Taiwan represents a very young arc-continental margin collision zone in a long subduction boundary. The collision started in Late Pliocene and is still vigorously taking place. Off coast of NE Taiwan a northward subducting slab, extending west at depth to northern Taiwan, can clearly be defined; although most parts of Taiwan have been rising steadily at about 5 mm/year for the last 8,500 years, northern Taiwan has had periods with no uplift. The intensity of the collision decreases toward the south off the island, and an east-dipping subduction zone can be delineated there. Thus Taiwan can be viewed as a transform zone in between two subduction zones with quite different geometries.

Seismically Taiwan is much more active than its neighbors, the Ryukyus and Luzon. Large earthquakes reveal the nature of the intense on-going intra-plate deformation; on land, EW compression or left-lateral shear occur along NNE faults and right-lateral shear occur along nearly EW faults; offshore to the southeast of Taiwan, left-lateral shear along NWW or NW faults and thrusts in several directions coexist; to the northeast, the focal mechanisms agree well with other subduction zones. The Ryukyus are terminated at about 123°E by a number of NNE striking right-lateral faults. Focal mechanisms to the southeast of Taiwan are consistent with a tectonic stress direction of S46°E to S76°E and plunging at −2° to 15°.

1. Introduction

The tectonics of Taiwan presents several points of interest. First, Taiwan is an anomalous member of the Ryukyu-Taiwan-Luzon-Philippine arc chain; whereas Ryukyu, Luzon and Philippines can be interpreted in terms of subduction of lithosphere albeit with complications in the case of Luzon-Philippines (Fitch, 1972), Taiwan represents a hiatus in a long subduction boundary; no systematic active andesitic volcanism along the trend of the Island and no recognizable bathymetric features exist in the vicinity of the Island that are usually associated with island arcs. Although a plate boundary here can be clearly defined in terms of a transition from oceanic crust to continent-like crust and other geological features indicating the past activities along this boundary, this boundary is not the only active feature and current intense shallow crustal deformation in this area extends to both sides of this boundary for more than 100 km. Secondly, the orogeny that produced a rather impressive mountain range (attaining a maximum height of 4,000 m), started sometime in Late Pliocene (Chai, 1972; Chou, 1973), increased its intensity throughout Pleistocene and is probably in the most active stage at present (Li, 1976); the incipience of the anomalous situation described above and the emergence of Taiwan were apparently concomitant and the present day tectonics is related closely to the orogeny that produced the Island.

The atypical nature of Taiwan as a member of an island arc chain has been recognized by many geologists. Biq noticed the apparently reversed convexity, the presence of nega-
tive Bouguer anomaly over the Island and positive anomaly in the eastern part of the Island (Biq, 1960, 1971) and concluded that Taiwan had been a continent-facing arc from Miocene to Pliocene and has undergone a transition to block tectonics since Pleistocene (Biq, 1971) but it may not yet have reached a steady state (Biq, 1972). On the other hand, Juan and Wang (1971) and Juan (1975) considered Taiwan a coastal range of the mainland Asia originally and had subsequently accreted ocean-ward by successively retreating subduction. While Biq (1964) hypothesized an east-dipping subduction zone under the Coastal Range, Juan (1975) proposed the rotation of a west-dipping subduction zone into a vertical zone in Early Pliocene. Chai (1972) was among the first to propose that the Coastal Range represents a former west-facing arc, i.e., with an east-dipping Benioff zone, and the westward migration of the trench line (with respect to the Asian continent) led eventually to the collision of the arc with the miogeosyncline on the margin of the Asian continent, whence the formation of the Island. Karig (1973) considered Ryukyu, the Philippines together with Taiwan and reached similar conclusions.

Due to a paucity of data, the polarity of the Bashi-Northern Luzon arc has not been ascertained (Katsumata and Sykes, 1969; Ludwig, 1970); but a short nascent subduction, perhaps west-dipping, may exist on the east side of Luzon (Fitch, 1972; Bowin et al., 1978). With more seismic data since 1968, we can now be quite sure that the Bashi-Northern Luzon arc is indeed east-dipping as suspected. On the other hand, the west-dipping Ryukyu arc has always been recognized (Katsumata and Sykes, 1969 among others). Thus between these two oppositely directed arcs Taiwan can, in a broad sense, be viewed as a transform zone. At present, this transform is accomplished by movements along definable boundaries and a substantial amount of intra-plate deformation under the ocean floor and on land.

Recently new data relevant to the interpretation of the tectonics of Taiwan have been accumulated rapidly. A newly established seismic network of 20 telemetered stations has been providing good location data for events down to magnitude 2 on the island and in the vicinity; a few detailed microearthquake surveys are now available mapping out active faults; and using local and teleseismic network data crustal and mantle structure in the vicinity of Taiwan have been determined (Lu, 1976). In addition, a new geological map has recently been published (Ho, 1974) and more geological data have been analyzed (see for example Juan, 1975 for a partial bibliography). Also of great importance is the acquisition of marine geological data in the seas around Taiwan (Bowin et al., 1978; R.S. Lu, personal communication, 1978).

In the present paper, we shall attempt to analyze the pattern of crustal movements on Taiwan and in its vicinity based on seismological as well as geological data, and from there we hope to understand the tectonics since Late Pliocene, when the Island as we now know came into being, by extrapolation. It should be remarked here that in most of the works cited, a two-dimensional interpretation of the tectonics was made, however there are clear variations in the north-south direction and they should be explained.

2. Geology

Most of the surface geological data have recently been compiled into a new geologic map for the Island (Ho, 1974); its explanatory text provides a comprehensive survey of the geological literature on Taiwan (Ho, 1975). In Fig. 1, we have presented a simplified version of it. Several detailed descriptions of the geology of Taiwan written pre-
Fig. 1. Simple geologic map of Taiwan. The numbers in the figure refer to the following rock units: 1+2+3, Late Paleozoic to Mesozoic metamorphics; 4, Eocene; 5, Eocene to Oligocene; 6+7, Oligocene to Miocene; 8, Miocene; 9+10, Early Miocene; 11, Middle Miocene; 12, Late Miocene; 13, Late Miocene to Pliocene; 14, Pliocene; 15 and 16, Pliocene and Pleistocene; 17, Pleistocene; blank, Holocene Alluvium. Major cities and off-shore islands are named.

vious to the publication of this map are still valid to a large extent (Ho, 1967; CHAI, 1972; among others). In this section, we shall merely outline aspects of geology of Taiwan that will be relevant to our discussions later.

Physiographically, four provinces can be distinguished on Taiwan. These are, from
Fig. 2. Physiographical provinces of Taiwan. The line indicates the position of the profile in Fig. 4.

Fig. 3. LANDSAT 1 band 5 image. Notice the prominent Longitudinal Valley of Taiwan and several N-S faults. Compare this figure to Fig. 15.
Recent Tectonics of Taiwan

East to west, the Coastal Range, the Central Range, the Foothills and the Coastal Plain (Fig. 2). The division of the provinces can be seen clearly on the LANDSAT (ERTS) image of Taiwan (Fig. 3).

The Coastal and the Central Ranges are separated by a striking feature, the Longitudinal Valley (Figs. 1, 2 and 3). The floor of the Valley is about 200 m above the sea level and has a width of 5 to 7 km. On the eastern side of this Valley the Coastal Range trends about N20°E and attains a maximum height of about 1,600 m. On the western side, the Central Range is parallel to the Coastal Range in the central and southern parts; toward the north however, it gradually turns to assume a northeastern trend. The ascent from the sea level to the respectable height of 4,000 m in the Central Range is achieved in a distance of not more than 35 km. The descent toward the west from the ridgeline is much slower; a belt of relatively low hills, the Foothills, separates the high mountains from the western Coastal Plains.

The oldest rocks crop out on the eastern side of the Central Mountains (Fig. 1): they are strongly deformed and metamorphosed and are probably Late Paleozoic to Mesozoic in age, based on a few Permian fossils. The rocks include graphite schists, quartz schists, marble, injection gneiss and paragneiss. Age dates are not yet available for most of these rocks, and, as a consequence, the Pre-Tertiary history is not clear. Some geologists believe that these rocks were metamorphosed during a Mesozoic orogeny (BIQ, 1971).

A sequence of slates, phylites, quartzite, locally with irregular bodies of graphite and carbonaceous shale form the record of Eocene and Oligocene; these rocks crop out either directly to the west of the Late Paleozoic-Mesozoic metamorphic sequence or to the west of the metamorphosed Miocene strata (Fig. 1).

The Miocene rocks on the western side of the Longitudinal Valley appear very extensively in the eastern Foothills, and also in thrust blocks on the western side of the Central Mountains (Fig. 1). Miocene rocks in the Central Ranges are generally metamorphosed and they include black schists, phylites and slightly metamorphosed shales. The Miocene rocks in the Foothills and in the thrust blocks have been studied in greater details. In general they consist of alternating sandstone and shale beds; in the northern part of the Island, they are quite often coal bearing but no coal seams are found to the south of an east-west line going through Peikang (Fig. 1), near the crest of the so-called “Peikang Basement High” (MENG, 1967), and shale beds dominate in the southern part of the Island.

Pliocene rocks in the western Foothills are composed of shale, siltstone, and thin lithic greywacke in the lower part and fine to coarse grained lithic greywacke with shale in the upper part. The upper part is deemed a flysch type of deposit and the commencement of deposition of this layer is considered the beginning of the present phase of the Taiwan Orogeny (BIQ, 1966; see later discussion).

Pleistocene rocks in the western part of the Island consist of two distinct facies; Early Pleistocene is represented by alternating lithic greywacke and shale, much like that of the upper formation of the Pliocene sequence, and the upper unit of the Pleistocene sequence is a thick conglomerate unit consisting mainly of boulders of Paleocene quartzite and other rocks, representing post or syn-tectonic Molasse.

East of the Longitudinal Valley the oldest rocks exposed belong to Miocene. Early Miocene andesitic agglomerates are widely distributed. They tend to form mountain ridges. Mid-Miocene and Early Pliocene rocks consist of a thick sequence (~3,000 m) of shale, siltstone, ill-sorted sandstones, and conglomerates. Late Pliocene rocks in the southern Coastal Range appear as a mélangé composed of serpentinite, basalts, andesites,
agglomerate, limestone and sandstone (Ho, 1978). Rocks of similar nature also appear in the southern tip of the Island; although the rocks there are not as well understood as the Coastal mélangé. A series of conglomerates forms the upper Pleistocene and Holocene outcrops.

As we have mentioned above that in the Coastal Range there are Miocene (K/Ar age of 17–22 my) andesites. These are the only extensive Pre-Pleistocene volcanics on the Island. The two islands to the southeast of Taiwan, Lanyu and Lutao are also composed of Miocene and Pliocene andesites (Ho, 1975). Judging from magnetic- and gravity-anomaly trends in the ocean (Lu et al., 1977; Bowin et al., 1978), these islands may link up with the andesites in the Coastal Range. In the Miocene strata of northwestern Taiwan, limited basaltic rocks and tuffs are present: the basalts occur as fissure flow or sills. Large exotic blocks of ultrabasics, serpentinites and basalts are found in the Coastal Range mélangé. The age of these rocks is not clear.

A series of basaltic, andesitic, and dacitic extrusions forms a group of volcanoes, agglomerates and small intrusions in northern Taiwan (Fig. 1). Although no age dates for them are yet available, based on the inter-fingering relation with the Pleistocene conglomerates (Ho, 1968), and the uniform normal polarization of the paleomagnetic poles (Hsu et al., 1966) we can give it a maximum age of 800,000 years.

It is generally agreed that the Coastal Range represents a compressed former island arc system that existed since early Miocene (Chai, 1972). During Miocene, the area west of the arc was probably a sedimentary basin not unlike the present continental shelf of Asia. The sense of subduction has been a subject of controversy; Jahn (1972) postulated an east facing arc, but later authors argued strongly for a west facing arc starting from Early Miocene (Chai, 1972; Karig, 1973). Because of the continued counterclockwise rotation of the Philippine Sea Plate, with respect to the Eurasian Plate, the trench moved westward relatively to the Asian mainland. During Miocene, as the island arc approached the continental margin, the crust near the continental margin was progressively depressed. Since the trench was apparently not parallel to the continental margin, the northern part of the depositional basin was on the continental shelf (north and northeast of the present Peikang Basement High) and the southern part was in the ocean basin; due to the thinner crust of the ocean basin, the depression was more pronounced. Such deposition and deepening continued—perhaps more vigorously—through Early Pliocene. The sources for the sediments during Miocene and Early Pliocene are mainly from the west-northwest, namely, in the area presently occupied by the Taiwan Strait (Chou, 1973).

The cessation of subduction in the vicinity of Taiwan and the collision of the island arc with the continental shelf occurred most probably in Late Pliocene. At that time, the source of sediments in the Taiwan depositional basin (both the southern and northern parts) switched from a western one to an eastern one signalling the rise of the Central Mountains. The deposition of the Pleistocene conglomerate probably represent the peak of the orogeny (Wu and Lu, 1976), although as we shall see later, the tectonics in and around Taiwan is still at a very active stage.

3. **Crust and Upper Mantle Structures**

The recently established telemetered seismographic network, the microearthquake surveys in various areas (e.g., Tsai and Liu, 1977) and gravity surveys in the Plains as well as along newly opened cross-island highways all provide data that allow us to determine
the crustal structure under and upper mantle structure around Taiwan. The present results were derived from data up to mid-1975, incorporating shallow (<7 km) reflection and refraction as well as deep well data; the details of data source and methods used can be found in Lu (1976).

Figure 4 shows the crustal structure based on gravity and seismic data. The rectangular boxes indicate seismically determined horizons. The line of profile is shown in Fig. 2 including the Peikang Basement High in western Taiwan and the middle section of the Coastal Range. The most prominent feature in this interpretation is the discontinuity
<table>
<thead>
<tr>
<th>Year-Month-Date</th>
<th>Locations affected</th>
<th>Nature of damage</th>
<th>Int.</th>
<th>Mag.</th>
<th>Remarks</th>
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<tbody>
<tr>
<td>1644-7-30</td>
<td>S. Taiwan</td>
<td>Tsunami in Anping Citadel wall damaged</td>
<td>IX-X</td>
<td>7</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Liquefaction of lowlands</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1654-12-14</td>
<td>Anping*</td>
<td>Felt aftershocks lasted for 7 weeks</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1655</td>
<td>Anping</td>
<td>Three weeks of aftershocks Liquefaction</td>
<td>VIII</td>
<td>6.5</td>
<td></td>
</tr>
<tr>
<td>1661-2-15</td>
<td>Tainan Yuchin Sanhua</td>
<td>Citadel wall cracked Many houses collapsed Ships in harbor rocked Wave breaks like clouds (Tsunami) Aftershocks lasted for 6 weeks Ground fissures</td>
<td>IX-X</td>
<td>7</td>
<td></td>
</tr>
<tr>
<td>1686-5-12</td>
<td>Chiayi Tainan Fengshan</td>
<td>Houses destroyed and damaged</td>
<td>VII-VIII</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>1694-4-24 5.23 (unsure)</td>
<td>N. Taiwan</td>
<td>Part of the Taipei basin subsided</td>
<td>IX</td>
<td>6.5</td>
<td></td>
</tr>
<tr>
<td>1715-10-11</td>
<td>Chiayi Fengshan Tainan</td>
<td>Houses collapsed</td>
<td>VII</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>1720-10-31</td>
<td>Tainan Chiayi Fengshan</td>
<td>Houses destroyed Surface faulting Liquefaction Aftershocks lasted for 10 days</td>
<td>IX-X</td>
<td>6.5-7</td>
<td></td>
</tr>
<tr>
<td>1721-01-05</td>
<td>Chiayi Fengshan</td>
<td>Twin shocks Destroyed many houses</td>
<td>VIII</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>1736-1-30</td>
<td>Changhua Chiayi Tainan Fengshan</td>
<td>Meizoso fault activated Foreshock on prev. day Seiches Extensive liquefaction Intense shaking, houses destroyed. Followed by fire Ground cracks Buildings destroyed</td>
<td>X-XI</td>
<td>7.5</td>
<td></td>
</tr>
<tr>
<td>1792-8-7</td>
<td>Tainan Chiayi Changhua Fengshan Tanshui</td>
<td>Liquefaction</td>
<td>VIII</td>
<td>6.5</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Houses destroyed</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>1711-10-22</td>
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<td>Liquefaction</td>
<td>XII-XI</td>
<td>6</td>
<td></td>
</tr>
<tr>
<td>1815-7-11</td>
<td>Ilan Tanshui Taipei Miaoli</td>
<td>Foreshock on prev. day</td>
<td>VIII</td>
<td>6.5</td>
<td></td>
</tr>
<tr>
<td>1839-6-27</td>
<td>Chiayi Tainan Fengshan</td>
<td>12 foreshocks felt in Chiayi Chiayi completely destroyed Limited in area</td>
<td>IX</td>
<td>6.5</td>
<td></td>
</tr>
</tbody>
</table>
beneath the Longitudinal Valley and the Coastal Range. The velocities under the Coastal Range do not correspond to any particular known type of crusts. The top layer probably represents a sedimentary (Mio-Pliocene) layer; the second layer velocity is within the range of that for a Pacific “transition layer” (SHOR et al., 1970). But the third layer in the “average” Pacific structure with velocity of about 6.8 km/sec is missing. This layer may be actually absent, but it is likely that either the refraction line is not well covered enough to detect that layer or the layer is fractured badly so that no distinctive discontinuity in velocity can be found.

The velocity structure under the main part of the Island is very similar to that of a continental crust; the mantle velocity of 7.75 km/sec is typical of a tectonically active region such as Basin and Range Province of western U.S. (HEALY and WARREN, 1968). Such value is also prevalent for the mantle under Japan (JAPANESE RESEARCH GROUP FOR EXPLOSION SEISMOLOGY, 1978).

Figure 5 shows a preliminary upper mantle structure in the area around Taiwan by inverting a $dT/d\delta$ curve derived from the telemetered network data with earthquakes located in Japan and Ryukyu to the northeast and Luzon and Philippines to the south. The details of data and methods used can be found in LU (1976). The upper part of the structure is fairly similar to that of KANAMORI (1967), but the “400 km” and “600 km” discontinuities are somewhat sharper.

4. Seismicity

There are several different sources of Taiwan seismicity data and depending on the nature of the data, they can be used in assessing the levels of activity for different regions, the activeness of a fault or mapping faults.

In Fig. 6(a) and (b), we have shown the damage areas of large historical earthquakes and the epicenters of large ($M>6$) earthquakes since 1900 (LEE et al., 1978) respectively. For the historical earthquakes, we determined the approximate intensities and magnitudes based on a collection of descriptions of the earthquakes in various ancient documents.
Fig. 7(a). Seismicity profiles parallel to the Philippines, Luzon, Taiwan and the Ryukyus (world-wide data from 1963 to 1974).
Fig. 7(b). Seismicity profiles perpendicular to the trenches or the main structural trend (in case of Taiwan).
Fig. 7(c). Definition of the regions; lines indicate the position of 0 ("trench") in 7(b) and direction of profiles each region.
In Table 1, we have listed the earthquakes parameters, their associated phenomena and the cities affected. The concentration of epicenters in western Taiwan is obviously related to the distribution of Han population; the aborigines who lived in the high mountain ranges in eastern Taiwan did not keep written records. It is likely that some of the epicenters should actually be located in the foothills. This is made evident in Fig. 6(b); here eastern Taiwan is seen to be more active. In Fig. 6(b), northern Taiwan is shown to be relatively inactive but a few earthquakes occurred in historical time; the large damage area and relatively light damage imply that the depths of these events are fairly large (perhaps 100–200 km). In Fig. 6 and in all subsequent seismicity figures, we have also plotted what we regard as active faults; these are either earthquake faults or geomorphically very prominent linear structures and geologically mapped faults. We shall refer to them in a later section, together with the possible significance of the historical seismicity.

To elucidate the nature of plate activity in the vicinity of Taiwan, we use the 1962–1974 hypocenters as contained in the Earthquake Data File (available from NOAA, U.S. Department of Commerce in Boulder, Colorado, USA).

In Fig. 7(a), we have plotted profiles parallel to Taiwan and the neighboring arcs and in Fig. 7(b) seismicity in profiles perpendicular to the island arcs are plotted. The regional definitions and the orientation of profiles for Fig. 7(b) are shown in Fig. 7(c). Toward the south end of the Philippines, the maximum depth of foci reaches 650 km; however, it becomes shallower going north, there are some gaps, but they may only be apparent ones resulting from a short time window for the data. The decrease in depth toward the north is however real and may be related to the change of rate of subduction of the Philippine Sea Plate (Wu, 1972). In the perpendicular profile, because of the superposition of the Sulu and the Philippines arc that have their Benioff zones dipping different ly, the thickness of the seismic zone is rather larger than usual. In Luzon, the maximum depth of foci is about 250 km and the profile perpendicular to the Manila Trench shows in general a deepening of foci from the Manila Trench area toward the east. This is consistent with normal fault mechanisms associated with earthquakes along the Manila Trench (Seno and Kurita, 1978). Therefore, going from the Philippines to Luzon we have a change of polarity of subduction; this point has been observed by Fitch (1972) and Katsumata and Sykes (1969) and others. There is a section of trench on the eastern side of Luzon, where thrust faulting has been found to occur. Fitch (1972) and others have hypothesized that a nascent west-dipping subduction has started there.

Going further north toward Taiwan, the foci becomes shallower and underneath Taiwan, between the latitudes of 23°N and 24.2°N, along the whole length of the Coastal Range, there are only earthquakes shallower than 70 km; deepening of foci occur at both ends of this section. This phenomenon can also be seen from Fig. 8, a plot of seismicity with the sizes of the symbols proportional to the depths rather than the magnitude, as is commonly done. It is also clear that shallow seismicity along the Manila Trench terminates at about 21°N, a large earthquake with normal fault mechanism occurred there in 1972; between that point and Pingtung Plain (Fig. 1) there is evidently a gap. Recent local Taiwan network data show that small earthquakes do exist south of the Pingtung Plain (Y.B. Tsai, personal communication, 1978); but whether they do continue to the Manila Trench is not yet known. Northeast of Taiwan the foci deepen toward the north and merge into the earthquakes easily identified with the Ryukyu island-arc. The area
Fig. 8. Seismicity along the western margin of the Philippine Sea plate in map view.
Fig. 9(c)–(d). Quarterly seismology map of Taiwan (1975) based on a local network of 70 stations. Note that there are two symbols with each point. The circle denotes the depth, the larger the deeper, and the size of the cross is proportional to magnitude. Here it is clear that deeper earthquakes can be found in northern Taiwan and also northeast of the island.
northeast of Taiwan is noticeably more active than its nearby regions (Fig. 8), but basically it shows a north-dipping Benioff zone. It is sometimes asserted that, at present, the Benioff zone under Taiwan dips to the east or becomes vertical (Lee, 1962; Juan, 1975; Bird and Dewey, 1970); from our study, it is evident that there simply is no intermediate earthquakes below the central section of Taiwan and that the deeper earthquakes to the north and to the south are associated with two Benioff zones with quite contrasting directions of dip. In this particular instance, if a cross section near Taiwan such as that in Fig. 7(b) is used, the conclusions can be misleading.

Figure 9(a)–(d) present seismicity maps based on 12 months (of 1975) data from the 22 stations telemetered network that is now in full operation in Taiwan. Several active faults shown as thick lines on land can be correlated with the location of the epicenters. The correlation will be discussed further in the section on active and inactive faults.

One interesting feature that emerges from Fig. 9 is that small but deeper earthquakes are found to occur under northern Taiwan, under the young but dormant Andesitic volcanoes. These hypocenters demonstrate that at present the Benioff zone may extend to that region. With the young age of the overlying volcanics, we may argue that the Benioff zone under it is established only recently.

Microearthquake surveys in various parts of the Island (Tsai et al., 1974, 1975; Liaw et al., 1973, 1974; Lu, 1974) show that shallow small earthquakes are quite common. In most cases, the seismicity does not correlate well with known faults. In one area (near Tsengwen reservoir of southwest Taiwan, see Liaw et al., 1974 and Wu et al., 1978) the density of foci decreases sharply below 8 km. It is conceivable that the high fluid pressure in the strata in the Tsengwen area (S.L. Chang, personal communication, 1978) and also in other areas in the foothills (Suppe and Wittke, 1977) is responsible for the microearthquake seismicity.

Table 2. Focal mechanism solutions.

<table>
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<tr>
<th>Date</th>
<th>Origin time</th>
<th>Lat.</th>
<th>Long.</th>
<th>Plane 1 Az, Pi</th>
<th>Plane 2 Az, Pi</th>
<th>Dep (km)</th>
<th>P Az, Pi</th>
<th>T Az, Pi</th>
<th>Type</th>
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<td>122.14</td>
<td>169.61</td>
<td>349.29</td>
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<td>169.16</td>
<td>349.74</td>
<td>T*</td>
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<td>23.09</td>
<td>120.58</td>
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<td>98.31</td>
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<td>171932.8</td>
<td>22.41</td>
<td>121.26</td>
<td>14.4</td>
<td>106.32</td>
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<td>6</td>
<td>95.4</td>
<td>185.4</td>
<td>SS</td>
<td>21</td>
</tr>
</tbody>
</table>

*Katsumata and Sykes (1969).
Fig. 10. Summary of focal mechanisms of large earthquakes ($M > 6$) from 1963 to 1975, superposed on bathymetry in the vicinity of Taiwan. The numbers beside the mechanisms correspond to those in Table 2.

5. Focal Mechanism Solutions

Two sources of data are available for focal mechanism solutions in this work. They are (1) world-wide Standard Seismograph stations plus the local Taiwan network of fifteen stations equipped with relatively old instruments (such as Wiecherts and Omori's), for
events with $m_0 > 5.8$ in the period of 1963 to 1975, and (2) microearthquake data for small events in the period of January 1973 to December 1974.

Some of the solutions in the vicinity of Taiwan, utilizing the WWNSS data have already been published (Katsumata and Sykes, 1969; Wu, 1970; Sudo, 1972). In Table 2 we have listed the previously published solutions as well as new data. It should be pointed out that 1972 is an important year in earthquake data for Taiwan, in that several mechanisms cleared up what had previously been only a conjecture. In Fig. 10, we have presented the data both in terms of interpreted slip vectors (in case of thrust faults), horizontal projections of pressure or tension axes (in case of 45° thrusts or normal faults) or displacement vectors (when the solution indicates strike-slip faulting). The new solutions themselves are presented in Fig. 11.

Earthquakes 6, 7 and 8 represent a distinct group at the junction of Ryukyu and Taiwan. Wu (1970) has inferred them to be right-lateral strike-slip fault(s) based on bathy-
mechanism solutions.

metry. Subsequently, results from a reconnaissance cruise in that area by R/V Hunt (Wageman et al., 1970) found right-lateral displacement in the shallow sediments; it is possible that these observations are related. Recently, more ocean bottom reflection data have been accumulated and a north-south zone of disturbed sediments and troughs have been found in the epicentral area of those earthquakes and indicate that a north-south structure underneath (Lu et al., 1977). Sudo (1972) on the other hand, believes that these earthquakes are associated with the extension of the left-lateral Central Philippine Basin Fault (Hess, 1946), which is apparently an inactive feature at present and has been interpreted, based on magnetic profiles as ancient ridges (Uyeda and Ben-Avraham, 1972). Sudo (1972) used G-wave radiation pattern from the March 12, 1966 event (No. 6 in Table 2) to support his conclusion. By using more stations for a similar study, we have found a dominant lobe of G-wave radiation pattern in the N 20°E direction (Fig. 12); this maximum represents the direction of growth of the fault, which is not probably parallel.
to the strike. Although the epicenter is located toward the north side of the aftershock zone, normally implying that the rupture had propagated southward, the observed radiation pattern could have resulted from a complex rupture process with the final pattern consistent with that of a northeastward propagating fault—the long period P waves of this earthquake indicate that more than three events occurred in a series.

Earthquakes 2, 10, 11, 12, 13 and 17 have thrust solutions. The fault planes for these solutions strike approximately north-south (event 17 shows NW-SE planes) and the dips of these planes range from 35° to 45°. Solutions 2 and 17 are for events on land. Solution 2 is without doubt associated with the east-dipping Chukou Fault (Fig. 1) although no fault break was observed in the field. Solution 17 belongs to an event immediately east of the Coastal Range and a thrust-type fault break was observed after the earthquake (Lu et al., 1976) in the Coastal Range.

Solutions 11 and 12 are in the vicinity of each other and are consistent; but events 10 and 13 occurred in the vicinity of other earthquakes that show an entirely different solution. Thus event 13 is next to event 14 that shows a pure strike slip mechanism; event 10 is located among events 8, 1 and 16, and these represent strike-slip fault (No. 8), and shallow thrusts (Nos. 1 and 16). At first this phenomenon may appear out-of-order, but this may reflect only that this region is populated by numerous faults or weaknesses oriented in various directions and under the current stress system the motions along these faults are controlled by the stress field and the orientation and/or the characteristics of the fault plane. For example, considering events 13 and 14, representing a thrust fault and a strike-slip fault, respectively, it is reasonable that it is not mechanically advantageous to have thrust motion taking place along the fault plane of a nearly vertical strike-slip fault, so that short-
ening has to take place elsewhere, while the reason that no strike-slip motion took place along the fault plane for the 45° thrust is that the thrust plane may be highly irregular along the strike as to inhibit horizontal strike-slip motion along it. In the next section, we shall try to find the directions of tectonic stress that is consistent with these fault plane solutions.

Events 14, 20 and 21 are all very well-defined strike-slip solutions. The two events in 1975 (20 and 21) have a large number of aftershocks recorded by the local network (Fig. 9(a) and (b)). The long axis of the aftershock zone coincide with the strikes of the NW-SE plane and are therefore deduced to be left-lateral strike-slip faults.

Event 3 is outside of the Ilan-Lotung Plain and has a normal fault solution. Interestingly enough, this solution agrees well with the composite solution of a microearthquake swarm in this area (C.C. Feng, personal communication, 1975). This solution indicates the dominance of N-S tension in this area and may be important in explaining the subsidence of the EW aligned Ilan-Lotung Plain. More will be said on the subject in later discussion.

Events 1, 16 and 19 have shallow thrust or high angle thrust solutions and they are consistent with other island arc situations, where subduction is inferred. Event 9 in an intermediate depth earthquake representing down-dip tension.

Events 4 and 15 are very important events. We have seen that Manila Trench as a submarine topographical feature terminates at about 21°N; event 15 is located very close to that point and the normal fault mechanism is consistent with that found by Stauder (1968) below the Aleutian Trench. Event 4, on the other hand, is located on the ridge south of Taiwan; it may have the same origin as event 15. Event 15 strengthens the argument that the Manila Trench is still active and the eastward subduction under Luzon is still taking place.

Many microearthquake surveys have been conducted in areas where seismicity data are required for planning or designing purposes. An extensive survey in the Longitudinal Valley, however, was conducted purely for geological reasons (Lu, 1976). These data have become an important supplement of seismicity in land areas where large earthquakes are infrequent. Due to the small number of stations used for each earthquake, it is not possible to obtain a solution for each event. Assuming that the focal mechanisms for a group of small events narrowly limited in space and time remain the same, then we can combine data from many earthquakes to obtain a composite solution. Such solutions are not as dependable as those from one large earthquake in the region both due to the spatial incoherence and the difficulty in ascertaining the takeoff angle when the crustal structure is not known in detail.

Figure 13 presents a summary of the available composite microearthquake focal mechanism solutions. The actual solutions are included in a series of papers and reports by Tsai et al. (1974, 1975), Liau et al. (1973, 1974), and Lu (1976). The quality of the solutions varies from good to very poor, depending on the location and number of stations, the number of earthquakes, accuracy of hypocenter determination, etc.

Solution E (Fig. 13) agrees well with solution 2 of Table 2 and Fig. 11 in being a thrust; Solution C agrees with the NE trending right-lateral surface faulting in that region during the 1935 Hsinchu earthquake (Allen, 1962). Solution H implies a fault almost perpendicular to the main fault (Lanyang Fault) in that region; Solution A together with aftershock data led Lu (1976) to conclude that the NNW plane is the fault plane and left-
lateral motion took place. The juxtaposition of thrust and normal solutions in the Tatun volcanoes region (Solutions F and G) is not easily explained. The events in or around the Longitudinal Valley (Lu, 1976) produce both thrust and partially strike-slip solutions; those with significant strike-slip components are consistent with an NNE striking left-lateral fault.

6. **Tectonic Stress in the Vicinity of Taiwan**

As McKenzie (1969) pointed out, individual focal mechanism solutions only put very mild constraints on the direction, not to mention the magnitude, of the earthquake-gener-
Fig. 14. Successive superposition of regions of admissible P axes to obtain a small region for P axes that satisfies all solutions. There are nine pairs of circles; the one on the left is the focal mechanism solution with region of admissible P as shaded areas; the one on the right represents the superposition of the shaded regions of the present and previous solutions.

...ating stresses. Depending on the strength of the fault, the exact location of the true P axis, for example, can be somewhere in the quadrant with downward first motion. In particular, if the strength of the fault is 0, then P can be located anywhere in that quadrant. When we have a number of focal mechanism solutions for a region that is tectonically homogeneous, i.e., no major change in crustal structure, and no major plate boundary crossing the area, and these solutions are sufficiently different, then we may find a region of P common to all solutions.

Figure 14 presents the result of such a superposition using nine solutions located to the southeast of Taiwan. The dark areas represent the area of P; the numbered circle represents the original solution and the circle to the lower right represents the result of successive superposition. We see that as more and more solutions are superposed, the common P region shrinks. The final result indicates that the tectonic stress to the southeast of Taiwan has a direction of S46°E to S76°E with a plunge between −2° and 15°.

Elsewhere, Wu (1978) has explored the further possibility imposing a reasonable failure criteria on faulting and thereby estimating the order of magnitude of tectonic stress. Using the same focal mechanism solutions and a combination of dehydration-related data (Murrel and Ismail, 1976), we have arrived at a tectonic stress of about 2 kbar.
Table 3. Active faults in Taiwan.

<table>
<thead>
<tr>
<th>No.*</th>
<th>Name</th>
<th>Remarks**</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Taipei</td>
<td>II, IV, VI, VII, I (1694?)</td>
</tr>
<tr>
<td>2</td>
<td>Chuchin</td>
<td>IV, VI</td>
</tr>
<tr>
<td>3</td>
<td>Median</td>
<td>II, IV, VI</td>
</tr>
<tr>
<td>4</td>
<td>Chuchin</td>
<td>IV, VI</td>
</tr>
<tr>
<td>5</td>
<td>Chihsu-Tuntuchio</td>
<td>I (1935), II</td>
</tr>
<tr>
<td>6</td>
<td>Chelungpu</td>
<td>I (1917), III, IV, V</td>
</tr>
<tr>
<td>7</td>
<td>Shuilieng</td>
<td>III, VI</td>
</tr>
<tr>
<td>8</td>
<td>Shuilieng</td>
<td>III, VI</td>
</tr>
<tr>
<td>9</td>
<td>Yichu</td>
<td>I (Inferred by the many earthquakes in the vicinity), VII, VIII</td>
</tr>
<tr>
<td>9a</td>
<td>Meizakeng</td>
<td>I (1792 (?), 1906), II, IV, VII, VIII</td>
</tr>
<tr>
<td>10</td>
<td>Chuko</td>
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</tr>
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<td>Chutouchi</td>
<td>IV, VI, VII, VIII</td>
</tr>
<tr>
<td>12</td>
<td>Hsinhwa</td>
<td>I (1946), III, VII</td>
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<td>Chaochou</td>
<td>II, VI, VII</td>
</tr>
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</tr>
<tr>
<td>17</td>
<td>Longitudinal Valley</td>
<td>I (1951), II, III, IV, VI</td>
</tr>
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</table>

* Refers to numbers in Fig. 15.
** Roman numerals refer to the following criteria used in judging the fault as being active: I, historical earthquake (date) occurred in the vicinity; II, mapped by microearthquake survey; III, quartery movements inferred; IV, geologically mapped; V, geodetic movement measured across the fault; VI, prominent on LANDSAT image; VII, mapped by reflection seismic method; VIII, encountered in borehole.

7. Active Faults and Recent Uplifts

In Fig. 15, we have shown what we judge to be active faults, and the basis of judgement is listed in Table 3.

One of the most discussed features of Taiwan is the Eastern Longitudinal Valley (Hsu, 1962; Allen, 1962; BiQ, 1965, 1971; York, 1976). It is no doubt one of the most important structural elements of Taiwan. It separates the Paleozoic-Mesozoic metamorphic rocks from the Post-Miocene island-arc-associated sedimentary and volcanic rocks. However, due to the rapid erosion rate and the thick alluvium cover in the Longitudinal Valley, exposures of actual fault contact is rare (Hsu, 1962; York, 1976). Consequently, the exact nature of the boundary is still being discussed.

Fault breaks associated with two earthquakes in 1951 proved that at least the part of the eastern side of the northern part of the Valley is fault-bounded and there are both strike-slip and dip-slip components along a high angled east-dipping plane (Hsu, 1962). Hsu (1962) reported on a segment of fault scarp on the eastern side of the Longitudinal Valley that has only thrust motion on it. York (1976) observed a thrust contact further south, and he postulated that one segment of a fault contact on the western side of the southern Longitudinal Valley is actually the continuation of the fault that broke during the 1951 earthquakes. Unfortunately, in 1951 the quality of the seismic stations around the world or in Taiwan was not yet good enough for dependable focal mechanism and since the establishment of WWNSS only one large earthquake occurred in the vicinity of the Longitudinal Valley (mechanism No. 17), and the solution shows a pure thrust.
Biq discussed the tectonics of the Longitudinal Valley in some detail. He accepted Hsu's (1962) description of the eastern side of the Longitudinal Valley; however, facing the difficulty of explaining the uplifting of two rock groups of very different ages, and the condition that at some time during Late-Pliocene the Pre-Tertiary rocks on the Central Range side and the Post-Miocene rocks on the Coastal Range side were coeval, he proposed a Central Range Fault on the western side of the Longitudinal Valley, a west-dipping thrust that would allow the Central Range to rise to the present height. Yen (1965) reported the observation of a fenster on the western side of the Longitudinal Valley, that revealed the sole of such a thrust; the condition of the outcrop however, does not allow exact determination of the nature of the contact. Recent seismicity data (Fig. 16) west of the Longitudinal Valley do not indicate the existence of such a fault; the data seem to show an east dipping boundary marking the more active eastern side from the less active west, but
Fig. 16. Local network seismicity data between 22.8°N and 24°N plotted in a profile perpendicular to the Longitudinal Valley. Note the absence of events deeper than about 60 km.

whether this boundary is a fault or not has to be clarified in the future. Thus, although Biq's reconstruction is logical, it has not received direct confirmation from actual observation.

The structural trend of the Coastal Range is also that for the rest of the Island; most of the through-going faults run north-northeast, except north of latitude 24°30'N, where the faults as well as the strikes of the folds start to turn eastward. In general, those faults that separate the major stratigraphic units are most probably deep thrusts with a left-lateral component and the myriads of low angle thrusts within the Miocene and Pliocene strata in the Foothills are gravity sliding planes (Biq, 1966).

Judging from the historical seismicity, the $M>6$ earthquakes from 1900 to 1976 (Fig. 6(b)) and the recent telemetered network data, the Median Fault (Fig. 15), is still active. Judged from the stratigraphic offset and the NWW-ESE orientation of the compression axis southeast of Taiwan, this fault should have a thrust component; however, one composite solution from a group of small earthquakes along the fault near the latitude of 24°N yield a normal fault mechanism with E-W tension axis (Lu, 1976). It is interesting that near the coastal area, along the northeastern extension of this fault which was recently mapped by a microearthquake survey (Tsai et al., 1973) the composite solution obtained is similar to that of an earlier large earthquake in the same region (No. 3 in Fig. 10): they both indicate normal faulting (C.C. Feng of CERC, personal communication, 1975). Although in this case the tension axis turns to N-S. This conclusion perhaps is not surprising in that this region is along the extension of the Okinawa Trough, a tensional feature, and that the Ilan-Lotung Plain is one of two areas in Taiwan that are undergoing relatively slow uplifting compared to the rest of the Island. A hinge action may take place somewhere south of 24°N.

The Median Fault is apparently offset by a (geologically inferred) east-west fault in the Central Range which was inferred from stratigraphic data and has not been directly observed (Ho, 1975). The southern section of the Median Fault is named Chao Chou Fault (Chiang, 1971). It is also a thrust fault and is shown to be active by telemetered
network data (Fig. 9), but no large historical earthquake is known to be associated with it.

On the western side of Taiwan, there are several faults that were associated with earthquakes. These are of two types: (1) NEE-SWW trending the right-lateral faults and (2) NNE-SSW trending thrust faults (Fig. 4). The 1935 earthquake, for example, was reported to have created both types; although whether these two faults (Fig. 4) were created during one continuous process or were two separate events cannot be assured. A combined fault length of about 50 km (NAKAMURA, 1936) can quite conceivably be produced by an $M=7.1$ earthquake. The 1941 Taiwan earthquake fault and the 1964 Tainan-Chiayi earthquake fault (also called the Chukou Fault) form another such pair; the 1964 earthquake however, did not show surface faulting, and the nature of the associated faulting is deduced from the focal mechanism solution (No. 2 in Fig. 9). The 1906 Chiayi earthquake was associated with an NEE-SWW right-lateral strike-slip fault (OMORI, 1907). An earthquake in 1792 (Table 1) near Chiayi probably occurred along the same fault, one person was reported to have fallen into a crack near Meitzukang (FANG, 1968). For an earthquake in 1917 in central Taiwan, no report on surface faulting had been found, but a leveling line across one of the major faults in the area showed a 15 cm maximum increase in elevation over the fault (IMAMURA, 1935).

A number of $M>6$ historical earthquakes cannot easily be correlated with exposed faults, but the estimated epicenters seem to line up in the vicinity of the Yichu marginal fault surrounding the Peikang basement high (MENG, 1967 and Fig. 15). The nature of faulting associated with earthquakes in this region has to be resolved in the future.

The gravity sliding faults are generally thought to be inactive, because they are not connected to deep-seated fractures. However, insofar as they are weak-zones, if they happen to overlie active unexposed faults or as the region is being strained they could be reactivated; the total displacement may not amount to much in this case.

That Taiwan and its vicinity is tectonically very active can also be judged from the rate of uplifted coastal terraces. Li et al. (1978) have recently dated by the C$^{14}$ method a number of new coral samples from the eastern side of the Coastal Range and around the Hengchun Peninsula; by adding these new terrace age data to data published by LIN (1969) and TAIRA (1975), the uplifting rates of the Coastal Range, the Hengchun Peninsula and northern tip of the island have been determined. Figure 17(a) shows the unreduced data set and the sea level data as determined by MORNER (1976).

Figure 17(b), (c) and (d) show the rate of uplift of eastern, southern (including southwestern) and northern Taiwan respectively. While eastern and southern Taiwan have a rate of about 5 mm/year, northern Taiwan underwent an episodic history for the last 8,000 years. We shall discuss the significance of these rates in the next section.

8. Plate Boundaries Near Taiwan

Taiwan represents an unique situation in the Circum-Pacific tectonics. For most of the cases where oceanic plate is subducting at the continental margin, the island arcs created on the non-subducting side are relatively low in topography and the highs are the isolated volcanic peaks. Taiwan, however, is composed mainly of miogeosynclinal sediments and they have been subjected to extraordinary compression to become high mountains.

Currently the region is seismically very active and it is difficult to define a single
Fig. 17. (a) $^{14}C$ dates and altitudes of the raised coral reef samples. Bottom half presents the sea level change curves (Peng et al., 1977). (b) Minimum uplift ($=$ altitude of coral reef $+$ eustatic sea level relative the present sea level) vs. the age of samples from Hengchung Peninsula and Taiwan area. The dotted line represents a proposed step-wise rise of the Hengchung Peninsula. (c) Minimum uplift vs. age for eastern coast of Taiwan. (d) Minimum uplift vs. age for northern coastal area and the Ryukyu Islands.
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continuous plate boundary based on seismicity alone. On the contrary, shallow earthquakes reveal a complex deformation pattern in a belt about 300 km wide; unlike a typical island-arc environment, where earthquake focal mechanisms have dominantly linear distribution patterns, where Benioff zone can be clearly defined and where bathymetric features correlate well with this distribution, in the vicinity of Taiwan, there is no clear display of either focal mechanism patterns nor trenches, ridges or interarc basins.

By combining all the available data however, we can define a boundary composed of several segments that are quite different in characteristics. Not all segments are typical boundaries of plates as commonly described.

It is clear that Ryukyu arc terminates about 100–120 km east of Taiwan. Not only the Ryukyu Trench stops there, but the presence of large strike-slip faults and the sudden increase in seismicity toward Taiwan all indicate the presence of a major boundary there. To the west of this boundary, a subduction zone similar to the Ryukyus exists; but judging by the offset in the E-W bathymetric ridge in the vicinity of 24°N, 122.8°E (Fig. 10), and the offset in the accretion lens as marked by the —100 m gal free-air anomaly (Bowin et al., 1978), this subduction is displaced to the north, relative to the Ryukyus. Further to the west, this subduction boundary connects to the thrust-left-lateral Longitudinal Valley fault zone. This fault zone is not a simple transform fault, but a collision boundary with a transform component. It is very difficult to define a plate boundary between 21°N and 22.8°N; here we have a hint of a subduction zone with deeper earthquakes under the offshore volcanic islands to the southeast of Taiwan, but there is an apparent lack of M>4 shallow earthquakes along the extension of the Manila Trench; perhaps subduction is now halted and collision is taking place involving a broad zone of deformation.

In Fig. 18, we have sketched the nature of the plate boundary in the Ryukyu-Taiwan-Luzon areas as it is understood.

The kinematics of the right-lateral fault that terminates Ryukyu is not entirely clear. Karig (1973) considers its formation to be a consequence of the opening of the Okinawa Trough. The difficulty with this explanation is that the Okinawa Trough evidently extends all the way to the Ilan Plain as shown by the focal mechanism and other data, and thus, if back arc basins do imply active spreading, there is no significant differential spreading between the Okinawa Trough behind the Ryukyus and its westward extension. Also, since the faults, as indicated by the focal depth of earthquakes are at depths of 50 km or so (Table 2), it is probable that only the subducting plates are involved; according to Karig's hypothesis, the upper plates are the ones in differential motion.

There is a possible alternative explanation. Because of the complex geometry and the changes in the nature of the plate boundaries near Taiwan, the tectonics here cannot be accounted for fully by purely rigid plate motions. Consider the generally north-westward motion of the Philippine Sea Plate here (Wu, 1972; Frigh, 1972, Seno, 1977), the northward component of this motion can proceed quite expectedly through subduction, the westward component is practically halted by the collision. This collision results in intense intraplate deformation near Taiwan; on the island the mountain building is vigorously taking place and in the ocean the plate is undergoing E-W compression and left-lateral strike-slip fault along NW planes before subduction. Both the compression and the strike-slip faulting would cause a congestion; as there is no prominent topography in the ocean immediately east of Taiwan, the rate of subduction northeast of Taiwan, west of longitude 122.7°E must be higher than that east of 122.7°E; thus, there is a differential motion along 122.7°E and the sense should be right-lateral.
The subduction zone and the strike-slip faults between Taiwan and the Ryukyus evidently are developed only recently, perhaps within 2 mybp; the absence of volcanic islands and wellformed interarc basins over this zone may be the result of this young age. As the subducted lithosphere is a part of the Philippine Sea Plate, the westerly component of the motion vector of the plate, the deeper part of this zone moves west. This motion is somewhat restricted, because the top of the zone is practically stopped by the collision. This western extension reaches under the northern extremity of Taiwan, and is obviously responsible for the andesitic volcanism there.

The collision-transform boundary is perhaps not as sharp as it is sometimes thought to be. Even though the Longitudinal Valley Fault is considered to be the main plate boundary, with the continuous distribution of foci east of the Coastal Range, this boundary can best be viewed as a boundary zone of tens of kilometers wide. The recently published triangulation data across the northern part of the Longitudinal Valley (Chen, 1974) indicates a left-lateral displacement rate of more than 6 cm/year, a value close to the rate of motion of the Philippine Sea Plate towards the north in this area; but it could also be reflecting, to a large part, displacement associated with the 1951 earthquake.

The right angle junction of the collision-transform and the subduction boundaries near Hualien has some interesting implications. For example, on the bathymetric map (Fig. 10), there is a rather prominent depression near 24°N and 122°E, and on the free-air gravity map of Bown et al. (1978) there is a corresponding low. It is tempting to conclude that these features are related to a nascent subduction zone. But the right angle
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Fig. 19. Mechanism for the bending of northern Taiwan in Pleistocene.

This junction of boundaries described above implies that in the vicinity of the bathymetric and free-air gravity low, the subducting plate is reaching deeper level, and while this plate is moving west and compressing Taiwan, the upper plate is not being pushed and may be decoupled from the subducting plate; thus a north-south tension feature develops. This trough is quite persistent; the high rate of sedimentation from a recently raised Taiwan has not been able to bury it.

This junction can also explain the bending of northern Taiwan toward the east as schematically shown in Fig. 19. As can be expected from the uplifting and mountain building in Taiwan, collision must have been taking place along the whole island in the early orogenic history. The commencement of the northward subduction sometime during perhaps Pleistocene causes the northern part of Taiwan to lose compression from then on, while the shortening of the rest of Taiwan is taking place; bending results from this differential compression. The modern revelation of this lack of compression in northern Taiwan is the episodic uplift of the northern coast against the steady uplift of the rest of the island in the last 8,000 years. It is also recorded in the change in fold axes orientation in the metamorphic complex near Suao (Tan, 1978).

With the normal fault type of earthquakes located in or near the Manila Trench south of 21°N and the apparently east dipping Benioff zone under the Bashi Strait and Luzon a more typical subduction plate boundary can be defined. Although complications do exist on the eastern side of Luzon, where a new subduction boundary may be forming.

9. Discussion

Taiwan is a remarkably young and active island. The present rate of uplift of Taiwan is one of the highest in the world and the seismicity is also high. Although the tectonics is not simple, the extrapolation back in time can perhaps be accomplished because of the relatively short geological history it involves.
The present orogeny of Taiwan began in Late Pliocene. That was the time when the sediments in the western Taiwan basin switched from a northwestern source to an eastern one (Chou, 1973). Collision of the island arc with the continental shelf occurred probably at that time or may be earlier. The extensive Late Miocene to Early Pliocene sediments accumulated in the trench were raised, deformed and eroded and some were redeposited in the southern part of the trench as it was probably still deep. At the same time, the Central Mountains were gradually formed; as the slope increased, some of the sediments were redeposited in the western Taiwan basin, and some of the Miocene and earlier sediments, some not yet fully consolidated, slid off or were pushed off the slope to form part of the western foothills.

Sometime in the Pleistocene, the northward subduction near Taiwan started. The commencement time of this event can be estimated as follows: assume a maximum depth of the subduction zone to be 130 km (Fig. 7(a) and Tsai et al., 1977), a lithosphere of about 50 km (Tsai et al., 1977) and an average subduction rate of about 5 cm/year (Seno, 1977), then the time it takes to develop the part of the subduction zone under the lithosphere is

![Fig. 20. Schematic time history from Miocene to present for Taiwan and its vicinity. Dotted outlines are used for location only.](image-url)
about 1.6 my. This corresponds also to the commencement time of the bending of Taiwan.

It is more difficult to reconstruct the sequence of events prior to the collision. Certainly northern Taiwan, north of the Peikang Basement High had been a depositional basin on the continental shelf continuously since early Tertiary. The southern Taiwan basin however, did not receive much sediments until after the Late Miocene, and it was a basin beyond the edge of the shelf. It is quite possible that the Late Miocene "marine transgression" of the northern as well as the southern basins was a result of the advance of the island arc toward the continental shelf, whereby the shelf and the surrounding area was depressed.

Although the idea of an arc, more or less parallel to the long axis of Taiwan, moving toward Taiwan during Miocene and Early Pliocene and eventually colliding with Taiwan is widely accepted (CHAI, 1972; KARIG, 1973; BOWIN et al., 1978), there is still a controversy as to the polarity of the arc. JAHN (1972) and recently Chen (J.C. Chen, personal communication, 1978) favored the subduction toward the west under the continental shelf while CHAI (1972), KARIG (1973) and others favored the subduction toward the east under the Philippine Sea. For sure, the subduction could not be toward the west just before the collision, otherwise we would expect to find traces of volcanic expressions associated with the subduction in the Central Mountains. Also it is much easier to explain the kinematics of a westward drifting arc with an east-dipping subduction zone; with the clockwise rotation of the Philippine Sea Plate (with respect to the Asian continent) an east dipping subduction zone will be pushed toward the west. An east dipping zone is consistent also with the plate boundary south of Taiwan.

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