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TECTONIC IMPLICATIONS OF THE 1944 TONANKAI AND THE 1946 NANKAIDO EARTHQUAKES

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The tectonic process taking place in the south-western part of Japan to the north of the Philippine Sea is discussed on the basis of the focal mechanism of the 1944 Tonankai earthquake ($M \approx$ 8.0 to 8.3) and the 1946 Nankaido earthquake ($M \approx$ 8.1 to 8.4), as well as on the basis of the spatial distribution of small earthquakes in the neighbouring region. It is found that these two earthquakes represent a low-angle thrust faulting whose slip vector is almost parallel to that of other great earthquakes in the north-western Pacific; the oceanic side is underthrusting beneath the continent. The spatial distribution of small earthquakes in the Kii peninsula region shows a remarkable two-dimensional wedge-like distribution to a depth of 70 km. The strike of this two-dimensional distribution is almost perpendicular to the slip vector of the Tonankai and the Nankaido earthquakes suggesting that the Philippine Sea plate started underthrusting about 2×10^6 y ago and has now reached a depth of 70 km or so. This picture is consistent with the local travel time anomaly, crustal structure, electrical conductivity anomaly, etc., observed in this region. The source parameters of the 1944 Tonankai earthquake are: latitude 33.70° N, longitude 136.05° E, depth 30 km, dip angles $\delta = 10°$ (fault plane) and $\delta = 80°$ (auxiliary plane), dip directions $\phi = 306°$ (fault plane) and $\phi = 126°$ (auxiliary plane), dislocation 3.1 m, stress drop 33 b, released strain energy 5×10^{23} erg. The source parameters of the 1946 Nankaido earthquake are: latitude 33.13° N, longitude 135.84° E, depth 30 km, dip angles $\delta = 10°$ (fault plane), and $\delta = 80°$ (auxiliary plane), dip direction $\phi = 310°$ (fault plane) and $\phi = 130°$ (auxiliary plane), dislocation 3.1 m, stress drop 33 b, released strain energy 5×10^{23} erg.

1. Introduction

The region along the south-western coast of the Japanese Honshu Island, between the Izu-Bonin arc and the Ryukyu arc and to the north of the Philippine Sea (fig. 1), is extremely interesting from geophysical point of view. The features and phenomena of particular interest are

(1) a series of great earthquakes with magnitude around 8 which have occurred in the Tokaido–Nankaido region with a remarkable temporal regularity (see fig. 1),

(2) the unique micro-earthquake activity beneath the Kii peninsula region (IMAMURA *et al.*, 1932; MIYAMURA, 1959; KANAMORI and TSUMURA, 1971),

(3) the Median Tectonic Line which apparently bridges the Izu-Bonin and the Ryukyu arcs,

(4) low seismic wave velocity in the upper mantle beneath the Philippine Sea (KANAMORI and ABE, 1968),

(5) high heat flows in the Shikoku basin (WATANABE *et al.*, 1970),

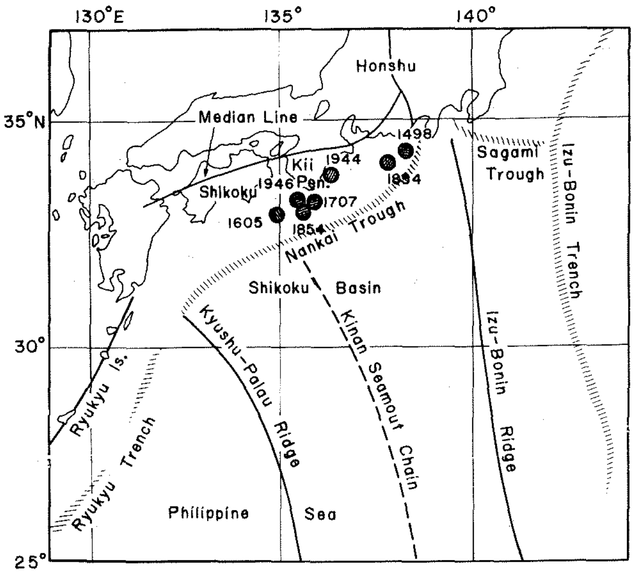

Fig. 1. A map showing the tectonic features in the south-western part of Japan. Epicentres of great earthquakes ($M \approx 8.0$) which occurred in the Tokaido–Nankaido region since 1400 are shown by hatched circles with the year of occurrence. The epicentres of old earthquakes are inevitably subject to large uncertainty.

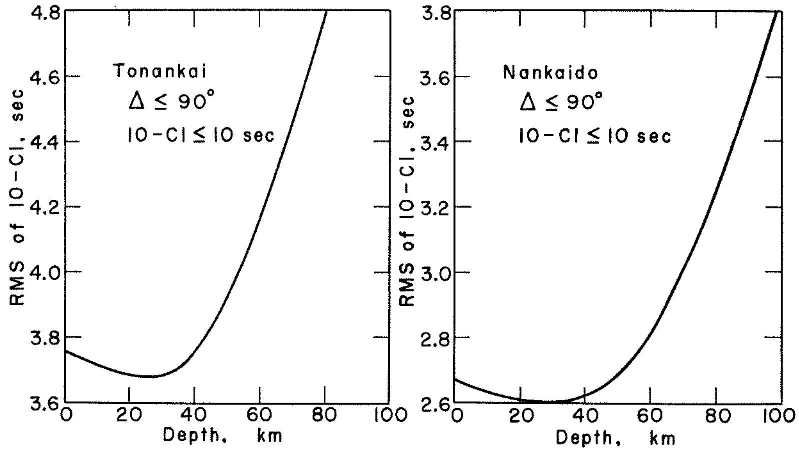

Fig. 2. The root-mean-square of the observed minus computed P times versus focal depth. The data at stations whose epicentral distance is less than 90° and whose O–C residual is less than 10 s are used.

(6) electrical conductivity anomaly centred at the Kii peninsula (RIKITAKE *et al.*, 1968).

In the previous study (KANAMORI, 1970), the occurrence of great earthquakes in the north-western Pacific was systematically explained in terms of a simple model which incorporates interactions between the oceanic and the continental lithospheres. In that model, however, the great earthquakes in the Tokaido–Nankaido region were not considered because no reliable data were available at that time for any of those earthquakes. Since then, old seismograms for the 1944 Tonankai and the 1946 Nankaido earthquakes have been collected for the purpose of elucidating the focal processes of these earthquakes and the tectonic processes taking place in the south-western part of Japan. This paper presents the results of such efforts and will show that the simple model previously presented can also explain in a systematic way the unique features and phenomena taking place in the south-western part of Japan.

Recently, KATSUMATA and SYKES (1969) discussed, on the basis of extensive data on seismicity and earthquake source mechanism, the tectonics of the western Pacific region including the Philippine Sea. FITCH and SCHOLZ (1971) discussed, primarily on the basis of the data on the crustal deformation associated with the 1946 Nankaido earthquake, the mechanism and the rate of the underthrusting in the south-western Japan. The present work is closely related to those of KATSUMATA and SYKES (1969), and FITCH and SCHOLZ (1971), but employs a different approach; data on great earthquakes in the north-western Pacific and the spatial distribution of micro-earthquakes beneath the Kii peninsula will be used.

2. Source parameters of the 1944 Tonankai and the 1946 Nankaido earthquakes

The epicentre coordinates and the focal depth of the 1944 Tonankai and the 1946 Nankaido earthquakes were redetermined on the basis of the data reported in the International Seismological Summary and in the Seismological Bulletin of the Central Meteorological Observatory. The method of these redeterminations is the same as that used by KANAMORI and MIYAMURA (1970). The hypocentre parameters thus estimated on the basis of the reported P times at stations whose epicentral distances are less than 90° and whose O–C residuals are less than 10 s are:

Tonankai earthquake: Origin time, $04^h35^m39^s$, December 7, 1944; latitude, 33.70° N; longitude, 136.05° E; depth, 30 km.

Nankaido earthquake: Origin time, $19^h19^m09^s$, December 20, 1946; latitude, 33.13° N; longitude, 135.84° E; depth, 30 km.

Fig. 2 shows the stability of the depth determination. The root-mean-squares (RMS) of the O–C residuals for various restrained focal depths are shown. It is evident that the focal depth of these earthquakes cannot be greater than 40 km; a depth of 30 km is most probable.

TABLE 1a

Surface wave magnitude of the 1944 Tonankai earthquake

Station	Distance (°)	Azimuth (°)	Amplitude (μm)	Period (s)	M
Bergen	78.0	337	2860	20	8.60
Bermuda	111.0	19	660	20	8.21
Lisbon	101.1	333	1320	19	8.47
Ottawa	96.1	22	780	18	8.24
Riverview	68.6	167	565	20	7.80
Stuttgart	85.3	328	1770	18	8.51
Victoria	71.8	40	377	20	7.66
Wellington	82.6	152	503	23	7.83
				Average	8.16

TABLE 1b

Surface wave magnitude of the 1946 Nankaido earthquake

Station	Distance (°)	Azimuth (°)	Amplitude (μm)	Period (s)	M
Belgrade	78.4	337	2400	20	8.52
Bergen	68.1	166	1000	20	8.04
Lisbon	101.5	332	1200	19	8.43
Ottawa	97.0	22	394	19	7.91
Riverview	82.7	319	590	17	8.04
				Average	8.19

The seismograms obtained by classical instruments at seismographic stations of the world were collected for determination of the magnitude, the distribution of P-wave first motions, and the polarization angles of S waves. The magnitude determination was made by the method described by KANAMORI and MIYAMURA (1970), and the results are summarized in table 1. The average surface wave ($T \approx 20$ s) magnitudes are 8.16 and 8.19 for the Tonankai and the Nankaido earthquakes, respectively. These values may be compared with the values 8.3 and 8.4 given by RICHTER (1958) and with 8.0 and 8.1 given by JMA (1958).

The P-wave first motions and the S-wave polarization angles on the lower half of the focal sphere are shown on the Wulff grid in figs. 3a and 3b. The hypocentres are placed at the Moho discontinuity (33 km). The stations whose epicentral distance is larger than 2° are used. The distributions of the first-motions and the polarization angles are surprisingly similar for these two earthquakes. In interpreting these data, we employ the double-couple model which is now receiving overwhelming support, and determine the two orthogonal nodal planes. The nodal planes thus determined are shown in figs. 3a and 3b. The dip direction and the dip angle are $\phi = 126°$ and $\delta = 80°$ for plane a and $\phi = 306°$ and $\delta = 10°$ for plane b for the Tonankai earthquake, and $\phi = 130°$ and $\delta = 80°$ for plane a and $\phi = 310°$ and $\delta = 10°$ for plane b for the Nankai-

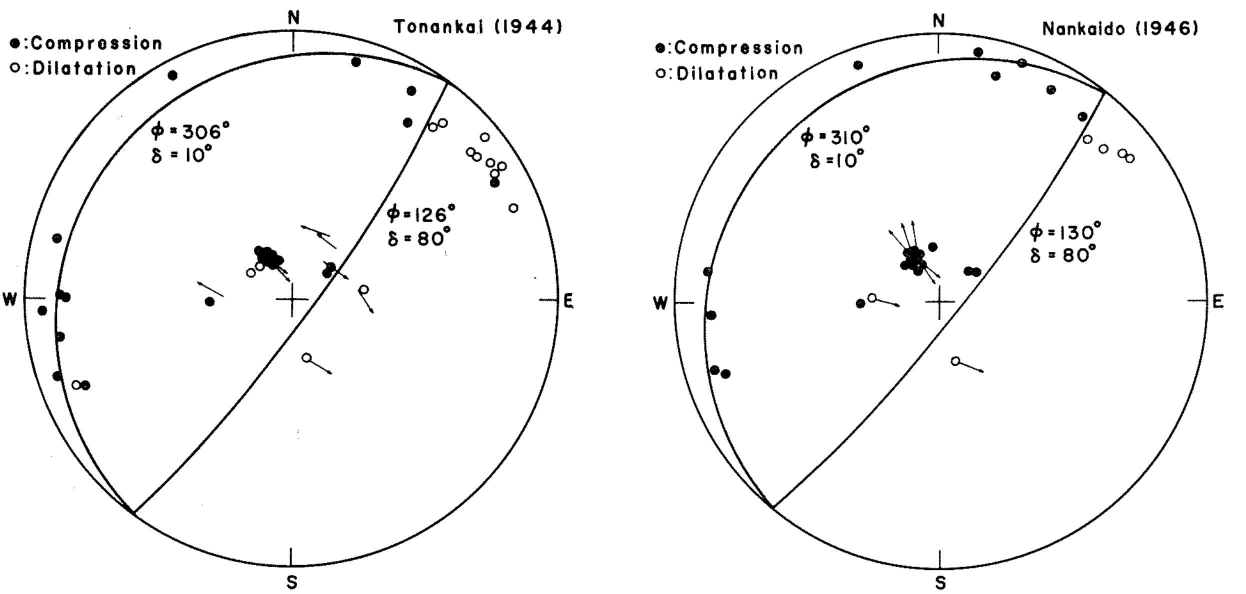

Fig. 3. The first-motion data and the S-wave polarization angles obtained for the Tonankai (a) and the Nankaido (b) earthquakes. The lower half of the focal sphere is projected on the Wulff grid. ϕ is the dip direction measured clockwise from the north and δ is the dip angle.

do earthquake. The nearly vertical nodal plane is constrained fairly well; it is approximately perpendicular to the arrows showing the polarization angles in accordance with the double-couple model. The dip direction of the gently dipping plane is more or less arbitrary because it becomes indeterminate when the dip angle is 0. Thus we conclude that the faulting of these earthquakes is essentially a reverse faulting with one nodal plane nearly vertical and the other nearly horizontal. The solutions shown in figs. 3a and 3b are not completely consistent with the first-motion data at near stations, but they are still tolerable in view of the uncertainties in the emergence angle of the ray leaving the focus for near stations. Although the selection of the actual fault plane out of the two P-wave nodal planes cannot be made from the data of the first-motions and the S-wave polarization angles, the spatial distribution of the aftershocks (fig. 4) strongly suggests that the gently dipping nodal plane is the actual fault plane; a nearly vertical fault plane is unlikely to yield an aftershock area which spreads out horizontally. Taking the gently dipping nodal plane as the fault plane, the slip vectors of the foot-wall side are determined as shown in fig. 5. The slip vector is nearly perpendicular to the eastern coast line of the Kii peninsula. It should be noted that the uncertainty in the dip direction of the gently dipping plane does not affect the direction of the slip vector.

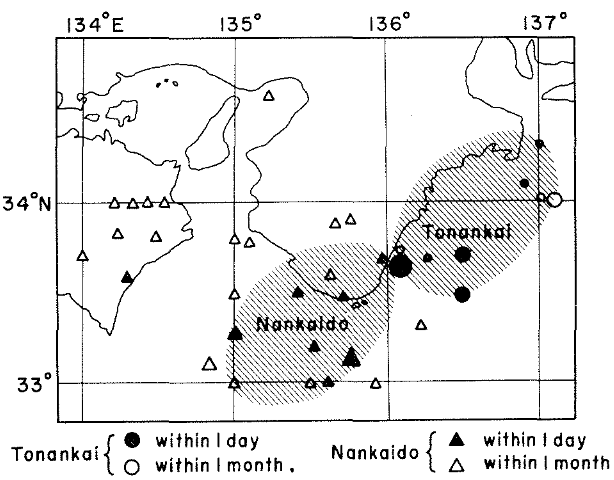

Fig. 4. *The aftershocks of the Tonankai (circles) and the Nankaido (triangles) earthquakes.* Filled circles and triangles are aftershocks within 24 h after the main shock, and open circles and triangles are those within 1 month after the main shock. The size of the circles and triangles corresponds roughly to the magnitude of earthquakes. The hatched zone shows the assumed fault surfaces.

Although the seismic moment of these earthquakes could not be determined precisely, a rough estimate was made as follows. The amplitudes of long-period (about 100 s) surface waves excited by the two earthquakes are found to be about the same; they are also comparable to those excited by the Kanto earthquake of 1923 (KANAMORI, 1971b), but are much smaller than that excited by the Tokachi–Oki earthquake of 1968 (KANAMORI, 1971c) and the Sanriku earthquake of 1933 (KANAMORI, 1971a). The Kanto earthquake and the Tokachi–Oki earthquake have a seismic moment of 7.6×10^{27} dyne · cm and 2.8×10^{28} dyne · cm, respectively, and both are characterized by a faulting having a considerable (about 50%) strike–slip component. The Sanriku earthquake has a seismic moment of 4.3×10^{28} dyne · cm and is characterized by a dip–slip faulting. In general, dip–slip faultings need a larger (about two times) seismic moment than strike slip faultings to yield the same amplitude of long-period surface waves (BRUNE and ENGEN, 1969). On these grounds we can very roughly estimate the seismic moment of the 1944 Tonankai and the 1946 Nankaido earthquake as 1.5×10^{28} dyne · cm. The corresponding dislocation can be estimated as 3.1 m if the fault area and the rigidity are assumed as 80×120 km^2 and 5×10^{11} dyne · cm^{-2}, respectively. The stress drop $\Delta\sigma$, the strain drop $\Delta\varepsilon$ and the released strain energy ΔW (the shear stress at the fault surface is assumed to vanish after the faulting) can be estimated, by the static dislocation theory, as follows: $\Delta\sigma = 33$ b, $\Delta\varepsilon = 0.6\times10^{-4}$ and $\Delta W = 5\times10^{23}$ erg. In the above, the estimate of the fault area is made on the basis of the aftershock area shown in fig. 4. The size of the aftershock area one day after the main shock has been shown to be approximately equal to the fault size estimated from the rupture length (e.g. KANAMORI, 1970). The aftershock area usually spreads out during the subsequent period of time, but this later aftershock activity seems to represent the readjustment of the stress caused by the main shock, but not the fracturing on the fault surface itself. On this ground, we take the aftershock area one day after the main shock to estimate the fault size. For the Nankaido earthquake, however, one problem arises in this regard. The aftershock activity of the Nankaido earthquake spread

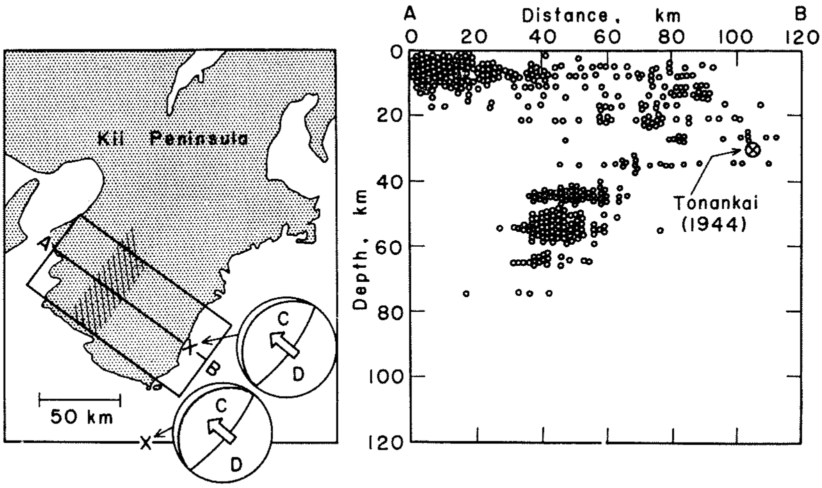

Fig. 5. The figure on the right shows the hypocentres of small earthquakes (for a period from January 1969 to July 1969) in the rectangular region shown in the figure on the left. The projection on the vertical plane at AB is shown. The hatched region shows the trend of earthquakes with depths from 50 to 70 km. The mechanism diagrams and the slip vectors of the Tonankai and the Nankaido earthquakes are shown in the figure on the left.

widely towards the west during the one month period following the main shock (see fig 4). Further, the land deformation associated with this earthquake extended even beyond the enlarged aftershock area (e.g. FITCH and SCHOLZ, 1971). Thus, the actual dislocation seems to have taken place over an area which is much larger than the one day aftershock area. On the other hand, the amplitude of long-period waves ($T \gtrsim 100$ s) excited by the Nankaido earthquake is about the same as that of the Tonankai earthquake and is anyway not exceptionally large; this observation suggests that the seismic waves of the Nankaido earthquake are unlikely to have originated from such a large area as that expected from the land deformation. Thus we suspect that the land deformation to the west of the one day aftershock area is associated with a creep-like deformation having a time constant longer than one minute or so. The lack of aftershocks to the west of 134° E seems consistent with the idea that the deformation which took place in this region is creep-like. Confirmation of this possibility, however, must await future studies.

3. Spatial distribution of small earthquakes

The spatial distribution of small earthquakes (magnitude $\leqq 4.5$) beneath the Kii peninsula shows a striking two-dimensional wedge-like feature (KANAMORI and TSUMURA, 1971). It is noteworthy that the strike of this two-dimensional wedge-like distribution is almost perpendicular to the slip vectors of the Tonankai and the Nankaido earthquakes and that the fault plane of the Tonankai earthquake marks the lower boundary of this activity as shown in fig. 5. The hypocentres of small earthquakes plotted in this figure are based on the data obtained during the period from January 1969 to July 1969 (Tsumura, personal communication, 1971). Whether a similar spatial relation between the Nankaido earthquake and small earthquakes exists towards the south is not known because such small earthquakes offshore cannot be detected. However, the distribution of moderate-size earthquakes located by the Japan Meteorological Agency Network for a longer period of time suggests an extension of the NE–SW trending distribution of earthquakes towards the south-west (fig. 6). In the framework of plate tectonics, this special relation between the shape of this two-dimensional distribution of small earthquakes and the directions of the slip vectors of such major thrust earthquakes as the 1944 Tonankai and the 1946 Nankaido earthquakes immediately suggests a model which represents the tectonics of this region. Fig. 7 illustrates such a model. The Philippine Sea plate underthrusts at the Nankai trough with a very low dip angle beneath the Kii peninsula. The clustering of small earthquakes at a depth around 50 km and the complete absence of earthquakes at depths deeper than 80 km suggests that the leading edge of the underthrusting

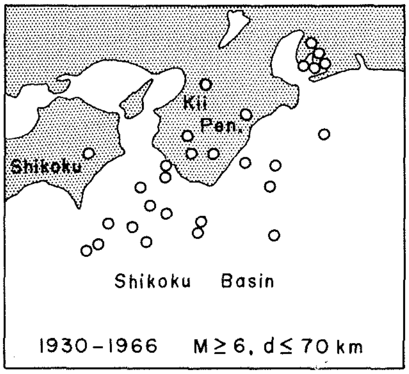

Fig. 6. Distribution of earthquakes whose magnitude is larger than 6 and whose depth is smaller than 70 km located by the Japan Meteorological Agency for a period from 1930 to 1966 (after KATSUMATA, 1970).

plate reaches to a depth of only 50 to 70 km and to a distance of 50 km inland. The Tonankai and the Nankaido earthquakes represent an elastic rebound which takes place intermittently at the interface between the upper surface of the underthrusting Philippine Sea plate and the continental lithosphere. The dip angle of this interface is very small in accordance with the fault-plane solutions shown in figs. 3a and 3b. Few earthquakes occur in the presumably rigid and homogeneous Philippine Sea plate. The earthquake activity shallower than 40 km represents fracturings of the relatively weak and inhomogeneous continental lithosphere caused by the stress associated with the interaction between the continental lithosphere and the underthrusting Philippine Sea plate. The source mechanism of shallow earthquakes belonging to the earthquake cluster in the upper left-hand corner of fig. 5 has been studied by WATANABE and KUROISO (1967), MIKUMO *et al.* (1970) and SHIONO (1970). They showed that the compression axes of these earthquakes are concentrated in the E–W to NW–SE directions while the tension axes are distributed uniformly on a plane perpendicular to the average pressure axis. The observation that the average compressional axis is oriented in a direction more or less parallel to the slip vectors of the Tonankai and the Nankaido earthquakes strongly favours the above model. Whether the clustering of earthquakes at depths around 50 km represents the activity in the continental lithosphere or in the underthrusting Philippine Sea plate is not clear, but it certainly seems to represent the collision of the leading edge of the underthrusting Philippine Sea plate against the continental lithosphere. The source mechanism of several earthquakes belonging to this group has been also studied by SHIONO (1970). The compression axes are all oriented in a direction more or less parallel to the slip vector of the Philippine Sea plate, again favouring the above model.

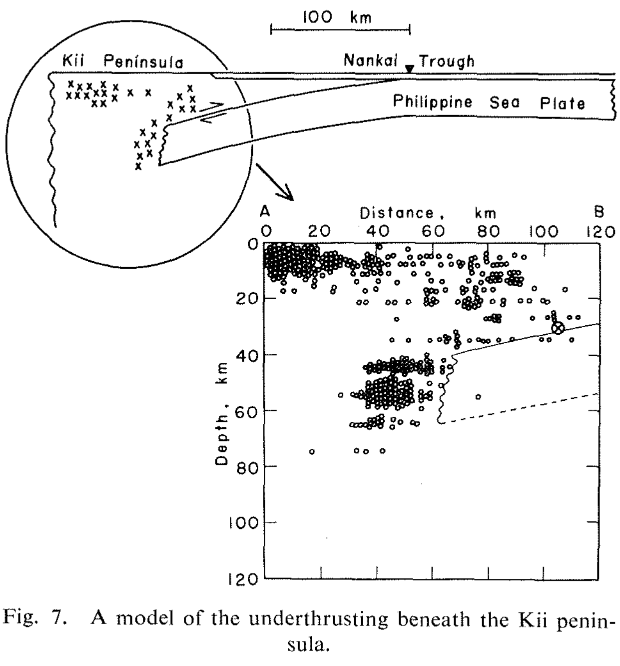

Fig. 7. A model of the underthrusting beneath the Kii peninsula.

Thus it may be concluded that the present-day tectonics of this region is characterized by the recent underthrusting of the Philippine Sea plate beneath the Kii peninsula. The lack of active volcanism in this region and of a well developed trench favours the idea that the underthrusting is a fairly recent event. This picture is in accord with that of HILDE *et al.* (1969) who found that the late Neogene to Recent turbidite is confined only to the Nankai trough, and suggested that the Nankai trough is a juvenile trench and that the Shikoku basin is underthrusting beneath the south-western Japan. The time T which elapsed since the underthrusting began can be roughly estimated from the length L of the underthrusting Philippine Sea plate, the slip $\bar{u}$ of the Tonankai and the Nankaido earthquakes, and the recurrence time τ of these great earthquakes. If we take the values $L \approx 60$ km from fig. 7, $\bar{u} \approx 3$ m from the previous discussion, and $\tau \approx 100$ y (see fig. 1), we have $T \approx (60 \times 10^5/3 \times 10^2) \times 100 = 2 \times 10^6$ y. This estimate is evidently subject to large uncertainty due to many factors, particularly the possible creep just after and

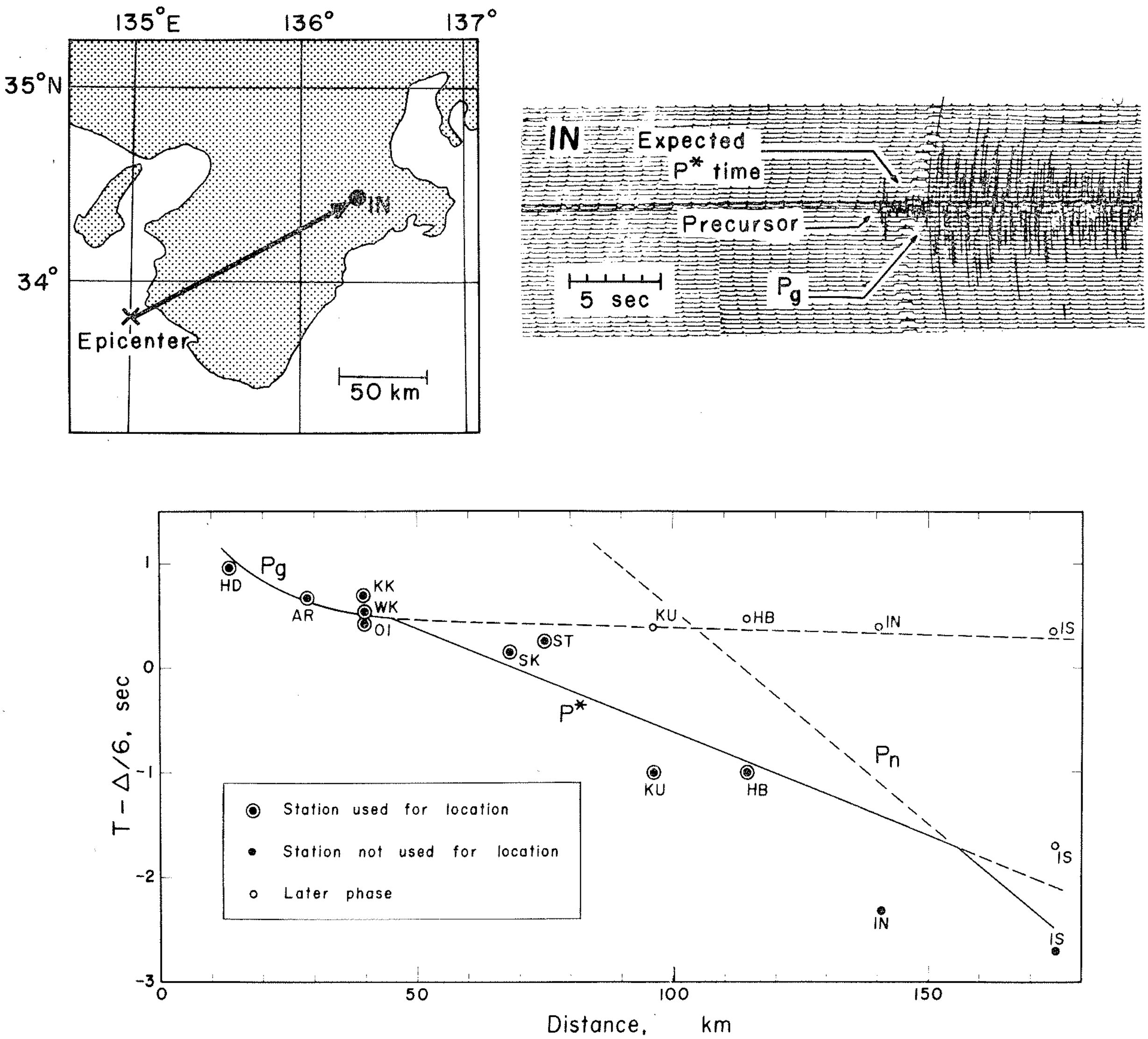

Fig. 8. An example of anomalous precursor observed at station IN for the events in the neighbourhood of the cross mark (after KANAMORI and TSUMURA, 1971).

between major earthquakes; nevertheless it can be used as an order of magnitude estimate. This estimate is consistent with that made by FITCH and SCHOLZ (1971) who interpreted the data on the crustal deformation associated with the 1946 Nankaido earthquake, and suggested that the underthrusting of the Philippine Sea plate must have started about 10^6 y ago. The present study, particularly fig. 5, provides a firm evidence that the underthrusting has reached only to a depth of 60 km or so. Thus, in terms of the model presented by KANAMORI (1970), the tectonic process now taking place in the south-western Japan can be considered to represent a stage prior to that for the Alaska–Aleutian region where the underthrusting of the oceanic lithosphere reaches a depth of 150 to 200 km.

4. Other phenomena supporting the underthrust model

4.1. *Travel time anomaly*

KANAMORI and TSUMURA (1971) found a very large local travel time anomaly for a certain epicentre–station combination in the Kii peninsula region. Fig. 8 shows an example. Kanamori and Tsumura showed that the local travel times in the Kii peninsula region can be fitted, in general, in terms of a uniform crustal structure within an error of 0.2 s. However, at station IN, a negative travel time residual as large as -1.5 s is consistently observed for all the earthquakes whose epicentres are in the neighbourhood of the cross mark shown in fig. 8. Since the epicentral distance is 120 to 150 km, this early arrival suggests the existence of a region with an extremely high seismic velocity at depths below 20 km. Since the model shown in fig. 7 suggests an injection of cold, and therefore high-velocity, material at such depths along the profile, this observation is consistent, at least qualitatively, with the underthrust model.

4.2. *Crustal structure beneath the Nankai trough*

The crustal structure beneath the Shikoku basin off the coast of Shikoku has been studied by DEN *et al.* (1968), and YOSHII *et al.* (1970). Yoshii *et al.* suggested, on the basis of a structure along a profile perpendicular to the coast, that the oceanic crustal structure extends across the Nankai trough towards inland. This picture favours the idea that the oceanic lithosphere is now underthrusting beneath the continent.

4.3. *Electrical conductivity anomaly (CA) and heat flow*

The anomaly of the ratio $\Delta Z/\Delta H$, where ΔZ and ΔH are the vertical and the horizontal components, respectively, of the short-period geomagnetic variations, particularly of the bay type variations of the period of one hour or so, has been suspected to be caused by the anomaly of the electrical conductivity distribution within the Earth (see RIKITAKE, 1966). One of the most striking anomalies is found in central Japan (RIKITAKE *et al.*, 1968) as shown in fig. 9. It is interesting to note that this anomaly is centred at the Kii peninsula and has a strike more or less parallel to that of the leading edge of the postulated underthrusting Philippine Sea plate. RIKITAKE (1969) interpreted this anomaly in terms of a depression of a high conductivity, and probably high temperature, layer beneath the eastern coast of Honshu island; this model is equivalent to that in which a low conductivity, and probably low temperature, layer exists at a relatively shallow depth beneath the eastern coast of Honshu island. This low-temperature layer may be accounted for, at least partially, by the cold Philippine Sea plate which is underthrusting beneath the Kii peninsula at a relatively shallow depth.

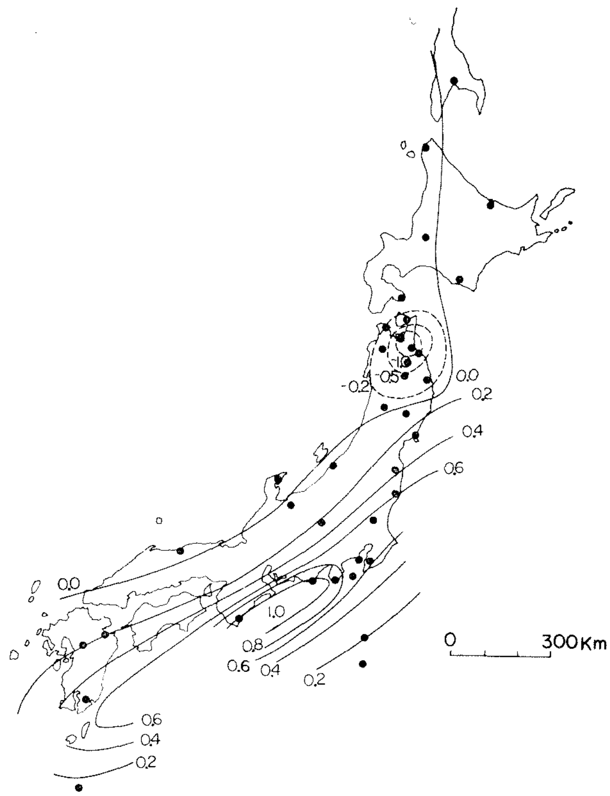

Fig. 9. Distribution of $\Delta Z/\Delta H$ for geomagnetic variation in Japan (after RIKITAKE *et al.*, 1968).

The present underthrust model may yield a rather complicated heat flow distribution, because the model involves, on the one hand, injection of a cold plate but, on the other, generation of frictional heat. Beneath the Kii peninsula which the underthrusting cold plate has just reached, the frictional heating may not have become so extensive that the effect of the injection of the cold plate may dominate. On the other hand, beneath the Shikoku basin off the coast, the frictional heating has continued for a fairly long time, 10^6 y or so, so that high heat flows may be expected. This model might partly account for the high heat flow values reported by WATANABE *et al.* (1970) for this region. For a construction of a dynamic model which explains the relation among CA, heat flow and the underthrusting plate, more quantitative analyses are obviously necessary.

5. Discussion

The underthrust model proposed in the preceding

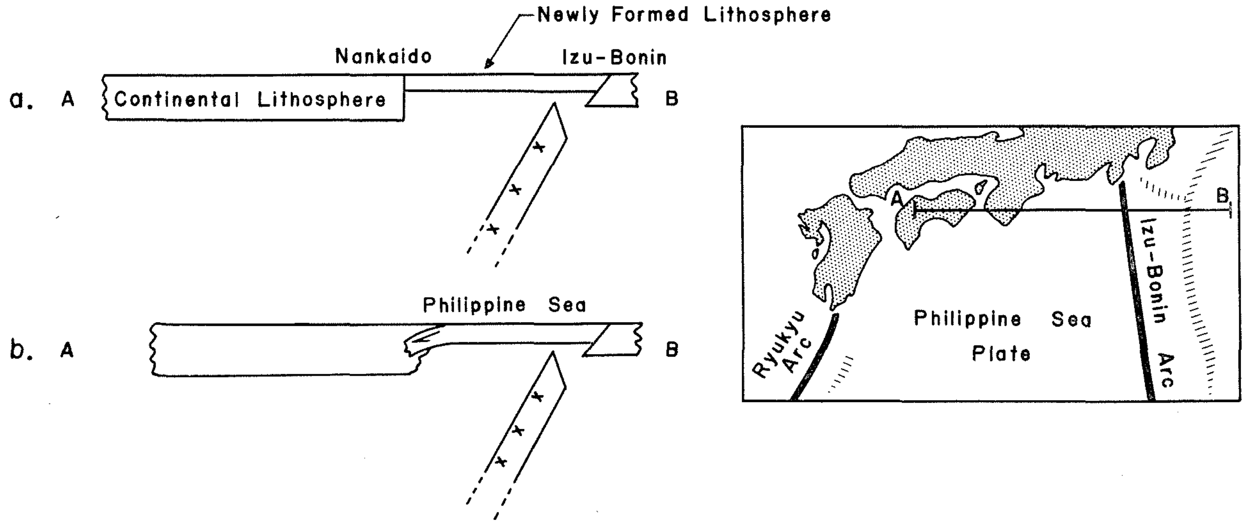

Fig. 10. A model of the underthrusting of the Philippine Sea plate in the Nankaido region. The figures on the left show the cross section at AB at two successive stages.

sections can be fitted in the framework of the model proposed by KANAMORI (1970). Essential to that model is a complete detachment of the lithosphere which is underthrusting beneath the continental lithosphere. Such detachment is caused by repeated large-scale normal faultings as evidenced by the 1933 Sanriku earthquake (KANAMORI, 1971a). It is suggested that when such detachment takes place, the remaining portion of the oceanic lithosphere loses the mechanical support from the continental lithosphere, and becomes likely to snap off. The Izu–Bonin arc is considered to be a surface manifestation of a lithosphere that snapped off in such a manner. The Philippine Sea is thus a relatively young feature which was formed by an overturn of the asthenosphere associated with the sinking of the Izu–Bonin lithosphere. The asthenosphere that rose to the surface eventually cooled down, and became a relatively rigid layer, lithosphere (fig. 10). KATSUMATA and SYKES (1969) suggested that a lack of seismic events in a large part of the Philippine Sea constitutes an evidence for the Philippine Sea being a single plate. The average thickness of the lithosphere beneath the Philippine Sea basin has been estimated to be about 30 km (KANAMORI and ABE, 1968; ABE and KANAMORI, 1970) which is considerably smaller than that beneath ordinary ocean basins, 70 km (see KANAMORI and PRESS, 1970). This newly formed lithosphere, though thinner than normal, then became capable of transmitting the stress caused by the relative motion between the Asian plate and the remaining Pacific plate east of the Izu–Bonin arc. The natural consequence of this stress transmission is the commencement of the underthrusting at the Philippine Sea–continent border towards the west (fig. 10). If the slip rate and the recurrence rate previously estimated are used, this underthrusting must have started about 2×10^6 y ago; prior to this epoch of underthrusting, the relative motion between the Asian and the Pacific plates had been absorbed by a more or less plastic deformation of the Philippine Sea basin.

The idea that the underthrusting started relatively recently, about 2×10^6 y ago, is consistent with the Early-to-Late Quaternary tectonics in this region. HUZITA (1969) suggested, from the structural geology in this region, that the orientation of the maximum compression axis of the tectonic stress must have changed from NS to approximately EW direction about 1 to 2×10^6 y ago. YOSHIKAWA *et al.* (1964) suggested, from the study of the coastal terraces on the southern tip of Shikoku island, that the uplift must have started not much earlier than 0.2×10^6 y ago.

In the model postulated above, no relative motion takes place between the Philippine Sea plate and the Pacific plate. The slip vector of the Philippine Sea plate should therefore be parallel to that of the Pacific plate. MCKENZIE and PARKER (1967) showed that the slip vectors of eighty North-Pacific earthquakes are parallel to one another. Comparison of the slip vectors of the 1944 Tonankai, the 1946 Nankaido earthquakes, and two earthquakes (nos. 4 and 15) reported by KATSUMATA and SYKES (1969) with those compiled by MCKENZIE and PARKER (1967) clearly shows that the parallelism is indeed the case (fig. 11). In fig. 11 are included the results for the 1968 Tokachi–Oki earthquake (KANAMORI, 1971c), the 1933 Sanriku earthquake (KANAMORI, 1971a) and the 1923 Kanto earthquake (KANAMORI, 1971b). The only exception is the 1933 Sanriku earthquake; the orientation of the slip vector is almost perpendicular to the trench axis and

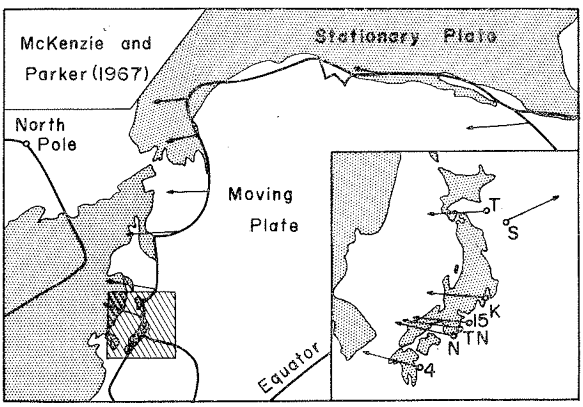

Fig. 11. A Mercator projection of the Pacific with a pole at 50° N and 85° W and the slip vectors of earthquakes in the Pacific (after MCKENZIE and PARKER, 1967). The insert which is an enlarged map of the hatched region shows the slip vectors of the Tokachi–Oki (T), Sanriku (S), Kanto (K), Tonankai (TN), Nankaido (N) earthquakes (after KANAMORI, 1971a, b, c), and the two earthquakes (nos. 4 and 15) reported by KATSUMATA AND SYKES (1968).

is quite different from that for other earthquakes. This difference, however, is quite natural if one admits the view that the Sanriku earthquake represents a fracture within the lithosphere and has nothing to do with the slip of the Pacific plate relative to the continental lithosphere (KANAMORI, 1971a).

As shown in fig. 1, the occurrence of great earthquakes in the Tokaido–Nankaido region is surprisingly regular. They occurred in 1498 (M = 8.6), 1605 (M = 7.9), 1707 (M = 8.4, suspected to be a twin earthquake), 1854 (M = 8.4, M = 8.4, the latter occurred 32 h after the former), and 1944 (M = 8.0), 1946 (M = 8.1). It is noteworthy that this remarkable temporal regularity is found in the Tokaido–Nankaido region where the underthrusting started relatively recently. Except in the Alaska–Aleutian and the Tokaido–Nankaido regions, the regions of large thrust earthquake activity is associated with intermediate and deep earthquakes which suggest that the underthrusting has continued for a relatively long time. In these regions, no such remarkable temporal regularity is found. According to the model presented by KANAMORI (1970), the mode of the occurrence of shallow thrust earthquakes is governed by the mode of the interaction between the oceanic and the continental lithospheres; the mode of the interaction changes as the time elapses after the commencement of the underthrusting. In the earlier stage of the underthrusting both lithospheres are more or less homogeneous, and the recurrence rate of the elastic rebound is controlled primarily by the slip rate; the elastic rebound therefore takes place more or less regularly. Thus the presence of the remarkable regularity of the recurrence rate of major earthquakes in this region of relatively recent underthrusting seems quite natural. In the later stage of the underthrusting, however, the continental lithosphere becomes inhomogeneous owing to the partial melting caused by frictional heating and to fracturings associated with the repeated rebounds. The recurrence rate, though basically controlled by the slip rate, is then affected by this inhomogeneity and therefore becomes irregular.

In this connection, the scarcity of aftershocks for these two earthquakes (see fig. 4), particularly for the Tonankai earthquake, should be noted; the aftershocks are surprisingly few for an earthquake with magnitude as large as 8. If this scarcity is not primarily due to a possible deterioration, caused by World War II, in the detection capability of the Japan Meteorological Agency Network, it again seems to reflect the relative homogeneity of the continental lithosphere in this region of recent underthrusting.

In the model presented above, the Median Tectonic Line is not a feature which primarily controls the present tectonics in this region, although it was a feature of primary importance in the past tectonics (e.g. MIYASHIRO, 1961). The right-lateral displacement recently reported at the Median Tectonic Line (KANEKO, 1966; OKADA, 1968) seems to represent a motion of the southern block caused by the interaction between the Asian plate and the underthrusting Philippine Sea plate. The direction of the motion is controlled not only by the slip direction of the Philippine Sea plate but also by the strike of the weak zone, the Median Tectonic Line.

6. Conclusion

The faulting of the 1944 Tonankai and the 1946 Nankaido earthquakes is found to be a low-angle thrust faulting, the oceanic side underthrusting beneath the continent. A two-dimensional wedge-like distribution of small earthquakes extending to a depth of 70 km and striking in a direction perpendicular to the slip vectors of the Tonankai and the Nankaido earthquakes is also found. These findings constitute a firm evidence for the Philippine Sea plate being at the initial stage of underthrusting. The observation that the slip vectors of these two earthquakes are parallel to those of other great earthquakes in the north-western Pacific suggests that no relative motion takes place between the Philippine Sea plate and the Pacific plate. It thus appears that the Philippine Sea plate acts simply as a stress guide between the Asian and the Pacific plate.

Because the commencement of the underthrusting is so recent (about 2×10^6 y ago), the zonal arrangements of such island-arc tectonic features as active volcanism, deep trench and negative gravity anomaly have not been developed yet in the south-western Japan.

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