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STRUCTURE AND TECTONIC EVOLUTION OF TAIWAN

BRUCE H. T. CHAI
Department of Geology and Geophysics, Yale University,
New Haven, Connecticut 06520

ABSTRACT. Although Taiwan is one link in the chain of island arcs in the West Pacific, the concave side of the Taiwan arc faces toward the Pacific, opposite all the other arcs of this chain. Structurally, Taiwan is a typical mountain range, which was formed from a preexisting geosyncline of Tertiary age. The pre-Tertiary history of Taiwan is little known, as most of the older rocks are strongly deformed and metamorphosed. Because of a few Permian fossils, they are believed to be Paleozoic in age. The Tertiary geosyncline formed probably during early Paleocene or latest Cretaceous (?), and a thick pile of monotonous argillaceous rocks was then deposited. Later the axis of this sedimentary trough migrated gradually toward the west, and thick alternating beds of sandstone and shale were deposited. In the Early Pleistocene, the orogenic paroxysm took place, and the Tertiary beds were folded into a mountain range. The Eastern Coastal Range is quite different from the rest of Taiwan and has its own peculiar structure and history. There was strong volcanism in that region during early Miocene time, and all later sediments seem to have been deposited in a highly unstable basin. Orogenic disturbance occurred there also in Early Pleistocene time, and the rocks were transported to their present position by gravitational gliding.

Tectonically, Taiwan can be divided roughly into infrastructure, transition zone, and suprastructure, proceeding from the west side of the subduction zone in the Longitudinal Valley westward. The pre-Tertiary crystalline rocks belong to the infrastructure and are characterized by flow folding. The Paleogene sequences belong to the transition zone, where shear folding predominates. The Neogene sequences belong to the suprastructure, and the folding is concentric or flexural-slip. Structurally, the Eastern Coastal Range is also suprastructure, but it is an independent tectonic province and can not be grouped with the province west of the subduction zone.

The direction of curvature, the tectonic break of the Longitudinal Valley, Bouguer gravity anomaly data, seismic data, volcanic activity, and submarine topography all seem to indicate that there is an east-dipping subduction zone (or Benioff zone) along the Longitudinal Valley. Within the framework of plate tectonics. Taiwan is an excellent example of a continent-arc collision structure, and the tectonic evolution of Taiwan is controlled by the underthrusting mechanism of the continental plate under the oceanic plate along the subduction zone of the Longitudinal Valley.

INTRODUCTION

Taiwan is situated in that part of the Circum-Pacific island chain that fringes the eastern margin of the Asian continent. Many geologists have considered that Taiwan is merely the junction of the Ryukyu arc to the north and the Samar arc of the Philippines to the south (Wilson, 1959). However, various lines of evidence discovered in recent years indicate that Taiwan itself is an island arc convex toward the west, opposite to the other west Pacific island arcs, which are all concave to the west toward the Asian continent (Biq, 1956, 1959, 1961, 1970, 1971b; Ho, 1967; Katsumata and Sykes, 1969; Dewey and Bird, 1970; Wu, 1970). That Taiwan faces in the opposite direction clearly indicates that its tectonic evolution should be rather different from that of other island arcs in the same region. Moreover, Taiwan is situated right on the west tip of the Philippine Sea plate where that plate hits the Eurasian continental plate. The main island is formed from Tertiary geosynclinal sediments lying on a metamorphic basement and was raised to its present elevation by late Cenozoic orogenesis.

Seismic data (Katsumata and Sykes, 1969; Dewey and Bird, 1970; Wu, 1970) show that the earthquake focus zone lies mainly off the island on the Pacific side, and Miocene andesites are found off the eastern coast too. These lines of evidence indicate that the Eurasian plate was once underthrust below the Philippine Sea plate at Taiwan and that the uplift of Taiwan is due to the collision of the Eurasian continent with an island arc. It is the purpose of the present paper to interpret the tectonic evolution of Taiwan from the point of view of plate tectonics.

STRATIGRAPHY AND GEOLOGICAL HISTORY

For convenience of description, the stratigraphy and geological history may be divided into four major parts, following Ho (1961, 1967) (fig. 1):

A. Pre-Tertiary basement,

B. Paleogene sequence,

C. Neogene sequence, and

D. East Coastal Range sequence.

A. Pre-Tertiary basement

Stratigraphy.—The oldest rocks of Taiwan are in the metamorphic belt extending along the eastern slope of the Central Mountain Range. This belt has a north-south length of 240 km and reaches in the north a maximum width of 25 km. It is made up mostly of schist and crystalline limestone with a small amount of gneiss. In the northern part, migmatite injections (quartz-diorite injection) were recognized. All these rocks are grouped under the general stratigraphic term Tananao schists, because schists are the dominant rocks in the sequence. Yen (1960, 1962, 1963, 1967) has differentiated the schists into three main types: (a) black schist—chiefly graphite schist; (b) green schist—chlorite schist, amphibole schist, amphibolite, metadiabase, and serpentinite; (c) siliceous schist—quartzite, schistose sandstone, chert, and vein quartz. In metamorphic grade, these rocks are in greenschist facies (quartz–albite–muscovite–chlorite subfacies, quartz–albite–epidote–biotite subfacies, quartz–albite–epidote–almandine subfacies) whereas the migmatite injections belong to the albite–epidote hornfels and hornblende hornfels facies. The black schist and siliceous schist are paraschists derived mainly from graywacke, sandstone, shale, and other siliceous rock. The green schist is derived from basic effusive and pyroclastic rocks as well as basic to ultrabasic plutonic rocks. At the eastern foot of the Central Range west of Yüli, glaucophane schist was found in the Yüli formation of the Tananao schist (Yen, 1959, 1966, 1967). The presence of such schist indicates low-temperature high-pressure metamorphism, although jadeite and lawsonite were not found and the associated carbonate was calcite, not aragonite.

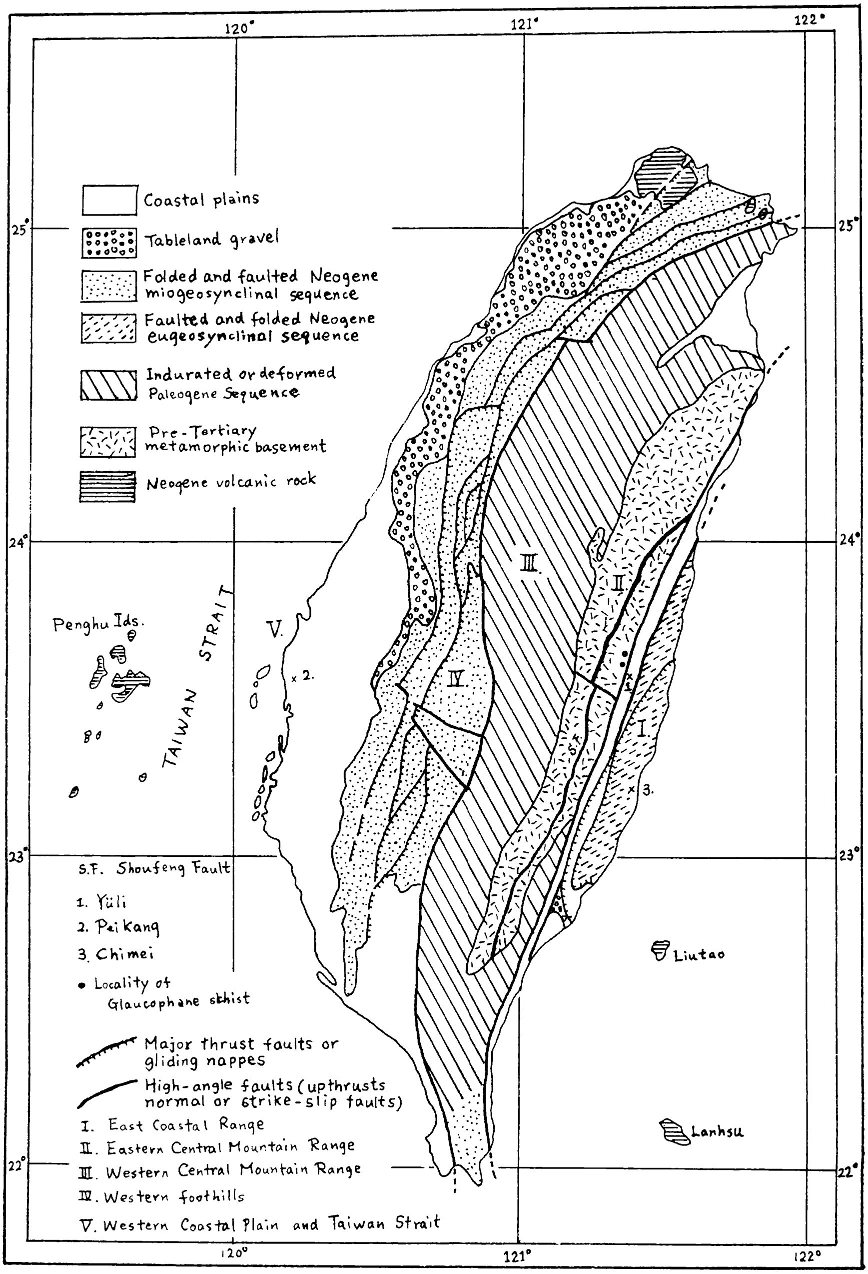

Fig. 1. Tectonic sketch map of Taiwan (modified from Ho, 1967).

Geologic history.—The structure of the pre-Tertiary metamorphic rocks is complex and little known. The succession of strata is obscure, and fossils are nearly absent or very poorly preserved. The exact age of most of the metamorphic rocks cannot be determined. The only fossil remains consist of some deformed specimens of fusulinids and corals found in crystalline limestone and are of Permian age, indicating that at least some of the original sediments of the metamorphic complex are of late Paleozoic age (Yen, 1953). All these metamorphic rocks are generally believed to be late Paleozoic to Mesozoic in age (Yen, 1960).

The time of the pre-Tertiary orogeny is also difficult to determine, because the time gap represented by the unconformity between the metamorphic complex (Permian) and the overlying sediments (latest Cretaceous(?) to Eocene) is large. It is generally believed that this orogeny took place in late Mesozoic time, either Jurassic or later. This orogeny is called the Nanao orogeny and may correspond to the Yenshan orogeny in China of the Nevadian orogeny (and possibly also the Sevier orogeny) in western North America. The metamorphic complex was mobilized again at a later time. Two radiometric dates (K-Ar method) on the micas in pegmatite and quartz diorite give ages of 33 and 86 m.y. respectively, roughly middle Tertiary and late Mesozoic (Yen and Rosenblum, 1964). The significance of these two dates is not clear; probably both micas partially lost their argon during the Plio-Pleistocene orogenic event.

The pre-Tertiary basement of the western Coastal Plain has also been revealed by drilling (Stach, 1958; Matsumoto, Hayami, and Hashimoto, 1965). It is composed of indurated sandstone, shale, and tuff. According to the fossil evidence, the basement here is of Jurassic to early Cretaceous age. Thus here also there is a big time gap between the basement and the disconformably overlying Miocene beds.

B. Paleogene sequence

Stratigraphy.—The Paleogene rocks occupy the crestal zone and a large part of the western flank of the Central Mountain Range. They are exposed west of the metamorphic complex and extend from north to south for approximately 340 km. The maximum width of this belt is 50 km. At the contact with the metamorphic basement, basal conglomerate composed of metamorphic pebbles has been found in a very few localities (Yen and others, 1956). In other places the contact is obscure and not clearly defined.

The Paleogene rocks of the Central Mountain Range form a thick monotonous sequence of dark indurated shale, which grades into phyllitic and argillitic shale and at the extreme into well-cleaved hard slate. The bulk of the shaly sequence was named the argillite sequence by Ho (1967). Lithologically, it is composed of dark gray, well-cleaved argillite. The degree of induration (or metamorphism) increases toward the east. Shear folding predominates, and there is considerable rock flowage and an extensive development of slaty cleavage. Detailed structural and stratigraphic analysis has not yet been possible, because neither the top and base nor the total thickness are clearly defined. The thickness must be over several thousand meters. Sandstone zones or sequences of interbedded sandstone and shale occur; limestone and marl beds or lenses are found in the higher parts of the Central Mountain Range. Interstratified or intrusive sills of basic to intermediate composition (basalt, dolerite, andesite, tuffs, and pyroclastic products) are found.

The dark gray shaly facies is typical of the early development of a geosyncline (pre-flysch facies) and indicates a considerable deepening of the basin while the border regions were being elevated. Other examples of this sedimentary facies include the undifferentiated Permian-Jurassic graywacke and argillite of the New Zealand geosyncline (Suggate, 1961), the Eocene Old Slates formation in the Barisan zone of Sumatra (Van Bemmelen, 1954), the Cretaceous to Eocene phyllitic rocks of Borneo (Roe, 1955), and the undiffereniated Mesozoic complex including the Muroto group (Cretaceous to Paleogene) and the Akigawa group (Triassic to Cretaceous of Southwest Japan (Murakoshi and Hashimoto, 1956). Furthermore, we can compare such facies with the "Schistes lustrés" or "Bündnerschiefer" in the geosynclinal trough of Alps (Biq, 1962).

Geologic history.—Fossils are very rare in most of the argillite. Fossils in the limestone or marl beds are mainly Eocene foraminifers (Yabe and Hanzawa, 1930). Later study shows that the foraminifers from the indurated shales are as young as Oligocene to Aquitanian (Chang, 1953, 1954, 1962, 1963). Two species of latest Cretaceous to Paleogene corals (*Astrocoenia* and *Elephantaria*) were found in the interbedded conglomerates at the base of the sequence (Yen and others, 1956). Thus the lower part of the slate sequence is possibly latest Cretaceous in age, but there is no other positive evidence to prove it. The Paleogene geosynclinal trough thus possibly formed in latest Cretaceous to Paleocene time and was well developed by the Eocene. Disturbances during deposition have been described by Yen and others (1956) and Chang (1963), because of the discovery of a few Cretaceous corals in the pebbles of some conglomerates and also the fossil hiatus between Middle Oligocene and Aquitanian. However, because of stratigraphic uncertainties and the lack of clearly defined structural breaks, those disturbances are questionable. Maybe there were some local fluctuations, transgressions, or regressions, but they were trivial as compared to the whole sedimentary history.

The Paleogene rocks are generally indurated and slightly metamorphosed, but the Neogene rocks are mostly non-metamorphosed; the two overlapped in Aquitanian time (Ho, 1967). A mid-Tertiary (or late Oligocene) orogeny has been suggested by many geologists. However, there is no convincing evidence of angular unconformities to substantiate this diastrophism. Only fault contacts are known between the Paleogene sequence and the Neogene sequence on the west. The base of the Neogene sequence is still not visible and is often obscured by faults on the eastern side. Kobayashi (1954) pointed out that no coarse orogenic sediments were found in the Neogene sequence to indicate a mid-Tertiary orogeny.

Drilling in the west coast area of south-central Taiwan shows that Miocene formations directly overlie the Mesozoic basement. Little or no basal conglomerate has been cored at the contact (Stach, 1958). Because direct structural evidence is lacking, the exact nature of the mid-Tertiary orogeny in Taiwan is not clear. Some geologists (Ho, 1967) suggest that Paleogene and Neogene sedimentation was continuous, but that the axis of the sedimentation trough was shifting constantly westward as embryonic folding upheaved the earlier formed sediments. The difference in degree of metamorphism between the Paleogene and the Neogene sediments is due to burial, which gives a higher degree of induration of the Paleogene rocks. It is postulated here that uplift of the sedimentary trough started far back of the middle Oligocene, that the rate of uplift was slow at first and then gradually speeded up, and that the axis of the sedimentary trough also moved gradually westward corresponding to the speed of uplift. On the eastern side of the sedimentary trough sedimentation might gradually lessen and then cease and erosion start. The speed of uplift may not have been uniform, but we still may consider it a continuous process. Thus a mid-Tertiary orogeny is not necessary. Even if it happened, it would have been a very slow and mild uplift associated with compaction and induration of the older strata.

C. Neogene sequence

Stratigraphy.—Neogene sediments are the most widely distributed rocks in Taiwan. They are exposed in the foothill region west of the Central Mountain Range, the region that was well mapped and studied after World War II. Paleontologic study in recent years indicates that this thick Neogene sequence begins with the Aquitanian, based on correlation with Caribbean faunal zones (Chang, 1962). Aquitanian rocks on the east side of the basin have different lithological facies and deformation history from those on the west side. On the eastern side they were indurated and deformed, on the western side slow subsidence and deposition continued throughout Neogene time, and a new sedimentary trough developed as a result of the westward migration of the geosynclinal axis.

The base of the western sedimentary trough dips eastward from a foreland area on the west. The maximum thickness of the Neogene sediments may reach well over 8000 m, but they thin toward the western foreland, forming a typical clastic wedge tapering off toward the western coastal plains and the Taiwan Strait. Exploratory drilling near Peikang (fig. 1) on the west coast shows a Neogene section less than 1500 m (Stach, 1958). A more recent borehole farther west on Penghu Island in the Taiwan Strait gives a Neogene column of approximately 200 m only.

The sedimentary surface in this Neogene basin also sloped toward the south, and the Miocene sediments change from shore facies to the north toward basin facies to the south. The shelf-type strata are characterized by white arkosic to orthoquartzitic sandstone, interlaminated shale, silt, and sandstone, and thin coal beds. They are well developed in northern Taiwan and interfinger southward with basin-type sediments. The basin-type sediments are exclusively marine and are mostly subgraywacke or calcareous sandstone and dark gray shale corresponding to a flysch-type sedimentation but lacking the characteristic sedimentary structures, such as flute casts or graded bedding. These basin-type sediments were developed only in the southern part of the basin during Miocene time but extended over the whole basin during Pliocene time. The coal-bearing facies is found only in the Miocene beds of northern Taiwan. The Pliocene in southern Taiwan is a thick mudstone series (Cholan formation or Neimem formation), nearly 3000 m thick, probably representing a mass flow from the neighboring highland (Ho, 1967; Tsan and Keng, 1968). A few limestone lenses are known in the Miocene of northern Taiwan. Neogene carbonates are more common in southern Taiwan, where reef limestones of varying size and extent were found at several levels in the upper Pliocene and Pliestocene formations. This is only a local phenomenon, as limestone beds are not known elsewhere in this region.

Volcanism (Kungkuan Tuff) occurred in the northern part of this basin. Basaltic tuffs, other pyroclastic materials, and lava flows form bodies of irregular shape. The tuff bodies differ widely in geologic occurrence, lateral extension, stratigraphic extension, and thickness but are not physically continuous. The volcanics are mostly submarine; they occur in almost all the Miocene units in northern Taiwan but diminish a great deal toward the south. They also shift from early Miocene to late Miocene southward (Ho, 1967, 1969).

Geologic history.—Slow subsidence and sedimentation continued throughout Neogene time, for no distinct and large-scale angular unconformity has been found, although minor depositional hiatuses or disconformities have been reported (Stach, 1958; Sun, 1965; Schreiber, 1962), possibly due to local upheaval, tilting, and erosion. Syn-tectonic sedimentation was common on the eastern side of the Neogene trough or fore deep, where earlier sediments were uplifted and eroded to feed the deepening basin. But the sediments in the western clastic wedge definitely come from the foreland on the western side. In early Pleistocene time, a thick conglomerate formation (Toukoshan formation) covered all the Neogene rocks. This conglomerate grades both northward and southward into fine clastics interbedded with conglomerate layers; it is a typical molasse-type sediment indicating the retreat of the sea and the initiation of orogenesis. The early Pleistocene was the period of orogenic climax of the Cenozoic movements in Taiwan. Strong orogenic disturbance occurred right after the accumulation of the thick conglomerate and equivalent beds, because the conglomerate follows the Pliocene beds without major break. The Neogene sediments were folded and thrust faulted as they were elevated into highlands. This orogeny is called the Penglai orogeny and can perhaps be correlated to the Pasadenan orogeny in North America. The present island was shaped by this orogeny.

The post-orogenic deposits are the laterite-capped tableland gravels, which lie unconformably on the eroded surfaces of the deformed Neogene formations. This angular discordance is the only conspicuous unconformity in the Cenozoic record of Taiwan.

D. East Coastal Range sequence

Stratigraphy.—The East Coastal Range is a small narrow range bordering the Pacific Ocean. It is separated from the Central Mountain Range to the east by a long and narrow rift valley—the Longitudinal Valley. The range is about 150 km long with an average width of no more than 15 km. The average altitude is around 1000 m, even lower toward the north. The East Coastal Range is composed mainly of marine sedimentary rocks and volcanic rocks of Neogene age. The total thickness of the Neogene sediments ranges from 4000 to 6000 m (Hsu, 1956). They are characterized by the coarseness and heterogeneity of the constituent grains. Facies gradation is sharp and rapid. The major stratigraphic units (modified from Hsu, 1956) are:

1. Tulanshan formation. The Tulanshan formation is a mighty sequence of andesitic lavas and pyroclastics. The latter are partly lava agglomerate (upper part), tuff, and lapilli. This formation is distributed extensively throughout the East Coastal Range and is over 1500 m thick. A thin bed of limestone, about 5 to 50 m thick, locally covers the uppermost part of the formation. Foraminiferal fossils in the limestone indicate a middle Miocene age (Hsu, 1956; Chang, 1967, 1968, 1969; Chang and Chen, 1970). Recent K-Ar ages of andesite extrusions in the Chimei igneous complex fall in the range from 17 to 22 m.y. (Ho, 1969), also Miocene. This is the oldest rock found in the East Coastal Range.

2. Takangkou formation. The Takangkou formation apparently lies conformably on the Tulanshan formation and also has a wide distribution. Lithologically, it is composed chiefly of fine-grained dark gray to black shale and conglomerate. The conglomerate occurs either as irregular bodies or lenses. The irregular bodies may reach 500 m in thickness and contain pebbles and cobbles derived from the metamorphic rocks west of the basin; they are distributed in the northern part of the range. The lenses may reach 100 to 150 m thick, they are composed of gabbro and serpentine in addition to the metamorphic rocks and are poorly sorted; they are more common southward. The average thickness of the formation is 2000 m, the maximum 2700 m. Fossils indicate middle Miocene age (Hsu, 1956). But Chang (1967, 1968, 1969; Chang and Chen, 1970) points out that most of this formation belongs to the *Sphaeroidinella dehiscens/Globorotalia crassaformis* subzone, thus being Pliocene, and that only a small portion of the lower part belongs to the Upper Miocene, so that there would be two disconformities (or hiatuses) in the section. He also points out that both biological and lithological evidence shows that the western part of this formation was deposited in a shallow sea.

3. Chimei formation. The Chimei formation is composed almost entirely of alternating thin beds of sandstone and shale. The sandstone is medium to coarse grained and locally contains pebbles of metamorphic rocks. Contemporaneous deformation occurred during deposition. Load casts, flute casts, convolute bedding, and slump bedding are very common in this formation (Hsu, 1954). It is a typical turbidite very similar to those of the Northern Apennines (Abbate and Sagri, 1970). It has a thickness of 1250 m and lies conformably on the Takangkou and Lichi formations. The age is upper Pliocene (Chang and Chen, 1970).

4. Lichi formation. Lichi formation is exposed at the southern end and southwestern edge of the East Coastal Range. It consists mainly of massive chaotic and non-stratified mud and clay and contains numerous exotic blocks of various sizes, shapes, and rock types, some of them hundreds or even thousands of meters in diameter. These exotic blocks include sandstone, limestone, gabbro, serpentinite, peridotite, dolerite, basalt, and andesitic agglomerate. The surface of these exotic blocks often show scratching and abrasions, and sole markings on sandstone also indicate overturning. All these lines of evidence indicate that the exotic blocks were transported and settled down in the clayey matrix. The thickness estimated by Hsu (1956) is 1560 m. A hole drilled by the Chinese Petroleum Corporation penetrated 1061 m of this clay formation without reaching the base (Meng and Chang, 1965). Moreover, the rootless basaltic and sandstone blocks encountered provide more explicit evidence for the exotic nature of the large rock bodies in this formation. It resembles either the Wildflysch of the Swiss Alps or the "argille scagliose" (olistostromes) of the northern Apennines in Italy (Hsu, 1956; Ho, 1967; Biq, 1969; Abbate and others, 1970a; Abbate, Bortolotti, and Passerini, 1970). The age of the formation has been given as middle Miocene (Hsu, 1956). Recent study shows however a wide range of age from middle Miocene (or earlier) to Pliocene (from the *Globorotalia mayeri* zone to the *Sphaeroidinella dehiscens* zone) (Chang, 1967; Huang, 1969), thus corresponding in age to all three formations already described. Clearly, it represents a different sedimentary facies.

5. Pinanshan conglomerate. The Pinanshan conglomerate occurs in the southwestern part of the East Coastal Range. It is composed of pebbles and cobbles of various kinds of metamorphic rocks including graphite schist, quartz-mica schist, and crystalline limestone. The pebbles range from 5 to 15 cm in diameter and are definitely derived from the Central Mountain Range to the west. The whole formation is estimated to be over 1400 m thick, although the lower boundary is not exposed. The age has been given as middle Miocene or lower by Hsu (1956), corresponding to the Tulanshan formation. However, micro-paleontological study gives an age of Pliocene (Chang, 1967) or later, possibly corresponding to the thick Pliocene conglomerate in western Taiwan (Toukoshan conglomerate—the molasse).

Geologic history.—The geologic history and tectonic evolution of the East Coastal Range is a subject that has long been disputed by

Taiwan geologists, because most of the original eugeosynclinal basin is missing now, and in the present East Coastal Range only a small fraction of the eugeosynclinal materials remain. Rocks of different facies but of the same age are now piled together, indicating that the original sedimentary basin was complex.

Pre-Miocene rocks are rarely exposed in the East Coastal Range. A single outcrop of quartzitic slate near Fuli is reported by Hsu (1956), but its relations are doubtful. The outcrop is too small to be mapped and is also isolated from outcrops of other formations by recent river deposits. Hsu suggests that it is in fault contact with the Lichi formation. I suggest that, if we believe that the whole Coastal Range is an allochthonous pile of eugeosynclinal materials, this outcrop might indicate a fault contact but does not belong to the East Coastal Range series, being autochthonous. On the other hand, it might also be just an exotic block within the Lichi formation. Thus the basement of the East Coastal Range is never exposed. However, the Bouguer gravity anomaly map of the Taiwan region shows that the anomaly is negative in the Central Mountain Range but positive in the Coastal Range and the Pacific Ocean basin off Taiwan (Liu, 1964; Yen, 1969; Wang, 1970). The seismic data also show that the crustal thickness is around 37 to 40 km under the Central Mountain Range but only 20 to 27 km under the East Coastal Range, and less than 20 km under the Pacific Ocean near Taiwan (Huang, ms; Yen, 1968). This suggests that a transition zone from continental crust to oceanic crust lies under the East Coastal Range.

Magmatic activities, mainly eruption of andesitic lavas and pyroclastics, undoubted occurred during the early Miocene in this region, corresponding to the volcanism and diastrophism of the Philippine arc during the Miocene (Ho, 1961) and also to the Miocene tectonic activity in the Philippine Sea basin (Karig, 1971). Large amounts of andesitic lava, agglomerate, tuff, and lapilli tuff accumulated on the eastern side of this basin. At the same time the relatively deep-sea sediments of the Lichi formation were also starting to accumulate on the western side. During the upper Miocene to earlier Pliocene, the Takangkou formation accumulated. The sedimentary basin was still rather deep. Disturbance occasionally occurred on the northeastern side of the Central Mountain Range (which was located northwest of the basin), and thick poorly sorted conglomerates were transported and accumulated in the northern part of the basin. Lithological features and field occurrence indicate that they are typical turbidites. Pebbles of metamorphic rocks clearly indicate that the conglomerate came from the west, from the Central Mountain Range. At the same time, the southern part of the basin was still relatively deep, and the Lichi formation was continuously accumulating.

After the accumulation of the Takangkou formation, during the middle to late Pliocene, the basin itself became unstable. Contemporaneous deformation is recorded in the Chimei formation, which was deposited at this time. Load casts, convolute bedding, and slump bedding are very common in this formation, but no cross-bedding, indicating that the basin was still very deep at that time. The Chimei formation is exposed mainly in the middle and southern part of the Coastal Range and seems to cover the Lichi formation conformably at the south (Hsu, 1956, geologic map and section). Thus the Lichi formation is probably older than the Chimei formation, providing an upper boundary for the Lichi formation up to middle Pliocene. But here we must remember that, as orogenic events migrate, in the areas that were first affected by movement a new depositional stage may set in, while at the same time in areas that were affected later, deposition may still have the character of the preceding phases. Consequently, one can find sequences of the same age that belong to different geosynclinal stages. I think that this may explain the age conflicts between these two formations.

Toward the end of the Pliocene and the beginning of the Pleistocene, the orogenesis reached its climax; the rocks were uplifted, crumpled, and broken against the interior of the cordillera. Large-scale submarine gravitational gliding took place during the later part of this orogeny. During the orogeny, the western border of the trough (that is, the present eastern flank of the Central Mountain Range) was also greatly uplifted. Thick conglomerates composed of metamorphic rock fragments piled up on the western side of the trough (the Pinanshan conglomerate). These conglomerates are in direct contact and indeed intertongue with the Lichi formation, which was moved westward to its present position by gravitational gliding. The exact date of the mass gliding is not known because no autochthonous rocks are exposed. Very possibly the Lichi formation slid more than once, to judge by the field evidence for intertonguing, as in the case of the allochthonous mud flows (the argille scagliose unit) of the Northern Apennines in Italy (Abbate and others, 1970a; Abbate and Sagri, 1970). More detailed sedimentologic and stratigraphic studies are required.

The last phase of tectonic movement was the development of the Longitudinal Valley (also named the East Rift Valley), which was formed by compression coupled with sinistral faults (Biq, 1965). Biq describes it as a ramp bounded by two upthrust faults, one on each side. Since the valley is very narrow, only about 5 to 7 km wide, and also the old crystalline metamorphic complex are significantly upthrown only along the western border of the valley, wheras the eastern margin shows no positive evidence of upthrow, it is doubtful if there is a great fault on the eastern border of the valley. Earthquakes (Hsu, 1954; Biq, 1965) do show sinistral movement along the eastern side of the valley. But I think that this is just the surface expression of a major wrench fault, a major subduction zone between the East Coastal Range and the Central Mountain Range. Since the material in the valley is Pleistocene to Recent, any part of the fault zone may show horizontal shifts or other surface expressions of the differential movement on the fault. This fault zone has a very long history and is still active now; it is probably the most important tectonic feature in Taiwan.

TECTONIC STYLES AND STRUCTURE PATTERNS

The structure of Taiwan has been studied by many geologists. It is generally agreed that Taiwan is a typical mountain range formed from a preexisting geosyncline by folding, thrusting, and uplift, although detailed interpretations differ from one another. Looking at Taiwan island, we recognize that Taiwan itself is an arc convex toward the Asian continent, opposite to all the other island arcs bordering the eastern margin of Asia (fig. 1).

Typical Alpine-type structural features are beautifully shown in the Taiwan fold-belt, more or less following the curvature of the island itself. Or we can simply say that the shape of Taiwan island is controlled by the structure patterns of the Taiwan fold belt. The tectonic provinces described here are almost exactly the stratigraphic provinces described in the previous section. They are (from east to west): (A) the East Coastal Range, (B) the Eastern Central Mountain Range, (C) the Western Central Mountain Range, (D) the western foothills, and (E) the western coastal plain and Taiwan Strait.

A. The East Coastal Range

The East Coastal Range is separated from the rest of the island by a major wrench fault—the Longitudinal Valley—and forms an unique independent tectonic province of exceptional interest. Its detailed tectonic features have been described by Ho (1967) and Biq (1965, and especially 1969). The main structural lines trend north-northeast generally parallel to the range. Neogene rocks are folded into a series of moderately dipping subparallel anticlines and synclines. Thrust faults are dominant in this belt and dip generally to the east. Recent studies show that the range is allochthonous and developed by gravitational gliding tectonics (fig. 2).

Gliding occurred twice. First was the accumulation of the Lichi formation, the "argille scagliose". Big exotic blocks were moved within the mudmatrix. The largest is a mass of pillow basalt 3.5 km by 1 km in size. The second largest is andesitic agglomerate, 1800 m by 700 m. The origin of the ophiolites has been disputed strongly for years. Juan (1958, 1964; Juan and others, 1953; Juan and others, 1959; Juan and Wang, 1960; Juan and Hsu, 1962) presumed that the ophiolites are plutonic intrusions, whereas Biq (1960), Ho (1967), and Wang (1966) think they are original ocean crust and mantle material that was squeezed up into the flysch sediments of the Lichi formation as solid broken blocks. The latter opinion is now favored by most Taiwan geologists. During accumulation of the Lichi formation, gliding was probably repeated several times, as proposed by Biq (1969). Wang and Chen (1966) proposed that the Lichi formation was formed by a diapiric squeezing. In my opinion, diapiric squeezing probably was as important as gliding, because there isn't very much room for gliding. Thus the tectonic accumulation of the Lichi formation was certainly chaotic.

The allochthonous Lichi formation was in turn overlain by a gravity nappe (Biq, 1965, 1969; Ho, 1967), which glided down from the east onto the clayey formation not long after the emplacement of the latter. In contrast to the Lichi formation, the nappe moved as a whole by gravity gliding; thus the original stratigraphic order is still intact. The movement was due to continuously increasing stress difference caused by gravity.

B. The Eastern Central Mountain Range

Across the rift valley westward, a large area of crystalline rock is exposed, made up mostly of schist and crystalline limestone with subordinate gneiss. The rocks are highly deformed into irregular patterns of folds and faults. Plastic deformation and flow or disharmonic folding predominate in these rocks. Since the original rocks may not have been homogeneous, this kind of ductile flow indicates that the rocks were probably deformed under comparatively high pressure and temperature. Biq (1966) uses the structure-level concept of Argand and Wegmann and infers that this region belongs to the "infrastructure". The structure levels are controlled by the thermal front. The migmatitic gneiss bodies in the northeastern part may represent this thermal front. Pegmatite intrusions in this region give K-Ar ages of 86 m.y. and 33 m.y. (mica dating), corresponding to late Cretaceous and middle Oligocene (Yen and Rosenblum, 1964). The first date is supposed to belong to an earlier cycle of orogeny (Nanao orogeny), and the later date to represent the begin-

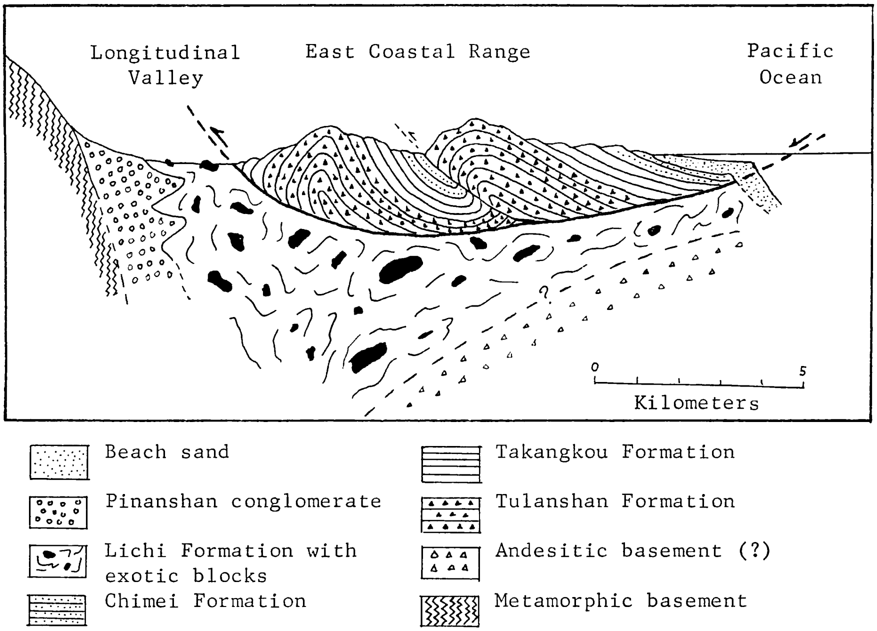

Fig. 2. Schematic tectonic cross section of East Coastal Range.

ning of remobilization and deep granitization of this crystalline basement. Whether this statement is true is highly doubtful, because no granite batholith has been found in this region. I suspect that both micas have lost some argon during the Plio-Pleistocene orogeny; thus the ages might not mean anything. More dating data is required to solve this problem. The structural style of this region may correspond to the Crystalline Appalachians (or Piedmont Plateau) in North America (King, 1959).

C. The Western Central Mountain Range

As mentioned in previous sections, the rocks of this region are mainly Paleogene argillites with subordinate sandstone, pyroclastics, and limestone lenses. These argillaceous beds display a different style of deformation from the crystalline basement. Since the rocks are rather ductile and more homogeneous in lithology, shear folding predominates with considerable rock flowage and development of slaty cleavage. Internal movements during folding are mostly along shear planes and are not restricted to bedding planes. The rocks are indurated and deformed by a very low-grade regional metamorphism, which is characterized by low temperature and moderately high pressure, as typical in a geosynclinal environment.

In general, the folds strike mostly north-northeast, and the thrust faults generally have the same strike and dip eastward. The cause of the westward movement was probably partly gravity and partly westward compression. Biq (1966) assigns this region to a transitional zone in structure level; it is similar to the Blue Ridge Province in eastern North America, which is characterized by the extensive development of slaty cleavage (King, 1959).

D. The Western Foothills

The western foothills region corresponds to the major part of the Neogene miogeosyncline of western Taiwan. The deformation is characterized by tight to broad superficial folding, thrust faults, and no metamorphism (Ho, 1967). As the rocks are mostly alternating beds of sandstone and shale that differ in competency, flexural-slip folding is predominant, and most folds are concentric or parallel. As we go eastward toward the Central Mountain Range, the fold style changes gradually to flexural-flow folding, owing to the increasing ductility of the rocks. On the extreme east of the Neogene fold zone, some accordion or chevron folds appear in the tightly folded rocks, indicating a transition from the western Neogene flexural-slip fold belt into the Paleogene shear-fold belt. Most of the folds are asymmetric or even overturned, and the axial planes dip southeast. The folds are generally bounded by thrust faults of the same strike, which can be traced for considerable distances in a north-south direction. The thrust faults generally dip southeast, forming a series of imbricated thrust blocks, each successively overriding the one northwest of it. Many thrusts cut through the surficial beds at steep angles so that they appear to be high-angle thrust faults on the surface, but deep drilling and underground mining show that these high-angle thrusts gradually flatten at depth and finally become very low-angle thrusts (Ho, 1959; Ho, Hsu, and Jen, 1962).

Cascade folds and gliding nappes have also been recognized in recent years (Biq, 1965, 1966, 1969). The cascade fold of Wenshuichi valley (fig. 3A) shows strong folding on the western side, decreasing intensity of folding eastward, and finally a few very broad folds to end the series. Two cascade folds occur in that region; both show downward and westward movement along fault planes. Thus instead of being east-dipping upthrusts, they were in reality gravity nappes sliding down and then

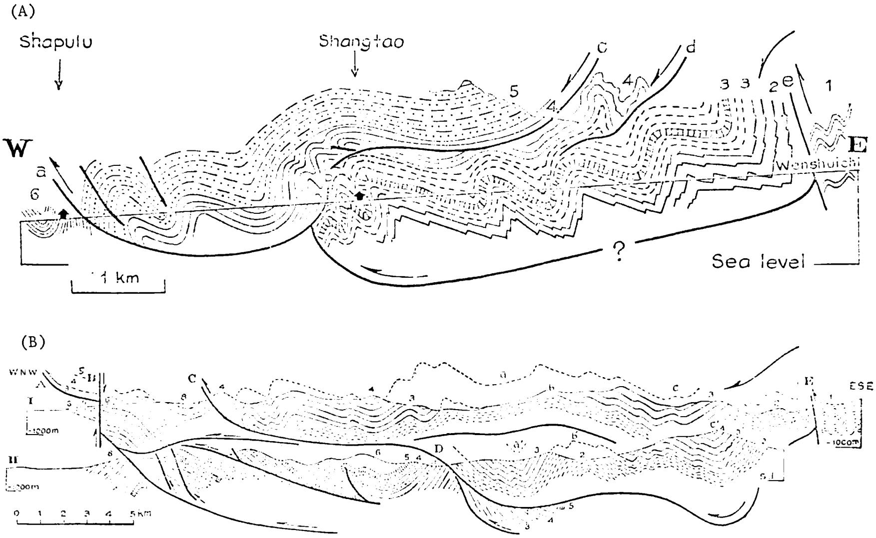

Fig. 3. Composite sections showing the nappe structures of the western-foothill region (from Biq, 1969).

A. The cascade-fold and nappe structure of the upper Wenshuichi valley. The central upper part of the section shows the structure of the southern part of the Nanchuang coalfield north of the valley. 1. Wulai group, 2. Wenshui formation, 3. Piling shale, 3′. middle sandstone member of the Piling shale, 4. Shihpi formation, 5. Talungshan formation, 6. Nanchuang formation. a. Tatungho fault, b. Maacchi fault, c. Luchang fault, d. Piling fault, e. the upthrust separating the foothill zone from the slate belt.

B. The nappe structure of the Alishan area. The upper section is a generalized structure-section of the down-thrown block of the Tatou fault, and the lower section is that of the upthrown block. The vertical slip of the fault is about 2 km, according to which the downthrown-block section is placed above the upthrown-block. This brings out the tectonic discordance between the Alishan nappe above and the autochthon below. Note the parautoshthonous Sub-Alishan nappe immediately below and immediately west of the Alishan nappe. Folds a, b, and c in the upper section correspond respectively to folds a′, b′, and c′ in the lower section; the fact that each of the three folds in the lower section is lower in the composite section than the corresponding fold in the upper section indicates that, owing to subsequent fault-folding, the eastern part of the Alishan nappe is much more down-buckled in the south block of the Tatou fault than in the north block. 1. Paleogene rocks, 2. Tapang formation, 3. Nanchuang formation, 4. Kuantaoshan sandstone, 5. Shiliufen shale, 6. Tawo siltstone, 7. Chinshui shale, 8. Cholan formation, 9. Toukoshan formation. A. Chukou thrust, B. Tatou fault, C. Luchu thrust, D. Chashan thrust, E. Changshan fault.

piled up into cascade folds before the momentum was no longer effective. This is a good example of décollement structure where only the surficial formations were stripped and deformed whereas the lower strata are practically intact. Another similar example is the Alishan nappe (fig. 3B) discussed by Biq (1969) in detail. It has a length of about 150 km in a nearly north-south direction and a maximum width of 26 km. This big nappe apparently has no root zone at all. Moreover, east-west across the area only a few broad folds are shown, indicating that this huge allochthonous mass was not pushed forward by compression. It seems that the interpretation as a gravity nappe is inevitable. The absence of a root zone indicates that the nappe is a décollement structure produced by tectonic denudation when the area to the east was uplifted during late Pliocene to early Pleistocene time.

In this region, only the Neogene rocks are involved in all these tectonic structures; no older Paleogene indurated rocks or the metamorphic complex have been affected or exposed. This again suggests that all the folding, thrusting, or nappes are relatively shallow. At the contact between Neogene rocks and underlying basement a décollement structure has been recognized. Thus Neogene rocks were stripped off the underlying basement (Ho, Hsu, and Jen, 1962). Biq (1966) thus infers that this region is a suprastructure in structural level and that the rocks were not metamorphosed and only deformed within a shallow region separated from below by a décollement. Tectonically this region is comparable with the Sedimentary Appalachians (Valley and Ridge province) of North America (King, 1959).

E. The western Coastal Plain and Taiwan Strait

The coastal region is topographically very flat, with only a few rolling hills and terraces, and is mostly covered by late Pleistocene to Recent sediments. Seismic and gravity data and deep drilling for petroleum supply plenty of sub-surface evidence (Stach, 1958; Chiu, 1970). Folds are gentler and broader toward the west. Faults also decrease westward in number and magnitude. Farther westward toward the shelf zone and Taiwan Strait, Neogene rocks wedge out abruptly. In a borehole on Penghu Island, the Neogene column is approximately 200 m thick only. The beds are very flat and more or less follow the shape of the basement. Small tensional block faulting has been recorded in the Taiwan Strait area. The structural pattern in this area was produced by compression from the east which dies out toward the west—a typical décollement structure as in the Jura Mountains (Bucher, 1955, 1956; Rodgers, 1964), but much gentler (Chiu, 1970).

In the very eastern part of this region, probably also including the western part of the region last discussed, imbricated thrusts and folds dipping toward the southeast were produced by compression from the east. Such thrust sheets have been described by Bucher (1955) as "peel thrusts". The broad gentle synclines and anticlines in strata in this region are very similar to those of the Foreland area (or the Allegheny synclinorium) in the Appalachians of North America. Thus as a whole the structure patterns of the West Taiwan area (not including the East Coastal Range) are almost identical to those of the Appalachians in North America, but the scale is much smaller in Taiwan. An east-west section across the whole island based on the above description is shown in figure 4.

TECTONIC EVOLUTION AND PLATE TECTONICS

In the early 1960's, concepts of sea-floor spreading (Hess, 1962; Dietz, 1961), continental drift, and transform faults (Wilson, 1965) were advocated to explain the large-scale tectonic movements and processes within the Earth. These ideas were expanded and refined in the later years of the decade, and a new theory of lithosphere plate tectonics, the so-called "new global tectonics", emerged (McKenzie and Parker, 1967; Le Pichon, 1968; Morgan, 1968; Isacks, Oliver, and Sykes, 1968; Dewey and Bird, 1970). An important change concerns the boundaries of individual plates. In the new theory, these boundaries are tectonic discontinuities, that is ridges, trenches, and transform faults. Six big global tectonic plates have been proposed by Le Pichon (1968). Thus continental drift is now merely a corollary of plate motion; continents as much as 70 km thick are superficial passengers on plates as much as 150 km thick bounded below by the low-velocity zone (asthenosphere). Thus plate tectonics is essentially ocean based. Plates are generated at oceanic ridges, cool, thicken, and subside as they move away from the seismically and thermally active "mountains" of the oceanic ridges (Dewey and Bird, 1970) and are finally consumed in marginal trenches.

Taiwan is a part of the Pacific island chain fringing the Asian continent. As mentioned in previous sections, Taiwan is not just a junction of two island arcs but is itself an independent arc concave toward the Pacific (Biq, 1959, 1961, 1970, 1971b; Ho, 1967; Katsumata and Sykes, 1969; Dewey and Bird, 1970). This unique attitude definitely indicates that the tectonic evolution of Taiwan is different from that of the other island arcs on this chain and has its own tectonic significance.

Several interpretations have been given for this area. Biq (1956, 1959, 1961) presumed that Taiwan is a double island arc, only much smaller in scale, and divided the region into (1) inner basin, (2) inner arc, (3) inner deep, (4) structural arc, and (5) foredeep, going from the ocean toward the Asian continent (fig. 5). Here I am not going to discuss Biq's papers but point out two points that seem not quite right. First, Biq assumed that the Inner basin (that is, the Philippine Sea basin) is sialic and so being incorporated into the growing continent of Asia. Recent studies (Katsumata and Sykes, 1969; Karig, 1971) prove that the basin is typically oceanic, like the rest of the Pacific Ocean. Second, he classified the Taiwan Strait as a foredeep. All recent evidence indicates however that the Taiwan Strait is on the continental shelf as part of the Asian continental plate. The average water depth is less than 50 m, and even the basement is shallower than in the inner arc; there seems no reason

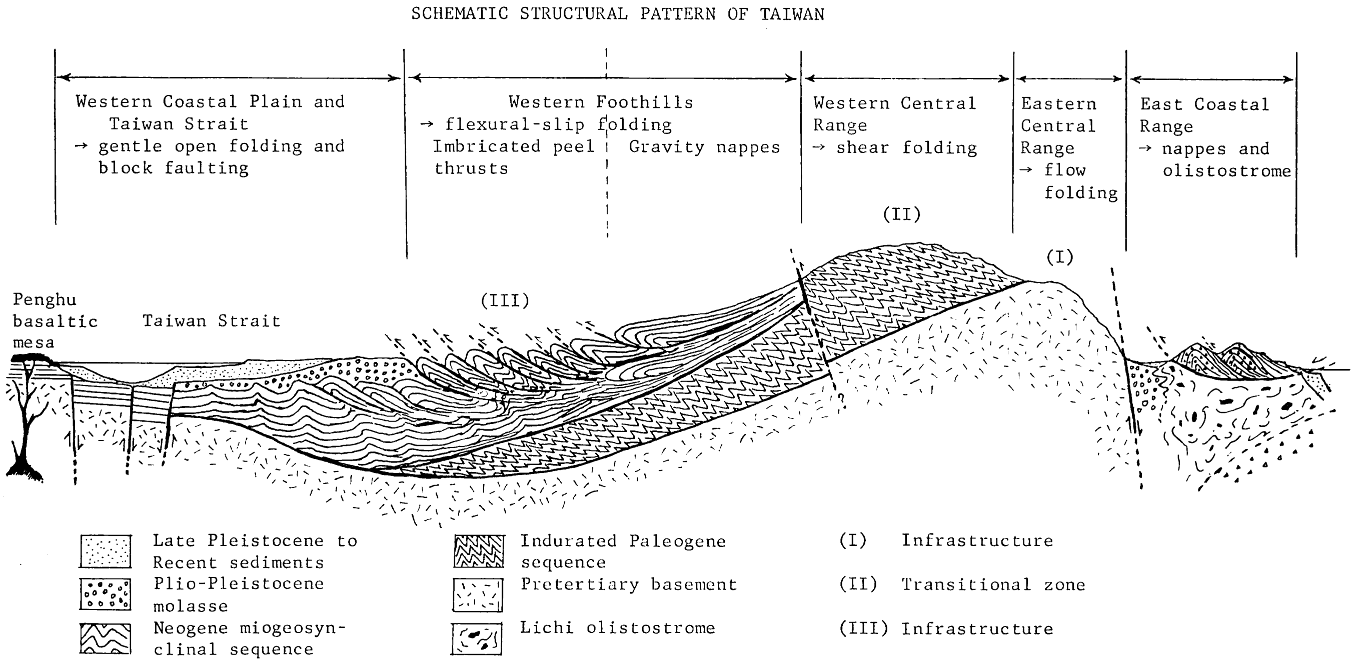

Fig. 4. Schematic structural cross section of Taiwan.

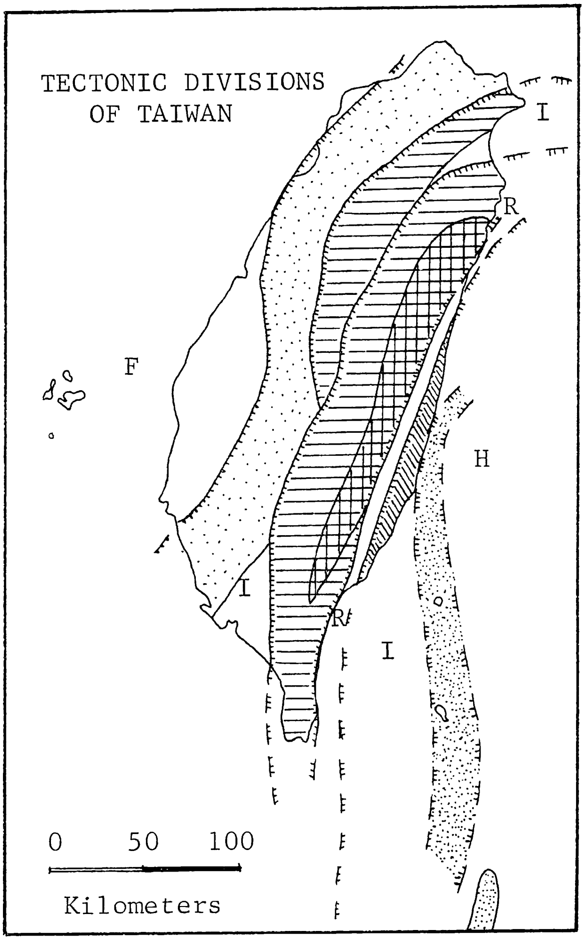

Fig. 5. Tectonic divisions of Taiwan (from Biq, 1961). H. inner basin; densely stippled, volcanic inner arc; I. interdeeps; diagonally ruled, East Coastal Range; R. rift valley; horizontally ruled, Central Mountain Range; cross-ruled, crystalline core in the Central Mountain Range; thinly stippled, Foothill Zone; F. foredeep; saw-teeth lines, major thrusts.

at all to class it as a foredeep. Structurally, Taiwan Strait may be classified as an exogeosyncline, which was not well developed just because it lies on the continental platform (Biq, personal commun., 1971). Juan (1958) has a completely opposite interpretation and declares that Taiwan is simply a Neogene marginal geosyncline. But it is very hard to explain the positive Bouguer gravity anomalies in the eugeosynclinal region of the East Coastal Range. Yen (1968) uses all available geophysical, geochemical, and geological data to show that the crust under most of Taiwan is of continental type but that the crust under the East Coastal Range is of intermediate to oceanic type. He concludes that the Taiwan region has both continental characters and distinguishable features of island arcs.

Yen's conclusion seems proper and more acceptable, but he did not point out what kind of mechanism might give these particular tectonic features. Here, in terms of plate tectonics, I suggest that Taiwan island is an example of continent-island arc collision, as proposed by Dewey and Bird (1970). The Eurasian plate is underthrusting below the Pacific plate under the Taiwan region (the names of the plates follow Le Pichon's terminology, 1968). This conclusion is based on several lines of evidence:

1. The shape of Taiwan island as well as the structure pattern, as mentioned previously, is concave toward east (the Pacific side), indicating that the underthrusting should be from west to east as compared with the other circum-Pacific island arcs and their direction of underthrusting.
2. One big tectonic break, the Longitudinal Valley, divides the island into two completely different geological provinces (Allen, 1962; Biq, 1965), suggesting that these two provinces were originally not neighbors but were brought together by later processes.
3. The Bouguer anomaly data (fig. 6) show a sharp change directly at the Longitudinal Valley; west of it anomalies are negative, east of it positive. This indicates probably two different types of basement—that west of it being continental and that east of it intermediate to oceanic (Yen, 1968).
4. According to Biq (1970), the seismic data show that earthquake foci are distributed on a surface that starts right below the Longitudinal Valley and dips eastward and may reach a depth as great as 200 km off the eastern coast in the Pacific Ocean. Katsumata and Syke's data (1969) (figs. 7, 8) are more scattered but still give a general trend dipping toward the east, although the slope is rather steep.
5. All evidence indicates that the thick pile of andesitic rocks in the East Coastal Range came from east of the present coast. Two andesitic islands (Lutao and Lanhsu) are now sitting off the eastern coast in the Pacific Ocean. The submarine maps of Heezen and Tharp (1965, 1971) show a volcanic ridge that extends southward from these islands to connect with Luzon Island in the Philippines. This is probably the source area of the thick andesitic rocks of the East Coastal Range. Thus, the distribution of andesite to the east of the subduction zone (or the Benioff zone) along the Longitudinal Valley also indicates that the plane of underthrusting dips toward the east.

Wang (1970) studied the K_2O versus SiO_2 variation of Pleistocene andesites of northwestern Taiwan with depth, but he picked the andesite of the East Coastal Range as one point. This seems wrong, because the andesite of the East Coastal Range belongs to a different andesitic province from the Pleistocene andesites he studied. K-Ar dating of the East Coastal Range andesite gives an age ranging from 22 m.y. to 17 m.y. (Ho, 1969), corresponding to a Miocene age, whereas the northwestern Taiwan volcanism belongs to Pliocene to Pleistocene time. Of course the andesite

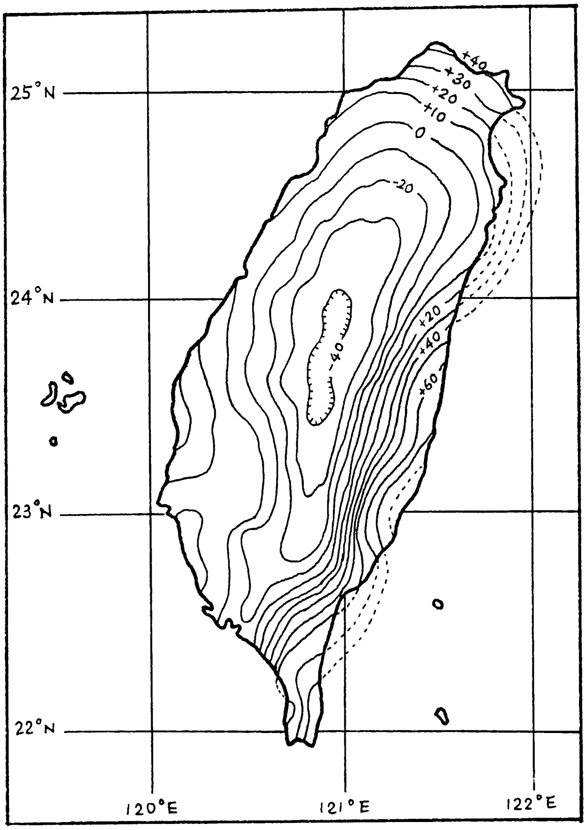

Fig. 6. Bouguer anomaly of Taiwan (after Liu, 1964).

near Chimei (in the middle part of the East Coastal Range) represents a shallow type, but the variation trend should be toward the east instead of toward the west. The chemical composition of andesite of Liutao and Lanhsu is not known so far, but an eastward trend of the variation seems more plausible.

6. A submarine trench extending southward from the Longitudinal Valley is also shown in Heesen and Tharp's maps (1965, 1971). It extends continuously southward far more than 800 km and finally forms the sediment-filled West Luzon Trough to the west of Luzon (fig. 9). This trench has an average depth over 3000 m and exceeds 5000 m in the deepest part. Possibly it is the relic of the original trench of the eastern island arc before collision. Dewey and Bird (1970) noticed that the position of this trench west of the northern part of Philippine arc might indicate that the oceanic crust of the Southern China Sea is being consumed there, and that the northern Philippine arc is approaching the Asian continent. A similar example is the approaching of the Australian continent to the New Hebrides. Two simple schematic sections to show this kind of

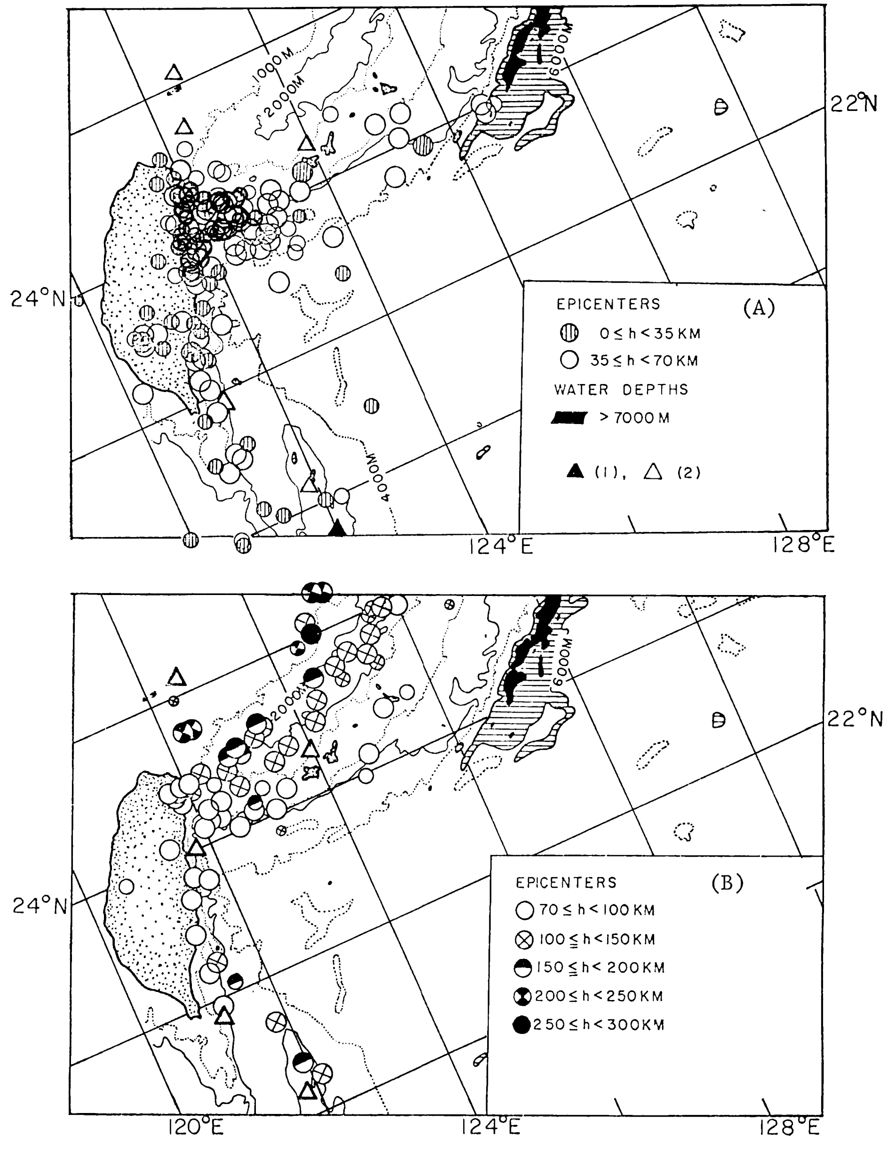

Fig. 7. The focus of earthquakes in the Taiwan region:
A. Shallow earthquakes; B. Intermediate-depth earthquakes (after Katsumata and Sykes, 1969).
1. Active volcanoes; 2. Extinct volcanoes.

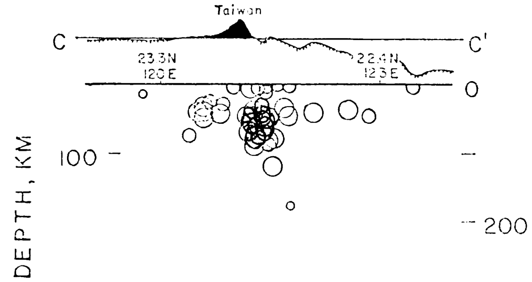

Fig. 8. Vertical section of epicenters in Taiwan region (from Katsumata and Sykes, 1969).

structure before collision are given below (fig. 10, after Dewey and Bird, 1970). Thus, where the trench is on the continental side of an arc and hence where marginal small ocean basins are contracting, island arcs may eventually collide with continental margins. Taiwan is, thus, an excellent example of this type of collision. Based on this idea a schematic tectonic evolution of Taiwan is proposed as following:

The pre-Tertiary history of Taiwan is not very well known because the rocks in the Central Mountain Range have been metamorphosed and highly deformed. The original sedimentary rocks were sandstone, siltstone, shale, and thick limestone. These rocks represent a near-shore shallow marine sedimentation on the continental shelf. Sedimentation probably started in early Paleozoic time and extended into early Mesozoic time. On the other hand, deep drilling on the Western Coastal Plain (Stach, 1958) reveals that a Cretaceous (?) indurated pyroclastic sequence directly underlies soft Miocene beds. Jurassic (?) sandstone and shale are also found. The drilling penetrates the whole sequence and gives a thickness of 530 m. The The ammonite *Holcophylloceras aff. mediterraneum* was identified within this sequence. A distinct angular unconformity underlies the basal conglomerate of the Jurassic, indicating that there was probably a tectonic disturbance before the Jurassic. A big hiatus that indicates emergency or non-deposition started after the Jurassic, but the dip in both the Jurassic and the Cretaceous sequence is practically the same. Thus the event might be merely a gentle uplift following a slight subsidence during which the water-laid pyroclastics were deposited. The Jurassic sequence was probably indurated by burial but not disturbed or metamorphosed. The pre-Jurassic rocks are highly indurated dark sandstone and shale which apparently are not metamorphosed or recrystallized. The correlation between these rocks and the pre-Tertiary crystalline rocks of the eastern Central Mountain Range is not known. The Cretaceous pyroclastics probably correlate with the late Cretaceous acid igneous activity on the adjacent mainland to the west (Stach, 1958). The relation between the Eurasian plate and the Pacific plate at this time is not known.

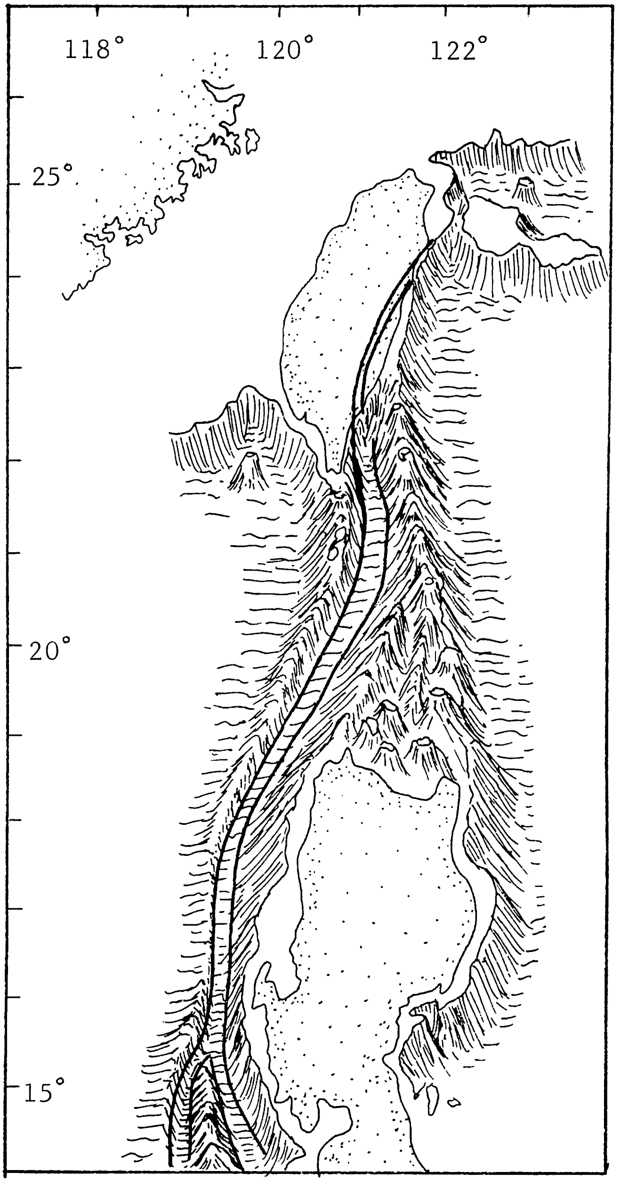

Fig. 9. Submarine topography map showing the extension of the Longitudinal Valley to the West Luzon trough (modified from Biq, 1965; Heezen and Tharp, 1965, 1971).

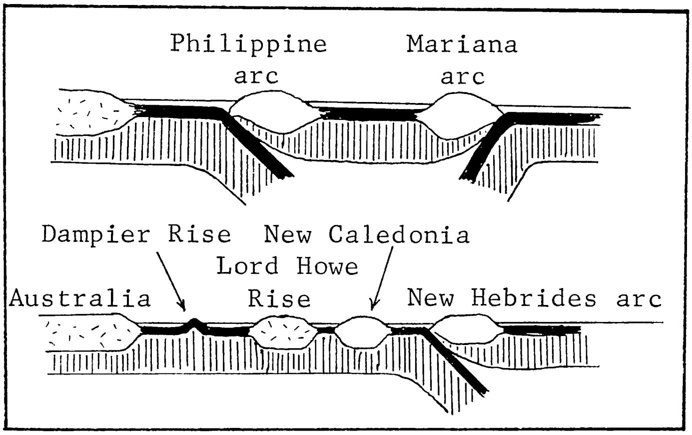

Fig. 10. Schematic sections of Philippine arc and New Hebrides arc to show the pre-continent-island arc collision structure (after Dewey and Bird, 1970).

Since glaucophane schist was found only within the Yüli formation (Yen, 1959, 1963, 1966), which is on the eastern side of the crystalline basement, and the Yüli formation is separated from the other formations by a large fault (the Shoufeng fault, fig. 1), Biq suggests that a subduction zone was formed along this fault during Mesozoic time (Biq, 1971a) and that the dip of the zone beneath Mesozoic Taiwan was westward, toward the continent. However, the K-Ar age of chromian muscovite from mica amphibole schist of the Yüli formation is only 6 m.y. old (Yen and Rosenblum, 1964). Biq interprets the apparently young age as due to burial, giving the time of uplift of the crystalline basement instead of metamorphism. In light of the two previously mentioned K-Ar dates of 33 and 86 m.y. obtained from micas in the pegmatite and quartz diorite, respectively, it seems unlikely that argon accumulation due to cooling began in the chromian mica 80 m.y. after accumulation began in the pegmatite. More probably this date gives the age of the later Plio-Pleistocene orogeny when the K-Ar clock was reset to zero. Furthermore, all the outcrops and the floating boulders of the glaucophane schist are located on the eastern side of Yüli formation, closer to the Longitudinal Valley rather than to the Shoufeng fault (Yen, 1963, 1966). Thus I suggest that the glaucophane schist is probably a product of the Plio-Pleistocene orogeny or at least is strongly affected by this orogeny. Whether this Mesozoic subduction zone existed is not very clear; more detailed field mapping and radiometric dating are required to solve this problem.

Submergence of the Central Mountain Range area probably started in late Cretaceous time or later, and a thick sequence of argillaceous beds was deposited during Paleogene time. This subsidence was probably due to pulling down of the Eurasian plate when this plate started to underthrust below the Pacific plate. A trench was forming on the continental side of the arc. As the arc came closer and closer, sinking of the continental plate was at first rapid and reached its maximum during Eocene time, then gradually decreased in speed because of the buoyancy of the continental rocks. Finally the force pulling the plate down was balanced by the force of buoyancy, the basement was in a steady state, and sedimentation decreased to nearly zero (after the Oligocene). To the west of the Central Mountain Range area the basement was buckling down as its eastern margin was buckling up and a thick sequence of Neogene miogeosynclinal sediments accumulated. Apparently this downbuckling was not very uniform and fluctuations were very common, but no distinct and large-scale angular unconformity was produced.

To the east of the Central Mountain Range area, sediments were still accumulating in a relatively deep trench. When the descending plate reached a depth over 100 km, the amphibolite and quartz-eclogite crust began to melt partially, and calc-alkaline magmas accumulated and rose (Ringwood, 1966; Dewey and Bird, 1970). Probably during early Miocene time, thick masses of andesitic lava flows and pyroclastics accumulated on the ridge of the island arc. When the continental margin finally reached the island arc, the deep trench sediments could not be swallowed into the mantle anymore and started to accumulate. Since this region was highly unstable, the sediments were made chaotic by mass slumping and were poorly sorted; contemporaneous deformation phenomena were also very common.

Since the buoyancy of the continental rocks prevented much underthrusting of the Eurasian plate (McKenzie, 1969), horizontal compression became a dominant factor and low-grade regional metamorphism started during this time (possibly late Miocene to Pliocene time). Since the geothermal gradient was low in this region (geosynclinal environment), metamorphism is characterized by low-temperature, high-pressure greenschist facies. The metamorphic grade increases toward the northeast (Yen, 1962). Glaucophane schist has been found by Yen (1966) near the eastern margin of the Central Mountain Range, near the Benioff subduction zone. Metamorphism decreased considerably both westward and upward; the Paleogene rocks were only indurated, folded, and slightly metamorphosed. Farther west the Neogene rocks were not metamorphosed at all.

When the compression reached a certain limit, the subduced part of the Eurasian plate (which was actually a plate of oceanic character) finally broke from the continental part. This breaking caused sudden release of the downward stress, and the continental plate bounced back by buoyancy, causing the upthrust of the Central Mountain Range along the western side of the Longitudinal Valley. This uplift on the eastern side of the island and shocks during uplifting produced a series of folds, imbricated thrust faults, and even several gravitational nappes which slid toward the west. This tectonic diastrophism probably started at the end of Pliocene or the beginning of the Pleistocene, and the tectonic structures of Taiwan island were formed at this time.

On the east side of the Longitudinal Valley, the "argille scagliose" materials of the Lichi formation were squeezed upward and slid toward the west as mélange (Hsü, 1968). Broken pieces of original oceanic crust (the ophiolite suite) were brought up by this mechanism. At the same time autochthonous molasse (the Pinnanshan formation) accumulated on the eastern flank of the Central Mountain Range and was intercalated with the Lichi formation. Probably right after the tectonic accumulation of the Lichi formation, a second gravitational gliding nappe arrived. This gliding nappe emplaced the thick piles of andesitic rocks and overlying rocks right over the Lichi formation, and the East Coastal Range was thus formed.

The last stage was probably the sinistral movement along the Longitudinal transcurrent fault, which is still active at present. This movement corresponds to the underthrusting of the Philippine basin below the Ryukyu arc. The northernmost part of Taiwan was subsequently underthrust by Philippine Basin plate. Thus the Pleistocene volcanism of northern Taiwan is a very young event and definitely cannot be related to the East Coastal Range volcanism. Figure 11 is a schematic diagram showing the various stages of the tectonic evolution of Taiwan.

Details of the space relations of Taiwan with the surrounding region are very complex, and until now no diagram has shown the relationships. Based on the idea of continent-arc collision tectonics and the data available (Biq, 1965, 1969, 1971b; Allen, 1962; Katsumata and Sykes, 1969; Dewey and Bird, 1970; Wu, 1970; Wageman, Hilde, and Emery, 1970), a schematic diagram is here presented (fig. 12). Judging by the descriptions, the transform fault of the Longitudinal Valley has its own unique tectonic significance. Before the Early Pleistocene orogenesis, it was a subduction zone (or Benioff zone); during the orogenesis it became an upthrust plane (same strike trace, but maybe a different dip); and after the orogenesis it became a transform fault (or wrench fault) with sinistral strike-slip movement.

CONCLUSION

The tectonic evolution of Taiwan is a very challenging problem, because of the unique tectonic features of the island. On the basis of the idea of tectonic collision of an island arc with a continental margin, most of the structural features can be explained in a more or less satisfactory way. If we accept this idea, then the main part of the island formed from a single miogeosyncline. The East Coastal Range shows several characteristics of eugeosynclinal structure, but its area is too small to distinguish it as an individual eugeosyncline. It is more properly explained as a deep trench structure, because most of the geosynclinal sediments have been lost under the subduction zone. Another advantage of this idea is that it greatly simplifies the tectonic evolution of Taiwan, for that evolution was probably all induced by the tectonic evolution along the Longitudinal Valley, which was once a subduction zone, then became an upthrust plane and finally changed to a transform fault that is still active at present.

The Pleistocene volcanism of northern Taiwan seems to be an event separate from the volcanism of the East Coastal Range, for the age gap is too large. The northern Taiwan volcanism can be compared with the

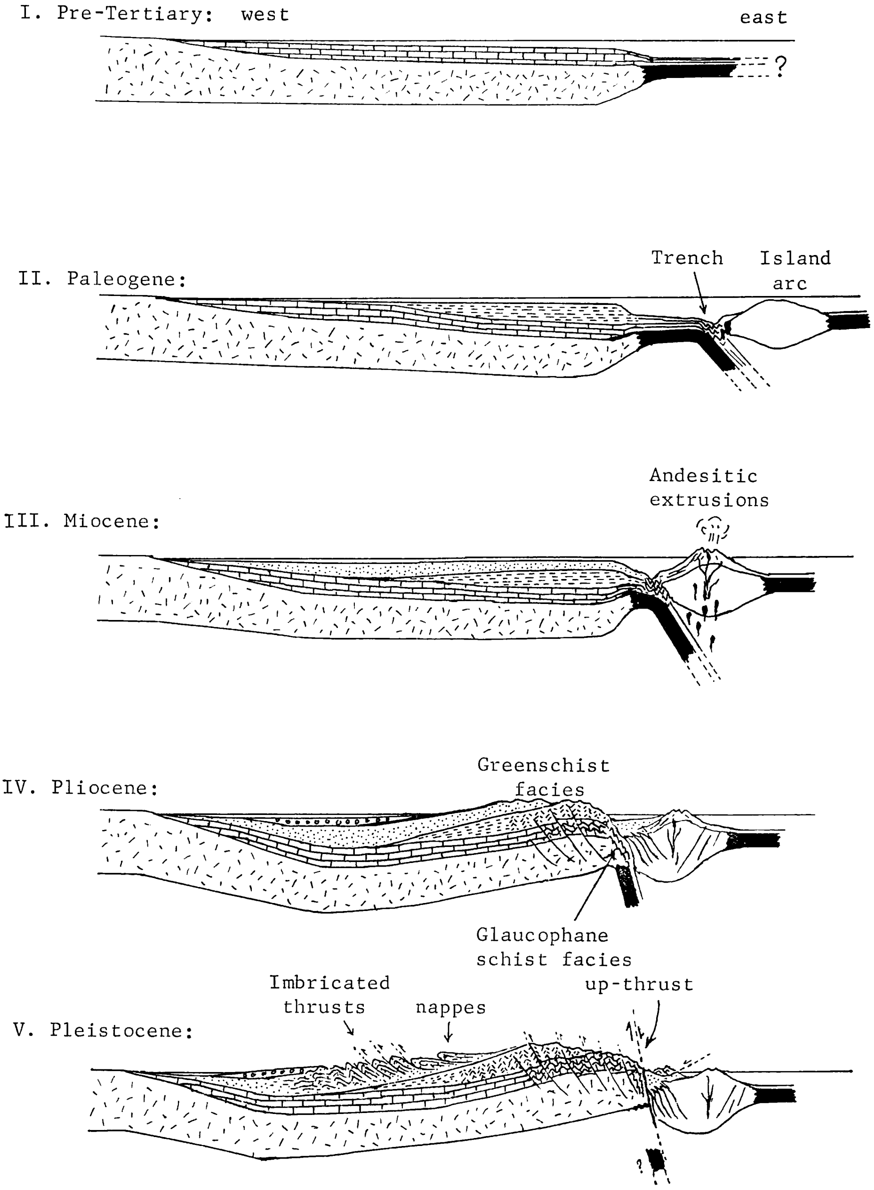

Fig. 11. Schematic sequence of sections illustrating the tectonic evolution of Taiwan by the collision of a continental margin and an island arc.

Circles: Molasse; dotted: Neogene sequence; dashed: Paleogene sequence; wall-pattern: Pre-Tertiary sequence; irregularly dashed: Crystalline basement.

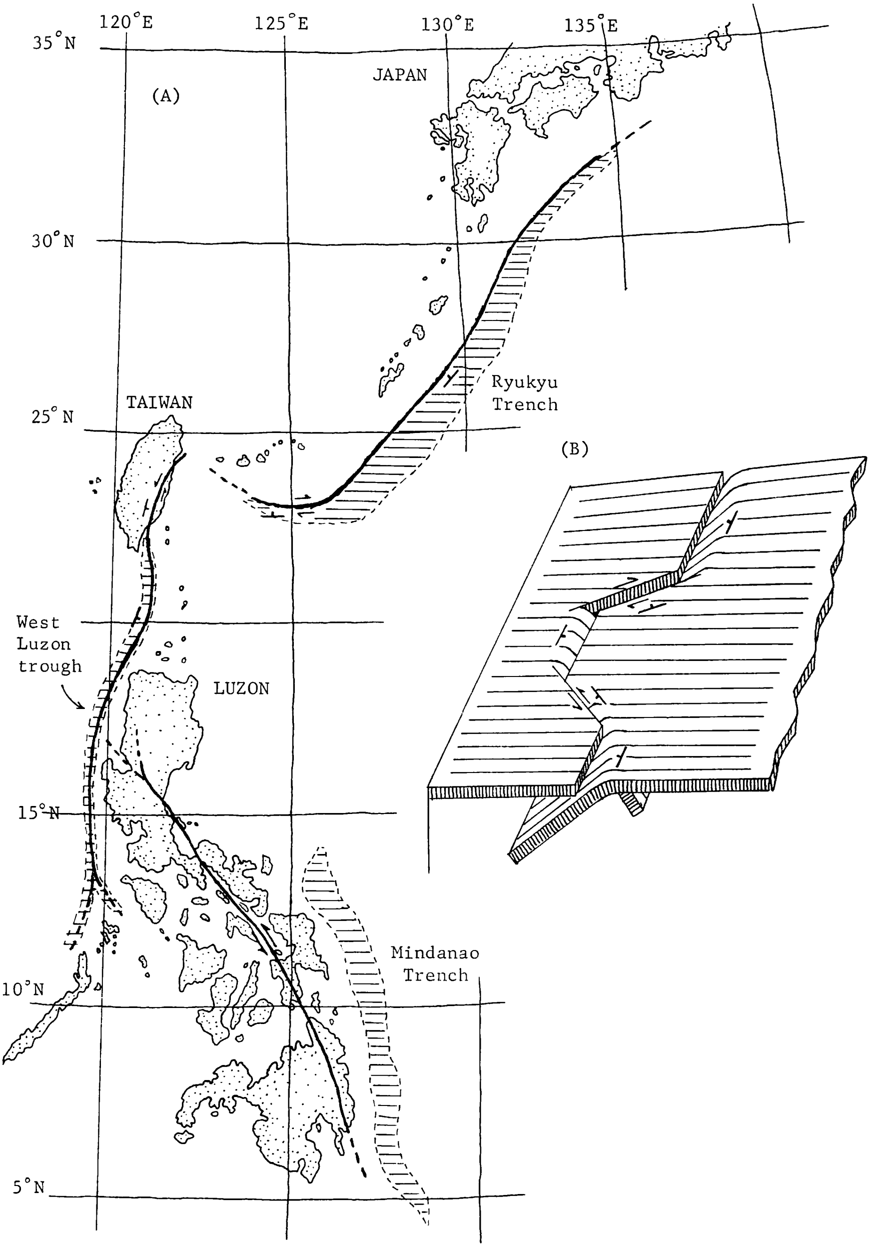

Fig. 12. Schematic tectonic relations of Taiwan and neighboring regions.

A. Map view of the structural relationships; B. Block diagram illustrating the three dimensional configurations. Note: The two transform faults have both strike-slip and dip-slip components.

volcanism of the Ryukyu islands, where the Pacific plate is underthrusting the Eurasian continental plate. This underthrusting started very early in the northern and central part of the Ryukyu arc, probably Miocene or even earlier, but the underthrusting below the northeastern part of Taiwan seems very young, probably after the upthrusting of the Central Mountain Range at the end of the Pliocene, which may have promoted such underthrusting. The underthrusting of the oceanic plate below northern Taiwan is clearly illustrated by the seismic data of Katsumata and Sykes (1969) (figs. 7, 8) and Wu (1970); once the underthrusting started it also pushed the continental plate upward, thus explaining why the crystalline rocks are more uplifted and more broadly exposed to the north whereas to the south they gradually disappear beneath the cover.

Much work is still to be done, especially age dating dating and determination of the facies relations of sedimentation. The rock sequence of the calc-alkaline suite is also very challenging. Trace element distributions may give more information on the magmatic history of Taiwan island. The mafic blocks in the Lichi formation are characteristically associated with zeolite minerals, very similar to those in the ophiolites in the "argille scagliose" of the northern Apennines in Italy (Abbate and others, 1970a), indicating that low-grade metamorphism of prehnite-pumpellyite facies occurred in the gabbro, diabase, and peridotite. Periodotite is quite commonly serpentinized, and chrysotile and lizardite are common polymorphs of serpentine. These ophiolites probably represent broken fragments of original ocean crust.

The picture of tectonic evolution of Taiwan presented here is more or less speculative because of lack of data. More study is required to refine and correct this preliminary report.

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