with the maximum compressive stress (Crampin, 1978). A shear wave travelling vertically is faster if the particle motion is in the direction of maximum horizontal stress. Shear-wave splitting studies for the continental crust in Northern Cascadia (Cassidy and Bostock, 1996) and Nankai (Kaneshima and Ando, 1989) both yielded a margin-parallel 'fast direction', consistent with a margin-parallel $\sigma_1$.

#### 2.1.4. *Neotectonic geological structures*

Quaternary faulting and folding and volcanic vent alignments constrain the regional state of stress at a much larger time scale than the above three types of evidence. Because of the large time window, this type of data is important in providing information on the regional tectonic background and the history prior to the modern stress regime.

Zoback and Zoback (1991) and Werner et al. (1991) summarized the neotectonic evidence for N–S compression in the Pacific Northwest, including the Cascadia forearc. More recent studies include E–W-trending reverse or thrust faulting in Washington (Bucknam et al., 1992; Johnson et al., 1994), fold and fault structures under N–S compression in Washington (McCrory, 1996), and kinematically constrained N–S shortening of the Cascadia forearc (Wells et al., 1998). Studies by Wells and Heller (1988), Wells (1990), England and Wells (1991) and Walcott (1993) indicate margin-parallel shortening as well as clockwise rotation of the Cascadia forearc earlier in Tertiary, which implies that the current margin-parallel compression has persisted for a long time.

Neotectonics of the Nankai forearc has been summarized by Tsukuda (1992) and Sugiyama (1994). South of the MTL (Fig. 2a), two parallel bands of deformation have been identified, a landward zone of forearc rise and a seaward zone of forearc basins. The forearc rise zone has a series of anticline and monocline structures trending perpendicular to the margin-parallel MTL. The forearc basin zone consists of five major basins divided by reverse faults and folds at high angles with the MTL. The deformation pattern of both zones indicates nearly margin-parallel shortening and compression.

In the active accretionary prism, especially in the frontal region, maximum compressive stress must be in the direction of plate convergence as indicated by numerous thrust and fold structures. If the mechanically coupled area of the subduction fault is relatively shallow, as is the case for Cascadia and Nankai, the frontal part of the forearc can be in large margin-normal compression while the rest of the forearc is in little compression or even in tension (Wang and He, 1999). In addition, the mechanical behaviour of the relatively soft frontal part of the accretionary prism is not representative of the continental rocks of the forearc. It is no surprise that margin-parallel compression is not prevalent in that area. In the rest of paper, we do not further discuss the stresses in the accretionary prisms.

### 2.2. On the origin of margin-parallel compression

The reason for the predominant margin-parallel compression in both places is not the main theme of this article. To the first-order, it can be understood as the consequence of a buttressed forearc sliver (Wang, 1996). Subduction at both margins is oblique with a dextral component. Characteristic of an oblique subduction zone is a margin-parallel shear zone in the upper plate that defines a forearc sliver (Fitch, 1972). The strike–slip fault MTL is a classic example of such a shear zone. Although a well-developed strike–slip fault is not present at Cascadia, there is strong geological evidence that a shear zone exists and is active along the volcanic front (Pezzopane and Weldon, 1993). The Nankai and Cascadia forearc slivers have a tendency to be transported along strike but are resisted at their leading edges (Fig. 3). The Cascadia sliver appears to be buttressed by the Canadian coast mountains and is shortened and slowly rotated clockwise, resulting in nearly E–W tension in the southern Cascadia arc (Fig. 3a) (Wells et al., 1998). The Nankai forearc sliver may still be moving ahead at the leading edge but against significant resistance because of the curvature of the margin and because subduction is no longer oblique farther east (Tsukuda, 1992; Seno et al., 1993; Kimura, 1996). In a buttressed forearc sliver with a weak strike-parallel shear zone, margin-parallel compression can develop (Wang, 1996). The forearc sliver may also be pushed at its trailing edge, causing additional compressive stress, as indicated by the focal mechanisms of some earthquakes near the Mendocino triple junction (Fig. 3a). At both Cascadia and Nankai, there is also roughly margin-

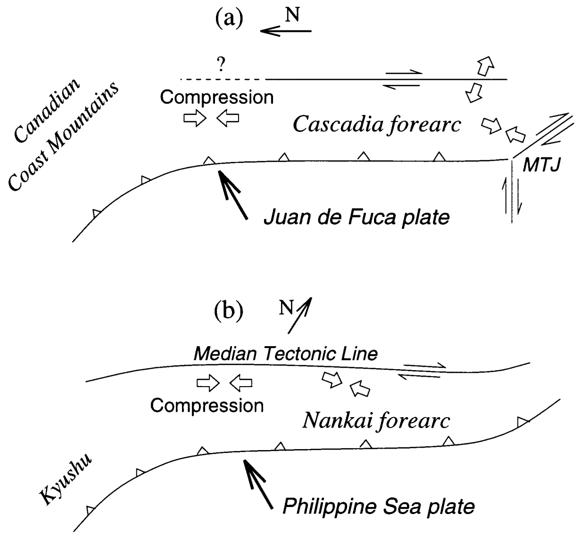


Fig. 3. Cartoon showing tectonics and stress regimes of the Cascadia and Nankai forearcs (not to scale). MTJ denotes Mendocino triple junction. Regions of compression or tension indicated by arrow pairs were delineated by crustal earthquake focal mechanisms. Oblique subduction in conjunction with a weak strike-parallel forearc shear zone can give rise to margin-parallel compression in the forearc sliver.

parallel compression landward of the slivers, which must be caused by regional plate tectonics. Further investigations are required to better understand the nature of the margin-parallel compression in these two places.

### 2.3. Contemporary strain rates

The directions of maximum crustal contraction determined by geodetic measurements are shown in Fig. 1b for Cascadia and Fig. 2b for Nankai. Most of the data for the northern Cascadia forearc have been reviewed by Dragert et al. (1994). Summaries for the entire Cascadia subduction zone were given by Snay and Matsikari (1991) and Murray and Lisowski (1999). Some of the measurements involved early triangulation surveys and did not give all the three independent components of a horizontal strain rate tensor. However, the direction of maximum contraction and the value of maximum shear strain rate could be determined. Later measurements were conducted using laser ranging or the global positioning system (GPS) which yielded strain rate tensor estimates. Published strain rates were determined mainly from campaign style measurements, and the values are averages over a time span of several years over which the measurements were made. They indicate crustal contraction in the direction of plate convergence. The crustal deformation is mainly uniaxial shortening (e.g. Savage et al., 1991). Results of recent GPS campaign survey conducted in and around southern Vancouver Island have yielded a similar direction of maximum contraction (Dragert et al., 1998). Continuous GPS monitoring has been conducted since 1991 mainly by the Geological Survey of Canada in British Columbia (Dragert et al., 1995). In addition, several continuous GPS stations have recently been installed in Washington and Oregon. Preliminary velocity vectors for individual stations were reported by Henton et al. (1998) (also http://www.pgc.nrcan.gc.ca, 1999).

Japan has a much longer history of using precise geodetic measurements to detect crustal deformation (Hashimoto and Jackson, 1993). Average strain rates are available for the whole country for the periods of 1883–1994 and 1985–1994 (http://www.gsi-mc.go.jp, 1999). The most relevant data for the subject of this paper are strain rates recently determined using GPS (e.g. Kato et al., 1998). Fig. 2b shows the strain rates in Southwest Japan derived from site velocities of the Japanese continuous GPS network GEONET for a 1-year period in 1997–1998 (www.gsi-mc.go.jp, 1988). I determined the strain rates by constructing overlapping strain nets of 50 km radius, assuming uniform deformation within each net. Each net includes at least six, but usually many more, GPS sites. The strain rates in Shikoku from the GEONET data are similar to those determined by Tabei et al. (1996) using campaign GPS measurements. Similar to Cascadia, the forearc deformation at Nankai is dominated by crustal contraction in the direction of plate convergence.

## 3. Small stress changes and large strain rates due to great subduction earthquakes

The apparent contradiction between the maximum compressive stress and maximum geodetic contraction can be easily understood by recalling that geodetic methods only detect strain changes between measurements, or average strain rates. A constant elastic stress is associated with a constant elastic strain $\epsilon$ that cannot be detected geodetically. If a change occurs in the elastic stress, the associated change in the elastic strain $\Delta\epsilon$ is measured. The direction of the stress change may bear no relation with the direction of the absolute stress itself, as is illustrated by the simple example of a stretched elastic band in Fig. 4. When deformation is elastic, geodetic measurements are not stress indicators.

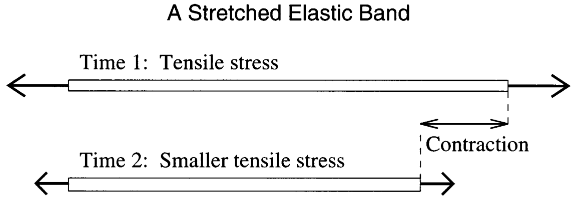


Fig. 4. An example to illustrate that an observed strain change does not indicate stress direction if the deformation is elastic. The absolute stress and strain remain tensile, but the deformation from time 1 to time 2 is contraction. Here contraction only means a smaller tension. Similarly, extension may only mean smaller compression.

Permanent deformation of a continuous medium is controlled by the absolute stresses. For example, in a viscous fluid, which is often used to approximate the lithosphere in modelling very long-term deformation, maximum contraction rate $\dot{\epsilon}_1$ and maximum compression $\sigma_1$ are in the same direction. In the case of fault slip, such as in earthquakes, the slip direction may be very different from that of the regional maximum shear stress, as discussed in Section 2.1, but to the extent that the collective effect of many faults approximates the permanent deformation of an isotropic continuous medium, the direction of average contraction approximates the direction of $\sigma_1$. However, the behaviour of the lithosphere in subduction earthquake cycles is mostly elastic, and $\sigma_1$ and geodetically observed $\dot{\epsilon}_1$ can be in very different directions.

For the Cascadia and Nankai forearcs, the maximum compressive stress is margin-parallel as is delineated by all stress indicators. Subduction earthquakes cause perturbations mainly to the compressive stress in the direction of plate convergence (referred to as "margin-normal" in this paper). When the subduction thrust fault is locked, elastic compression increases in this direction, causing crustal contraction. When a subduction earthquake occurs, compression decreases in this direction, causing crustal extension.

Stress changes associated with great subduction earthquakes are small. The drops of shear stress along subduction faults in these earthquakes are $<10$ MPa, and often 1–2 MPa (Kanamori and Anderson, 1975; Kanamori, 1980). As will be shown in the following section, the average shear stress along the Cascadia or Nankai subduction fault cannot be much higher than this level. The total fluctuation of shear stress on the thrust fault throughout subduction earthquake cycles is limited to a small value by the weakness of the fault. This is true even if significant stress drop occurs in aseismic fault slips between great earthquakes (Heki et al., 1997; Wang, 1995). Such small shear-stress changes on the fault cause small perturbations to margin-normal stresses in the forearc, such that the pattern of margin-parallel compression remains unchanged throughout subduction earthquake cycles. However, these perturbations occur sufficiently fast to cause rates of elastic strain detectable by geodetic measurements.

It is not a trivial matter to estimate the crustal deformation rates if the fault slip rates are not kinematically prescribed, because the deformation involves the interaction between the elastic lithospheric plates and the more viscous asthenosphere. However, for this discussion, it suffices to demonstrate that with reasonable crust and mantle mechanical properties, the small stress perturbations can cause large strain rates similar to the observed ones. Parts of the model results by Wang et al. (1994) are reproduced here for this purpose. This is a cross-sectional finite element model using viscoelastic rock rheology to simulate crustal deformation in hypothesized Cascadia subduction earthquake cycles. The model profile is across southern Vancouver Island perpendicular to the subduction zone. An earthquake is produced every 500 years by releasing the previously locked fault. The fault is locked after the earthquake. Interested reader is refer to Wang et al. (1994) for the details of the modelling, such as the model geometry, boundary conditions, rock property values, and the finite element technique. The modelled deviatoric stresses in the forearc observed at different times after a subduction earthquake are shown in Fig. 5. It should be kept in mind that the maximum compressive stress is normal to the cross-section. The surface margin-normal strain rates obtained from the same model are shown in Fig. 6 and compared with the observed values that are projected to the model profile. The stress changes associated with the subduction earthquake cycles are small (Fig. 5), but the strain rates are large (Fig. 6). Models of crustal deformation and stresses associated with 1944–1946 great Nankai earthquakes lead to the same conclusion of small stress changes (Miyashita, 1987; Yoshioka and Hashimoto, 1989).

Although there is large margin-parallel compression and perhaps permanent crustal shortening in that direction over a long time, the rate is overshadowed by that of the elastic deformation due to subduction earthquakes. According to values summarized by Wells et al. (1998), the margin-parallel contraction rate averaged along the Cascadia forearc is ca. 0.01 μstrain year$^{-1}$ (a few mm year$^{-1}$ over the forearc length), one order of magnitude less than the observed and the theoretically predicted margin-normal component.

## 4. Shear stress level on the thrust fault: plate coupling

### 4.1. Low plate coupling stress

Why is the margin-normal compressive stress so small at Cascadia or Nankai? The reason is that the subduction fault is very weak. The strength of plate coupling is defined as the level of long-term (or static) shear stress along the plate interface, as if the subduction fault would slip continuously at the plate convergence rate. From the small stress drops in subduction earthquakes, we do not know whether the fault is strongly or weakly stressed in the absolute sense on average (see Fig. 7). However, the level of coupling shear stress

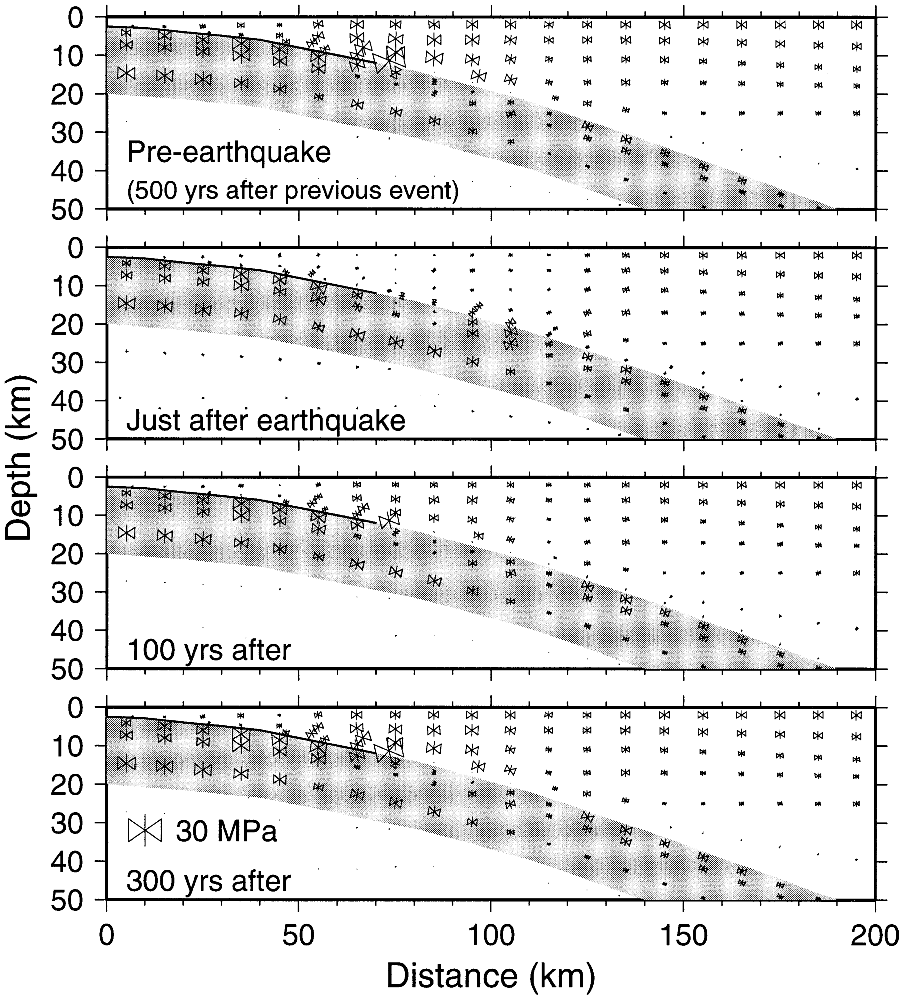


Fig. 5. Evolution of the upper plate stresses throughout a hypothesized Cascadia subduction earthquake cycle as predicted by the viscoelastic finite element model of Wang et al. (1994). The model profile is across southern Vancouver Island. The stick–slip (intermittent locking and unlocking) portion of the subduction fault is indicated by a solid line. Further downdip there is a 30 km free-slip zone, and the rest of the fault is viscously coupled. Shown are deviatoric stresses. An hour-glass represents compression, and a bar represents tension. Note that the stress fluctuations are very small. The subducting plate is shaded.

can be inferred from near-field intraplate stresses and the amount of frictional heating, as in the case of a strike–slip plate boundary (Mount and Suppe, 1987; Zoback, 1987; Brune et al., 1969; Lachenbruch and Sass, 1980).

The continental crustal earthquakes at Cascadia are a mixture of thrust and strike–slip events (Fig. 1a), and therefore, assuming no preferred fault orientations, the margin-normal and vertical stresses must be similar in magnitude, as shown in Fig. 8a. The strike–slip focal mechanisms of the forearc earthquakes at Nankai (Fig. 2a) indicate that the margin-normal stress is the least compressive ($\sigma_3$), and the intermediate stress $\sigma_2$ is vertical, as shown in Fig. 8b.

The mechanics of low margin-normal compression in a thrust environment is quantitatively studied by Wang and He (1999). In brief, in the presence of margin topography, the continental forearc has a tendency to collapse under its own

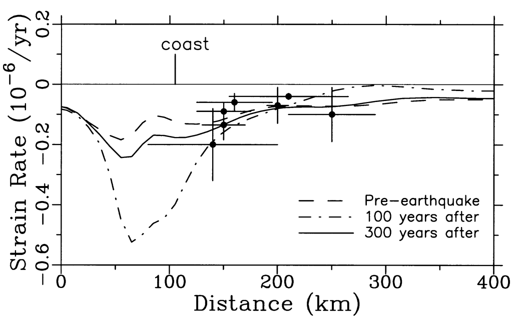


Fig. 6. Surface horizontal strain rates (margin-normal contraction) calculated from the same model as shown in Fig. 5. The small stress fluctuations shown in Fig. 5 result in these large strain rates. Strain rate values determined from geodetic measurements for six sites near the model profile are also shown, with uncertainties (1 standard deviation) and sizes of the strain networks indicated by vertical and horizontal bars, respectively. The strain rates just after the earthquake are too large to be shown at this scale.

weight, but traction along the subduction fault provides lateral support to hold it together. Whether the forearc is in lateral compression or tension relative to lithostatic depends on which effect is stronger. This is illustrated in Fig. 9 using a numerical example of two converging elastic plates in frictional contact. The strength of the subduction fault is described using a static friction law $\tau = \mu' \sigma$, where $\tau$ and $\sigma$ are shear and normal stresses along the fault, and $\mu'$ is an effective coefficient of friction that approximately includes

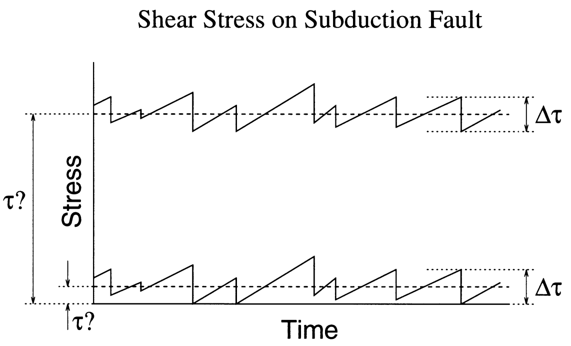


Fig. 7. Average shear stress ($\tau$) on the subduction fault and fluctuations ($\Delta\tau$) in great earthquake cycles. In this schematic illustration, the linear stress increase is interrupted by abrupt stress drops in earthquakes. In reality, the stress increase between earthquakes may not be linear, and the stress drop may not always be abrupt.

the effect of pore fluid pressure. A larger plate coupling force can overcome the gravitational effect and result in greater compression (Fig. 9c). A very small plate coupling force may just balance the gravitational effect or result in margin-normal tension such as currently in Cascadia and Nankai, respectively. A very weak subduction fault may rupture even when the forearc is in margin-normal tension. This is obvious by considering the end-member case of a frictionless fault (Fig. 9a).

The fault stress and hence margin-normal compression may exceed the present level before the next subduction earthquake or slip event, but the increase cannot be very large. Wang et al. (1995b) showed that the contribution to surface heat flow from frictional heating along the Cascadia subduction fault was very small, which limited the long-term average shear stress on the fault to be within 20, and probably, 10 MPa. Their preferred model is shown in Fig. 10, which incorporates no frictional heating along the thrust fault. When the average shear stress is $>10$ MPa, the calculated heat flows will be too high compared to the observed ones. Therefore, the currently locked fault is unlikely to be able to sustain a high shear stress. Details of the modelling procedure and sensitivity tests can be found in Wang et al. (1995b). The stress field in the Juan de Fuca plate before subduction (Wang et al., 1997) and the large-scale regional stress field of western North America (Geist, 1996) are also consistent with very low coupling stress at the subduction zone boundary. The possible causes for the apparent low strength of the subduction fault have been discussed by Wang et al. (1995b) (see also references therein). Compared to Cascadia, the state of stress in the Nankai forearc argues more strongly for a very weakly stressed (although currently locked) subduction fault. Heat flow data from Nankai are also consistent with low frictional heating on the subduction fault (Wang et al., 1995a), although they are not as diagnostic as in Cascadia.

It is interesting to notice that the margin-normal stress in the Nankai forearc is smaller than that in Cascadia, perhaps for an entire interseismic period, indicating an even lower coupling force on the Nankai subduction fault. This may partially explain why the recurrence time of subduction

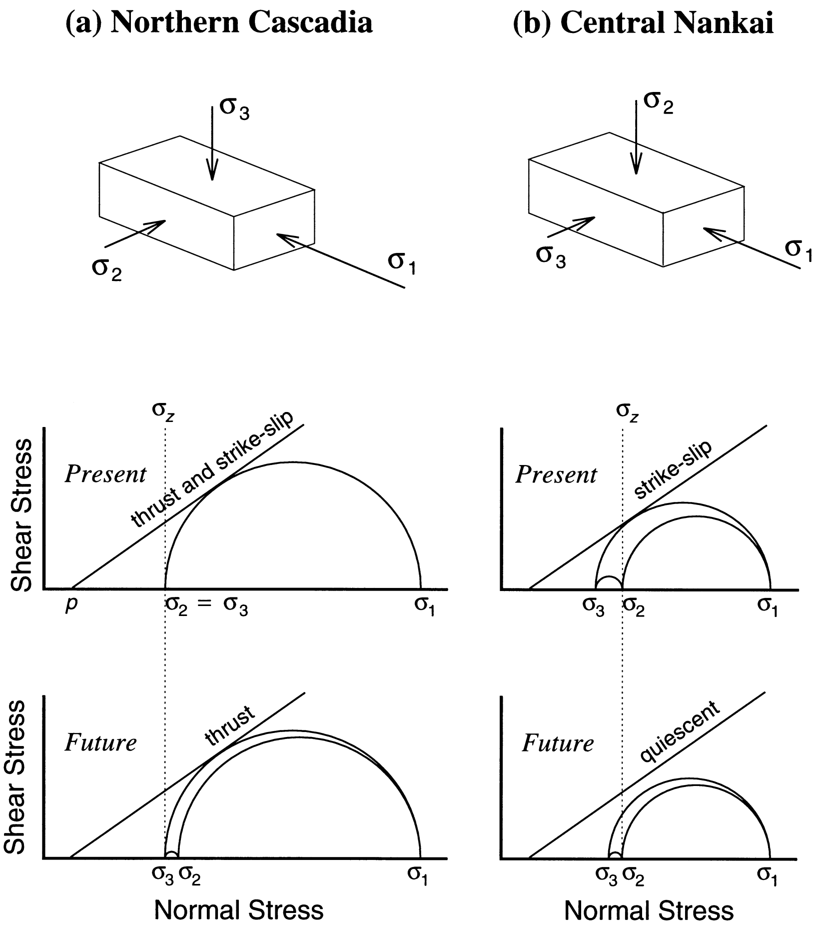


Fig. 8. The state of stress in Cascadia and Nankai forearcs. In both places, maximum compressive stress $\sigma_1$ is margin-parallel. At present, the margin-normal stress has the same magnitude as the vertical stress in Cascadia but is less than the vertical stress in Nankai. 'Future' means sometime later in the interseismic period, as discussed in Section 5. The straight line tangent to the Mohr circle represents the strength of the continental crust. Pore fluid pressure $p$ is unknown.

earthquakes in Cascadia appears to be longer than Nankai. It is not known at this time what causes the difference in the fault strength in the two places. Given the same failure criterion for crustal rocks and the same pore fluid pressure, as implicitly assumed in the diagrams of Fig. 8, an immediate inference is that the margin-parallel compressive stress in the northern Cascadia forearc is larger than in central Nankai, as represented by a greater Mohr-circle diameter. This may be useful information for understanding the tectonic processes responsible for the predominant margin-parallel compression in both places.

### 4.2. *Weak fault and great earthquakes*

There have been many great subduction earthquakes at the Nankai subduction zone (Ando, 1975; Kumagai, 1996), the latest ones being the 1944 Tonankai and 1946 Nankaido earthquakes, of magnitude 8.0 and 8.2, respectively (Kanamori, 1986). There is strong evidence that subduction earthquakes also have occurred repeatedly at the Cascadia subduction zone [see reviews by Atwater et al. (1995), and Clague (1997)], and the last event was probably in 1700 (Satake et al., 1996). Therefore weak subduction faults can produce

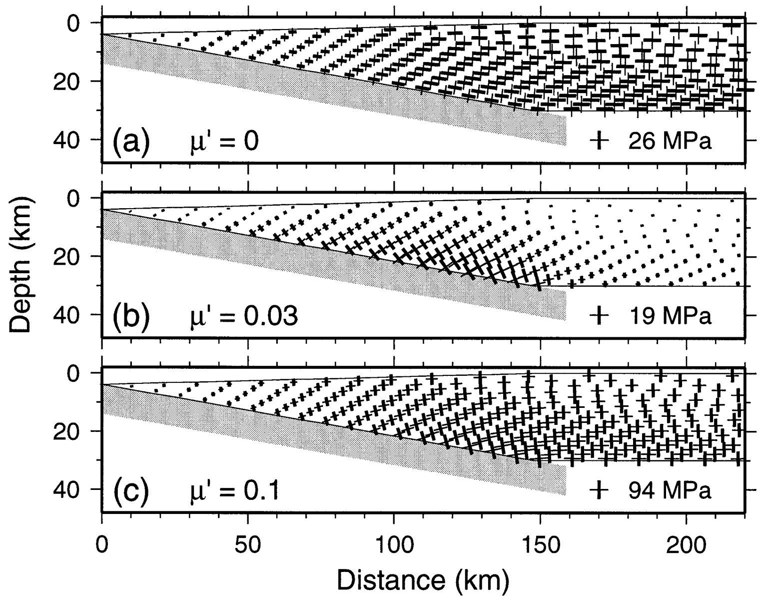


Fig. 9. A finite element model of two converging elastic plates, showing the effect of plate coupling on forearc stresses. Shown are static deviatoric stresses (not to be confused with stress perturbations in Fig. 5), with a thin bar indicating compression. Bar length represents the value of maximum shear stress. The subducting plate is shaded [from Wang and He (1999)].

great earthquakes. Considering the well known small stress drops in these earthquakes (Kanamori, 1980), this conclusion is by no means a surprise.

### *4.3. Weak fault and large compression*

Although small margin-normal compression indicates a weak subduction fault, large compression does not necessarily indicate a strong fault. The integrated effect of the small shear stress along the downdip dimension can be large and can cause margin-normal compression. In both Cascadia and Nankai, the subducting plates are very young, roughly 10 and 20 Ma, respectively, with thick insulating sediment covers before subduction. Consequently, the subduction zones are relatively hot (Lewis et al., 1988; Yamano et al., 1984), limiting the slab and interplate seismicity to shallow depths. As argued by Hyndman and Wang (1993) and Hyndman et al. (1995, 1997), the interplate seismogenic zone is limited by a temperature ca. 350°C. Above ca. 450°C, the continental crustal rocks of the upper plate deforms plastically, with shear strength decreasing rapidly with depth. Therefore for 'hot' subduction zones such as Cascadia and Nankai, the portion of the fault that can sustain a few tens of megapascals shear stress is limited to shallow depths and is relatively narrow (Wang et al., 1995a,b). The integrated coupling force is then small, causing little margin-normal compression in the upper plate. An older and/or faster subducting plate with little sediment cover will make the subduction zone much colder, and the coupled region will extend to greater depths. Shear strength of the fault increases with depth, assuming Coulomb frictional behaviour. Because of this effect and the greater downdip width of the coupled region, the total plate coupling force in these colder subduction zones can be large even for a weak fault, causing margin-normal compression in the upper plate. An example is the Northeast Japan subduction zone (Wang and Suyehiro, 1999).

For similar reasons, small plate coupling stresses can cause large margin-parallel compression in a buttressed forearc sliver in the case of oblique subduction as mentioned in Section 2.2. The tangential component of the plate coupling stress is

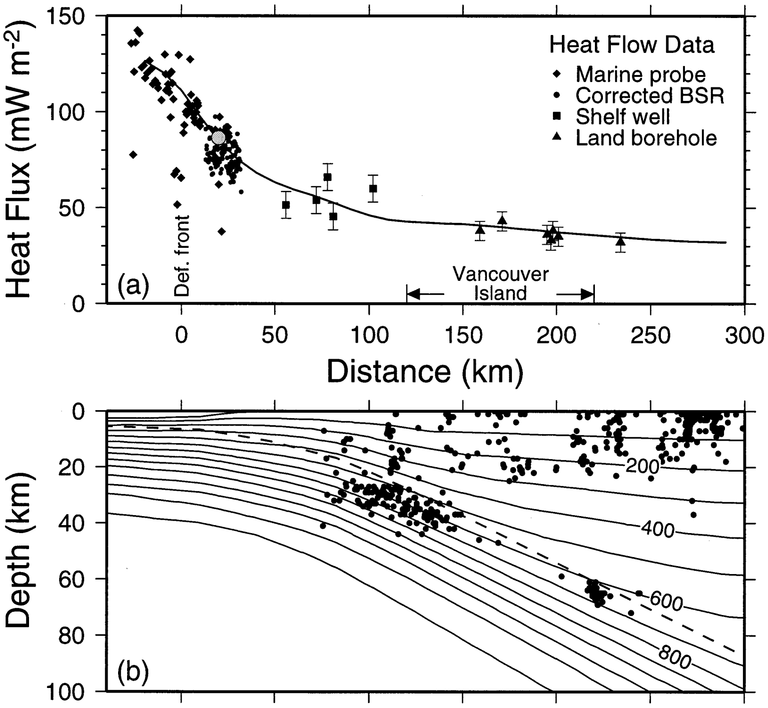


Fig. 10. Two-dimensional finite element thermal model of the northern Cascadia subduction zone across southern Vancouver Island, showing that heat flow data do not indicate significant frictional heating along the subduction fault. (a) Calculated and observed heat flows. The borehole heat flow value from ODP Leg 146 is shown as a large shaded circle. Heat flows estimated from the position of methane-hydrate bottom simulating reflectors (BSR) have been corrected for sediment thickening and fluid expulsion (Wang, 1994). No frictional heating is incorporated in this model. (b) Temperature contours for the same model. Dashed line shows the position of the plate interface. Dots are earthquake hypocentres, excluding those >45 km offshore because of their poor depth determination [modified from Wang et al. (1995b)].

integrated over a great distance along strike, such that the low plate coupling stress that does not cause margin-normal compression at Cascadia and Nankai may be responsible for some of the margin-parallel compression.

## 5. Forearc seismicity related to subduction earthquakes

### 5.1. Cascadia

Wang et al. (1995b) speculated that the small variations of forearc stresses associated with subduction earthquake cycles might cause changes in the style of crustal earthquakes. Because of ongoing plate convergence, a locked subduction fault should cause the compressive stress in the direction of plate convergence to increase with time. For the Cascadia subduction zone, this ‘margin-normal’ stress may have been smaller than the vertical one just after the previous great subduction earthquake, a state of stress inducive to strike–slip crustal events under margin-parallel compression. The margin-normal stress may have gradually increased to the present level, resulting in a mixture of strike–slip and thrust crustal events, shown as ‘present’ in Fig. 8a. If the margin-normal stress continues to increase and exceeds the vertical stress in the future prior to the next great subduction earthquake, the continental earthquakes will consist mostly of thrust events, shown as ‘future’ in Fig. 8a. For simplicity, normal subduction has been assumed in Fig. 8. Because of oblique subduction (Fig. 3), the present elastic stress increase and crustal contraction has a small margin-parallel component (see Section 2.2), as indicated by geodetic data in Washington and Oregon (Fig. 1b). However, as the stresses evolve from the present to a future state, the maximum differential stress between the margin-parallel and vertical components ($\sigma_1$–$\sigma_3$) is limited by the strength of the continental crust which is represented by the straight line in Fig. 8a. Because the vertical stress is a constant, the margin-parallel $\sigma_1$ is not expected to exceed greatly the present level.

### 5.2. Nankai

Similar reasoning can be applied to the Nankai subduction zone. However, at Nankai, both margin-parallel and margin-normal stresses appear to be smaller than their counterparts at Cascadia, as shown in Fig. 8 and discussed above. In particular, the margin-normal compressive stress is smaller than the vertical stress. It is the difference between the two horizontal stresses ($\sigma_1$–$\sigma_3$) that causes strike–slip earthquakes in the continental forearc at present. The intermediate vertical stress is constant. The margin-normal $\sigma_3$ increases with time as a result of the locking of the subduction fault. Whether and when $\sigma_1$ increases or decreases depends on slips along the MTL (Fig. 2) which accommodates some of the margin-parallel plate motion. Generally, we do not expect $\sigma_1$ to increase faster than $\sigma_3$. In the future prior to the next great subduction earthquake, the differential stress ($\sigma_1$–$\sigma_3$) may become too small to cause strike–slip earthquakes. The continental forearc will then be seismically quiescent, shown as 'future' in Fig. 8b.

Changes in forearc seismicity have been observed at Nankai. Okano and Kimura (1979) and Kimura and Okano (1995) compiled sequences of earthquakes ($M>4$) that occurred from 1930 to 1994 in the overriding and subducting plates at the Nankai subduction zone (Fig. 11). From the upper plate sequence (Fig. 11a), the continental forearc was seismically quiescent before the 1944–1946 great Nankai subduction events, in exactly the same situation shown as 'future' in Fig. 8b. After the great subduction earthquakes, strike–slip style seismicity sharply increased in the continental crust, indicating a state of stress described as 'present' in Fig. 8b.

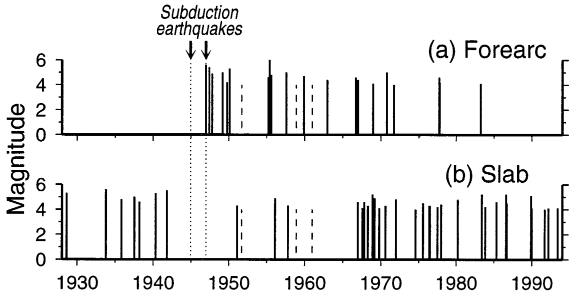


Fig. 11. Earthquake sequences reported by Kimura and Okano (1995). The 1944–1946 great Nankai subduction earthquakes affected the continental forearc seismicity pattern (a). The seismic quiescence (at this magnitude level) before the subduction events is described in Fig. 8b as the 'future' state prior to the next great subduction earthquake. The dashed lines are events that could not be clearly identified to be in the upper or lower plate.

Kimura and Okano (1995) offered the same explanation that the change was induced by a reduction of N–S compressive stress caused by the great Nankai earthquakes, although they did not consider the Nankai margin to be a subduction zone. It is reasonable to expect the decrease in seismicity that happened before the previous great subduction earthquake to happen again before the next one. In fact, by examining the earthquake sequence in Fig. 11a, it appears that the continental forearc has been fairly quiescent at this magnitude level since 1983. The Nankai observations substantiate the speculations in the preceding section on Cascadia seismicity. Hori and Oike (1996) showed that in most of Southwest Japan, north of the study area of Kimura and Okano (1995), crustal seismicity suddenly increased just after the 1944–1946 subduction earthquakes, consistent with Fig. 11a.

The earthquakes within the subducting slab at Nankai indicate downdip compression and margin-parallel (E–W) tension (Okano et al., 1985; Shiono, 1988; Konomi et al., 1992), opposite to those in the overriding plate. The slab earthquakes also tend to be larger and much more frequent (Kimura and Okano, 1992). Compared to upper plate events, the pattern is much less affected by great subduction earthquakes (Fig. 11b). The stress in the slab is probably controlled mainly by slab-mantle interaction, an issue that will not be pursued here.

### 5.3. Discussion

The above simple history of forearc stress and seismicity requires some provisos before it can be generalized:

(1) A great subduction earthquake must rupture a very large portions of the margin along strike, so that it reduces the margin-normal stress regionally in the forearc. Along the Nankai subduction zone, continental forearc seismicity in the past few decades has been higher in eastern and central Shikoku and the Kii Peninsula, which appears to be correlated with the main rupture areas of the two great subduction earthquakes in 1944–1946. If the rupture areas had been much smaller, they would have only locally affected the forearc stresses.

(2) Similarly, for the increase in the margin-normal stress to be uniform along strike, the locked portion of the subduction fault should also be very long. If only patches of the subduction fault are locked, and other parts slip aseismically, stress increase occurs around those locked portions, and the resultant seismicity pattern will be quite complex. In the models of Hyndman and Wang (1995) and Flück et al. (1997), the entire Cascadia subduction fault is assumed to be locked at present, but that model assumption remains to be tested against more geodetic data.

(3) Finally, the failure stress of the subduction fault may not be the same for each great earthquake even for the same subduction zone. An earthquake may occur before the state of stress evolves to the 'future' stage in Fig. 8. As discussed in Section 4.1, a weak fault can rupture even if the margin-normal stress is less than the vertical stress in the forearc.

## 6. Conclusions

(1) For elastic crustal deformation such as a forearc in response to subduction earthquake cycles, geodetic measurements reflect the temporal stress changes not the absolute stress. Maximum contraction and maximum compressive stress can be in different directions. Only when the observed deformation is permanent, do observed strain changes provide information on absolute stresses. The same principle applies to other tectonic settings.

(2) At Cascadia and Nankai and perhaps all subduction zones, subduction faults are very weak. The low coupling stress in these two places results in low margin-normal compression in the continental forearc and low frictional heating on the subduction fault. This does not mean that high margin-normal compression in some other subduction zones must indicate strong subduction faults.

(3) Weakly coupled subduction zones such as Cascadia and Nankai can produce great or giant subduction earthquakes. The magnitude of subduction earthquakes is not an indicator of the strength of subduction faults. The magnitude depends on the size of the fault rupture. How far the rupture propagates from the nucleation area must be controlled by the properties of the fault itself, which may depend on temperature, fluids and sediments in the fault zone, and the roughness of the subducting seafloor.

(4) Great subduction earthquakes cause relatively small perturbations to forearc stresses. Although the associated elastic strain changes are small, they occur quickly to result in high strain rates detectable by geodetic measurements. Margin-normal crustal contraction occurs when the subduction fault is locked, regardless of the absolute background stresses.

(5) Small stress changes throughout subduction earthquake cycles may affect the seismicity pattern in the forearc, depending on the regional stress. At Nankai, the increase in margin-normal stress due to a locked subduction fault slightly reduces the differential stress responsible for strike–slip earthquakes (with a margin-parallel $\sigma_1$), and the continental forearc may become seismically quiescent until a subsequent great subduction earthquake. Seismic quiescence in the Nankai forearc was observed before the 1944–1946 great Nankai subduction earthquakes. At Cascadia, a similar increase in margin-normal stress may change the forearc earthquakes from a mixture of strike–slip

and thrust events (also with a margin-parallel $\sigma_1$) into mostly the thrust type.

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