EVOLUTION OF ARC SYSTEMS IN THE WESTERN PACIFIC

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INTRODUCTION

Very shortly after their discovery arcuate island chains, with their associated volcanoes, trenches, and seismicity, were recognized as critical tectonic complexes in some way related to the development of orogenic zones (e.g. 102). However, technical inability to adequately investigate these primarily submarine features, and even worse, the often misleading conclusions drawn from the few segments above water, delayed an adequate understanding of island arc systems until after the Second World War. Since then, high precision echo-sounders, low-frequency acoustic profiling, and better bottom sampling apparatus have clarified arc system morphology and shallow structure. These observations, coupled with seismological and other geophysical studies, have now directly and indirectly indicated the role of island arc systems in the orogenic process and in global tectonics.

Differences in opinion exist concerning many, or most, details of arc system character and evolution, but most workers in western Europe, Japan, and the Americas accept the underlying concept of plate tectonics. In this framework, island arc systems result from the underthrusting, or subduction, of one lithospheric plate beneath another. The following discussions assume the validity of plate tectonics and concentrate on processes by which subduction is manifested in the geologic record. Different concepts of island arc evolution, which deny the concept of subduction, are discussed by Meyerhoff, Meyerhoff & Briggs (74), Carey (13), Beloussov (7), and others.

This paper reviews recent data and the resulting ideas which bear on evolution of arc systems in the western Pacific. Results and conclusions from arc systems in other regions have been introduced where pertinent. Geological and crustal geophysical aspects are stressed, and there is no attempt to review either data or theoretical studies from the subcrustal regions beneath the arc systems, or petrologic studies of island arc igneous activity.

STRUCTURAL FRAMEWORK OF AN ARC SYSTEM

Unfortunately, island arc terminologies which have been presented over the years have not reflected the uniformity of morphology and structure which has generally

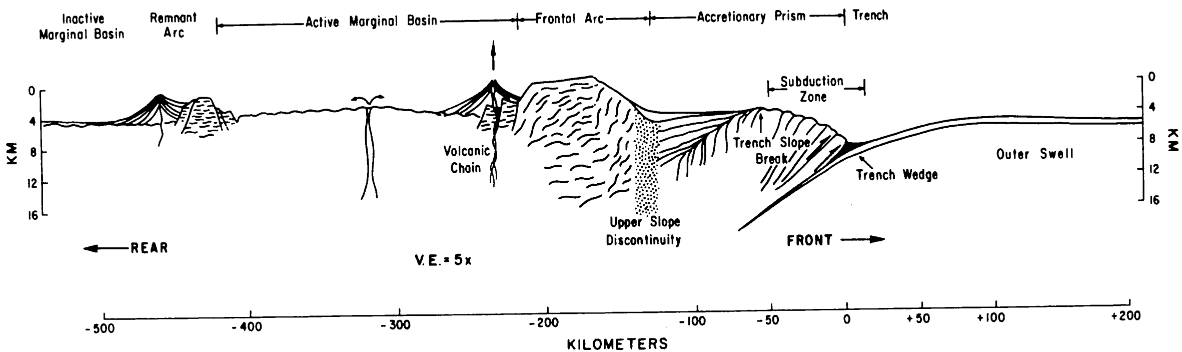

Figure 1 Generalized framework of a western Pacific island arc system after Karig, Le Pichon & Sharman (57) showing the terminology used in this paper.

been observed in arc systems. For instance, the Indonesian terminology as presented by Vening Meinesz (108) called for an outer or tectonic arc between volcanoes and trench and could not be applied successfully to the Tonga system. Although there are islands trenchward of the Tongan volcanoes, these are part of the volcanic arc as defined in Indonesia, and the tectonic arc is deeply submerged and poorly developed.

A new terminology (Figure 1) has been developed from recent geological and geophysical studies in the western Pacific, which employs generalized and descriptive terms although some have genetic connotations. A fully developed island arc system can be divided into three sections across its trend. On the forward side of the system is the inner slope of the trench, which apparently overlies the area of active subduction and accretion. Behind the volcanic chain an extensional terrane may form in which previously developed sections of the arc system are separated from the active system. Between them lies the frontal arc, a relatively passive block of older, thicker crust which undergoes primarily vertical displacements, although important plutonic and metamorphic processes may occur at deeper levels.

This subdivision stresses the processes of arc system growth and destruction. Differences among the arc systems of the western Pacific stem from variations in the subduction process along the leading edge of the system and from the degree to which extension has operated along the trailing edge. Confusing this pattern are secondary effects of oblique subduction, collisions, and arc polarity reversals.

THE LEADING EDGE AND SUBDUCTION PROCESSES

The slope leading down into the trench axis from the frontal arc is one of the most complex areas within an arc system, showing broad variations in morphology, lithologic content, and structural style (Figure 2). It can generally be divided into an upper, undeformed, and sediment-covered section and a lower, steeper slope underlain by deformed sediments or acoustically opaque material. The boundary between the two segments can be expressed as a change in slope, a shelf or terrace edge, or a discrete ridge. This range in aspect has led to a corresponding range in descriptive terms, including tectonic, outer, or nonvolcanic arc, midslope basement high, and trench slope break.

Variations in the inner trench slope, rather than being random, can be reasonably correlated with differences in supply of sediment to the subduction zone from the few major sources. In arc systems where a plate carrying a relatively thin pelagic sediment cover is being subducted and where there are few sources of terrigenous sediment, the inner trench slope is in two sections separated by a simple slope break (19, 47, 56). Good examples of this type are the Tonga, Kermadec, and Mariana trenches. Very limited dredging of the lower slope has recovered predominantly igneous rocks (26, 35, 87). Pelagic constituents have not been reported, but their selective removal during dredging might be expected. The single seismic refraction profile over this variety of trench slope (83) indicated that it is underlain by a crust having seismic velocity units similar to those of oceanic

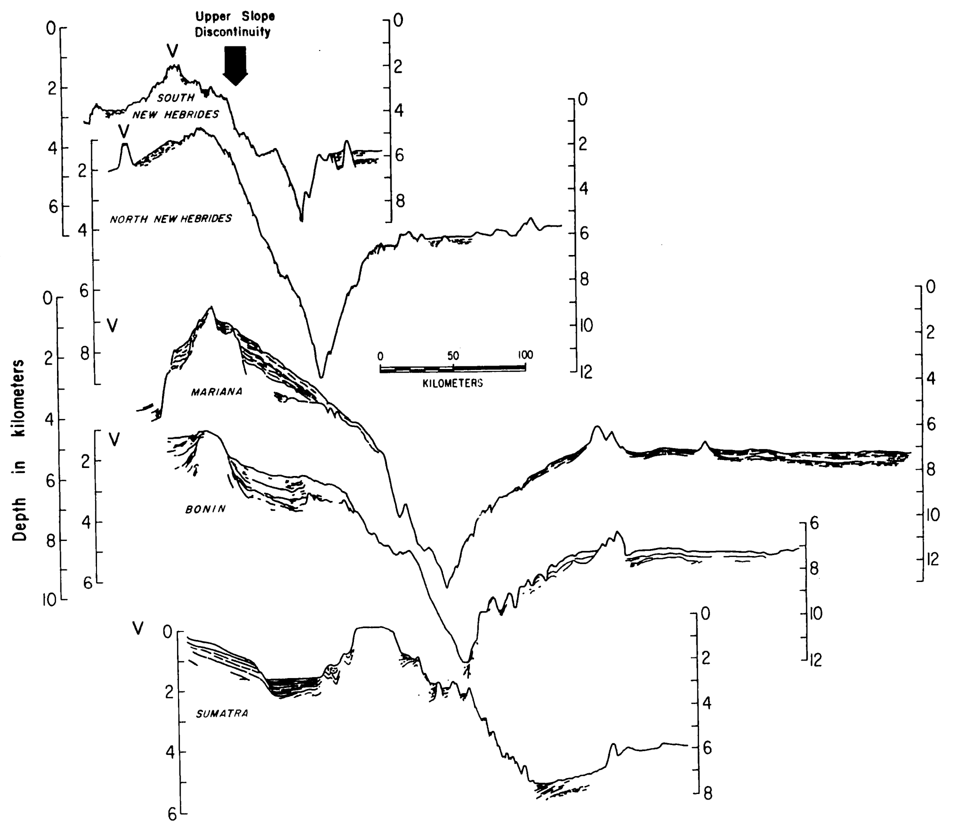

Figure 2 Seismic reflection profiles across inner trench slopes of arc systems having varying ages and sedimentation characteristics (57). These profiles are normalized to the upper slope discontinuity in order to show the wide range in width and shape of the accretionary prism.

crust, but somewhat thickened. On gravity profiles over these trenches, the free air minima nearly coincide with the topographic axis (103, 104, etc). These data suggest that material underlying the inner trench slope is composed of oceanic crustal components, although probably in a structurally disordered state.

With a second variety of inner trench slope, a thick sediment section is fed to the trench, but relatively little is derived from the frontal arc near the point of subduction. The sediments represent either the cover on the downgoing plate or a turbidite wedge in the trench axis which has been transported along the axis from a distant source. The Central Aleutian, Luzon, and Sumatra arc systems represent this style, which displays a strongly developed ridge between the upper and lower slope sections. The distinct trough behind the ridge results from inability of sediment sources on the frontal arc to keep pace with accretionary upbuilding of the trench slope break.

Exposures on this ridge form of the trench slope break (57), together with bottom sampling (110) and seismic refraction profiles (65, 90), indicate that the lower slope, trench slope break, and much of the upper slope trough are underlain by a basement of deformed sediments. The strong landward displacement of free-air minima on gravity profiles across these arc systems also suggests low-density, sedimentary basement (30, 103).

In arc systems where terrigenous sediment influx from the frontal arc is high, the upper slope area is kept filled, forming a shelf or terrace. Often sufficient material passes across the upper slope and into the trench to form a turbidite wedge which effectively supplies the subduction zone with abundant sediments, regardless of the nature of sediments on the downgoing plate. In this case, well represented by the eastern Aleutian area, basement characteristics are similar to those of the second variety (109). In other cases, as in the Japan arc, a smaller turbidite contribution to the trench and thin pelagic sediments on the downgoing plate produces a denser basement which may contain igneous rocks (66, 103).

These categories of inner trench slope are artificial in that all transitions can be found. It should also be noted that the morphologic style and lithologic content of the inner trench slope is independent of its size (57), which closely reflects the duration of subduction and absolute feed rates to the subduction zone.

Structural Style of Accretion

Structures within the accretionary prism which result from subduction have only begun to be investigated. A single Deep Sea Drilling Project (DSDP) hole (110), a few good reflection profiles, reconnaissance geology of the trench slope break on several islands, and a hazardous transfer of observed structures from orogenic zones to their original setting (e.g. 80, 81) constitute the bases for several available models.

The stripping off of sediment cover from the downgoing plate in fold or thrust units and its accretion to the trench wall is suggested by some seismic reflection profiles (6, 14, 99, 100, 109). Where they are next observed, at the trench slope break, the sediments are isoclinally folded with steeply dipping axial surfaces. Most of this deformation is thought to occur on the lowermost trench slope, because sediments usually become acoustically opaque at the base of the slope and because DSDP site 181, part way up the lower slope, recovered highly deformed sediments (110).

The structural process by which material is transferred from oceanic to arc plate probably varies according to the thickness and type of sediment in the trench and also the rate of subduction. In trenches where a turbidite section, deposited in the trench, overlies pelagic or hemipelagic sediments, deposited on the downgoing plate, the turbidites may be systematically removed and accreted while the rest of the section may be underthrust to deeper levels (57). Shearing should be localized in the top of the pelagic section because these sediments have minimum strength and maximum porosity (31). Bedding thrusts, aided by expected high pore pressures would place highly deformed turbidites over the surficial turbidites on the trench floor (Figure 3a), and would explain the observed sharp juxtaposition of very opaque acoustic basement and undeformed sediments at the base of the inner trench wall (30). Slow subduction appears to favor a greater degree of folding in the accretionary process (100; M. S. Marlowe, personal communication).

The mode of deformation in subduction zones involving downgoing plates with thin pelagic covers is more difficult to model. The weak, high porosity cover could again be utilized as a zone of shear, and all but the uppermost sediments might be carried deep below the inner slope of the trench. However, growth of the inner slope area with time in this variety of subduction zone (57) indicates that pieces of oceanic crust and even upper mantle are sheared off the downgoing plate (Figure 3b). This might occur intermittently when large-scale topography, such as one of the normal fault scarps which develop on the flexed outer trench slope (66), or a seamount, impinges on the inner trench slope.

Apparently deformation is most intense at the base of the inner trench slope but continues, at a decreasing rate, to the trench slope break, where the greatest cumulative effects are seen. Sediments of the upper slope which overlie the trench slope break are openly folded and moderately faulted (30, 37, 57, 67).

This deformation and limited geodetic measurements on pertinent islands demonstrate uplift of the trench slope break relative to both sea level and the upper slope area (48, 110). There is little or no evidence to support earlier views of major subsidence in the area of the trench slope break (39, 82).

On the other hand, there is substantial evidence that the upper slope area is subsiding relative to the frontal arc, to the trench slope break, and often, to sea level. Many reflection profiles show faulting or folding on both flanks of the upper slope sediment pile (e.g. 37, 48, 57, 67) and often reveal a downward increase in displacement or in rearward tilt (e.g. 91, 110). The sharp contrast between this zone of subsidence and the emergent frontal arc (e.g. 28, 89) thus requires that a structural discontinuity separate the two units.

The boundary between the two regimes has been designated the upper slope discontinuity (57). In part, the upper slope discontinuity appears to represent the upper continental or insular slope which existed before a pulse of subduction began. With time, an accretionary prism builds the lower slope outward from the remnant upper slope and sediments fill in behind and over a basement of accreted material, but there is also demonstrable faulting and flexure along the upper slope discontinuity (30, 47, 110), especially in the broader accretionary prisms. Other than being steeply dipping and downthrown toward the trench, however, the nature and origin of the faults is not yet known.

Available geodetic and structural evidence can be combined with bathymetric profiles which show changes in the width of the inner trench slope with time to produce a reasonably satisfactory growth model of subduction zones. Initiation or

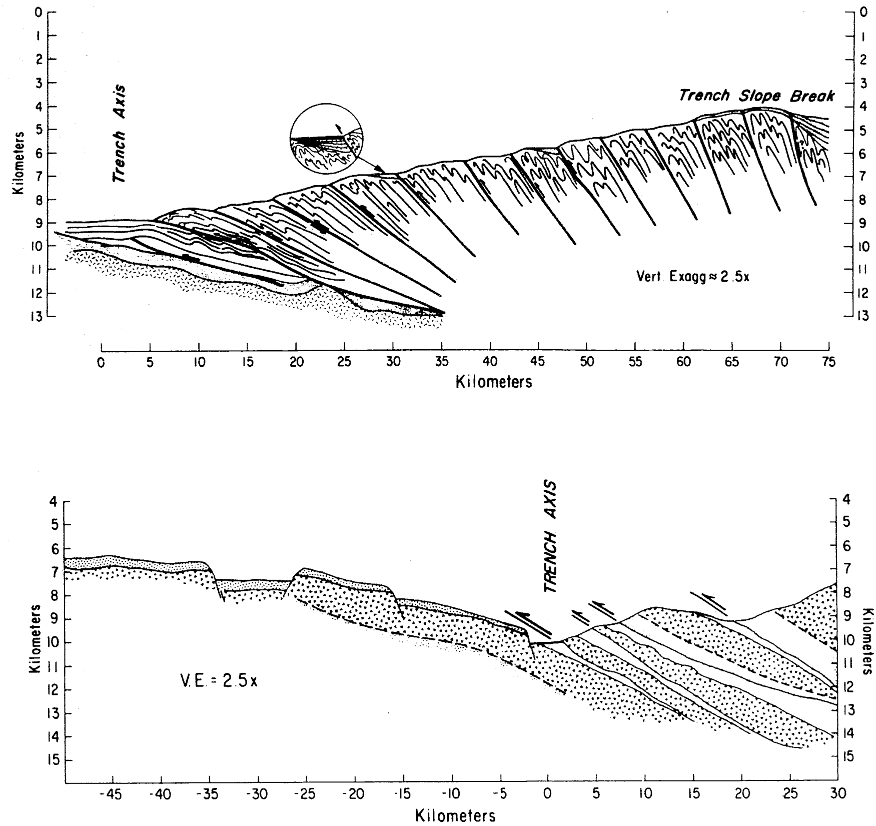

Figure 3 A. Possible subduction mechanism where turbidites are accreted to the trench wall and the pelagic sediments and basement carried to deeper levels. Bedding plane step thrusts sided by high pore pressures in the weak pelagic sediments produce the observed juxtaposition of highly deformed sediments next to and overlying undeformed turbidites. B. Possible subduction mechanism where a very thin sediment cover is involved. Intermittent accretion of basement slabs, intercalated between highly deformed pelagic sediments, is postulated.

rejuvenation of subduction produces a trench near the foot of a pre-existing continental or insular slope. If a significant rise sedimentary apron is present, it is uplifted and deformed into an initial accretionary prism, as in the Shikoku arc system (37, 57).

As material is accreted, either in fold packets or in slabs, previously accreted material is lifted up the lower slope to form the crest of the trench slope break. Then, however, the material subsides and the trench slope break moves forward (Figure 4).

Uplift is attributed to the effects of underplating and internal deformation within the accreted material beneath the lower slope. Subsidence of the upper trench slope is apparently a reflection of the depression of the flatter, upper section of the Benioff zone by the weight of accreted material (57). The cause of the change from subsidence to uplift across the upper slope discontinuity is unknown, but permanent uplift of the frontal arc has been attributed to flexure or irreversible effects of compression (28).

During growth of an accretionary prism, or of several such prisms (11), the volcanic chain remains fixed, or moves slowly to the rear, relative to the crust below

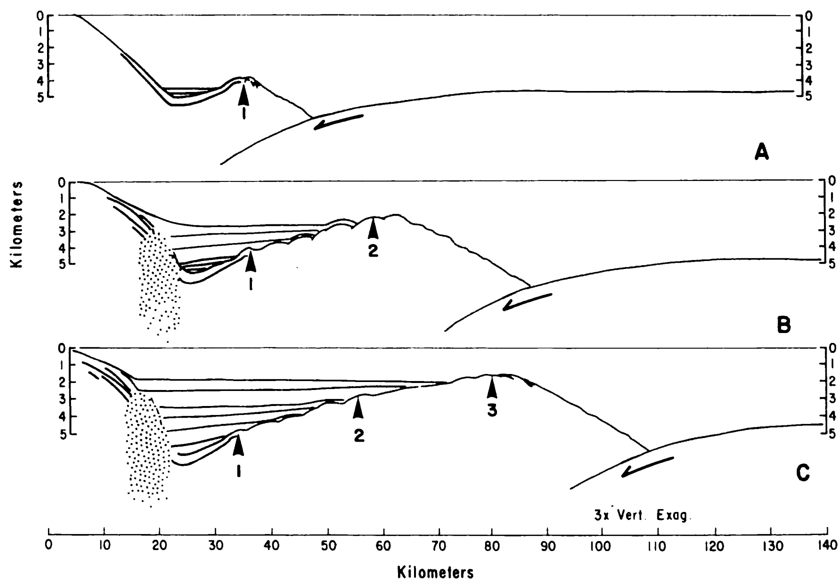

Figure 4 Model of accretion along arc system where a moderate to thick sediment cover occurs (57). Accreted material rises to occupy the trench slope break, then subsides to form basement for upper slope sediments. At the same time, the trench slope break is migrating outward.

(21). This leads to an increasingly broad volcano-trench separation and, because the Benioff zone remains at a fairly stable depth beneath the volcanoes, to a very broad, flattened upper Benioff section (30, 32, 57, 62). The trenchward overstepping of the low P/T metamorphic regime over the high P/T zone, as proposed by Matsuda & Uyeda (72), Oxburgh & Turcotte (85), and others, appears to require a different process; either a longer term, discrete shift of the volcanic chain and Benioff zone, or some form of tectonic disruption.

Tectonic erosion rather than accretion has been suggested to explain the igneous rocks and the morphology on the walls of many western Pacific trenches (30, 82). Growth of the accretionary prism with time in these arc systems and volume calculations indicating the need for basaltic crust within the accretionary prism (57) obviate the need for such erosion.

THE TRAILING EDGE AND CRUSTAL EXTENSION

Behind the volcanic chain in an active island arc system lie one or more marginal basins, whose character and origin are still subjects of debate. Some marginal basins are volcanically and structurally active; these always lie directly behind the volcanic chain of the arc system and are called inter-arc basins (47), or active marginal basins (49). The few investigations in the small number of active marginal basins in the western Pacific have revealed only their general characteristics.

Pillow basalts are erupted, without the formation of central shields, along the axial zones of the active Mariana (33, 48) and Lau (97) basins. In these larger, active marginal basins the axial zone is elevated relative to the basin flanks (50, 51), but shows no clearly defined central rift or axial symmetry (Figure 5). The locus of active volcanism is restricted by the sediment distribution to the central 40 km or so, but there are no data delineating the extent or form of volcanic centers within this band. Fissure eruptions have been suggested (48). Basin morphology consists of linear ridges and troughs which most likely reflect tilted fault blocks, perhaps modified by volcanism. No transform faults have yet been recognized, but this may be due to lack of data.

Although there is a remarkable dearth of sediment cover in active marginal basins, the sedimentation rates are very high. Volcaniclastic aprons, building out from the volcanic chain, accumulate at rates exceeding 100 m/m.y. (12, 25, 48). Montmorillinitic clays and finer volcanic detritus are being hemipelagically deposited further away from the volcanoes at rates decreasing with distance (29, 47, 48). Foraminiferal and calcareous nannofossil oozes are the dominant sediments furthest from the volcanic chain at active marginal basins having no continentally derived sediments. In these basins the overall sediment pattern is one of thicker deposits along the basin flanks, especially on the flank adjacent to the volcanoes, and thinnest deposits in the axial zone. In active marginal basins which lie adjacent to continental margins, as the Okinawa Trough (111), thick sediment sections may be involved in faulting and volcanic activity along the axial zone.

Fault scarp systems generally bound active marginal basins on both flanks,

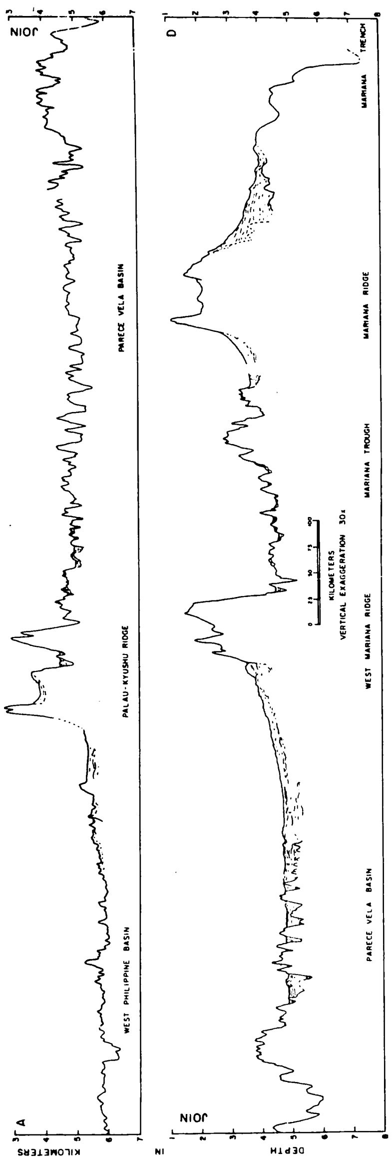

Figure 5 Seismic reflection profile across the Mariana arc system showing the character of the active marginal basin (Mariana Trough) and inactive basins to the west (48).

although the volcanic chain often conceals the scarps on the forward side. The rear flank forms part of a submarine ridge, which is defined as the third arc (107) or, in more general terms, a remnant arc (50). In the simplest examples, remnant arcs appear to represent the rear flanks of frontal arcs, and display primarily volcanic chains and volcaniclastic aprons (Figure 5). In other, more complex situations, remnant arcs may represent entire arc systems which have been abandoned or which have collided with other remnant arcs (50). Some remnant arcs are relatively wide and display broad, flat tops, while others are little more than truncated volcaniclastic aprons. Where more fully developed, they show a basement core, fault bounded on both sides, with the rear faults modified by volcanic activity. There is accumulating evidence that remnant arcs have subsided up to several kilometers from sea level or from much shallower depths (50, 94).

Inactive marginal basins, separated by remnant arcs and lying behind the active basin, if one is present, show basement morphology similar to the active basins (Figures 5 and 6). In most older basins, however, a thick sediment cover obscures the basement. This sediment cover ranges from a few hundred meters of pelagic muds in basins protected from continental sediment sources, like the Philippine Sea, to several kilometers of terrigenous turbidites where a continent forms a basin flank (73, 98).

The depth to basement in marginal basins, after the effects of gravitational loading by sediment are removed, increases with age from approximately 2.5 km in active basins to more than 5.5 km in older basins (49). This increase in depth with age, accompanied by a fall in crustal heat flow values from over 3 HFU to normal (e.g. 96) is qualitatively similar to that observed in the main ocean basins.

Although this similarity suggests that both marginal basins and major oceanic basins are generated by a similar process, there are minor, but perhaps significant, differences in their crustal and lithospheric character. Marginal basins of the Philippine Sea and several other regions are deeper than oceanic basins of the same age and some have higher heat flow at both the equivalent age and depth (96). This may be attributed to a thinner than oceanic lithosphere, as suggested to exist beneath the Philippine Sea (46), which might result from sublithospheric shear and erosion (96). Most of the difference in depth could also be accounted for by thinner than oceanic crustal layers which are suggested by seismic refraction studies in the arc system.

Identification of the basaltic basement as a product of extensive volcanic activity occurring after the basin was originally trapped has been postulated by Scholl et al (93) for the Kamchatka Basin and by Edgar, Ewing & Hennion (23) for some Caribbean basins. The idea that marginal basins originated through the subsidence of continental crust (8, 58, 105) has largely been abandoned. Generation of island arc systems from transform faults has been suggested in some areas (56, 106).

Very few data other than drill holes of depths beyond the present capability of the Deep Sea Drilling Program will unambiguously settle the problem. Closely controlled seismic refraction profiles can place limits on the possible thickness of a basaltic cover and masked sediments which might overlie normal ocean crust. Discrimination between trapping and extensional origins might be made by looking

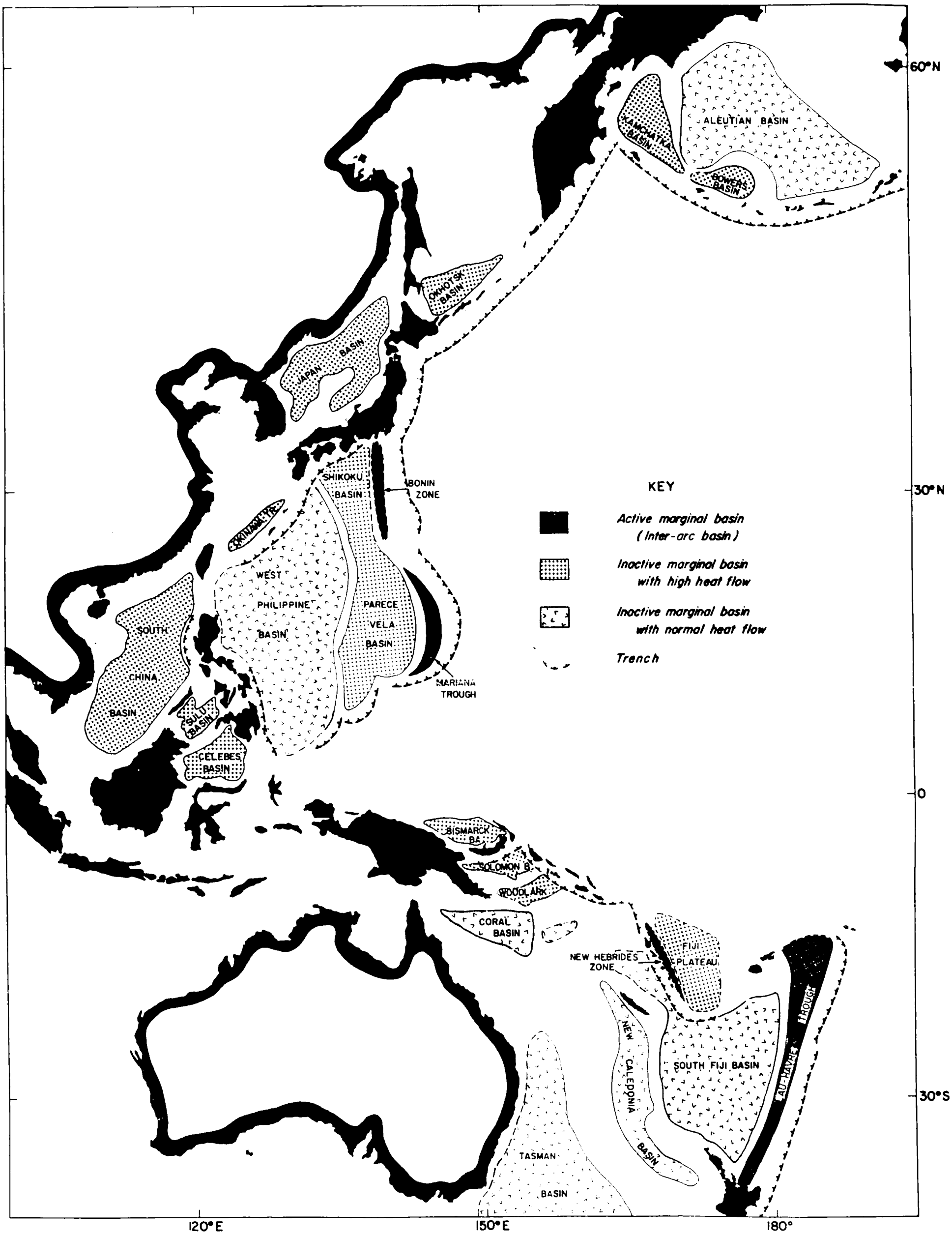

Figure 6 Distribution of marginal basins in the western Pacific, after Karig (49) showing the classification based on increasing age, depth, and crustal heat flow.

for systematic changes in basement age which would result from axial crustal spreading.

Magnetic anomalies in marginal basins do not show well-developed, symmetrical linear patterns which can be correlated with those generated at oceanic rises. Linear anomalies do exist, but in most cases they are not coherent or identifiable. Intensive surveying of the Sea of Japan (41) reveals low amplitude (200 gamma peak to peak), high frequency nonsymmetrical anomalies in comparison to those of the Pacific Ocean east of Japan. Murauchi (82) reports similar low amplitude anomalies, often averaging less than 100 gamma, in the Philippine Sea.

If the magnetic anomalies of marginal basins and oceans are to be compared correctly, however, the remnant fields must have been generated at the same time along spreading ridges with similar orientations and magnetic latitudes. On this basis, the data are ambiguous and scanty. Spreading in both the active Mariana basin (48) and the East Pacific Rise off Central America (36) is producing anomalies with 200–300 gamma amplitudes. In contrast, anomalies of late Oligocene age in the Kamchatka basin (60, 94) have amplitudes only about half that of anomalies generated under similar conditions along the Mid-Atlantic Ridge. Spreading rates, which were not considered in these comparisons, may affect the anomaly amplitude in marginal basins.

While magnetic lineations have often been noted in marginal basins, there are relatively few cases where symmetry or identification of these lineations has been claimed. Luyendyk, MacDonald & Bryan (69) and Ben Avraham, Bowin & Segawa (9) report identifiable anomalies in the Woodlark basin near New Guinea and the West Philippine Basin, respectively. A much more convincing set of anomalies, out to anomaly 3 (5 m.y.) has been reported from the Lau Basin (J. G. Sclater, personal communication), but this organization breaks down further from the spreading axis.

The apparently different magnetic field characteristics in marginal basins and mid-ocean ridges might result from differences in the magnetic properties of the extruded basalts, from degradation of remnant magnetism in marginal basin basalts by injection into thick sediment covers or by hydrothermal alteration (M. Marshall, personal communication), or from a different process of crustal extension.

Basalts from the Mariana basin, however, have as high remnant magnetizations and Konigsberger ratios as do mid-ocean ridge tholeiites (M. Marshall, personal communication). Furthermore, there is very little sediment cover near most of the extensional zones of active marginal basins. Effects of hydrothermal alteration are unknown but, to date, thermally altered basalts are uncommon in marginal basins and have been dredged only from fault zones. A mode of crustal extension in marginal basins different from the passive spreading at mid-ocean ridges emerges as the most reasonable cause for the differences in magnetic field characteristics.

The underlying mechanism responsible for the young crust, high heat flow, and apparent extensional tectonism in marginal basins is presently the focus of considerable attention. Oxburgh & Turcotte (85), Hasebe, Fujii & Uyeda (34), Karig (49), and others favor a rise of heated mantle material from the shear zone

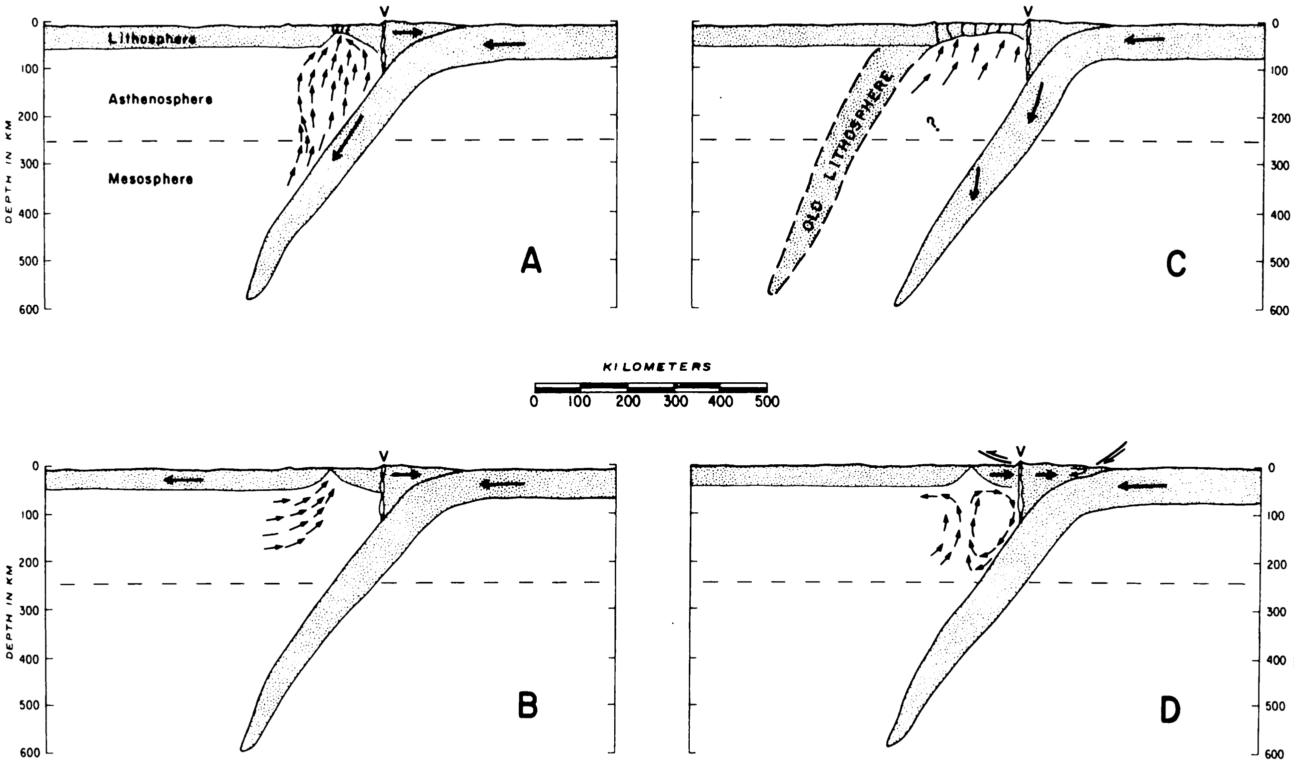

Figure 7 Hypothetical models advanced to explain the young crust and high heat flow in marginal basins. A. Active diapirism resulting from heat and/or water generated along the Benioff zone. B. Passive diapirism in response to regional extensional stress in the lithosphere. C. Stepwise migration of arc system with collapse of the oceanic crust behind the new arc (45). D. Subsidiary convection cells driven by drag along the Benioff zone (38, 101).

along the top of the downgoing plate (Figure 7A). A variant of this model, suggested by studies on the role of water in the mantle (114), would have water, derived from the downgoing plate, cause partial melting and diapiric rise of the immediately overlying mantle. Either stated or implied in both schemes is the active nature of the diapirically rising mass, which would cause extensional features at the surface but produce and transmit compressional stresses at deeper levels of the lithosphere.

A more passive upwelling of asthenospheric material into an extensional zone generated by some sort of plate rearrangement or by cessation of subduction has been suggested by Packham & Falvey (86), Scholz, Barazangi & Shor (95), and Kanimori (45) (Figure 7B, C). Hydrodynamic schemes which utilize the viscous drag along the Benioff zone to drive asthenospheric convection cells (e.g. 101) have also been proposed (Figure 7D).

Hydrodynamic schemes, which recirculate rather than increase the mass of material beneath the island arc region, find little support in geologic data. Although Sleep & Toksöz (101) cite a compressional focal mechanism behind the Tonga arc, other workers have found no similar earthquakes and there is no geologic evidence of sufficient crustal shortening within the arc system to balance the extension in the inter-arc basin. Moreover, no explanation for the seaward migration of the arc system or the increase in volume is offered.

The studies of seismic energy attenuation and the Q structure beneath island arc systems hold great promise for discrimination among possible mechanisms. There already appears to be a close correlation between low Q (high attenuation) zones and active or recently active marginal basins (4, 77, 79; M. Barazangi, personal communication). Preliminary results indicate that the forward edge of the low Q zone, near the volcanic chain, is steep and fairly sharp (4, 77) and that the rear edge is probably also quite steep. These steep boundaries, and the continuity of the low Q zone to the Benioff zone at depths greater than 250 km are difficult to explain in the hydrodynamic models.

The zone of high attenuation can better be explained by diapiric rise, either active or passive. With passive upwelling, the adiabatic rise of asthenosphere up a relatively steep geothermal gradient could cause partial melting, but it is doubtful whether the low Q zone produced would be as steep or extend as deep as observations indicate.

General occurrence of extension within arc systems during rapid subduction (49) rather than following cessation of subduction is further evidence in favor of active diapirism. Only in the Scotia arc system does extension appear to be associated with slow subduction (5). However, there is no low Q zone observed beneath this postulated extensional zone (M. Barazangi, personal communication), which might suggest an origin different from that of western Pacific marginal basins.

Although the extension in the Mariana trough might be related to subduction of the Philippine plate into the second series of trenches, this type of passive extension cannot be generally applied. Overlap of the Tonga and New Hebrides trenches and the geometrically required extension has been suggested as the reason for the Lau and New Hebrides inter-arc basins (4). However, the Lau basin is only the northern end of an extensional zone which continues into New Zealand. In this part of the arc, as in the Japan, Ryukyu, and other arc systems, the frontal arc and trench must be forced forward, increasing the distance between trench and continent.

If rise of mantle material beneath active marginal basins is the cause, rather than the effect, of crustal extension, then low Q zones should be observed where the hot, low-density material has not yet forced its way into the lithosphere to cause extension. Large-scale uplift and silicic volcanism in the Central Andes (84, 88, 92) are hints that mantle diapirism exists below that area. Seismological results are conflicting. James (42) finds no low Q material beneath the volcanic chain and Altiplano, but both the earlier study of Molnar & Oliver (79) and more detailed recent work (M. Barazangi, personal communication) indicate a substantial zone of high attenuation.

Major problems with an active diapir model are that the rate of crustal extension in active basins implies too great a rate of supply of mantle from the Benioff zone (49) and that the zone of extension apparently remains in the center of the active basin, which would be far behind the Benioff zone and volcanic chain in the larger marginal basins.

The most acceptable model calls for collection of heated mantle below the lithosphere behind rapidly subducting arc systems. Whenever the hydrodynamic forces of the diapir overcome regional compressive forces, the material would rise into the lithosphere, and cause crustal extension. Cessation of subduction would be only one example of reduction of regional compression. Locus of upwelling, strongly controlled by existing weakness zones, would begin near the volcanic chain and migrate away as new lithosphere was symmetrically generated in the inter-arc basin.

Active diapirism may be sufficiently different from passive diapirism to produce the variations in tectonic and magnetic characteristics between marginal basins and ocean basins. Moreover, the transition from subduction-related extension to inter-plate extension in the Gulf of California (55) should serve as a warning that extension begun by active diapirism might be continued by other mechanisms.

Still to be satisfactorily explained is the discontinuous nature of extension in marginal basins, which produces the series of basins and remnant arcs. The direct correlation of extension with subduction pulses (49) is too simplistic, and cessation of extension might better be related to exhaustion of available mantle material, perhaps by reduction of subduction rates. Sleep & Toksöz (101) suggest a cyclic process where extension causes flattening of the Benioff zone which, in turn, causes resistance to extension and cessation of spreading.

Analogous crustal extension within continental margin arc systems or island arc systems with very broad frontal arcs of continental nature seems to be expressed in volcano-tectonic rift zones (49, 55). Geologic syntheses in these zones (2, 49, 70) indicate that extension, in the form of basin-range faulting, is preceded by widespread eruption of ignimbrites. Silicic magmas apparently result from melting within the lower crust (24), which in turn reflects a rise in geothermal gradient. Active diapirism above the Benioff zone is a reasonable source for the excess heat required. In island arc systems, where there is a thinner, more mafic crust, this early volcanism occurs as much smaller eruptions of volatile-rich dacite and rhyolite (54). Following the rupture of the lithosphere in the volcano-tectonic rift zones, bimodal eruptions of both silicic and basaltic volcanic rocks takes place (64).

At least some island arc systems have developed from continental margin systems by evolution of volcano-tectonic rift zones into marginal basins. The Japanese arc is the most obvious example. The spectrum of arc systems around the Pacific suggests a process by which a broad strip of continental crust is intially separated from the continental core. Continued subduction and pulses of extension split this broad arc and create additional basins. The long-term result is to preferentially remove the continental crust from the rear of the frontal arc and leave it behind in the older remnant arcs (51) and to accrete new crust, with oceanic affinities, along the forward edge.

One of the most intriguing of current problems is the nonuniform geographic distribution of marginal basins. Rapid subduction may be a necessary prerequisite for extension, but it is not sufficient, because neither active nor inactive marginal basins exist along the west coast of South and Central America. In some of these areas volcano-tectonic rift zones or low Q zones occur, but since late Mesozoic, marginal basins have been developed most dominantly in the western Pacific region. Prior to that time, there is little evidence for marginal basins in that area and growing evidence of late Paleozoic-early Mesozoic marginal basins along the west coast of North America (e.g. 15).

Over the last 50 years, a number of proposals link the formation of marginal basins with a general westward shift of the crust or lithosphere over the deeper levels. Recent proposals of relatively fixed hot spots (plumes) has led to quantitative analyses of plate motions over the asthenosphere. From one of these, Wilson & Burke (113) deduced that marginal basins develop behind subduction zones flanking plates which are stationary or slow-moving relative to the asthenosphere. An important test of the validity of this and other models is to determine the distribution of marginal basins among the various subduction zones with time and to compare this distribution with plate motions deduced from hot spots.

SECONDARY TECTONIC PROCESSES

The effects of oblique subduction, collisions, and polarity reversals are considered secondary to accretion, extension, and associated processes. They, and perhaps other tectonic effects not yet recognized, are responsible for the development of varied and complex compressive plate boundaries.

Where the trajectory of the downgoing plate is not perpendicular to the trench, oblique subduction occurs. Very often in such cases, the strike-slip component of subduction is not accommodated in the trench, but occurs within the arc system, most commonly within active extensional zones (27, 56) (Figure 8). Longitudinal shearing may combine with extension to produce en-echelon extensional structures (as in the New Hebrides, Tonga, Bonin systems) or less commonly occur as discrete strike-slip faults (as in the Taupo Graben of New Zealand). Strike-slip faulting initiated in this manner may have continued after extension ceased in the Proto-

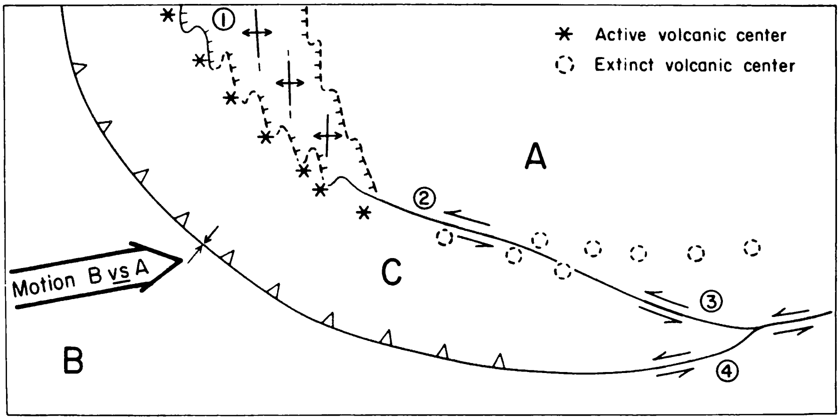

Figure 8 Model showing various methods by which the strike-slip components of oblique subduction can be expressed within the arc system. 1. En echelon extension. 2. Strike-slip faults near the volcanic chain. 3. Strike-slip faulting between the trench and volcanic chain. 4. Strike-slip faulting in the trench. Also shown by light arrows are the distribution of inter-plate motions when a third small plate (C) is created between the major plate (A and B).

Gulf of California (55) and Semanko rift (59, 112). Where no extensional zones exist, strike-slip faulting can occur between the volcanic chain and the trench, as on Median tectonic line of Japan (44), Longitudinal Valley of Taiwan (40), and Atacama Fault (1) or may not be separated from the under-thrust component.

The longitudinal shear component is separated from underthrusting because resistance to rupture along vertical planes within the lithosphere, especially within the extensional zone, is less than frictional resistance along the low-dipping Benioff zone. Thus, between the two major plates, a narrow plate is created which migrates along the subduction zone and causes the direction of underthrusting to be more nearly perpendicular to the plate boundary than the motion vector between the two major plates.

Most island arc systems in the western Pacific have normal polarity (face away from the continents), but a fair number have the opposite, or reversed, polarity. Either the geometry of reversed polarity was an initial condition, generated within an ocean basin (16, 68) or the arc system has since reversed its polarity. Evidence for polarity reversal has been presented for the New Hebrides (56, 76), Solomon (50, 61) and Luzon arc systems (52). To date, there is no substantial evidence that the reversed polarity of any arc system was generated from an initial condition.

In some instances, polarity reversal may have resulted from collision between the arc system and another tectontic unit, perhaps another arc system, a mid-plate rise

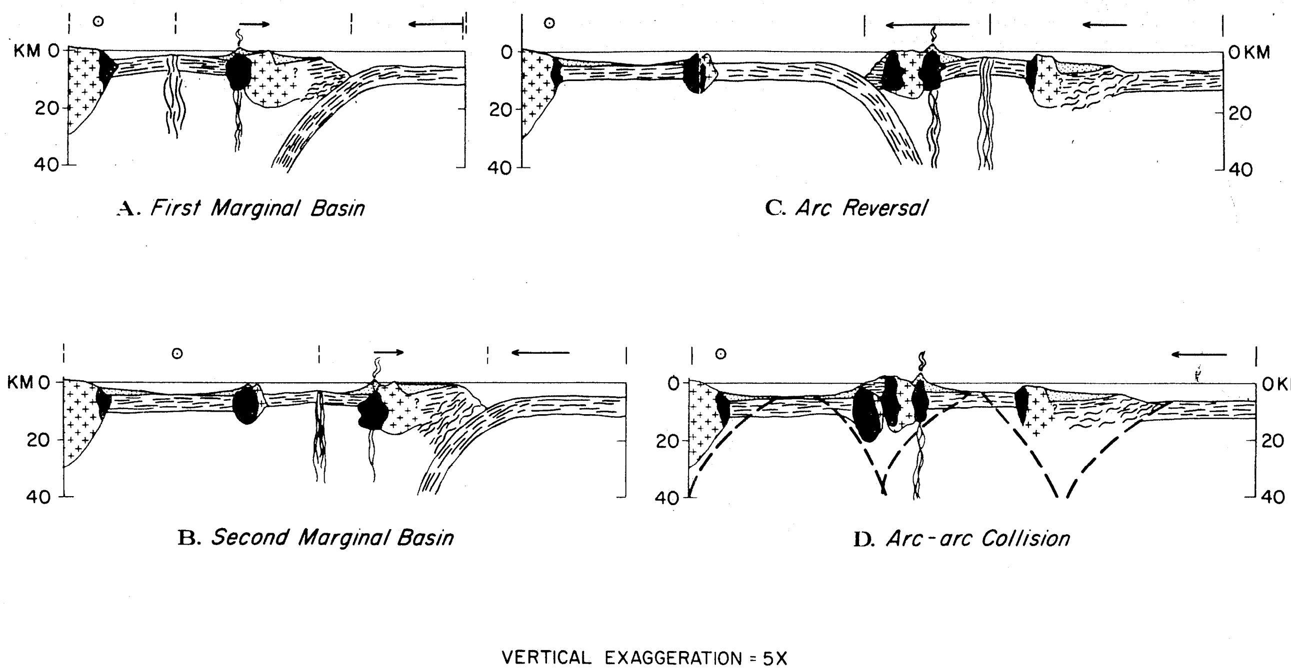

Figure 9 Hypothetical development of a marginal basin sequence, including arc polarity reversal and the emplacement of the ultramafic overthrust sheets. The arrows show qualitative motions of crustal blocks relative to the continental block at the left edge of each figure. Stage A illustrates the opening of the first marginal basin, which, as mentioned in the text, may not be as well organized as shown. Spreading is shown to be symmetric and localized but may be diffused across the basin. In stage B, a remnant arc has been created and a new marginal basin developed. In this and other figures, the consumption rates in the trenches are higher than the closure rates between major blocks by the amount of extension in the active marginal basin. In stage C, arc system polarity has reversed. Extension is shown behind the reversed arc system but need not be present. Collision between the reversed arc system and a remnant arc (stage D) leads to the overthrusting of the oceanic type crust on the leading edge of the frontal arc and requires that continued crustal consumption shift to a new locus, as indicated by the heavy dashed lines (50).

(61), or a very large seamount chain. More often, no such unit can be identified. A relative rearward motion of the lithospheric plate carrying the arc system over the asthenosphere may rotate the Benioff zone into and past the vertical so as to require more energy to continue normal polarity subduction than to reverse polarity (50). An apparent polarity reversal, now beginning at the south end of the Luzon arc system (27, 52), may be related to collision of the northern part of that arc system with Taiwan.

Subduction on reversed polarity arc systems results in the consumption of marginal basin crust, and ultimately in collision with a continent or remnant arc (Figure 9). The Hidaka-Kamuikotan orogen of Hokkaido and Sakhalin (78), much of New Caledonia (10), and several belts along the northern side of New Guinea (17, 43, 50) are examples of this type collision, which result in the high level emplacement of ophiolite sheets and high pressure metamorphic rocks.

These and other varieties of collisions involved in subduction processes (18) lead to further rearrangement of plate boundaries, since continental or arc crust is relatively light and resists subduction (71). If the relative motion between two plates is to remain unchanged, then subduction must shift to another locale.

In better documented examples of rejuvenation or initiation of subduction in the western Pacific (e.g. Shikoku, Manila, New Hebrides, Solomon, and New Britain arc systems), the new trench develops near an interface between oceanic crust and a continental or arc block.

The initiation of a subduction zone or rejuvenation following complete cessation of activity, remains an enigmatic but important process. Inactive subduction zones do not have the mass imbalances which are dynamically sustained in active arc systems, implying that tectonic and structural readjustments have taken place. Upon rejuvenation of subduction the new trench may form slightly forward of the final position of the old oceanic downgoing plate, and account for the large and relatively undeformed sheets of oceanic crust associated with the Franciscan complex of California (3) and with other subduction complexes.

CONCLUSIONS

Patterns of island arc and continental margin systems show superficially broad variations, but an increasing mass of data supports the early ideas which suggested a rather uniform underlying process. There seems to be no reason to create a classification of different types of subduction zone. Variations are important in that they reveal subsidiary or related tectonic processes, but to use them to subdivide arc systems into types masks the underlying continuum in the fundamental processes.

Variations among the relatively few parameters which control the character of arc systems can produce very complicated overall tectonic patterns. In most old orogenic belts these patterns would be further complicated by collisions which close the marginal basins and juxtapose the ridges. The complexities of recent arc patterns in the Melanesian and Indonesian-Philippine regions strongly suggest that unraveling a complex orogenic zone may be possible only in very general terms.

Identification of arc system units within orogens is critical for this interpretation, but depends on criteria that are now only poorly developed. Some indication of arc polarity has been obtained in simple cases from the K_2O-SiO_2 ratios (22), from the contrasting metamorphic regimes (78), and from the spatial relationships between subduction zone and upper slope rocks (63). Marginal basins probably play a significant role in most orogens, and if diagnostic sediment facies and facies patterns could be determined, might be valuable aids in determining ages of subduction pulses and arc polarity at that time. Undoubtedly, a combination of sedimentological and structural data, including the spatial relationships among rock and structural units will be necessary to successfully interpret any orogen. The task must begin with a better understanding of these parameters in contemporary arc systems.

Another major remaining task is to investigate temporal and spatial ordering of all events associated with consuming plate margins. Previous geotectonic concepts assumed orogenic cycles, based on frequently observed sequences of rock units and structures. Many of the proposed causes of these observations have been shown to be erroneous, but the consistent sequence and pattern of geologic units maintain their validity.

Several large-scale geologic syntheses of the subduction process, using both the observations in orogenic zones and the reconnaissance marine data, have been presented (18, 20, 75) as replacements for the orogenic or geosynclinal cycle. However, only intensive, problem-oriented geological and geophysical investigations in active arc systems will produce accurate and sufficiently detailed correlations between the geology observed in orogenic zones and processes active along subducting plate boundaries, and will permit the construction of an adequate orogenic flow sheet.

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