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EVIDENCE FOR PLATE-TECTONIC REGIMES IN THE ROCK RECORD

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ABSTRACT. The plate-tectonic theory requires a fresh evaluation of the rock record of geologic history in terms of the characteristic petrotectonic assemblages associated with different kinds of plate junctures and modified by the tectonic consequences of the evolution of plate margins. As the oceanic lithosphere formed at divergent plate junctures is destroyed systematically by plate consumption at convergent junctures, the fragmentary record that remains from older plate-tectonic regimes is preserved only within the present continental blocks. The key petrotectonic assemblages are ophiolitic sequences of oceanic crust from divergent junctures and volcano-plutonic orogens of magmatic arcs from convergent junctures, but present distributions of these rocks are the result of complex plate motions, juncture migrations, and crustal collisions in the past.

Ophiolitic sequences of peridotite, gabbro, dolerite, basalt, and various sedimentary strata within orogenic belts represent residual upper mantle, partly metamorphosed igneous crust, and overlying sediments of oceanic lithosphere from vanished ocean basins. The rifted margins of continental fragments produced by continental separations that initiated the development of ocean basins are marked by continental terraces of miogeoclinal strata and their offshore facies equivalents deposited on adjacent oceanic crust. Exogeosynclinal wedges spread by dispersal of sediment backward across miogeoclinal assemblages are evidence for orogeny involving juxtaposition of a continental margin and a convergent plate juncture by either activation of or collision with an arc-trench system.

In arc-trench systems, melanges and imbricate slabs are formed by subduction at the trench, both basins and uplifts occur within the arc-trench gap, and volcanic chains are underlain by batholith belts along the magmatic axis; either thrust belts and foreland basins or marginal seas with oceanic structure occur in the backarc area. The belt between the trench and the arc is a sliver plate thermally detached from the main plate behind the arc; transform shear may occur along the arc, and divergent spreading or convergent subduction may occur at the rear of the arc. Although arc orogens include both oceanic and continental varieties, there is a consistent relation between increasing potash content (K) in the magmas with increasing depth (h) to the inclined seismic zone in the mantle beneath. The polarity of volcano-plutonic orogens is indicated by transverse gradients in K for coeval igneous rocks and also by the varying intensity of blueschist metamorphism, which increases toward the arc at the structural levels exposed in uplifted melange belts of old subduction zones. Melange belts may grow by lateral accretion as subduction continues, and migration of arc activity may shift the magmatic axis over prior sites of the subduction zone. Crustal collisions which represent successive steps in continental assembly bring sialic blocks that inhibit plate consumption against arc structures across the melange belts and ophiolitic shreds of subduction zones, which thus become suture belts along which strike-slip is common during and following the collision orogeny.

The nature of the oceanic lithosphere present during different times in the past was dependent on secular trends in the evolution of the crust-mantle system. The evolution of continental lithosphere preserved from consumption may have depended also upon unknown processes of aging with time. Plate-tectonic behavior is the unifying mechanism for crustal evolution by the formation of oceanic crust from mantle-derived magmas melted beneath rises and the formation of continental crust from arc magmas melted primarily off the tops of slabs of oceanic lithosphere descending along the inclined seismic zones.

INTRODUCTION

According to plate-tectonic theory (McKenzie, 1972), major tectonic activity is concentrated within elongate belts of juncture between nearly rigid slabs or plates of a strong outer rind of the Earth. Tectonism is confined mostly to these plate junctures because the plates are in relative motion with respect to one another and also with respect to a weaker and softer layer beneath. The outer rind is the lithosphere, and the undermass is the asthenosphere (Daly, 1940 after Barrell, 1914-1915). The base of the rigid lithosphere is probably coincident with the top of the low-velocity zone, a partly molten and probably mobile region of the mantle (Anderson, Sammis, and Jordan, 1971). The base of the crust at M is everywhere contained within the slabs of lithosphere.

The configurations of present plate margins are defined by belts of seismicity marking active deformation of the crust. The relative motions of slabs of lithosphere in contact along plate junctures can be described in terms of components of divergence, convergence, and shear. These three extremes of behavior have characteristic geologic expression as intra-oceanic rise crests, arc-trench systems, and transform faults, respectively. The fast rates (1-10 cm/yr) of relative plate motions and the unsystematic orientations of plate junctures imply that the geometry and nature of plate margins are not stable with time. The interpretation of geologic history in terms of the plate-tectonic model requires that past plate margins or junctures be identified in the rock record. This can be done by relating certain petrotectonic assemblages to sets of tectonic and associated geologic processes that occur only under specific plate-tectonic regimes (Dickinson, 1971d).

The importance of bringing plate-tectonic logic to bear on the rock record of tectonic history is shown by the fresh insights that the exercise brings to the following major topics of geology (*also see* Dewey and Horsfield, 1970):

1. The nature of the tectonic entities and events called geosynclines (Dickinson, 1971b) and orogenies (Dickinson, 1971c) and their mutual relations in geotectonic cycles or sequences (Coney, 1970) responsible for the development of mountain belts (Dewey and Bird, 1970).

2. The separation of continents by drift (Stewart, 1972), the growth of continents by accretion of materials to active continental margins (Hamilton, 1969), and the assembly of composite continents by crustal collision (Hamilton, 1970).

3. Paleogeographic patterns of landmasses, epeiric seas, and ocean basins with their implications for the history of organic evolution (Valentine and Moores, 1970) and the relations between epeirogeny and orogeny (Armstrong, 1969; Johnson, 1971).

4. The origins of igneous magmas (Dickinson, 1970; Green, 1971) and controls on the distribution of different kinds of magmatic provinces (Lipman, Prostka, and Christiansen, 1971).

5. The geochemical evolution of the crust-mantle system during geologic time (Ringwood, 1969; Dickinson and Luth, 1971).

As well as influencing the most fundamental concepts of geological science, these topics also bear directly on two significant aspects of economic geology: (A) the distribution of metallogenic provinces in space and time, and (B) the occurrence of sedimentary receptacles for fossil fuels.

The purpose of this paper is to review the facets of plate-tectonic theory most pertinent for interpreting the geologic history of crustal rock masses. The major points covered include the different kinds of plate junctures with their geologic expressions and salient patterns of evolution of plate margins with their geologic consequences.

PRINCIPAL INTERPRETIVE CONSTRAINTS

Plate-tectonic theory stems from a kinematic analysis of current tectonophysical features, especially (A) orientations of sets of transform faults as families of small circles on the spherical globe (Morgan, 1968), (B) colinear patterns of geomagnetic anomalies at sea (Le Pichon, 1968), and (C) slip vectors deduced from first-motion solutions for earthquakes (Isacks, Oliver, and Sykes, 1968). Plate-tectonic logic can be extended directly to Tertiary events because axisymmetric patterns of colinear geomagnetic anomalies at sea indicate that fresh oceanic lithosphere has been generated along spreading centers or axes by the same style of movements of rigid plates receding from one another throughout the Cenozoic. As the dating of Mesozoic sea-floor is imprecise and pre-Mesozoic sea-floor is absent or rare, the full geometric logic of plate tectonics cannot be applied in general to pre-Tertiary events. Fragmentary evidence for older plate junctures must be sought mainly on the continents.

The oceanic lithosphere, with its scum of oceanic crust from rise magmatism, is thought to be formed at divergent plate junctures and to be consumed at the arc-trench systems of convergent plate junctures, presumably by inversion to dense phases and disappearance into the mantle. Minor subtractions of crustal materials from the descending slabs of oceanic lithosphere feed substance to (A) arc magmatism which occurs near the over-riding lips of the consuming plate margins and may be the principal means by which continental crust is formed, and (B) belts of melanges and imbricate slices caught in crustal subduction zones at trenches. Consumption of continental lithosphere, with its thick capping of light sialic crust, is not required by plate-tectonic theory and seems unlikely on gravimetric grounds. Once they are formed, continental blocks, and the slabs of lithosphere beneath them, presumably remain forever above the asthenosphere. Plate-tectonic events that affect continental margins thus leave a permanent record, although this rock record may be spoiled by deep erosion, concealed by burial to unexposed crustal levels, or even obliterated locally by crustal deformation or partial subduction. Continental rock masses now adjacent to one another may have formed in places far apart, and others formed side by side may now be found far apart, but the continental record is not destroyed systematically by the complementary growth and consumption of lithosphere in the surrounding oceanic realm.

Evidence for past plate-tectonic regimes depends upon analogies between Cenozoic rock assemblages, for which the plate-tectonic settings are known, and older rock assemblages for which the plate-tectonic settings must be inferred. Mesozoic and Paleozoic rock assemblages are so similar to their Cenozoic counterparts that no severe obstacles to this method of reasoning arise for at least much of the Phanerozoic. For the Precambrian, opinions differ as to the degree of similarity and dissimilarity of rock assemblages formed then and now. Where differences appear, in Precambrian rocks or even in somewhat younger rocks, two alternate but mutually compatible turns of thought must be considered: (A) there is a time before which the plate-tectonic model is inappropriate, or (B) plate-tectonic behavior has undergone a series of evolutionary stages, or an evolutionary progression of changes, through geologic time.

PLATE-TECTONIC REGIMES

The petrotectonic assemblages most diagnostic of specific plate-tectonic regimes are (A) ophiolitic sequences representing oceanic crust formed along the spreading axes of divergent plate junctures, and (B) volcano-plutonic orogens representing magmatic arcs standing beside convergent plate junctures. For each of these key petrotectonic assemblages, there are associated assemblages, like the miogeoclinal wedges of inactive continental margins and the subduction zones of active continental margins, arranged in orderly geographic patterns. No distinctive petrotectonic assemblages of similar large dimensions are formed along transform junctures of shear, but the positions of transforms connecting segments of divergent or convergent junctures strongly influence the distribution of the assemblages related to those kinds of junctures.

The pattern of juxtaposition of different petrotectonic assemblages within present continental blocks depends upon the past sequence of plate-tectonic regimes that affected the blocks. The sequence in time is controlled by the evolution of plate boundaries. Both the progressive modification of given plate margins and the development of new patterns of plate junctures are involved in this evolution. As plates inherently adjoin more than one other plate, the complex but rigorous rules that govern the evolution of triple junctions, where three plates are in mutual contact, are important logical tools (McKenzie and Morgan, 1969).

The history of present ocean basins suggests that many divergent plate junctures develop from an intracontinental rift, which evolves through the stage of a narrow oceanic trough, to the status of an intra-oceanic rise lying within a broad expanse of newly formed oceanic lithosphere flanked by rifted continental margins over which sediment aprons are draped. The surprising continuity of the world rift system, which branches along the rise crests, suggests that most new divergent junctures are born as extensions and bifurcations of pre-existing ones. These new branches penetrate previously intact plates and locally pry continents apart.

Convergent plate junctures at trenches lie both along continental margins and along island arcs standing between regions underlain by oceanic crust and lithosphere. Many island arcs are descendants of fringing arcs that migrated away from formerly attached continental margins as marginal seas opened by a special form of divergent spreading in their wakes (Karig, 1971; Packham and Falvey, 1971). It is not demonstrated, however, that all intra-oceanic island arcs began as fringing arcs marginal to continents; two features that may be incipient island arcs, the Macquarie Ridge and the Ceylon-Cocos seismic zone (Sykes, 1970), trend across oceanic regions, although both are linked at one end to projections of continental blocks.

Two arguments show that plate junctures can move with respect to one another. First, some plates, like the present African one, are nearly surrounded by long-active divergent junctures with no convergent junctures present in the central region between. It follows that the intra-oceanic rises along the divergent junctures move away from one another, as well as from the center of the plate, as new oceanic lithosphere is added to the plate margins. Second, the expansion of the ocean basins lying between the dispersed fragments of Pangaea and Gondwana implies the net contraction of the vast Pracific basin, despite the presence of divergent plate junctures within it. The arc-trench systems marking long-active convergent junctures that nearly ring the Pacific must move closer together to achieve this contraction. Elsasser (1971) has used the term retrograde motion to describe the back-bending required in a descending slab of lithosphere to accommodate the advance of a consuming plate margin across the original line of flexure in a plate being consumed. Encounters between migrating plate junctures can occur if at least one is convergent; in general, encounters will produce one or more triple junctions (for example, Atwater, 1970).

Especially significant for geologic history are the crustal collisions that occur when continental blocks on plates being consumed at convergent junctures meet the crustal elements of the associated arc-trench systems. The continental blocks, which resist consumption, probably choke the subduction zones and force rearrangements of plate boundaries and motions. The special influence of continents on gross plate configurations may be even greater if possible contrasts in thickness and structure between suboceanic and subcontinental lithosphere make continent-ocean interfaces preferred sites for plate consumption. The continental margins blocked out by continental separations along divergent junctures may in due time serve as the agents to terminate consumption along some convergent junctures and to initiate consumption along others. Although nearly all plates consist of both continental and oceanic portions that move together as a unit, the distribution and motions of the continental blocks thus may set constraints on the development of the entire global tectonic system (for example, McKenzie, 1969).

OPHIOLITIC SEQUENCES

Ophiolites now exposed within continental orogenic belts are apparently fragments of the crust and uppermost mantle from oceanic lithosphere that was not destroyed completely by plate consumption in the past (Cann, 1970; Coleman, 1971; Dewey and Bird, 1971). Where structurally intact slabs of ophiolitic sequences are exposed, they display a pseudostratigraphy that mimics the layering known for the present oceanic crust (fig. 1). The basal peridotites, commonly harzburgites, are residual mantle, from which the basaltic magmas that built the oceanic crust were bled, and their tectonite fabrics were imposed during emplacement in the upper lithosphere by solid flowage in the hot regions beneath rise crests or other axes of spreading. The overlying gabbroic rocks and some cumulus peridotites represent frozen basaltic magma chambers near the crust-mantle interface beneath an axis of spreading. The doleritic rocks represent compact swarms and sheeted complexes of dikes and minor sills that connected the magma chambers at lower crustal levels with the lavas at upper crustal levels. The pillow basalts and associated lavas represent the erupted magmas at the top of the magmatic crust. The sedimentary rocks, nearly always including argillite

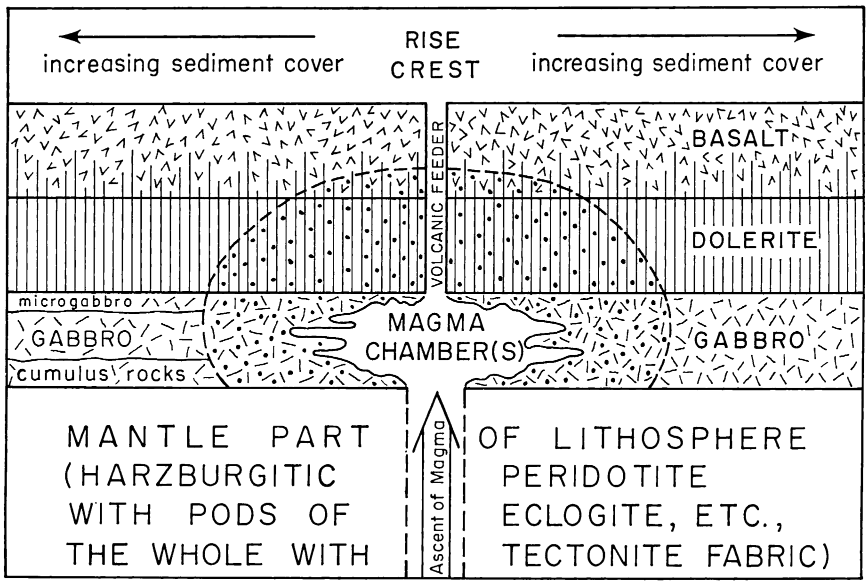

Fig. 1. Diagrammatic sketch showing generalized layers of an ophiolite sequence to illustrate formation of oceanic crust (see text) along divergent plate juncture between two accreting plate margins receding from rise crest (center). Basalt layer includes fresh pillow lavas in upper part but also altered pillow lavas and massive intrusive(?) sheets in lower part. Dolerite layer is typically a sheeted dike complex gradational into basalt layer above and gabbro layer below. Gabbro layer includes chilled rocks (microgabbros), especially in the upper part, and also cumulus rocks, both gabbros and peridotites, especially in the lower part (see left for idealized distribution). Stippled region inside dashed line is subject to thermal and hydrothermal metamorphism; igneous rocks of all crustal layers except the fresh, upper part of the basalt layer thus pass through this region of potential alteration following initial magmatic crystallization. Oceanic layer 2 is probably the unaltered part of the basalt layer, and oceanic layer 3 probably extends down to the base of the gabbro layer, or to the top of potential cumulus peridotite layers representing floors of crustal magma chambers. Modified after Cann (1970 and personal commun., 1971), Kay, Hubbard, and Gast (1970), and Davies (1971). Uncommon dioritic and keratophyric rocks are not shown.

and pelagites such as chert or micrite and commonly including turbidites and tuffs as well, were added as a cover when the mafic igneous crust was rafted away from the divergent plate juncture into an ocean basin in the interior of a plate. Basaltic seamount piles of large bulk can also be added to the tops of ophiolitic sequences during their sojourns in plate interiors. Many ophiolitic sequences also include relatively silicic but potash-poor differentiates, commonly described as trondhjemite or soda-granite where intrusive and as keratophyre or quartz keratophyre where extrusive. These rocks occur as part of the main igneous complex below the sedimentary cover, and intrusive dioritic counterparts are known from the modern oceans.

Each of the component parts of an ophiolitic sequence has a sheet-like form, but each kind of sheet is a time-stratigraphic facies equivalent of the others if both vertical and lateral relations are considered. Even the intrusive components of the axial magma chambers and their feeder systems evolve in a semicontinuous fashion. Older increments of the intrusive complex are crystallizing and cooling along the flanks of a rise, as sediment begins to accumulate above the capping lavas, even as younger magmas are emplaced to form younger increments closer to the rise crest. Magmatic temperatures are thus maintained within the crust along the axis of spreading. As temperatures decline to either side, lower levels of the igneous succession are exposed to potential thermal and hydrothermal metamorphism during the formation of an ophiolitic sequence. Metabasalts and metagabbros of greenschist and amphibolite facies, and also serpentinites and spilites, have been dredged from appropriate horizons along the walls of median rift valleys and transverse fracture zones where submarine fault scarps expose partial sections through oceanic crust (for example, Cann, 1969; Miyashiro, Shido, and Ewing, 1969, 1970b, 1971; Barrett and Aumento, 1970; Aumento and Loubat, 1971). Metamorphic effects observed in the ophiolitic sequences of orogenic belts may date in some cases from the times of origin at sea rather than from the times of structural emplacement into continental blocks.

Details of the vertical succession and lateral facies within intact ophiolitic slabs may reveal information about varying mechanisms of formation (for example, Dewey and Bird, 1971, p. 3186-3187). Although all modern oceanic crust has roughly the same overall structure, the processes of formation probably differ somewhat from fast-spreading rises to slow-spreading rises, from rises in open ocean to marginal seas behind migrating island arcs, and from simple axes of spreading to the vicinity of fracture zones along transform faults. Successive stages in the development of divergent plate junctures within the ocean basins are reflected by complicated patterns of geomagnetic anomalies and by changes in the trends of fractures zones along transforms (Menard and Atwater, 1968; Atwater and Menard, 1970). These features record intricate changes in the rates and orientations of divergent plate motions that might influence the genesis and ocean-floor metamorphism of ophiolitic sequences.

Unfortunately, the preservation of old oceanic crust is fragmentary at best, and the presence of ophiolitic sequences in an orogenic belt commonly indicates little more than the former existence of a divergent plate juncture within an ocean basin of indeterminate width. The remnants of once extensive ophiolitic suites in old ocean basins are exposed on the continents by isostatic uplift following one of two general modes of structural emplacement within orogenic belts: (1) incorporation of dislocated ophiolitic scraps into the deformed terrane of a subduction zone where the plate containing the ophiolitic sequence is being consumed, or (2) subduction of light crustal materials beneath an intact ophiolitic slab where the ophiolitic sequence is part of the consuming plate margin.

CONTINENTAL SEPARATIONS

The configuration of a divergent plate juncture at the time it caused a continental separation is indicated directly by the shapes of the rifted continental margins. These are marked by miogeoclinal wedges of sedimentary strata deposited along the attenuated and foundered edges of the continental fragments. Together with parallel continental rise aprons deposited on adjacent oceanic crust, these sedimentary successions form continental terraces that largely mask the underlying basement structures formed by rifting during continental separation (Rona, 1970).

The rifted continental margins formed by continental separation can have complex configurations, and the age of the two margins need not be synchronous along the whole length of the separation (fig. 2). Various cases can be distinguished only by careful attention to the comparative ages of the basal strata present locally along the trends of paired miogeoclinal sequences. Not illustrated is the case where similar, "pseudomiogeoclinal", deposition might mask an extinct arc-trench system whose activity along a continental margin was terminated without destruction of the adjacent ocean basin. Sedimentary successions in this case would lack the characteristic miogeoclinal property of resting, along the continental side, on the deeply eroded roots and across the sharply truncated trends of much older basement rocks.

MIOGEOCLINAL ASSEMBLAGES

The petrologic succession in typical miogeoclinal sequences is similar (for example, Rodgers, 1968; Stewart, 1972), and idealized but distinct stages of development can be recognized (King, 1969, p. 11). A thick basal clastic sequence, largely sandstone with minor conglomerate and shale, is deposited rapidly as a wedge thickening seaward across the continental margin. Deposition is triggered by wholesale foundering of the raw continental edge following the attenuation of continental crust during continental separation. Basaltic lavas related to the initial rifting may occur below or within the basal clastic wedge. When the configuration of the continental terrace assumes a more mature form, the basal clastic wedge is succeeded by a thinner cover of carbonates and shales whose deposition builds the terrace upward and outward along the tectonically quiescent continental margin. During both these petrologic stages, sediment dispersal is off the adjacent continent toward facies equivalents in coeval rise aprons and abyssal

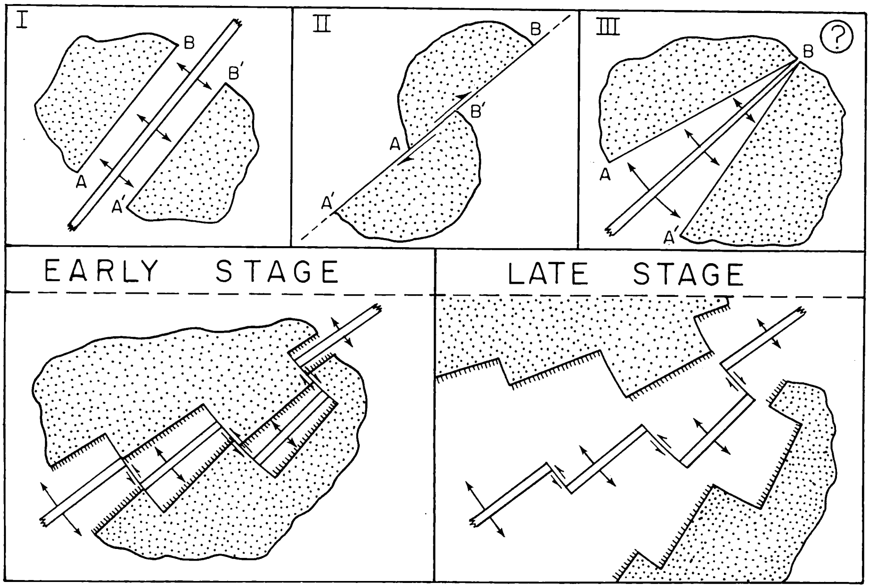

Fig. 2. Schematic maps to illustrate various patterns of continental separation. Continental fragments are stippled. Upper three diagrams show three extremes of behavior:

I. Separation by a rise segment that is far from the geometric spreading pole and is, therefore, spreading at a nearly uniform rate along its whole length; the onset of miogeoclinal deposition along the full lengths of AB and A′B′ can be synchronous.

II. Separation by a transform whereby miogeoclinal deposition would begin earlier at A′ and B than at A and B′; disruption by strike-slip along the continental slopes would continually offset miogeoclinal strata deposited on the edges of continental basement from offshore facies of the continental rises and abyssal plains deposited on oceanic crust.

III. Separation by wedging apart along a rise segment whose geometric spreading pole is near or within the initial continent; the onset of miogeoclinal deposition might be diachronous, beginning earlier near A and A′ than near B (query indicates region where other types of plate interactions would be required by the geometry shown).

Lower diagrams show successive stages in a complex continental separation along a rise with offsets and associated transforms; aspects of all three extreme cases of behavior are present; hachured rift margins of continental fragments shown could receive miogeoclinal wedges from initial time of rifting, which might be earlier in the southwest than in the northeast; rifted margins not hachured would have an earlier history of transform shear prior to deposition of any miogeoclinal wedge (see LePichon and Hayes, 1971, fig. 1, for stages in early history of such marginal offsets and associated transforms). Complex rifting and gashing of the continental fragments near the angular concave reentrants, where the hachured (initial) miogeoclinal segments meet the marginal offsets along initial transforms, may form wedge-shaped, semioceanic but intracratonic, depressions penetrating to interior regions of the continental blocks: the Benue depression (Burke, Dessauvagie, and Whiteman, 1971) off the Gulf of Guinea is a Phanerozoic example, and the so-called aulacogenes (Hoffman, 1971) may be Precambrian examples.

plains deposited in deeper water farther offshore. Except at the start, miogeoclinal deposition thus takes place along quiescent continental margins in an intraplate setting and does not monitor the evolution of any plate margin.

Later stages of geologic evolution in typical miogeoclinal regions involve the deposition of exogeosynclinal clastic strata piled across the miogeoclinal wedge from sources on the side away from the continent. These deposits reflect either (A) activation of plate convergence along the adjacent continental margin with consequent development of a magmatic arc shedding clastic debris behind the arc orogen, or (B) collision of the adjacent continental margin with an arc-trench system along a separate consuming plate margin that approached from offshore and provided a fresh sediment source. Either the activation or the collision scenario for the termination of miogeoclinal deposition implies that the position of a convergent plate juncture is fixed in place and time by the sedimentary record.

Exogeosynclinal sequences above miogeoclinal sequences commonly include a lower, "flysch-like" phase of fine, deep-marine clastics and an upper, "molasse-like" phase of coarse, shallow-marine or continental clastics (King, 1969, p. 11; Hoffman, 1971). This pattern may be evidence for collision, with the fine clastics shed into deep water where the continental edge was depressed by partial subduction. The coarse clastics might be shed later from uplifts formed along the collision zone after subduction was choked. The collision scenario might also account well for the thrusting of offshore turbidite facies across the seaward edge of some miogeoclinal assemblages during the events that terminated deposition. Both the sedimentary and structural features suggestive of collision might be explained, however, by thrusting and deposition behind a magmatic arc. The inherent ambiguities between the activation and collision scenarios cannot be resolved in a given case without detailed knowledge of a specific region.

ARC-TRENCH SYSTEMS

The main tectonic elements of arc-trench systems along convergent plate junctures are the trench itself, the arc-trench gap, the magmatic arc or orogen, and the backarc area (Dickinson, in press). The rock mass of the system grows in five ways (fig. 3): (1) Oceanic materials, mainly ophiolitic rocks and graywackes in varying proportions but including layers of ash from the arcs, are added to melange belts in the subduction zone beneath the acoustic basement barrier between trench and arc-trench gap as oceanic lithosphere is consumed. (2) In many systems, but not in all, significant quantities of arc-derived detritus are deposited as thick clastic prisms in elongate sedimentary basins within the arc-trench gap (Dickinson, 1971a). (3) Volcanic and volcaniclastic rocks accumulate in great piles, foundered masses, and flanking aprons along the trend of the arc. (4) Plutons are emplaced in the roots of the arc orogen, either within the base of the volcanic sequence or in the crust beneath. (5) In most backarc areas, there are either (A) thrust belts made of stacked

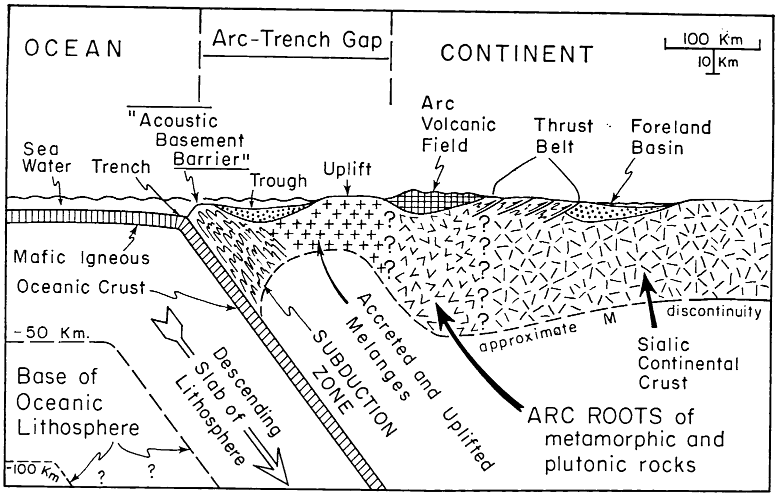

Fig. 3. Hypothetical and generalized sketch of an arc-trench system along a continental margin; vertical exaggeration about 2.5X. Undeformed sediments in the arc-trench gap and foreland basin are stippled. Plutonic rocks of the arc roots intrude the base of the cogenetic volcanic pile and at deeper levels may intrude either old sialic crust and deformed strata above it or accreted melanges and oceanic crustal scraps within or beneath them. Protoliths of metamorphosed country rocks among the plutons vary depending upon how the site of arc magmatism migrates with respect to the preexisting continental margin and the positions of the subduction zone at earlier stages in the evolution of the arc-trench system.

thrust sheets of crustal rocks which move across continental crust behind the arcs and spread clastic wedges into shallow foreland basins beyond; or (B) oceanic crust and lithosphere newly formed in marginal seas opening behind the arcs, from which clastic wedges are shed into the deep water present. The two alternative features in backarc areas pose the most direct contrast possible. In the first case, typical of many arcs on continental margins, some convergence of lithosphere behind the arc can be invoked to explain the crustal shortening involved in the thrusting of crustal rocks at the back of the arc structure, and in its rear, over adjacent continental crust. In the other case, typical of many intraoceanic arcs, some divergence of lithosphere behind the arc must be invoked to explain the development of a basin with oceanic structure. To complicate matters further, there is strong evidence for strike-slip displacements at the surface along or parallel to the magmatic axes of some arcs. Most notable is the Semangko fault zone in the Barisan Range down the spine of Sumatra (Katili and Hehuwat, 1967; Katili, 1970). These diverse features reflect major departures from plate rigidity within arc-trench systems. Apparently the integrity of consuming plate margins is locally spoiled by weakening and softening of the lithosphere along the belt of high heat flow beneath and behind the magmatic arc (for example, Barazangi and Isacks, 1971, fig. 17).

The local dismemberment of consuming plate margins within or just behind arc-trench systems can be represented schematically by small sliver plates centered on the arc-trench gap with the trench as one plate boundary and the arc or backarc area as the other plate boundary (fig. 4). The total vector convergence between the two large plates to either side of the arc-trench system as a whole thus becomes the sum of several potential vectors. At the trench, convergence may be normal or oblique to the trend of the subduction zone (McKenzie and Parker, 1967), within which strike-slip components of displacement thus may be significant as well as dip-slip or thrust components. Along the axis of the arc, strike-slip may occur, and it seems conceivable that oblique convergence for the total arc-trench system could be accommodated, through complex adjustments of stresses and strains in the lithosphere, by nearly normal

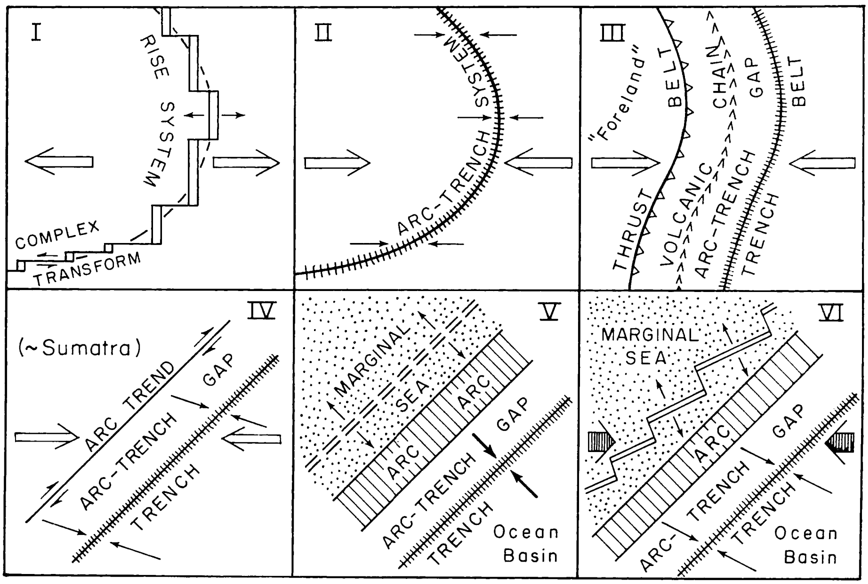

Fig. 4. Diagrams to illustrate compound plate boundaries: single lines are transform shears, double lines are spreading axes, crossed lines are subduction zones, large arrows show overall plate motions, small arrows show local plate motions.

I. Complex rise-transform network in typical rectilinear pattern; dashed line indicates gross arcuate trend of plate juncture.

II. Arcuate convergent plate juncture showing local variations from normal to oblique subduction common for arc-trench systems.

III. Arc-trench system on continental margin showing relative positions of "primary" subduction zone at trench and thrust belt ("secondary" subduction zone?) between arc and "foreland basins" behind arc.

IV. Complex arc-trench system in which oblique subduction induces strike-slip transform shear along a zone of heated lithosphere under the volcanic chain, and the arc-trench gap can be treated kinematically as a sliver plate (idealized interpretation of Sumatra where the Semangko dextral fault system follows the volcanic chain of the Barisan Range; movement vectors speculative).

V. Intra-oceanic arc-trench system with marginal spreading sea behind it (overall plate motions uncertain but rate of consumption at trench presumably exceeds rate of spreading behind arc).

VI. Complex intra-oceanic arc-trench system similar to Case V in which spreading azimuth in marginal sea behind arc is oblique to arc trend (D. Karig, personal commun., 1971); movement vectors speculative.

convergence in the subduction zone at the trench coupled with strike-slip of essentially transform character along the arc. Where major thrust belts behind arcs lead to the development of so-called two-sided orogens (Burchfiel and Davis, 1968), some auxiliary plate convergence and secondary subduction may take place behind the arcs. Spreading along discrete axes or within diffuse regions in marginal seas behind arcs may be in a direction normal or oblique to the trend of the arc-trench system.

MAGMATIC ARCS

The magmatic arcs of arc-trench systems appear in the geologic record as volcano-plutonic orogens made of granitic batholith belts with cogenetic volcanic covers. Enclosing sheaths of metamorphic rock form the high-temperature, low-pressure regions of paired metamorphic belts. Derivative sedimentary strata of dominantly volcano-plutonic provenance may be deposited along flanking belts in troughs or basins of arc-trench gaps and in foreland basins or marginal seas behind the arcs. Sediments in these settings may also be derived, however, from other sources in uplifts to either side of the magmatic arcs (see fig. 3). In principle, the full record of the orogens is preserved, except where lost by erosion, but in practice the overprinting of the results of complex processes makes interpretation difficult.

The petrologic and structural variants of magmatic arcs are incompletely known but include oceanic and continental extremes. Intra-oceanic arc structures are built on thin oceanic, or ophiolitic, crust of amphibolitic character when metamorphosed and are made largely of basaltic and andesitic rocks which include marine volcaniclastic strata deposited in blocky intra-arc basins. Associated plutons are commonly small and may be largely gabbroic or dioritic masses with relatively minor granitic associates. The distinction between true ophiolitic suites and intra-oceanic arc assemblages thus may prove difficult (for example, Jakes and Gill, 1970), particularly as the latter inherently overlie the former. Arc structures on continental margins are built upon and within sialic crust and are made largely of andesitic and dacitic rocks, including continental volcaniclastic beds, associated with extensive underlying granitic batholiths with relatively minor gabbroic and dioritic associates. The two idealized extreme types of arc assemblages are doubtless only end-members of a spectrum of behavior. Intra-oceanic arcs may evolve, through the progressive addition of crustal materials to the semicontinental arc structure, into arcs of continental aspect. Continental arcs may also evolve into arcs of semi-oceanic aspect through the progressive accretion of subduction zones to the continental margin and consequent migration of arc magmatism away from the old sialic block.

The polarity, or direction toward the associated trench, in volcano-plutonic orogens is indicated by the level of potash content, which increases in the igneous rocks going away from the trench. The potash level can be correlated with the vertical depth from the surface to the position of the inclined seismic zone which roughly marks the top of the lithosphere descending at an angle of 15° to 75° into the mantle beneath the arc. If a potash index, K, is defined as 10^3 times the K_2O/SiO_2 ratio indicated for a silica content of 57.5 percent on a variation diagram of K_2O versus SiO_2 for an igneous suite, then the depth, h, to the seismic zone beneath active arcs is given approximately by the equation, $h(km) = 6K + 7$ (Dickinson, in press, fig. 6). For the active arcs, h ranges from 75-100 km to 250-275 km (Dickinson, 1970, 1971d). As the K-h relationship does not depend on the varying thicknesses of subjacent crust beneath the arcs, the magmas are probably generated at greater depth near the seismic zones, either from the mantle or from oceanic crust in the lithosphere descending beneath the arcs. In either case, the construction of the arcs appears to be a means to build new blocks of sialic crust derived from the mantle or the oceanic realm.

An apparent pulsing of igneous activity is typical in many batholith belts (Dickinson, 1970) and finds no ready explanation in plate-tectonic theory. The encounter of a spreading rise along a divergent plate juncture with a trench along a convergent plate juncture might trigger a round of intense magmatism in the associated arc (Grow and Atwater, 1970). A number of such encounters would be required to explain the observed episodicity of magmatism in various local areas of an orogenic region like western North America. The plot of figure 5 suggests a more general explanation. In 3 of the 6 segments of the magmatic belt depicted, areas that represent subparallel bands lying side by side display apparent plutonic episodes that are systematically staggered in time to form possibly complementary arrays of magmatic pulses (for example, Smith and others, 1971). The local episodicity shown for each area may stem only from a selective sampling of more continous magmatism whose axis wavered back and forth across the orogen. Prograde migration of the arc activity toward the trench may be caused by accretionary growth of the melange belt of the subduction zone, and there is evidence that retrograde migration of the arc activity away from the trench may be characteristic where this accretionary growth is insufficient to dominate the evolution of the system (James, 1971; Dickinson, in press), or where subduction may actually abrade and wear back the edge of the overriding plate (Rutland, 1971). In terms of the K-h relationship, migration of the arc activity can occur either by a change in K and h or by a change in the position or the dip of the inclined seismic zone with K and h fixed.

SUBDUCTION ZONES

The subduction zones that form along trenches on the frontal flanks of active arc-trench systems appear in the geologic record as belts of tectonic melanges and imbricate thrust slabs within the high-pressure, low-temperature blueschist regions of paired metamorphic belts. The melange belts have the potential to grow steadily in width by lateral accretion so long as plate convergence and consumption continue to stuff shreds of oceanic crust with its sedimentary cover into the subduction zones. Accretionary growth may not occur if the subducted materials are carried to deep crustal levels that are unexposed beneath adjacent sialic crust in the consuming plate margin of a marginal arc-trench system, or if they can invert to phases dense enough to allow them to disappear into the mantle as part of a descending slab of lithosphere. Actual accretionary growth of a melange belt may be masked if consequent migration of the trench causes sympathetic migration of the axis of arc magmatism (Matsuda and Uyeda, 1971). In this case, older increments of the melange belt will become arc roots and may be modified almost beyond recognition by metamorphism and intrusion.

Among the melanges and thrust sheets of several subduction zones, Ernst (1971) has shown that pressure-controlled metamorphic isograds generally parallel the trends of the belts. The metamorphic assemblages indicative of highest pressures lie closest to the associated arc orogen. Evidently the process of subduction drew the deformed materials downward into a roughly wedge-shaped crustal mass that was thickest on the side next to the flank of the arc. When subduction ceased and uplift

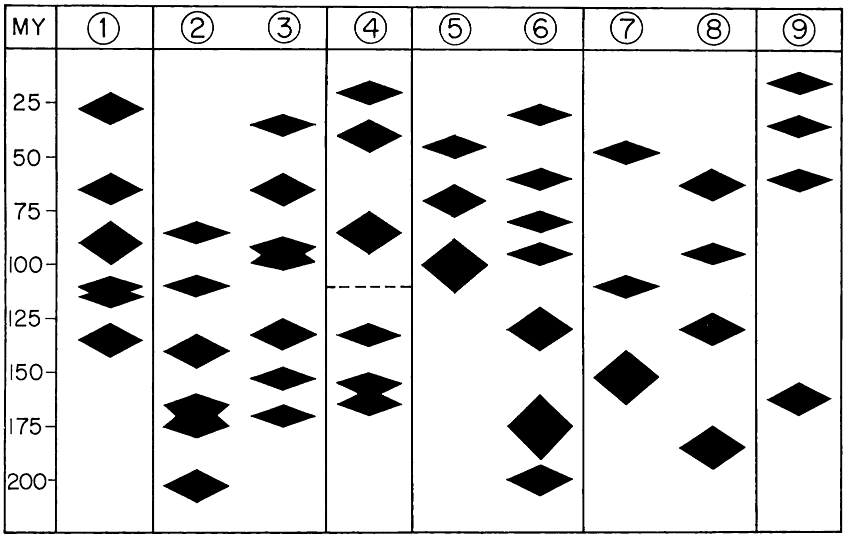

Fig. 5. Apparent spacing in time of major intrusive episodes in Mesozoic-Cenozoic arc-type orogens in western North America as indicated by clustering of radiometric dates in the literature. Widest parts of diamonds indicate inferred peaks of plutonic activity, and vertical extents indicate inferred durations of intense plutonism or uncertainties in timing of peak activity. As most events shown are controlled by K-Ar dates only, some may record merely tectonic uplift of arc terranes; however, as cogenetic volcanic rocks accompany the plutonic rocks for 25 to 40 percent of the events shown, most are probably magmatic episodes as inferred here. Vertical lines separate sequences of magmatic episodes in six different segments of the western cordilleras from Mexico to Alaska. For three of these segments, double columns represent sequences of magmatic episodes in parallel belts side by side. Areas represented include: (1) Baja California, Peninsular Ranges, Arizona, and Sonora; (2) Sierra Nevada; (3) Great Basin in Nevada; (4) Klamath Mountains (below) and Cascade Range (above); (5) western British Columbia (Coast Range); (6) southeastern British Columbia; (7) southeastern Alaska panhandle; (8) northern British Columbia and southern Yukon; (9) southern Alaska (Gulf), Alaska Peninsula, and Aleutian Islands. Plot includes data published through 1971.

occurred to restore isostatic balance, the thickest part of the wedge rose the most and exposed the deepest pressure levels observable in surface outcrops today. This relationship serves to emphasize the fact that the chaotic deformation of melanges and related rocks in subduction zones may stem partly from processes related to diapiric uplift as well as from slicing during plate consumption. The exceptional structural disorder and the commingling of incompatible metamorphic facies in belts of these rocks warn us that their history has been more complex than we can describe adequately with present information (for example, Suppe, 1970). Where accretionary growth of melange belts occurs, each increment of the belt may undergo a continuous cycle of deformation beginning with depression to deep crustal levels and ending with elevation to shallow crustal levels (fig. 3).

The thrust belts that separate marginal arc-trench systems from foreland basins behind the arc structures are apparently intracontinental subduction zones (Roeder, in preparation). The thrusting in this setting is commonly ascribed to gravitational gliding, but this explanation seems unsatisfying for thrust belts that extend for thousands of kilometers. Such distances are known for the Cenozoic Subandean belt east of the present Andean arc-trench system in South America and for the Mesozoic Cordilleran foreland east of the Mesozoic volcano-plutonic orogen along the western continental margin of North America (Hamilton, 1969). Explanations based upon the concept of gravitational spreading of a highland belt along the trend of the orogen are more satisfactory (Price and Mountjoy, 1970) but cannot account for the apparent loss of continental basement beneath the region where the thrust sheets are stacked (Price, 1971). The tilt of the sloping basement beneath the thrust sheets (Oriel, 1971) is compatible with partial subduction of sialic crustal material beneath the rear flank of the orogen. The conditions required to foster this behavior are uncertain but may be simply the active motion of continental lithosphere driving past the underlying asthenosphere toward a marginal arc-trench system which is advancing by retrograde subduction over the adjacent ocean basin (Coney, 1971).

CRUSTAL COLLISIONS

When sialic crustal blocks of continental fragments or island arcs are brought together at a convergent juncture by plate motions, melanges and associated ophiolitic shreds of the subduction zone mark the suture belt along which the blocks of sialic crust are welded together. Suture belts thus mark successive steps in the tectonic assembly of a composite continent. In general, collisions involving continental margins and arc-trench systems will not be synchronous along the full length of the curvilinear features involved (Moores, 1970). Instead, the timing of collision along any suture belt will vary in some semi-progressive manner (fig. 6). Collision points, or tectonic transitions, will migrate in a manner analogous to the migration of triple junctions. The diagram does not attempt to show the complex relations that would develop in both space and time if the continental margins depicted had irregular shapes, or if the arcs were segmented by connecting transforms. Orogenies caused by crustal collision will vary in timing, and probably in detailed geologic expression, along the length of a suture belt. Tests for synchroneity of similar orogenic phases in different parts of an orogenic belt are not a valid means for correlating or distinguishing plate tectonic events of this kind.

In general, the approach vector of relative motion between two crustal masses about to collide will include both normal and strike-slip components with respect to one or both colliding margins. The resulting oblique contraction across the suture belt has been called transpression (Harland, 1971) in recognition that there is a potential for transcurrent faulting along the trend of the suture belt during collision. Some component of strike-slip motion may well continue along the suture belt for some indeterminate period after collision. Strike-slip along such suture transforms may account for the complex history of some orogenic fault zones, like the Brevard in the Appalachians for which the relative significance of thrust and transcurrent movements is uncertain. Recall also that roughly analogous longitudinal strike-slip faulting may occur, in the absence of crustal collision, along the subduction zone or the magmatic axis of arc-trench systems where plate convergence is oblique to the plate juncture (see above).

The full effects of crustal collisions are uncertain. Where the crustal suturing of a continental collision welds large areas of two separate plates together, plate interactions at all other margins of the two plates involved may be changed. A full global reorganization of plate motions thus might be promoted by the collision. Alternatively, continental collision and choking of subduction at one margin of a continent might trigger compensatory initiation of subduction and plate consumption at another margin of the continent. Speculations of this kind are inherent in the concept of plates as intact entities and have sweeping implications for establishing sequential relationships among major tectonic events in widely separated areas.

COMPLEX TRANSFORMS

By convention, a transform involves only simple shear, with full conservation of surface area and neither convergence nor divergence across the transform juncture. Except for the evidence for offset of rock masses, the geologic record of such motion is confined to shear zones of mylonitic or dynamothermally metamorphosed rocks and to clastic wedges dumped where highlands are faulted adjacent to lowlands or basins. In a larger view, however, major transform systems of shear may include minor or local components of divergent or convergent plate motion. Convergent components will give rise to transpressive folding in adjacent rocks in patterns en echelon to the trend of the transform at an orientation diagnostic of the sense of the transform motions (for example, Dickinson, 1966).

Divergent components lead typically to more complex behavior including the opening of rhombic depressions at intervals along the trend

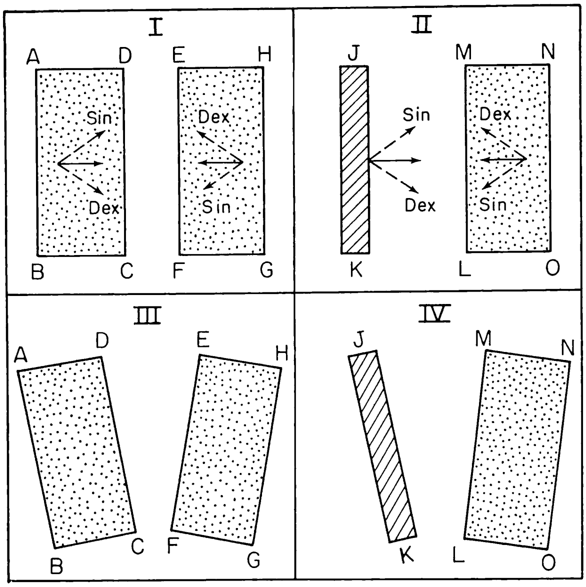

Fig. 6. Schematic maps to illustrate hypothetical patterns of continental collision (I, III) and arc-continent collision (II, IV). Configurations shown are just prior to encounters between colliding continents (stippled) and island arcs (hatched). Arrows denote relative motions of colliding crustal elements prior to collision. For cases I and III, either continental margin, CD or EF, or both, may lie along convergent plate junctures with marginal arc-trench systems. For cases II and IV, island arc JK must mark a convergent plate juncture with an active arc-trench system facing the continent *if* continental margin LM is inactive (that is, miogeoclinal) within the interior of a plate; but if continental margin LM harbors a marginal arc-trench system along a convergent plate juncture, then arc JK may be dormant or may face either way if active. Simplistic scenarios for orogenic events accompanying collision in different cases are as follows:

I. Collision will cause synchronous suturing of parallel continental margins CD and EF as subduction and arc activity along either or both is choked. Adjustments of plate boundaries and motions may trigger (A) plate consumption and new arc-trench systems along margins AB or GH, (B) transform shears across formerly coherent plates (that is, along ADEH or BCFG trends if the vector of relative plate motion is initially normal to CD and EF as denoted by solid arrows), or (C) plate rearrangements elsewhere to compensate for the CD-EF collision. If the vector of relative plate motion is not normal to the colliding margins (that is, dashed arrows), dextral or sinistral transform shear may occur along the suture belt during collision. Such transform motions may persist along the suture belt after all local plate consumption has been braked by collision.

II. Synchronous collision and possible transform motions along the suture belt during collision, as for Case I, will cause accretion of arc structure JK to continental margin LM through one of two possible sequences of events: (A) where continental margin LM is active, accretion of arc JK, whether active or dormant, will cause marginal arc-trench activity to shift to the left from initial trend LM to accreted marginal trend JK; or (B) where continental margin LM is inactive, collision with right-facing arc JK will probably reverse polarity of arc JK as subduction zone flips from right side to left side of trend JK. Either alternative scenario thus leads to a left-facing marginal arc-trench system along accreted trend JK following collision and accretion to trend LM, the continental margin prior to collision.

III. In the general case of collision of non-parallel continental margins, encounter of C and F at a point contact may brake plate motions and prevent wholesale continental collision. More likely, progressive slight adjustment in plate motions or boundaries may allow diachronous suturing of continental margins CD and EF in time sequence from C-F to D-E as triple junction or collision point migrates. In general, as for Case I, dextral or sinistral transform shear probably will occur along of the transform. Divergent plate junctures tend to orient themselves at right angles to local vectors of separation. If this direction is not parallel to the gross trend of a plate juncture, the juncture tends to break down into a sawtoothed, rectilinear pattern of short rise segments joined by short connecting transforms (fig. 3). In principle, all gradations should exist between (A) complex divergent junctures along which the divergent motion between the separating plates has no component of strike-slip with respect to the gross trend of the juncture, and (B) others along which the strike-slip component is dominant in a regional view, with long transforms connecting short rise segments, as in the Gulf of California. Wherever there is divergence as well as shear involved in a continental separation, the construction of ophiolitic oceanic crust between the continental fragments, and of miogeoclinal wedges on their trailing edges, will tend eventually to overwhelm, as evidence of behavior, any local clastic wedges or fault zones related to associated transforms. Only if divergence is essentially absent will the effects of the transform motions be prominent in the geologic record. In the special case where pure transform motion is concentrated along a continental margin, sediment delivered to the adjacent ocean basin as turbidites will be separated continually from deposits on the edge of the continent, and development of a full continental terrace with both miogeoclinal and oceanic facies is frustrated.

The recognition of paleotransforms offers the most serious challenge to plate-tectonic analysis of the geologic record because no distinctive petrotectonic assemblages of large dimensions are formed along them. Regionally anomalous distributions of more diagnostic petrotectonic assemblages may be the only firm clues to their presence.

LITHOSPHERIC EVOLUTION

Plate tectonic behavior is the unifying mechanism that links together into a coherent system the formation of oceanic crust by rise magmatism and the formation of continental crust by trench tectonism and arc magmatism. Plate tectonics thus implies an irreversible geochemical evolution of the crust-mantle system (Ringwood, 1969). In the simplest model for the overall system, residual lithospheric mantle depleted of crustal constituents by successive episodes of partial melting beneath rises and arcs descends through the asthenosphere to form an accretionary mesosphere or inner mantle (Dickinson and Luth, 1971). Fresh asthen-

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the lengthening suture belt during collision and again may continue after collision is complete if residual components of the pre-collision plate motions causing transform shear persist after the convergent components are suppressed by collision.

IV. Diachronous collision causes diachronous accretion of arc JK to continental margin LM in a manner analogous to the continental collision of Case III. A tectonic transition point, which marks a change in tectonic style, as for Case II, migrates along the accretionary continental margin from K-L to J-M during collision and accretion. The potential also exists for transform shear along the lengthening suture belt during collision and accretion, following which the obliquity of subduction and the consequent potential for transform shear within the complex accretionary continental margin is dependent upon the behavior of the lithosphere to the left of arc JK.

osphere is thus displaced upward and tapped for the production of more crust and lithosphere. Variations in the chemical and isotopic compositions of oceanic basalts suggest, however, that the asthenosphere is too inhomogeneous to have had the uniform history implied by this simple model (Peterman and Hedge, 1971). Similar mantle inhomogeneities evidently occur also beneath continental regions (Hedge and Noble, 1971). Episodes of local depletion or removal of easily fusible constituents by partial melting in irregular patterns at times deep in the past apparently are required by the data. Some inhomogeneities in the asthenosphere could also arise by contamination from either crust or mantle within slabs of lithosphere sinking through it. Geochemical studies of characteristic petrotectonic assemblages in crustal masses of different age may have the potential to define, by inference, secular trends in the composition of asthenosphere and lithosphere at different times in the past (for example, Hart and others, 1970). These trends may have been influenced by declining heat production from radioactive decay through time.

The contrast between oceanic and continental crust may not be the only contrast between oceanic and continental lithosphere. The history of a segment of oceanic lithosphere typically runs the course from creation to consumption within 1 to 2.5×10^8 years. Beneath many continental blocks of Precambrian rocks, however, the mantle in the lithosphere has been preserved in place for roughly ten times as long, or 1 to 2.5×10^9 years. Conceivably, unknown aging processes may modify lithospheric mantle beneath continental blocks with the passage of time. Mass may be lost if entrapped fusibles are transferred upward to thicken the overlying continental crust with the passage of time. Such a process may be reflected by the fact that the depth of M in the United States and the Soviet Union can be expressed as a function of the age of the continental crust (Woollard, 1972). Also compatible with the suggestion is the observation that increases in crustal thickness are commonly accompanied by increases in the mean seismic velocity of the crust or upper mantle or both (Woollard, 1972).

On the other hand, mass may be gained by the acquisition of material from the underlying asthenosphere. In the Andean region, the zone of low seismic velocity and high attenuation of shear waves that may mark the base of the lithosphere is about 250 km deep beneath the continent but only about 50 km deep beneath the adjacent ocean (James, 1971). General models developed by Sclater and Francheteau (1970) to account for patterns of terrestrial heat flow require subcontinental lithosphere to be about 200 km thick whereas suboceanic lithosphere is only about 100 km thick. Their data also show that the time after crustal formation required for thermal decay to a surface heat flux of 1 to 1.25 HFU is about 10^9 years in continental regions as opposed to only 10^8 years in oceanic regions. As the plate tectonic theory emphasizes that the whole lithosphere, rather than the crust alone, is the fundamental entity in tectonics, these and other hints at differences in the overall structure of subcontinental and suboceanic lithosphere have broad implications.

CRUSTAL EVOLUTION

The compositions of basaltic magmas erupted along divergent plate junctures at different stages of structural development can be explained by the partial melting of pyrolite or lherzolite at different depths (Green, 1971). The undersaturated alkalic magmas commonly erupted along intracontinental rift systems could be generated by small amounts of partial melting at depths of 100 km or more within the low-velocity zone beneath a full thickness of ruptured lithosphere. With continued extension, thinning and eventual separation of lithosphere plates would allow elongate diapirs of asthenosphere to rise into the developing slot within the lithosphere. This upwelling and consequent release of confining pressure leads to the segregation of magmas from the mantle source at progressively higher levels with successively greater degrees of partial melting as the superheating increases. Experimental results suggest that the abyssal basalt magmas of the ocean floors could be produced in this fashion by partial melting of 25 percent or more at depths of magma segregation less than 25 km.

The experimental results are applied to this model of magma generation with the tacit assumption that the basaltic magmas of the various tectonic provinces are essentially primary melts which last equilibrated with residual or cognate crystals at the site of initial segregation from mantle rock. There is also the alternative of extensive magmatic evolution by partial differentiation in magma chambers within the mantle or crust at pressure levels intermediate between the site of initial segregation and the site of eruption at the surface (O'Hara, 1965). In this case, the compositions of the abyssal basalts may not represent the primary magmas (O'Hara, 1968). Two related lines of evidence appear to support this currently unpopular contention.

First, the compositions of porphyritic abyssal basalts with either olivine or plagioclase phenocrysts can be used to infer a single presumed olivine-plagioclase cotectic curve in the position expected for crystallization at low pressures near atmospheric (Miyashiro, Shido, and Ewing, 1970a). The bulk compositions of the rocks cluster close to the trend of this cotectic, as if the magmas were liquids derived at low pressures: perhaps from partial melting not far below the crust, or perhaps by partial crystallization of parent magmas in magma chambers at crustal levels where the gabbros of ophiolitic sequences form. In the ophiolitic complex of Papua, Davies (1971) has shown that a large part of the gabbro layer consists of cumulates, and that the compositions of the basalts and gabbros are systematically different. The basalts closely resemble modern abyssal basalts, but the gabbros contain significantly more magnesia and lime and less iron and alkalis. If the gabbros and basalts were derived from the same parent magma, as a plate-tectonic interpretation of their structural relations implies, that parent magma did not have the composition of abyssal basalt. The gabbros in several other well-known ophiolites also have consistently lower Fe/Mg ratios than the associated basalts and appear to represent a differentiated composition roughly midway between the residual mantle below and the erupted lavas above (Davies, 1971, p. 31). This relation suggests that even the most primitive abyssal basalts may be derivatives of more primitive magmas that are nowhere erupted at the surface owing to systematic entrapment in the magmatic "surge chambers" at deep crustal levels beneath oceanic spreading axes.

The compositions of arc magmas erupted along convergent plate junctures can be explained by the partial melting of eclogitic rocks formed from the inversion of basaltic crust as it descends into the mantle at the upper surface of slabs of lithosphere (Green and Ringwood, 1968; Raleigh and Lee, 1969; T. H. Green, personal commun., 1971). According to Armstrong (1971), lead-isotopic compositions of arc lavas indicate that oceanic sediments on the slabs being consumed also contribute to the magma batches, but Compston (personal commun., 1971) has observed that strontium isotopic compositions of arc lavas indicate that only small amounts of oceanic sediments entered the magmas. The transverse increase in the content of potassium, and other incompatible elements (J. Gill, personal commun., 1971; Jakes and white, 1972), in the igneous rocks across the arcs in the direction away from the trench, is expected in terms of partial melting at progressively greater depths along the inclined seismic zones through any one of three controlling mechanisms: (1) melting of different phase assemblages controlled by pressure (Jakes and White, 1970; Fitton, 1971), (2) lesser degrees of partial melting at successively greater depths (for example, Jamieson and Clarke, 1970), or (3) greater amounts of scavenging of overlying mantle as the length of travel paths to the surface increases.

Significant contributions of substance to the rising magmas from mantle or crust above a descending slab cannot be excluded in some cases but are unlikely in many cases and unnecessary in most (Dickinson, 1970). Strontium-isotopic ratios in the igneous rocks of volcano-plutonic orogens are so low as to preclude much contamination of the arc magmas by ancient sial. If anatectic melting of crustal materials in the hot roots of the arcs generates some of the magmas erupted or emplaced near the surface, the materials involved must themselves have been recently emergent from the mantle. Cycling of rubidium from suboceanic asthenosphere to basalt in oceanic crust, to eclogite in a descending slab, to volcanogenic strata in arc roots, to anatectic melts could be fast enough to avoid a significant build-up of radiogenic strontium at any point in the cycle.

Whatever the details in the arc-building process, arc magmatism appears to be the main means by which sialic continental crust is formed (Taylor, 1967). However, the upper continental crust above the Conrad discontinuity is more siliceous and contains more of the incompatible elements than the bulk composition of intra-oceanic island arcs. To account for this discrepancy, Jakes and White (1971) have suggested that magmatic arcs only evolve a mature continental structure late in an evolutionary history. They argue that partial melting of thickening arc roots would lead to gross stratification of the whole crustal structure into the upper and lower tiers characteristic of continental crust. The upper crust would in this way be suitably enriched in the more readily fusible elements. Further thickening and maturation of sialic crust may occur in the interior of plates if aging of continental lithosphere is an important phenomenon.

Precambrian greenstone belts and granitic belts can be interpreted as complex magmatic arcs (for example, Goodwin, 1971). Precambrian radiometric belts of igneous and metamorphic rocks may represent some combination of (A) accretionary progressions of volcano-plutonic orogens marginal to early shield nuclei, and (B) peripheral orogenic belts where radiometric clocks were reset during crustal collisions. Truncated radiometric belts may reflect either transform offsets or continental separations at various times in the past. Analysis of the distribution and paleomagnetism of different radiometric belts on the various continents may reveal patterns of plate tectonic regimes responsible for the assembly and fragmentation of successive composite continents and paired continental fragments during the Precambrian.

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