EARTHQUAKE MECHANISMS AT THE HEAD OF THE PHILIPPINE SEA PLATE BENEATH THE SOUTHERN KANTO DISTRICT, JAPAN

Paul Somerville*

Earthquake Research Institute, University of Tokyo, Tokyo, Japan
(Received May 7, 1979; Revised June 10, 1980)

An analysis has been made of the rupture mechanism of two earthquakes which occurred in the vicinity of the oblique collision beneath the tip of the descending Philippine Sea plate and the Pacific plate beneath Chiba, Japan. The $M=6.5$ 1956 Chiba earthquake occurred in a dense nest of seismicity at a depth of between 70 and 80 km beneath Chiba city, and may represent deformation at the tip of the Philippine Sea plate. The mechanism of the $M=5.5$ 1965 Chiba earthquake indicates that the stress field in the lower seismic plane of the Pacific plate beneath Chiba is undisturbed by the collision.

The $M=6.1$ 1968 Saitama earthquake may be regarded as an interplate event occurring on the upper surface of and near the tip of the Philippine Sea plate. The high stress drop of several hundred bars of this earthquake and of the 1956 earthquake in the Chiba nest may reflect the penetration of the tip of the Philippine Sea plate into the Eurasian and Pacific plates, respectively.

1. Introduction

The rupture mechanism of two earthquakes which occurred beneath Chiba has been examined in order to improve our understanding of the tectonic stress there. This region is tectonically complex because of the oblique collision of two subducting slabs, the Pacific plate and the Philippine Sea plate, beneath the southern Kanto district. The existence of this collision has been inferred by many Japanese seismologists.

The Izu block, which constitutes the northern tip of the Philippine Sea plate, is colliding with the Eurasian plate in central Honshu (Fig. 1). The Sagami trough, which lies between the Izu block and the southern Kanto district, is a segment of the boundary between the Philippine Sea plate and the Eurasian plate. The convergence of these two plates is accompanied by great earthquakes on the Sagami trough (Matsuda et al., 1978), collision between the Izu block and Honshu in the southern Fossa Magna (Matsuda, 1978), and internal deformation of the Izu block (Somerville, 1978).

While the dominant component of motion between the Philippine Sea plate
and the Eurasian plate along the Sagami trough is strike-slip, there is a significant component of underthrusting (KANAMORI, 1971). Thus the Philippine Sea plate underthrusts the Eurasian plate along the Sagami trough, and dips roughly NNE at a shallow angle beneath the Kanto district. At the same time, the Pacific plate underthrusts the Eurasian plate along the Japan trench, and descends westwards at a shallow angle beneath the Kanto district. Thus the two plates must be colliding beneath the south Kanto district. The purpose of this paper is to shed some light on this collision process by studying in detail the rupture mechanisms of two earthquakes in the vicinity of the collision zone.

2. Seismicity of the Southern Kanto District

The south Kanto district has been seismically very active throughout historic times. USAMI (1976) has compiled a catalog of earthquakes, beginning in 799, which caused damage in the Kanto area. He distinguished two major groups of seismic activities among the earthquakes which caused damage in Tokyo. The first group consists of earthquakes taking place west and south of Tokyo,
which are presumably interplate earthquakes on the Sagami and Suruga troughs, and intraplate earthquakes in the Izu Peninsula. It is usually thought that the great earthquakes which have struck Tokyo occurred on the Sagami trough (the interface between the Philippine Sea and Eurasian plates) and not on the interface between the Pacific and Eurasian plates.

The second group, which is more numerous, consists of earthquakes which occurred near Tokyo. The epicenters of this latter group, as plotted by Usami, are distributed within a very narrow zone extending along the west shore of Tokyo Bay from Yokohama to Matsudo. This zone coincides precisely with the modern

![Fig. 2. Seismicity of the southern Kanto district in the depth range 40–80 km, 1971–1977 (after Maki et al., 1980). Epicenters of the 1968 Saitama earthquake and the Chiba earthquakes of 1956 and 1965 are shown. The fault plane of the 1923 Kanto earthquake, and its extension downdip to a depth of 50 km, are shown by long dashed lines and short dashed lines respectively. The profile shown in Fig. 6 is along BB'. Seismograph stations used are shown with station code and triangular symbol.](image)
metropolitan area.

If we study the distribution of instrumentally located earthquakes in the Kanto region (Tsumura, 1973; Maki et al., 1980—Fig. 2) we see that the concentration of epicenters on the west shore of Tokyo Bay, which is so prominent in Usami's map, is not apparent in the instrumentally located seismicity map. There are many possible explanations for this discrepancy, for example a temporal change in seismicity, or a difference in the seismicity of small and large earthquakes. However, it appears that there may have been bias in the location of historical earthquakes. Several factors may have contributed to this bias. Firstly, the population distribution favors the reporting of earthquakes in areas that have been heavily populated in historical times, as Tokyo was. Secondly, alluvial deposits along the west shore of Tokyo Bay amplify ground motion and thereby distort the intensity distribution.

A distorted intensity distribution is apparent for most of the earthquakes plotted by Usami (1976) which have instrumentally located epicenters. A remarkable example is the earthquake of 21 May 1928. Although the instrumental epicenter is in central Chiba, the largest intensity is on the west shore of Tokyo Bay. This earthquake would probably be located in Tokyo if the intensity distribution alone were used. It therefore seems reasonable to conclude that many of the Tokyo earthquakes in Usami's map really occurred elsewhere, perhaps in regions that can be recognized in the modern-day-instrumental seismicity.

The most conspicuous feature of the modern day seismicity is the nest of seismicity beneath Chiba city in the depth range 70–80 km (Fig. 2) to which the previously-mentioned 1928 earthquake belongs. It has been suggested by Maki et al. (1980) that this nest is a manifestation of the collision zone between the descending Pacific and Philippine Sea plates. It is therefore important to study this nest, since it seems likely that a large number of the historical earthquakes that brought destruction to Tokyo occurred in this nest.

3. Seismogram Analysis

In this study we investigate the rupture mechanism of two earthquakes. The first is a moderately-large ($M=6.5$) earthquake that occurred within the nest beneath Chiba city. This earthquake is of interest since it is the largest event to have occurred in the nest in recent years. Our objective in studying this earthquake is to estimate the state of stress at the center of the nest. The second event, having a magnitude of 5.5, is the largest earthquake to have occurred recently within the lower seismic plane of the Pacific plate beneath the Chiba nest, and provides an opportunity to determine to what extent this zone is disturbed by the collision between the two descending slabs.

3.1 Nodal plane solution

P-wave first motion data reported in the Bulletins of the Japan Meteorological
Earthquake Mechanisms at the Head of the Philippine Sea Plate

Table 1. Nodal plane parameters of the 1956 and 1965 Chiba earthquakes.

<table>
<thead>
<tr>
<th>Event</th>
<th>Location</th>
<th>Fault dip dir</th>
<th>Plane dip dir</th>
<th>Slip angle</th>
<th>Auxiliary dip dir</th>
<th>Plane dip ang</th>
</tr>
</thead>
<tbody>
<tr>
<td>1923</td>
<td>Kanto</td>
<td>24°</td>
<td>25°</td>
<td>140°</td>
<td>123°</td>
<td>84°</td>
</tr>
<tr>
<td>1968</td>
<td>Saitama</td>
<td>6°</td>
<td>30°</td>
<td>122°</td>
<td>150°</td>
<td>65°</td>
</tr>
<tr>
<td>1956</td>
<td>Chiba</td>
<td>90°</td>
<td>60°</td>
<td>90°</td>
<td>270°</td>
<td>30°</td>
</tr>
<tr>
<td>1965</td>
<td>Chiba</td>
<td>310°</td>
<td>80°</td>
<td>—118°</td>
<td>58°</td>
<td>30°</td>
</tr>
</tbody>
</table>

Agency (JMA) and the International Seismological Centre (ISC) were used to obtain nodal plane solutions. The solutions obtained (Table 1; Fig. 3) are almost identical to those of ICHIKAWA (1971).

3.2 Hypocentral locations

The 1956 event was followed within one hour by three aftershocks. These events were relocated using the HYPO77 program of T. Maki. The travel time table of ICHIKAWA and MOCHIZUKI (1971) for local earthquakes in and near Japan were employed. Arrival times at JMA stations within approximately 500 km of the epicenter were used, and data with large residuals were excluded from the calculations until stable solutions were obtained. The results are shown in Table 4.

In Fig. 4 the relocated hypocenters are plotted on a vertical plane normal to the strike of both nodal planes of the mechanism solution, together with a vertical section of the fault model derived later in this section, centered on the mainshock hypocenter. Although the uncertainties of the relocations are considerable, the chosen fault plane is clearly preferred over the orthogonal auxiliary

Table 2. Instrument constants of seismographs.

<table>
<thead>
<tr>
<th>Station</th>
<th>Year</th>
<th>Instrument</th>
<th>CPT</th>
<th>V</th>
<th>T</th>
<th>h</th>
<th>l</th>
</tr>
</thead>
<tbody>
<tr>
<td>HGG</td>
<td>1956</td>
<td>Omori</td>
<td>EW*</td>
<td>1.5</td>
<td>120</td>
<td>0</td>
<td>**</td>
</tr>
<tr>
<td>(Hongo)</td>
<td></td>
<td></td>
<td>NS*</td>
<td>1.5</td>
<td>120</td>
<td>0</td>
<td>**</td>
</tr>
<tr>
<td>HGI</td>
<td>1956</td>
<td>Hagiwara</td>
<td>EW*</td>
<td>162</td>
<td>0.44</td>
<td>7.66</td>
<td>17.0</td>
</tr>
<tr>
<td>(Hongo)</td>
<td></td>
<td></td>
<td>NS*</td>
<td>185</td>
<td>0.54</td>
<td>8.66</td>
<td>19.7</td>
</tr>
<tr>
<td>TOK</td>
<td>1956</td>
<td>S</td>
<td>EW</td>
<td>1.0</td>
<td>6.0</td>
<td>0.5</td>
<td>30</td>
</tr>
<tr>
<td>(Tokyo)</td>
<td></td>
<td></td>
<td>NS</td>
<td>1.0</td>
<td>6.0</td>
<td>0.5</td>
<td>30</td>
</tr>
<tr>
<td>TOK</td>
<td>1965</td>
<td>S</td>
<td>EW</td>
<td>10.0</td>
<td>5.0</td>
<td>0.5</td>
<td>15</td>
</tr>
<tr>
<td>(Tokyo)</td>
<td></td>
<td></td>
<td>NS</td>
<td>10.0</td>
<td>5.0</td>
<td>0.5</td>
<td>15</td>
</tr>
<tr>
<td>YOK</td>
<td>1965</td>
<td>S</td>
<td>EW</td>
<td>1.0</td>
<td>6.0</td>
<td>0.55</td>
<td>**</td>
</tr>
<tr>
<td>(Yokohama)</td>
<td></td>
<td></td>
<td>NS</td>
<td>1.0</td>
<td>6.0</td>
<td>0.55</td>
<td>**</td>
</tr>
<tr>
<td>MTJ</td>
<td>1965</td>
<td>Hagiwara</td>
<td>EW</td>
<td>34.2</td>
<td>5.0</td>
<td>0.60</td>
<td>15</td>
</tr>
<tr>
<td>(Tsukuba)</td>
<td></td>
<td></td>
<td>NS</td>
<td>34.4</td>
<td>5.0</td>
<td>0.64</td>
<td>15</td>
</tr>
</tbody>
</table>

* true orientation is 11° counterclockwise, ** no correction made. 

V, static magnification; T, seismometer period; h, damping constant; l, arm length of pen in cm. Station locations are shown in Fig. 2.
Fig. 3. Nodal plane solutions of the 1956 and 1965 Chiba earthquakes.
Earthquake Mechanisms at the Head of the Philippine Sea Plate

Table 3. Assumed shallow crustal model for Tokyo.

<table>
<thead>
<tr>
<th>$\alpha$ (km/sec)</th>
<th>$β$ (km/sec)</th>
<th>$ρ$ (gm/cc)</th>
<th>$h$ (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.83</td>
<td>0.58</td>
<td>2.0</td>
<td>0.25</td>
</tr>
<tr>
<td>1.83</td>
<td>0.70</td>
<td>2.0</td>
<td>0.78</td>
</tr>
<tr>
<td>1.83</td>
<td>1.50</td>
<td>2.0</td>
<td>0.47</td>
</tr>
<tr>
<td>2.80</td>
<td>1.50</td>
<td>2.2</td>
<td>0.80</td>
</tr>
<tr>
<td>5.50</td>
<td>3.00</td>
<td>2.6</td>
<td></td>
</tr>
</tbody>
</table>

$α$, P wave velocity; $β$, S wave velocity; $ρ$, density; $h$, layer thickness.

Table 4. Hypocentral coordinates of the 1956 and 1965 Chiba earthquakes.

<table>
<thead>
<tr>
<th>Event</th>
<th>Year</th>
<th>$M$</th>
<th>Day</th>
<th>HR</th>
<th>Min</th>
<th>Sec</th>
<th>Lat</th>
<th>Long</th>
<th>Depth</th>
<th>$M$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mainshock</td>
<td>1956</td>
<td>9</td>
<td>29</td>
<td>23</td>
<td>20</td>
<td>53.9</td>
<td>35.51</td>
<td>140.23</td>
<td>79</td>
<td>6.5</td>
</tr>
<tr>
<td>Aftershock</td>
<td>1956</td>
<td>9</td>
<td>30</td>
<td>00</td>
<td>03</td>
<td>29.4</td>
<td>35.59</td>
<td>140.19</td>
<td>72</td>
<td></td>
</tr>
<tr>
<td>Aftershock</td>
<td>1956</td>
<td>9</td>
<td>30</td>
<td>00</td>
<td>11</td>
<td>10.5</td>
<td>35.47</td>
<td>140.33</td>
<td>93</td>
<td></td>
</tr>
<tr>
<td>Aftershock</td>
<td>1956</td>
<td>9</td>
<td>30</td>
<td>00</td>
<td>15</td>
<td>45.8</td>
<td>35.58</td>
<td>140.24</td>
<td>77</td>
<td></td>
</tr>
<tr>
<td>Mainshock</td>
<td>1965</td>
<td>5</td>
<td>31</td>
<td>17</td>
<td>38</td>
<td>07.0</td>
<td>35.82</td>
<td>139.94</td>
<td>124</td>
<td>5.5</td>
</tr>
</tbody>
</table>

Fig. 4. Locations of the hypocenters of the 1956 Chiba earthquake and its three aftershocks, plotted on a vertical E–W profile. The fault plane of the earthquake is shown. Bars indicate one standard deviation in the hypocentral locations.

We therefore select the eastward dipping plane as the fault plane of the 1956 earthquake. This has important consequences which we will discuss in the tectonic analysis. No aftershocks of the 1965 earthquake were reported in the Seismological Bulletin of JMA, so the identity of the fault plane of that earthquake remains ambiguous.
3.3 Data analysis

Low-gain displacement S-wave seismograms recorded in the Kanto district were used for analysis. All of the recordings used were sufficiently close to the epicenters that the upgoing waves impinged subcritically on the Moho.

For the 1956 event, only three recordings in Tokyo were available, but the seismographs have different responses and together provide a broad-band data set. For the 1965 event, only 6 sec period seismograms are available, but the recordings at three stations (Tsukuba, Tokyo and Yokohama) provide a geometrically well distributed data set. The instrumental constants are shown in Table 2, and locations of seismographic stations are shown in Fig. 2.

The original seismograms were digitised using the Tokyo Computer Control instrument of the Earthquake Prediction Observation Center, E.R.I., by sampling extrema and inflection points. Corrections were then applied for the departure of the pen from purely transverse motion, and for the finite arm length of the pen. The resulting time function was interpolated using the weighted average slope interpolator of WIGGINS (1976), and then resampled at a rate of 25 samples per second. A least-squares line was then removed from the time series.

3.4 Synthetic seismograms

SAVAGE's (1966) model of radially spreading rupture on a plane fault surface was chosen as the source model since it is probably a realistic description of the evolution of rupture on small faults that do not grow so large as to intersect the free surface or a major discontinuity in stress. It was considered that there were not enough recordings to resolve the shape of the fault boundary, so a circular shape was assumed.

Crustal structure causes a considerable distortion of the impinging waveform at periods of the order on one second. Earthquakes of moderate magnitude have a maximum ground velocity in this period range, and so some means of taking the crustal structure into account must be devised if we wish to study the earthquake source mechanism.

The shallow crustal structure beneath Tokyo is well known from wells (GOTO et al., 1978) and seismic refraction profiles (SHIMA et al., 1976), and the deeper structure is also known (e.g. MIKUMO, 1966). Thus we can include the effect of the layered crust in the calculation of the synthetic seismograms. The Thompson-Haskell matrix formulation for plane waves impinging at a subcritical angle at the base of a horizontally layered crust was employed. All multiple reflections and conversions are included in the calculations. According to the assumed crustal and upper mantle structure, all of the recording stations are at sufficiently short epicentral distances that the rays impinge subcritically at the base of the crust. The foci are sufficiently deep that the plane wave assumption may be an acceptable approximation.

The shallow crustal model chosen for Tokyo and Yokohama is that obtained
Earthquake Mechanisms at the Head of the Philippine Sea Plate

for Yumenoshima (in Tokyo Bay) from the third Yumenoshima and Yoshikawa explosions by Shima et al. (1976). In that study, separate P and S velocity profiles were obtained which have different layered structures. These were combined (Table 3), preserving all of the interfaces of both models with the exception that the depth of the bases of both models were constrained to coincide with that of the P velocity profile (this interface is roughly 0.23 km deeper in the S velocity profile). This model is expected to be a good representation of the structure beneath the stations in Tokyo since they are quite close to Yumenoshima. The overall thickness may be somewhat different beneath Yokohama but the structure is unlikely to be grossly different. Computational tests using this shallow structure and the crustal model E-3A3 of Mikumo (1966) showed that the shallow structure completely dominates the amplitudes of the synthetic seismograms, especially in the first few seconds. For this reason, synthetic seismograms were calculated for waves impinging on the shallow structure only. Only far field terms need be included in the calculation of the S waves, since the recording stations are sufficiently distant from the source (roughly twenty wavelengths) for this to be a reasonable approximation.

The use of S waves enables better resolution of the rupture velocity than P waves allow because the variation of waveform with zenith angle (measured from the fault plane) increases rapidly as the rupture velocity approaches the wave speed.

When the locations of aftershocks are unavailable and the fault is sufficiently buried that no surface deformation is observable, there exists no method for resolving the fault plane of the mechanism solution other than by the analysis of waveforms. Because of the finite speed of seismic waves, the apparent evolution of rupture varies with the orientation of the observer, producing different waveforms at different stations. In this study, we attempted to resolve the fault plane from the two nodal planes by waveforms analysis, although aftershock data for the 1956 event identifies the fault plane with reasonable confidence.

Figure 5 shows the comparison between observed waveforms and calculated waveforms for fault surfaces on the two nodal planes which fit the observed waveforms best. In the case of the 1956 event, comparison is made with the E-W components only since they are much larger in amplitude than the N-S components. The N-S components are nodal and therefore cannot be used to obtain a reliable estimate of source characteristics. The fault plane (chosen on the basis of aftershock locations) is preferred because the width of the pulse recorded on the Omori seismograph is correctly modelled. In order to match this width with the auxiliary plane model, the waveforms at the other two stations become too wide. However, the difference between the waveforms of the two models is very slight and does not allow a conclusive resolution of the fault plane.

A similar situation exists in the case of the 1965 event. The ultimately chosen fault plane model agrees better with the observations in the backswing following the first motion, but the difference between the two models is slight.
Fig. 5. Comparison of observed and synthetic seismograms computed from Savage's fault model assuming rupture on the fault plane and on the auxiliary plane. A ramp time function is assumed. Synthetic seismograms are scaled to match the peak amplitude of the larger observed component.
Table 5. Fault parameters of the 1956 and 1965 Chiba earthquakes.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>1956 event</th>
<th>1965 event</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fault radius value</td>
<td>mean min max</td>
<td>mean min max</td>
</tr>
<tr>
<td>Radius (km)</td>
<td>4.0 3.5 4.5</td>
<td>2.5 2.0 3.0</td>
</tr>
<tr>
<td>Rupture velocity (km/sec)</td>
<td>4.0 3.5 4.5</td>
<td>3.5 3.0 4.5</td>
</tr>
<tr>
<td>Rise time (sec)</td>
<td>0.8 0.7 0.9</td>
<td>0.6 0.5 0.7</td>
</tr>
<tr>
<td>Dislocation (cm)</td>
<td>107 136 86</td>
<td>30 44 21</td>
</tr>
<tr>
<td>Moment ($\times 10^{26}$ dyne·cm)</td>
<td>3.6 3.5 3.7</td>
<td>0.39 0.37 0.39</td>
</tr>
<tr>
<td>Stress drop (bars)</td>
<td>240 350 175</td>
<td>110 200 60</td>
</tr>
<tr>
<td>Effective stress (bars)</td>
<td>210 330 140</td>
<td>85 165 45</td>
</tr>
<tr>
<td>Dislocation velocity (cm/sec)</td>
<td>135 200 100</td>
<td>50 90 30</td>
</tr>
</tbody>
</table>

Since station MTJ is on bedrock, the synthetic seismograms were made assuming a half-space having the seismic velocities of the half-space in Table 3. It can be seen that in both the observed and calculated seismograms, the shallow crustal layering produces a significant effect, since the waveforms at TOK are more complex than the waveforms at MTJ.

3.5 Fault parameters

The fault parameters of the best-fit models of the two earthquakes, which were obtained from the synthetic seismogram study, are summarized in Table 5. The range of uncertainty of the parameters is indicated by the values obtained for the parameters for the cases of the largest and smallest allowable fault radius. The seismic moment of the 1965 event was calculated from the 40 sec G wave at BAG, assuming a point source at a depth of 80 km in earth model 8–3–2 of SAITO and TAKEUCHI (1966), and the attenuation coefficients of MITCHELL et al. (1976). Details of the calculation are similar to those of SHIMAZAKI and SOMERVILLE (1979). The value of $1.6 \times 10^{24}$ dyne·cm which was obtained is approximately half the value obtained from seismograms at near distances, indicating that the latter value is not grossly in error. Since the 1956 earthquake occurred before the widespread recording of long period seismograms, the Omori seismogram at Hongo, Tokyo was used to the constrain the seismic moment of this event. These moment determinations allowed the stress drops of the two events to be estimated.

The stress drops of both events are quite high. The stress drop of several hundred bars for the 1956 event is comparable to that of the Saitama earthquake of 1968 at a depth of 52 km (Abe, 1975; Abe's value should be roughly doubled since a shear modulus of roughly $7 \times 10^{11}$ dyne·cm², appropriate to the mantle, should be used in place of the value of $3.4 \times 10^{11}$ dyne/cm² that he used).

4. Tectonic Setting of the Two Earthquakes

We now wish to examine the tectonic setting in which the two earthquakes occurred, and relate the focal mechanisms of the two earthquakes to this tectonic
First we identify the location of the upper surface of the Philippine Sea plate beneath Chiba. For this purpose we use seismicity data, and also the seismologically and geodetically determined fault-plane of the 1923 Kanto earthquake which we assume to represent the upper surface of the Philippine Sea plate.

Fault model II of Matsu'ura et al. (1980) is chosen to represent the fault plane of the 1923 earthquake. This model was derived from the corrected triangulation data of Sato and Ichihara (1971). The orientation of the fault plane and the slip vector of Model II are similar to those derived from first motion data by Kanamori (1971), the largest differences lying in the shallower dip angle and the larger dip-slip component of Model II.

One disadvantage of the fault plane of Model II is that it was not constrained by the vertical deformation accompanying the earthquake. The vertical deformation calculated from Model II differs significantly from the observed deformation in the Shonan area.

The fault strike of Model II, N114°E, is roughly 20° from the strike of Ando's model (N135°E) which was chosen to follow the trend of the Sagami trough in Sagami Bay. However, a line drawn from Kozu (where the Sagami trough intersects the coastline) to the triple junction between the Japan trench, the Sagami trough and the Izu-Bonin trench has a strike which is almost parallel to that of Model II (Fig. 1). The Sagami trough deviates considerably from this line in Sagami Bay and along the Kamogawa submarine cliff off the southeast shore of the Boso Peninsula. However, we propose that this deviation from linearity is a superficial feature. Kasahara et al. (1973) pointed out the probable presence of an active fault (the Kamogawa thrust) along the steep submarine slope off the SE coast of the Boso Peninsula. The steep gradient of the uplift of the southern tip of the Boso Peninsula during the 1703 earthquake (Matsuda et al., 1978) suggests that slip on this thrust fault is restricted to shallow depths, and may not extend as deep as the model proposed by Matsuda et al. In summary, we suggest that the fault plane of Model II, while not modelling accurately the irregular shallow part of the plate interface, is nevertheless an average representation of the orientation of the interface at depth.

The existence of a fault plane with an eastward component of dip beneath the Kanto region was noted by Tsumura (1973) in an EW profile extending 50 km south from Latitude 36°N. In this profile, and also in the EW profile of Maki et al. (1980) over a similar width centered on Latitude 35°30′N, there is an apparently eastward dipping zone whose upper surface intersects the surface at Longitude 139°E, and the westward dipping Pacific plate at Longitude 140°E.

This apparently eastward dipping zone is not evident on the more northerly EW profiles centered on Latitude 36°N (Maki et al., 1979) and extending 50 km north from Latitude 36°N (Tsumura, 1973) except for a concentration of seismicity deeper than 50 km. This suggests that the true dip of the zone is not due east but has a northerly component. This is in accord with the roughly NNE-dipping
Fig. 6. Seismicity profile in the NE-SW direction (profile BB' in Fig. 2) for 1971–1977. The inferred upper surface of the Philippine Sea plate is shown by a dashed line. The hypocenters of the 1956 and 1965 Chiba earthquakes are shown (after MAKI et al., 1980).

The orientation of the 1923 earthquake fault plane deduced from seismological and triangulation data.

The SW-NE profile of MAKI et al. (1980) (Fig. 6) is almost parallel to the dip direction we have assumed for the fault plane of the 1923 earthquake. The upper surface of this plane intersects the surface in the vicinity of the Sagami trough. Further, the apparent dip of this zone is $22^\circ$, which agrees well with the dip of $25^\circ$ of Model II. This suggests that the upper surface of the seismic zone corresponds to the interface between the Eurasian and Philippine Sea plates. The profile includes hypocenters within 25 km of the plane of the profile. Inspection of Fig. 2 indicates that the nest at the center of the profile is the nest underneath Mount Tsukuba, not the nest underneath Chiba.

Epicenters in the depth range 40–80 km (MAKI et al., 1980) are shown in Fig. 2, on which the fault plane of Model II and its down dip extension to a depth of 50 km are superposed. The NW corner of this extended fault plane coincides with the hypocenter of the 1968 Saitama earthquake (ABE, 1975). The orientation of the fault plane of the Saitama earthquake on Table 1 is so similar to that of the 1923 Kanto earthquake that we may consider them to be an extension of the same fault plane. From the absence of seismicity north of the Saitama earthquake, we assume that the head of the plate is in the vicinity of the hypocenter of the Saitama earthquake.

The NE corner of the extended fault plane is at the top of the intersection
of the two planes at Longitude 140°, and lies above the nest of earthquakes beneath Chiba city. This suggests that the nest reflects the collision between the Philippine Sea plate and the Pacific plate. The details of this collision, and its possible eastward continuation along the Sagami trough, are beyond the scope of this study.

The selection of the eastward-dipping fault plane of the 1956 event on the basis of the aftershock distribution has important consequences in the interpretation of the tectonics of the nest, because it rules out the possibility that the fault plane was the upper surface of the westward-dipping Pacific plate. Similar mechanisms are commonly observed oceanward of the aseismic front (Yoshii, 1975), with focal depths not exceeding 60 km, and are usually interpreted as occurring at the interface of the plates. However the focal depth of the 1956 event is deeper than 60 km and lies within, not at the upper surface, of the seismic zone, thus discounting the possibility that it is an interface event.

It is of interest to determine whether the distortion of the stress field in the collision zone extends as far as the lower surface of the Pacific plate. The mechanism of the 1965 event, indicating down-dip tension, is similar to mechanisms of events in the lower seismic zone off Tohoku (Yoshii, 1979; Hasegawa and Umino, 1978). Since there is no other mechanism solution available in the vicinity of the nest, a study was made of the mechanism solutions determined by Ichioka (1971) for events located in the lower seismic zone in the region 34.75°-36.25°N in Latitude, 139.25°-140.75°E in Longitude. Among the seven events, five have normal-faulting mechanisms, of which three have down dip tensional axes. It thus seems that the 1965 event in the lower seismic zone beneath the Chiba nest shows the focal mechanism commonly found in the lower zone and is not anomalous. We conclude that although the tectonic field is strongly disturbed in the nest, it is not seriously disturbed at the lower seismic zone of the Pacific plate.

5. Conclusions

The mechanism of the 1956 Chiba earthquake in the dense nest of seismicity at a depth of between 70 and 80 km beneath Chiba city appears to represent internal deformation in the collision zone between the Pacific and Philippine Sea plates. The stress field at the lower surface of the Pacific plate, as inferred from the mechanism of the 1965 earthquake, appears to be undisturbed by the collision.

The 1968 Saitama earthquake may be regarded as an interplate event occurring on the upper surface of and near the tip of the Philippine Sea plate. The high stress drop of several hundred bars of this earthquake and of the 1956 earthquake in the Chiba nest may reflect the penetration of the tip of the Philippine Sea plate into the Eurasian and Pacific plates respectively.
The author wishes to thank Drs. T. Maki and K. Kudo for their assistance and advice throughout this study, and Dr. T. Iwata for his help in obtaining seismograms. This research was done while the author was in receipt of Research Fellowships from the Japan Society for the Promotion of Science, and subsequently from the Kajima Foundation.

REFERENCES


Somerville, P., The accommodation of plate collision by deformation in the Izu block, Japan,


Yoshii, T., A detailed cross-section of the deep seismic zone beneath northeastern Honshu, Japan, Tectonophysics, 55, 349–360, 1979.