PLATE TECTONICS AND SEDIMENTATION

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ABSTRACT

The theory of plate tectonics offers a fresh opportunity to interpret the evolution of sedimentary basins in terms of changing plate interactions and shifting plate junctures. Although plate-tectonic theory lays primary emphasis on horizontal movements of the lithosphere, large vertical movements are also implied in response to changes in the thickness of crust, in the thermal condition of lithosphere, and in the isostatic balance of lithosphere over asthenosphere. As thick sedimentation requires either an initial depression or progressive subsidence to proceed, the auxiliary vertical movements largely control the evolution of sedimentary basins. Ancillary geographic changes related to the governing horizontal movements also affect patterns of sedimentation strongly.

The geosynclinal terminology used prior to the advent of plate tectonics is inadequate to describe fully the plate-tectonic settings of sedimentary basins. Basins can be described instead in terms of the type of substratum beneath the basin, the proximity of the basin to a plate margin, and the type of plate juncture nearest to the basin. Intraplate settings of oceanic or continental character contrast with zones of plate interaction, which include those of divergent, convergent, and transform motions and within each of which the underlying crustal structure is or may be complex. The evolution of a sedimentary basin thus can be viewed as the result of a succession of discrete plate-tectonic settings and plate interactions whose effects blend into a continuum of development.

Oceanic basins contain an assemblage of diachronous facies whose relations are controlled by thermal subsidence of the lithosphere as it moves away from midoceanic rises. Rifted continental margins undergo successive stages of structural evolution as the following features are formed: prerift arch, rift valley, proto-oceanic gulf, narrow ocean, and open ocean. Sedimentary phases related to each stage are components of the rifted-margin prism of strata that masks the continent-ocean interface beneath a continental terrace-slope-rise association or a progradational continental embankment. Marginal fracture ridges along marginal offsets and aulacogens along failed arms of triple junctions locally break the continuity of rifted-margin prisms. Sedimentary basins associated with arc-trench systems where oceanic lithosphere is consumed include trenches beyond the subduction complex beneath the trench slope break, forearc basins in the arc-trench gap, intra-arc basins within the magmatic arc, and interarc basins or retroarc basins in the backarc area. Interarc basins are oceanic basins between a migratory intraoceanic arc and a remnant arc, whereas retroarc basins rest on continental basement adjacent to a foreland fold-thrust belt behind a continental margin arc. Peripheral basins adjacent to suture belts formed by crustal collision occur in an analogous foreland setting between orogen and craton, but in front of a colliding magmatic arc. Retroarc basins and peripheral basins both imply partial subduction of continental lithosphere. Intracontinental basins include infracontinental types, beneath which incipient continental separation gave rise to crust of transitional thickness, as well as supracontinental types.

INTRODUCTION

The theory once termed the *new global tectonics* (Isacks and others, 1968), which postulates a segmented and mobile lithosphere, is no longer new. Most geologists apparently accept its main tenets as valid, together with the corallary concepts of continental drift and seafloor spreading, transform faults (Wilson, 1965) and subduction zones (White and others, 1970), although some geologists, notably Beloussov (1970) and Meyerhoff (Meyerhoff, 1972), have challenged these fundamental concepts. *Plate tectonics* (McKenzie, 1972a; Dewey, 1972) has become an alternate designation for the new global tectonics because the discrete spherical caps of essentially rigid lithosphere inferred to be in relative motion with respect to one another, and to the softer and weaker asthenosphere beneath, are commonly called *plates* (McKenzie and Parker, 1967). The characteristic patterns of lateral motion of these surficial slabs, curved to conform to the spherical outline of the earth, were described by Morgan (1968) as motions of crustal blocks. He indicated, however, that these fundamental tectonic entities are actually blocks of *tectosphere*, a layer thicker than crust in the ordinary sense of the layer above M and essentially synonymous with the lithosphere of others. Because of the large lateral dimensions of the main intact pieces of lithosphere, which are of the order of only 100 km thick, the passage of time has favored usage of the word plate, rather than block, as the basic descriptor.

As a comprehensive theory that purports to explain the global distribution of all belts of tectonic deformation within the crust as the loci of the boundaries or junctures between plates of lithosphere, plate tectonics has the flavor of a fresh paradigm that must be accepted or rejected almost in its entirety with only modest allowances made for deviant behavior. The evident tectonic complexity of the earth admittedly forces the recognition that unusually broad zones of deformation occur along some plate junctures (Atwater, 1970), that the motions of some small plates are controlled partly by the interaction of adjacent large plates (McKenzie, 1970, 1972b; Roman, 1973a, 1973b), and that intraplate deformation is possible on a limited scale or to a limited degree (Sykes, 1970; Doyle, 1971). These adjustments within the framework of plate-tectonic theory dilute its elegance somewhat but do not challenge its fundamental premises. Moreover, the history of mountain belts is better illuminated by plate tectonics than by any preceding theory (Coney, 1970; Dewey and Bird, 1970a).

Concepts derived from plate tectonics are used here as the basis for a discussion of general relations between tectonics and sedimentation. Plate tectonics offers fresh ways to explain the evolution of sedimentary basins, and many concepts of the past can be discarded or must be modified to conform to the new point of view. The development of plate-tectonic interpretations and models of sedimentary basins thus entails the mental exercise of changing outworn interpretations and unjustified conclusions without denying established facts. Application of plate-tectonic analysis to the evolution of a specific sedimentary basin also requires the uniformitarian assumption that present styles of plate-tectonic behavior are useful keys to plate-tectonic behavior during the time span represented by the evolution of the basin.

Unfortunately, there seem to be no clearcut means yet to judge when plate-tectonic behavior of the modern sort began, or whether somewhat different forms of plate interactions prevailed at different times in the past. Events that could conceivably mark times of tectonic transition when plate tectonics could have been initiated or could have undergone some change in kind include (a) the breakup of Pangaea starting roughly a quarter of a billion years ago (Dietz and Holden, 1970), (b) the formation of the oldest recognized blueschist belts (Ernst, 1972) and ophiolite sequences (Burke and Dewey, 1972) about a half billion years ago, (c) a null in the reported frequency of radiometric dates for orogenic granitic and metamorphic rocks at about three-quarters of a billion years ago, (d) the development of the oldest lasting cratons during the Precambrian, or (e) the formation of the first cratonic nuclei deep in the Precambrian (see also Burke and Dewey, 1973).

Perhaps the most revolutionary facet of plate-tectonic theory as applied to sedimentary basins is the startling light it sheds on the tempo of major geologic events. At ordinary rates of spreading along midoceanic rises and of plate consumption at trenches (Le Pichon, 1968; Chase, 1972), oceanic basins 5000 km wide can form or, once formed, can disappear within 50 to 100 my, a span of time representing only one or two periods of the standard geologic column. It follows that no sedimentary basin with a long history of deposition is likely to have remained in the same plate-tectonic setting throughout its evolution. Realization of this principle is a vital guard against oversimplified versions of local geologic history in terms of plate tectonics. From a plate-tectonic standpoint, the Phanerozoic alone is an immense span of time, nominally long enough to open and close an ocean as broad as the Atlantic five or ten times!

Among the many things that might be written about plate tectonics and sedimentation, this paper discusses the following topics: (a) the vertical movements of lithosphere that are inherent in plate tectonics and required to set the *conditions for sedimentation*, (b) the *ancillary effects* of horizontal movements of lithosphere described as continental drift and seafloor spreading, (c) the problem of translating the *basin terminology* employed by geosynclinal theory into terms compatible with plate-tectonic theory, (d) the main *plate-tectonic settings* important for sedimentation, and (e) the gross outlines of *basin evolution* implied by the concept of plate tectonics.

CONDITIONS FOR SEDIMENTATION

Thick sedimentation in a given place implies the prior existence of a deep hole into which sediment can be dumped or progressive subsidence of the substratum to accommodate successive increments of strata. The formation of either kind of sediment trap on a large scale requires pronounced vertical movements of the earth's crust. Plate-tectonic theory as a geometric paradigm to explain tectonic patterns lays special emphasis instead on grand horizontal translations of lithosphere with its capping of crust. However, major vertical motions of crust and lithosphere are required to accompany the horizontal motions by any feasible geologic interpretations of the mechanisms of plate motions and interactions. The vertical motions are related to changes in crustal thickness, in thermal regime, and in the conditions for isostatic balance. These three facets of plate-tectonic theory postulate inherent vertical motions of an order and on a scale that no previous tectonic theory can match in overall scope. Despite its quite proper formal emphasis on horizontal translations of lithosphere, plate-tectonic theory thus also affords the best theoretical basis yet devised to account for grand vertical movements of crust and lithosphere. From the standpoint of sedimentation, therefore, it is a mistake to view the emphasis on horizontal motions as a potential weakness of plate-tectonic theory. The ancillary vertical motions induced by plate interactions are fully equal to the demands of data on sedimentary basins and their provenances.

Crustal thickness.—New oceanic crust and lithosphere is formed continually at spreading centers along midoceanic rise crests and within marginal or interarc basins behind migrating arc structures. When continental blocks are rifted apart by extensions of spreading centers, thin oceanic crust of igneous origin is thus created adjacent to thick continental crust by submarine volcanism and associated intrusions. Frequency curves of crustal elevation show that the floor of such a new oceanic basin should stand typically about 4 km below the mean surface level of the two continental fragments formed by the rift. The process of continental separation can thus form a new crustal depression capable of serving as a receptacle for sediment, and in principle such a sediment trap can form adjacent to any part of a continental block as potential provenance. In detail, continental separation involves the development of a belt of transitional crust and lithosphere between each continental fragment and the adjacent oceanic basin. The width of the transitional region is not well known, but is probably 100 to 250 km wide in typical cases. Prior to sedimentary loading, the transitional crust will presumably stand at elevations intermediate between those of the continental block on one side and the oceanic basin on the other. Off Norway, Talwani and Eldholm (1972) described a broad transitional region in water depths of 1000 to 2500 m.

Studies of incipient continental separations suggest that two types of processes contribute to the development of belts of *transitional crust* of intermediate thickness (fig. 1):

(1) Attenuation of continental crust by stretching is accomplished by extensional faulting at upper crustal levels accompanied probably by pseudoplastic flowage at deep crustal levels (Lowell and Genik, 1972); sediment deposited on this type of transitional crust will rest upon basement rocks of continental character, but not upon a continental block of ordinary crustal thickness.

(2) Crust with oceanic affinities but unusual thickness forms where sedimentation contemporaneous with volcanism within an incipient rift depression helps construct a crustal profile of mingled sedimentary and igneous components in a complex of lavas, dikes, sills, and sediments (D. G. Moore, 1973); sediment deposited later on this type of transitional crust will rest upon a substratum of oceanic character having a crustal profile that may be of nearly continental thickness.

Plate interactions related to the consumption of lithosphere at arc-trench systems, rather than its creation at rise crests, can also produce crust of anomalous thickness that differs from both the normal oceanic value of 5 to 10 km and the normal continental value of 30 to 40 km. Such anomalous crust can be either of fundamentally oceanic or of fundamentally continental character in terms of its rock components. *Anomalous crust,* in the sense the term is used here, can form by either of two basically different mechanisms that are tectonically linked only as two contrasting facets of the geologic processes that operate within arc-trench systems (fig. 2):

(1) Igneous materials are added to the crustal structures of magmatic arcs, either as volcanic components of the surficial edifice or as intrusive components within the crustal roots of the arcs. Presumably by this process, the crustal thickness beneath intraoceanic arcs like Tonga-Kermadec (Shor and others, 1971, fig. 3) and the Marianas (Murauchi and

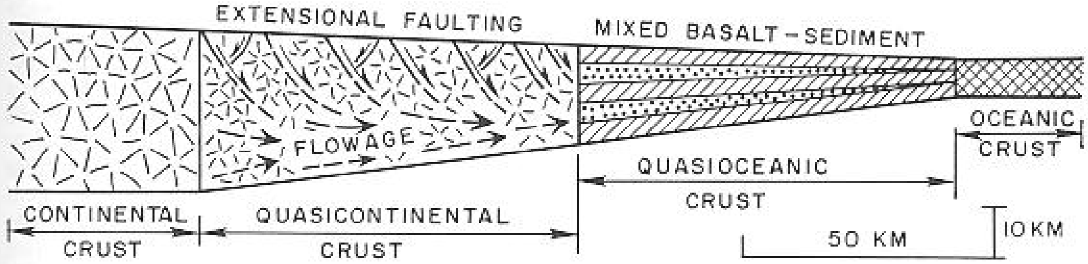

FIG. 1.—Schematic diagram at true scale to illustrate concepts of quasicontinental and quasioceanic types of transitional crust along a rifted continental margin (see text for discussion). Either type of transitional crust may be dominant to the near exclusion of the other type in specific cases (*e.g.,* see Lowell and Genik, 1972, figs. 3, 5 for quasicontinental and Moore, 1973a, fig. 10 for quasioceanic).

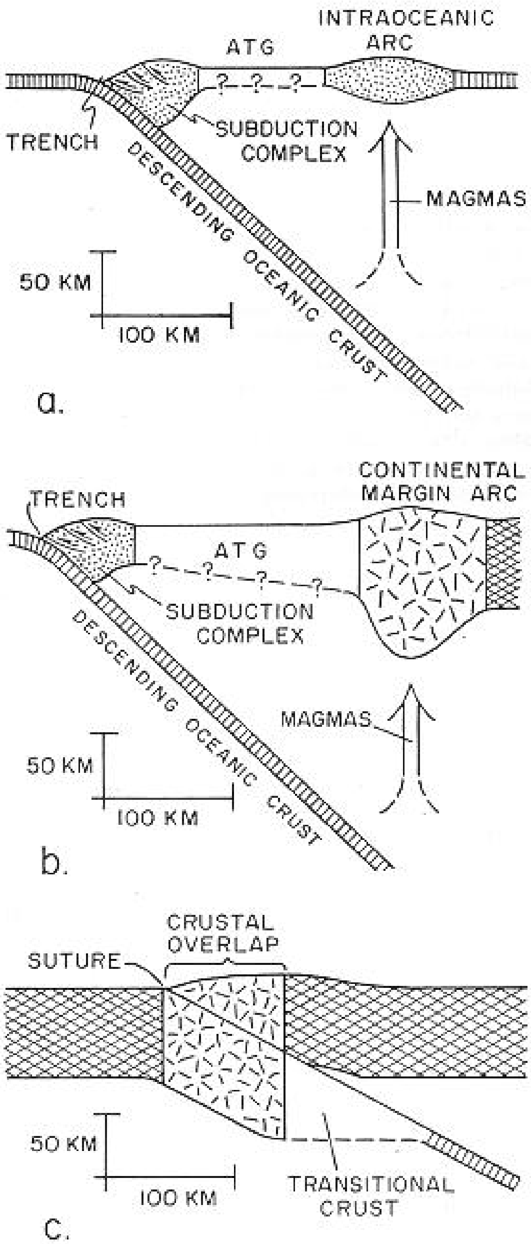

FIG. 2.—Schematic diagram at true scale to illustrate concepts of paraoceanic and paracontinental types of anomalous crust (see text for discussion): *a*, intraoceanic arc-trench system; *b*, continental margin arc-trench system; *c*, suture belt formed by crustal collision when continental block on descending plate encountered subduction zone. Oceanic crust is ruled and continental crust is cross-hatched; paraoceanic crust formed of overthickened oceanic elements is stippled and paracontinental crust formed of overthickened continental elements is jackstrawed. ATG denotes crust of variable and uncertain thickness within arc-trench gap.

others, 1968, fig. 3), whose deep underpinnings are probably oceanic crust, can be increased from the normal oceanic range to a thickness of 12 to 15 km, or perhaps even to 15 to 25 km, as argued by Markhinin (1968) for the Kuriles. Beneath the continental margin arc of the central Andes (James, 1971), the unusually thick crust of nearly 75 km likely also includes major contributions of magmatic rock injected from below into pre-existing continental crust during arc activity.

(2) Oceanic crustal slabs are stacked tectonically in subduction zones to produce thickened crust, or subduction can cause actual overlapping of continental blocks. In California, the subduction complex of the Franciscan assemblage (Ernst, 1970) is a structurally scrambled terrane of oceanic materials (Hamilton, 1969). Included are ophiolitic scraps, deformed seamounts, oceanic pelagites, and turbidite graywackes in mélanges (Hsu, 1968) and thrust slices with a total apparent tectonic thickness of 20 to 30 km (Hamilton and Myers, 1966). No basement rocks of continental character have been detected either within or beneath the complex. In Tibet, the unusual crustal thickness of as much as 75 km has been attributed by Dewey and Burke (1973) to crustal thickening that was effected by essentially doubling up the continental crust. A northern extension of the Indian subcontinent apparently was carried beneath the Tibetan plateau from a subduction zone marked now by the suture belt of ophiolitic mélanges and other deformed oceanic materials along the Indus line between the Himalayas and the Trans-Himalayan ranges.

Changes in crustal thickness within arc-trench systems where lithosphere is consumed thus tend uniformly, although in diverse fashion, to produce thicker crust. This trend fosters isostatic uplift and thus, potentially, the creation of elongate highlands as major sources of sediment. Intraplate crust can also thicken with time—certainly in oceanic regions and possibly in continental ones. In the oceans, the construction of volcanic chains like that of Hawaii on top of previously formed lithosphere can roughly double the thickness of the crust locally. By contrast, continental separations and arc migrations promote crustal attenuation and produce thinner crust associated with newly formed lithosphere. From considerations of the isostatic balance of crust, taken in isolation from other factors, spreading centers are thus generators of sites of potentially thick sedimentation.

Thermal regime.—Plate-tectonic behavior involves convective motions of asthenosphere and lithosphere (Elsasser, 1971), regardless of whether some sort of triggering perturbation of the system is induced primarily by tidal forces governed by astronomic relations (Bostrom, 1971; Knopoff and Leeds, 1972; G. W. Moore, 1973). Convectional overturn of mantle material causes relative uplift and subsidence of the surface of the lithosphere in places whose locations are partly independent of local crustal thickness.

The magnitude of the thermal effect is best understood for the elevation of oceanic crust, which stands at shallow depths beneath active rise crests and at progressively greater depths down the flanks of the rises (Sclater and others, 1971). As the age of the oceanic crust can be inferred from the correlation of magnetic anomalies, rates of subsidence can be estimated empirically. The crests of midoceanic rises have depths of 2.5 to 3 km, but all ocean floors that are roughly 75 my old and lack much sediment cover have a depth of about 5.5 km; oceanic crust of intermediate age stands at intermediate depths related in a regular fashion to age. Rates of subsidence are initially almost 100 m/my but decline with time towards a figure of 10 m/my. The observed subsidence can be explained well simply by the thermal contraction of a cooling lithosphere that is about 100 km thick. The calculations assume that isostatic compensation takes place at the base of the slab of lithosphere where it is in contact with the asthenosphere. Various assumptions for the conductivity and basal temperature of a slab of lithosphere 75 to 100 km thick allow subordinate contributions to crustal elevation from phase changes in the slab and from convective bulging of the asthenosphere beneath the slab.

Thermal tumescence along intracontinental rifts prior to continental breakup, and the succeeding thermal decay following continental separation, also cause major uplift and subsidence of continental basement rocks (Sleep, 1971). An initial thermal uplift of the order of 1 to 2.5 km can be inferred, and is matched well by the observed domal uplift of 1.5 km during the late Tertiary in central Kenya along the East African rift system (Baker and others, 1972). Crustal thinning by erosion of the arched region along an incipient rift may contribute significantly to the net crustal attenuation associated with continental separations (Hsu, 1972). The duration of purely thermal subsidence along a fresh continental margin newly formed by rifting is unlikely to persist for more than 100 my, and for typical continental ruptures the major effects probably occur within 50 my while the continental margin is within 1000 km of the rise crest (fig. 3).

Potential activators of crustal uplift and subsidence traceable to changing thermal regimes

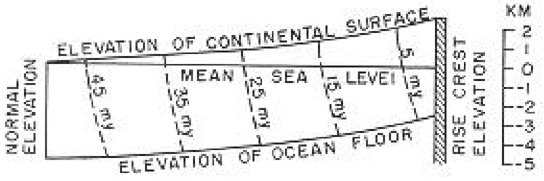

FIG. 3.—Schematic diagram to illustrate subsidence of a rifted continental margin as it moves away from a rise crest. Dashed lines show successive idealized positions of the continental slope at the intervals of time indicated. Width of diagram is about 1000 km for a half-spreading rate of 2 cm/yr; effects of sedimentation are ignored.

include the poorly understood hotspots, whose positions Morgan (1972) ascribed to fixed advective plumes or columns of hot material rising from the deep mantle. If hotspots form bumps on the upper boundary of the asthenosphere, the motions of mobile lithosphere may cause parts of the plates to bob up and down as they cross over the sites of the underlying hotspots (Menard, 1973a). Epeirogenic warping amounting to hundreds of meters vertically, and with wavelengths of the order of 1000 km or more, may be attributable to such a phenomenon. The eventual impact of recent analyses (Burke and others, 1973; Molnar and Atwater, 1973) showing that all supposed hotspots cannot be fixed in position relative to one another is not yet clear. In discussing midplate tectonics, Turcotte and Oxburgh (1973) have offered alternative explanations for the origin of the linear island and seamount chains whose relations the concept of hotspots purports to explain. With fixed hotspots, the unidirectional extension of these volcanic chains is interpreted as a result of the passage of plates of lithosphere over fixed hotspots below. However, the same general effects can be achieved in theory by postulating the development of propagating cracks in the lithosphere induced by thermal stresses from the cooling of slabs of lithosphere and by membrane stresses from changes in the radii of curvature of spherical caps of lithosphere as they change latitude on the globe. Regardless of how the hotspot controversy is resolved, the possibility of epeirogenic warping of lithosphere in irregular patterns as plates pass over a bumpy asthenosphere remains open (Menard, 1973b), unless the boundary between lithosphere and asthenosphere is assumed arbitrarily to be uniformly smooth.

Isostatic balance.—In the past, isostatic reasoning commonly has been applied by assuming the base of the crust at M to be the level of compensation. To the extent that slabs of litho-

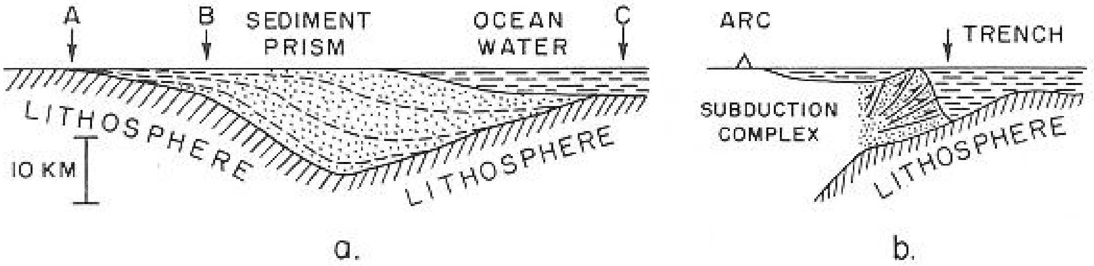

FIG. 4.—Schematic diagrams to illustrate subsidence of substratum by flexural bending of lithosphere under surficial loading: *a*, downbowing of continental margin owing to load of sediment prism deposited offshore (after Walcott, 1972, fig. 2) where A is inland line of flexure and B is initial edge of continental block before marginal subsidence (note that substratum at B is depressed to a depth at roughly the same level as that of normal ocean floor at C); *b*, hypothetical downbowing of ocean floor offshore from load of subduction complex stacked tectonically in subduction zone associated with trench beneath which oceanic substratum is depressed prior to final descent into the mantle (after Hamilton, 1973, fig. 1 and note added in proof); vertical exaggeration 10 ×.

sphere are literally rigid, the base of the lithosphere is a more appropriate level of compensation to choose. In reality, perhaps, given the limited strength of rock masses, assumption of partial compensation at both those levels, and likely at others as well, may prove the most useful stance to adopt. In any case, our past tendency to think of isostasy in terms of crustal balance alone is invalid.

If we may speak, nevertheless, of crustal isostasy in isolation from the broader context, then depressions associated with plate consumption and subduction zones are anisostatic, in the sense that the elevation of the top of the crust is not there a function of crustal thickness and density alone. Instead, the overall motion of a descending slab of lithosphere compels the crustal elements in the top tier of the slab to follow downward. Where oceanic lithosphere is consumed, the trenches are 2.5 to 5 km deeper than the floors of the adjacent oceanic basins; despite this marked difference in elevation, the crustal profile beneath the oceanward slope of the trench is demonstrably the same as in the open ocean. Beneath the axis of the trench, the ponding of turbidites may even increase the thickness of the crust there, although in cases where this effect is dominant a bathymetric trench may not be present as an expression of subduction.

McKenzie (1969) has argued convincingly that the presence of continental blocks prevents plate consumption by simple gravimetric resistance to plate descent. Even so, the attempted subduction of a continental margin, or of continental lithosphere generally, though arrested at a stage short of actual plate consumption, may be able to accomplish appreciable anisostatic subsidence, as that term was used above with reference to oceanic trenches. The local subsidence of a continental surface standing initially near sea level to nearly oceanic depths along a linear belt seems conceivable if the excess depth of 2.5 to 5 km noted for trenches can be extrapolated to this roughly analogous setting. Even if conditions of deep water were never attained, a linear belt of thick sediments deposited on subsiding continental basement might develop as a result of partial subduction. Such a region would presumably appear in the geologic record as a mobile pericratonic fringe bordering an otherwise stable craton.

Walcott (1972) has shown also that flexural bending of lithosphere under sedimentary loading of oceanic and transitional crust just offshore from a rifted continental margin can cause marked subsidence of continental basement along the adjacent edge of the continental block (fig. 4a). As sediments accumulate off the continental margin, the isostatically compensated lithosphere sags downward, and the upper surface of the continental block tilts seaward. The depressed belt along the continental margin may be 200 km wide inward from the initial continental edge to the line of no vertical displacement within the continental block, and the substratum at the initial edge of the continent can become buried beneath as much as 4 km of sediment deposited as an elongate wedge thickening seaward within the depressed belt. Landward of the depressed belt is a gentle linear upwarp (not shown on fig. 4a) parallel to the axis of the linear sag in the lithosphere offshore.

Elevation changes related to crudely analogous flexures of the lithosphere may occur around arc-trench systems in response to tectonic loading represented by the buildup of tectonically stacked subduction complexes. Hamilton (1973) argues that the weight of a seaward-thinning wedge of mélanges forming the subduction complex bows down the descending plate of lithosphere oceanward of the subduction zone (fig. 4b). This action would tend to reduce the angle of plate descent near the surface, as accretionary lateral growth of the welt of mélanges covered the initial site of plate consumption, and perhaps to increase the depth of the trench as a result of the sag developed in the lithosphere.

A complication of uncertain significance colors all detailed considerations of the behavior of lithosphere. Several kinds of data, and especially those on terrestrial heat flow, suggest that continental lithosphere is thicker than oceanic lithosphere and perhaps as much as twice as thick (Scalater and Francheteau, 1970). If so, important questions about the origin of continental lithosphere and about the motions of plates of lithosphere over the asthenosphere are raised. In any case, rigorous treatment of the isostatic balance of lithosphere over asthenosphere clearly cannot be attempted until possible variations in the thickness of the lithosphere are better known.

ANCILLARY EFFECTS

The sedimentary record is only partly a result of paleotectonic conditions suitable for sedimentation. Paleogeographic relations govern to a large extent the nature of the sediment accumulated in a given place at a given time. The factors that govern geographic variations with time are largely ancillary side effects of plate tectonics. The main influences of this kind are related to changes in latitude, changing patterns of geography, and eustatic changes.

Research on paleomagnetism (McElhinny, 1973) indicates that drifting continents have changed latitude drastically during the course of geologic history. Unless one supposes a wholly uniform climate from equator to pole at times in the past, this conclusion implies that each continental block has moved through fundamentally different climatic belts during its history. In general, reconstructed paleolatitudes also cross each continental block in different directions for different times. It follows that an adequate analysis of any sedimentary basin must include the recovery of the paleolatitudes of the basin for the times of interest. Where long periods of time are represented by the sedimentary sequence, a graph showing the trend of changing paleolatitudes for the center of the basin, or for the ends of an elongate basin, may well prove to be essential for a full interpretation of sedimentation.

Changing geographic patterns arising from continental drift may exert important influences on the distribution of potential sediment sources. Patterns of oceanic circulation should also be affected, as well as patterns of atmospheric circulation related to rain shadows and other important effects. Unfortunately, a full assessment of these types of influences on sedimentation in a particular basin at specific times in the past must await the development of a sequential atlas of paleogeography on a globally integrated basis. For much of the Phanerozoic our knowledge is still inadequate to shape this goal even with respect to the positions of all the parts of the present continental blocks. We may never be able to reconstruct well the configurations of the floors of vanished ocean basins, which may have harbored rises for which no clear evidence remains.

Eustatic changes in sea level stemming from ice storage in polar regions are probably modulated, in the long view, by the movement of continental blocks into and out of positions where they can support large glaciers; distributions of other continents so as to block latitudinal circulation in the oceans probably also favor extensive glaciation on the continents located at high latitudes (Crowell and Frakes, 1970). The remarkable display of cyclic sedimentation in the late Paleozoic sequences deposited at low paleolatitudes on coastal plains and in epeiric seas of North America and Europe can be ascribed tentatively to fluctuations of the glaciation at high paleolatitudes in Gondwana (e.g., Wanless and Shepard, 1936).

The glacial explanation of eustatic effects relies upon changing volumes of ocean water coupled with constant volume of the ocean basins. In recent years, various authors (*e.g.*, Valentine and Moores, 1972) have speculated that changes in the globally integrated spreading rate along midoceanic rises can cause eustatic effects by changing the volume of the ocean basins while the volume of ocean water remains constant. The root of the idea rests upon the principle that oceanic crust subsides with age; therefore, if the mean age of the oceanic crust changes with time, the mean depth will also vary. Evaluation of the effect is difficult (e.g., Johnson, 1971), both because we lack sufficient data to estimate globally integrated past spreading rates closely on an areal basis and because the available worldwide data on areal flooding of continental blocks at specific times in the past remains partly equivocal (but see the paper by Sloss and Speed in this volume for a fresh synthesis and a unique interpretation). The whole question of continental freeboard through time is discussed provocatively by Wise (1972).

BASIN TERMINOLOGY

Any field of human inquiry requires a sort of code of simple words or phrases to denote complex concepts. Without this aid to brevity all communication becomes too tedious to pursue. When the underlying framework of concepts changes, the established code begins to lose meaning. Something of this sort has occurred in the past few years as the geosynclinal theory for sedimentary basins and orogeny has given way to plate-tectonic theory. Although the rocks, which are the substance of the geologic record that we discuss, remain the same, the way that we view their evidence has changed. During the present period of transition in concepts, there are two fundamentally opposed ways to proceed with description. One course is the adoption of a wholly new terminology for sedimentary basins. The other course is the adaption of the old terminology to reflect the new concepts. In practice, the most likely path of thinking is a middle course that blends the two approaches by borrowing from the old where convenient and inventing the new where necessary. In practice, also, a quick consensus of views is unlikely, for many thoughtful workers will offer conflicting terminological schemes, each with its own flavor, strengths, and weaknesses (e.g., Mitchell and Reading, 1969; Dewey and Bird, 1970; Dickinson, 1971a).

The most challenging obstacle to a clean translation from geosynclinal to plate-tectonic terminology stems from the two meanings of the word *basin* in geological science. In one sense a basin is merely a bathymetric or topographic depression, but in a more significant sense a basin is the prism of rock forming a thick sedimentary succession. Various types of bathymetric basins can be related readily to the current global pattern of plate tectonics, but sedimentary basins can be related to plate tectonics only by deductive reasoning that postulates the historical dimensions of plate interactions. On the other hand, the existence of sedimentary basins is the starting point for geosynclinal theory, and their relationship to bathymetric basins must be inferred from inductive reasoning.

The dominant theme of geosynclinal theory is the recognition of parallel and adjacent miogeosynclinal and eugeosynclinal belts (Kay, 1951). Although several other kinds of geosynclines have been named, their designations arise as extensions of terminology to encompass sequences whose history does not conform to the ruling concepts for the recognition of these two basic stratigraphic elements of orogenic belts. In general, miogeosynclinal terranes are characterized by clearcut depositional contacts with continental basement and by a lack or paucity of turbidites and interstratified volcanic rocks. By contrast, eugeosynclinal terranes are characterized by equivocal contact relationships with continental basement and by an abundance of turbidites and volcanic rocks within the sedimentary sequence. As a first approximation, the former can thus be interpreted as thick accumulations of strata on the margins of continental blocks and the latter, as strata formed somewhere within an adjacent oceanic basin including its island arcs. This loose approach to the translation of geosynclinal terminology into terms compatible with plate tectonics is not entirely satisfactory. It does not allow for the considerable sophistication of geosynclinal theory in full flower, and results in the unnecessary lumping of things that the full geosynclinal terminology accords different status. Nor does it meet the need to relate various types of sedimentary basins to different kinds of plate interactions, rather than just to the two main kinds of substratum.

PLATE-TECTONIC SETTINGS

In terms of plate tectonics, the settings of basins can be described with reference to three fundamental factors: (1) the type of crust and lithosphere that serves as substratum for the basin; (2) the proximity of the basin to a plate margin, and (3) the type of plate juncture or junctures nearest to the basin.

Types of substratum.—In terms of immediate substratum, normal *continental* crust and standard *oceanic* crust are clearly two end members. For the *transitional* crust discussed in an earlier section, the term *quasicontinental* is applied here to the type characterized by attenuated continental basement, and the term *quasioceanic* is applied to the type characterized by an overthickened profile of plainly oceanic elements including both igneous and sedimentary materials (see fig. 1). The terms *paracontinental* and *paraoceanic* are applied, respectively, to *anomalous* crust of previously continental or quasicontinental, and previously oceanic or quasioceanic, character where crustal thickening has occurred by the addition of igneous materials in magmatic arcs and hotspot chains or through processes of tectonic stacking related to subduction zones (see fig. 2). There is inherent ambiguity between arc-related and subduction-related subtypes of paracontinental and paraoceanic anomalous crust because both magmatic arcs and subduction zones can migrate with respect to the intervening sliver of lithosphere (Dickinson, 1973). The different categories of crust are largely conceptual at present, for the operational means to distinguish between them with geophysical observations remain elusive for the most part. Moreover, the terms are unlikely to suffice as a full catalogue of significantly different types of crust and associated lithosphere. They would prove to be especially inadequate, and perhaps misleading to some extent, if it develops that significant exchange of substance can occur between crust and mantle, or between lithosphere and asthenosphere, in settings that lie within intact plates.

Proximity to junctures.—The degree of proximity of a basin to a plate margin must be understood in relative rather than absolute terms. The point is the extent to which tectonic effects related directly to plate interactions influence the setting of the basin. The thermal decay of the lithosphere as it moves away from a spreading center is one such effect, which will be confined within a certain distance from the rise crest depending upon the spreading rate. Similarly, the locus of arc magmatism along a certain trend parallel to the associated trench will occur at a distance that depends upon such parameters as the rate of plate consumption and the dip of the inclined seismic zone. In broad terms, basin settings can thus be divided into *intraplate* settings as opposed to *zones of plate interaction*.

Plate junctures.—There are three basic varieties of plate junctures: (1) *divergent,* where old lithosphere separates at spreading centers, and new lithosphere is built along midoceanic rise crests to fill the opening gap by accretion of material to the retreating edges of the separating plates; (2) *convergent,* where plate consumption carries old lithosphere downward into the asthenosphere along inclined seismic zones, and the processes that operate in the subduction zones and magmatic arcs modify the lithosphere of the overriding plate by adding both magmatic and tectonic increments to its profile; and (3) *transform,* where two plates slide past one another along a lateral fault zone without either forming new lithosphere or destroying old lithosphere.

The three kinds of junctures are geometric end members and are analogous, in terms of the geometry of strain indicated, to the three familiar classes of faults: normal (extensional), reverse-thrust (contractional), and strike-slip (lateral). There are variants of the three types of plate junctures where motions oblique to the trends of the junctures occur (Dickinson, 1972). Obliquity of convergence, as indicated independently by slip vectors determined for earthquakes and by calculations of relative plate motions from correlations of magnetic anomalies at sea, is common along modern subduction zones. Obliquity of divergence, however, is commonly resolved into a rectilinear lattice of rise crests and connecting transforms for apparently mechanical reasons (Lachenbruch and Thompson, 1972). Along transforms where the relative motion of the plates has a component of divergence, the same mechanical tendency evidently fosters a similar rectilinear lattice of transform segments connected by short rise segments. Along transforms where the relative motion of the plates has a component of convergence, the combined effect has been called transpression (Harland, 1971).

Some incipient plate junctures may become inactive before developing the full characteristics of their class. For example, an aborted divergent juncture within a continent might never undergo sufficient plate separation to develop a fully oceanic crust and lithosphere. A sedimentary basin with an unexposed floor of transitional crust might well form above the site of such a feature. Although clearly intracontinental in its setting, such a basin can be described as *infracontinental* as opposed to *supracontinental* basins deposited on a full complement of continental basement. Similarly, plate consumption might be arrested at some stage of partial subduction before a magmatic arc was developed. As the confident application of plate-tectonic logic depends upon the recognition of the full display of geologic features characteristic of each type of plate juncture, such partially developed junctures are apt to foster ambiguous interpretations.

Combined parameters.—Considering jointly the parameters of crustal substratum, proximity to plate interaction, and type of plate juncture, the gross settings of sedimentary basins can be grouped into a hierarchy consistent with plate tectonics. Initially, intraplate settings are contrasted with settings within zones of plate interaction, which include zones of divergent, convergent, and transform motion. For each of the four broad classes of plate settings so derived, the nature of the crustal substratum may vary, and subclasses of the settings can be recognized on the basis of the variations:

(a) For intraplate settings, the substratum need not be normal continental or standard oceanic in nature, for transitional or anomalous crust inherited from extinct plate junctures can be present.

(b) Zones of divergence include both intracontinental and intraoceanic types, although the two commonly are merely sequential stages in the evolution of a single plate juncture responsible also for the formation of transitional crust during intermediate stages of its evolution.

(c) Zones of convergence include types, or phases of development, in which either oceanic or continental (or transitional) crust is drawn into the subduction zone, and in parallel also include types in which the anomalous crust of the arc-trench system develops from crust of either oceanic or continental (or transitional) character initially; consequently, arc-trench systems embrace multiple settings with diverse overall relations.

(d) Transform zones include three basic types in which the two plates sliding past one another are both oceanic, both continental, or one continental and one oceanic, but transitional or anomalous crustal blocks may also be involved; moreover, each of the three basic types includes two variants where some auxiliary motion is either divergent or convergent.

The most important classes of geosynclines represent the net development of sedimentary successions over spans of time long enough for the plate-tectonic settings of the sites of deposition to change. Typical steps in the evolution of different classes of geosynclines thus apparently represent characteristic patterns in the evolution of plate interactions and the consequent changes in plate-tectonic setting. Examples of apparently anomalous geosynclinal evolution then represent the results of less common sequences of plate interactions with the different changes in plate-tectonic setting implied thereby.

BASIN EVOLUTION

Geosynclinal theory is forced by its inductive basis to address the analysis of basin evolution in terms of type examples. Where deviations from supposed norms of evolutionary trends occur, the theory is unable to offer clear insights. Plate tectonics, by contrast, approaches the problem of basin evolution in terms of alternate sequential patterns of plate interactions. Given the overall framework of varied plate motions, deductive reasoning from the theory has the potential to shed some insight on quite unfamiliar evolutionary trends, as well as to explain in coherent fashion a range of events that might issue in different circumstances from any particular stage of the evolutionary development of a given type of basin. By treating basin evolution as a function of plate motions and interactions, plate tectonics thus broadens the scope of theoretically controlled analysis and reduces the need for wholly intuitive suggestions.

Plate tectonics as a framework of thinking thus precludes the possibility of a neat catalogue of basin types. Each basin, seen in a developmental perspective of space and time, to some extent partakes of a unique flavor in principle. The constants in the equation of basin evolution are the types of plate interactions and settings, but the order in which they may be arranged in space and time is variable within wide limits. Geosynclinal theory, by contrast, assumes the overall trend of development as the standard, and views the incremental events that occur during evolution as variable within wide limits. In an analogical but very real sense, geosynclines as conceived by classic theory thus have an ontogeny, or life history, driven by tendencies akin in their effects to a vital force. Plate tectonics casts a more prosaic light on sedimentary basins, but allows naturally for more shadings of behavior without the need to modify any of its fundamental tenets.

From the standpoint of plate tectonics, the evolution of sedimentary basins is incidental to the formation and consumption of lithosphere. The major perturbations of a stable and level earth's surface are related to the opening of oceanic basins accompanied by the rifting and fragmentation of continental blocks, and to the closing of oceanic basins accompanied by the collision and assembly of continental blocks. The principal trends of basin evolution can thus be described as they pertain to the following realms of interplay between tectonics and sedimentation: (a) *oceanic basins* underlain by oceanic lithosphere, (b) *rifted continental margins* along the transitional interface between oceanic and continental lithosphere, (c) *arc-trench systems* where oceanic lithosphere is consumed beneath island arcs or continental margins, (d) *suture belts* where continental blocks are juxtaposed by crustal collision, and (e) *intracontinental* basins in the interior of continental blocks. For none of these realms of interplay should one assume invariant modes of evolution, but a discussion of each in order does afford the means to indicate the salient relationships of plate tectonics to basin evolution.

Oceanic Basins

Depending upon their size, the distribution of divergent plate junctures within them, and the distribution of convergent plate junctures within or around them, oceanic basins may lie at any given time either wholly within zones

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Oceanic Basins

Depending upon their size, the distribution of divergent plate junctures within them, and the distribution of convergent plate junctures within or around them, oceanic basins may lie at any given time either wholly within zones of plate interaction or in wholly intraplate settings; most typically, they lie partly within zones of plate interaction but otherwise in an intraplate setting. Each segment of oceanic crust and lithosphere typically experiences the following succession of plate-tectonic settings in order: (1) the zone of plate interaction along a divergent plate juncture where the oceanic substratum is formed; (2) the intraplate setting of a deep oceanic basin; and (3) the zone of plate interaction along a convergent plate juncture where the bulk of the oceanic lithosphere is consumed while variable and uncertain proportions of the oceanic crustal elements are caught up in the subduction zone. During either the initial or final phases of evolution in zones of plate interaction, selected segments of the oceanic crust may be subjected also to deformation along transform plate junctures associated with the divergent or convergent plate junctures.

Ignoring features related to rifted continental margins and to arc-trench systems, the principal settings of oceanic facies controlled by tectonic relations are the following (fig. 5): (a) rise crests where the layered igneous succession of ophiolite sequences (Vine and Moores, 1972) are formed along the trends of the spreading centers, (b) rise flanks where the oceanic substratum gradually subsides as it cools in moving away from spreading centers, and (c) deep basins beneath which the thermal contraction of lithosphere is essentially complete. This gross picture must be modified to allow for special conditions of shearing along active transforms near rise crests and for sharp topographic contrasts across and along fracture zones that mark the inactive extensions of transforms down the rise flanks. The outline of settings may also be inappropriate in detail for the oceanic basins generated by spreading centers within marginal seas or interarc basins behind migrating island arcs.

Nevertheless, the ideal triad of rise crest, rise flank, and deep basin serves to emphasize the main characteristic trends of evolution for an oceanic basin (see fig. 5). The igneous components of the ophiolite sequence formed at the spreading center are the first of a series of diachronous facies typical for oceanic sequences. The pelagic sediment that covers the igneous portion of the ophiolite sequence has a stratigraphy with facies relationships that reflect changing water depths (Berger, 1973). While the rise crest and flanks are above the carbonate compensation depth, calcareous sediment is deposited, but is succeeded by siliceous sediment deposited lower on the rise flanks and in the deep basins. In basins that tap turbidity currents from landmasses, the pelagic layers are covered eventually by turbidites of abyssal plains.

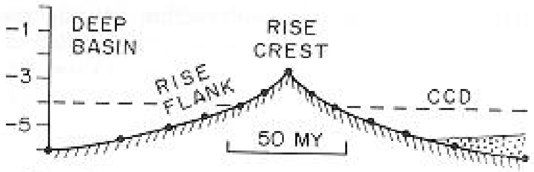

Fig. 5.—Sketch to illustrate principal settings of oceanic facies on transverse profile of typical mid-oceanic rise (depths after Sclater and others, 1971). Vertical scale is in km of water depth, but horizontal scale is in my because lateral dimensions of a mid-oceanic rise are dependent upon the spreading rate. CCD (dashed line) is typical level of carbonate compensation depth (Berger, 1973). Abyssal plain of turbidites indicated schematically, without showing isostatic compensation of substratum, by stippled area on right.

Where the oceanic basin changes latitude during its history or otherwise encounters different oceanic provinces, complexities are introduced into the diachronous sedimentary succession (Frakes and Kemp, 1972; Heezen and others, 1973). For example, owing to the high equatorial productivity of calcareous plankton, the carbonate compensation depth is lowered significantly below the lysocline in a narrow belt along the equator. This phenomenon has potential consequences for an oceanic sedimentary sequence formed on one side of the equator as a doublet of rise-crest calcareous pelagites overlain by siliceous pelagites reflecting later deposition in deeper water. If the segment of the oceanic basin bearing this doublet then crosses the equator, its transit may be marked by the deposition of a layer of equatorial calcareous pelagites. After the segment of the basin has moved away from the equator into the other hemisphere, it will then carry two calcareous-siliceous doublets. The two successive calcareous horizons, each overlain gradationally by siliceous sediment, record the times of positioning at the rise crest and equatorial transit, respectively. Whether details of this kind can ever be read clearly from the deformed oceanic facies of orogenic belts is a moot question at present, but avenues for inquiry are surely open.

A special set of oceanic facies is associated with islands and seamounts built as isolated mounds or in linear chains across oceanic regions. The thick volcanic piles themselves may be capped by reefoid sediment and flanked by archipelagic aprons of volcaniclastic turbidites derived locally. In certain instances, widespread and thick carbonate platforms like those in the

Bahamas may also be built within oceanic regions, probably on the quasioceanic crust of marginal fracture ridges (see below) or the submarine ridges of hotspot-generated island-seamount chains.

One of the most remarkable corollaries of plate-tectonic theory is the inference that all the old oceanic crust—igneous rocks and sediments alike older than the present ocean floors—has been placed into one of three non-oceanic repositories: (1) the mantle, into which crustal materials capable of pressure-induced inversion to suitably dense phases could be swept together with the bulk of the plates of lithosphere consumed through time by descent into or through the asthenosphere; (2) subduction complexes, into which crustal materials could be scraped from the tops of descending plates and thus welded by accretion to the flanks of continental and island-arc crust; or (3) magmatic arc structures, into the roots of which crustal materials melted from the upper levels of descending plates could be fed from below. Given the ages currently estimated for the present ocean floors, this inference means that the presumably immense bulk of all the turbidites in all the subsea fans and abyssal plains of all pre-Jurassic oceanic basins have met one of those fates, of which the second seems the most likely at present.

Rifted Continental Margins

Rifted continental margins form in pairs when continental separations occur along divergent plate junctures, and form singly when magmatic arc structures are rifted away from the margins of continental blocks by spreading behind the arcs. In the former case of simple continental separation, each rifted continental margin presents the juxtaposition of a high-standing continental block with sediment sources against a newly formed oceanic basin to serve as a sediment sink. The resulting sedimentation forms a characteristic sedimentary prism spanning the interface between continental and oceanic crust. Different components of the prism, here called *rifted-margin prism,* rest on continental, transitional, and oceanic crust. The prism thus contains strata of both miogeosynclinal and eugeosynclinal affinities. The nearshore assemblage of mainly paralic and shelf facies resting on continental basement has been aptly termed the miogeocline (Dietz and Holden, 1967) in recognition of the fact that these strata form, in transverse section, a wedge thickening seaward toward the shelf edge in existence at the time of deposition. Similarly, the offshore assemblage of turbidites and other deposits in deep water near the foot of the continental slope can be termed the eugeocline in analogous recognition of the asymmetric form of the thick accumulation, whose site is controlled by the position of the continental margin. However, as these latter deposits may grade imperceptibly into those of the broad oceanic basin nearby, the designation eugeocline is commonly less useful in practice than the term miogeocline.

The rifted-margin prism, when completed, includes a number of distinctive sedimentary phases within a complex assemblage of deposits. Each of the phases reflects either deposition in a particular plate-tectonic setting during the time when the rifted continental margin still lay within the zone of plate interaction along a divergent plate juncture, or else deposition during a particular stage in the growth of the prism during the time when the rifted margin was later in an intraplate setting. Variations arise within the total sedimentary assemblage as rates of spreading during different continental separations vary in relation to rates of sediment delivery to the rifted margins. In principle, the process of accumulation of a rifted-margin prism can also be terminated during any given phase of sedimentation by orogeny. Such orogeny may be related either to the activation of an arc-trench system along the continental margin, beneath which the offshore oceanic lithosphere thus begins to be consumed, or to crustal collision with an arc-trench system that approaches the continental margin by consuming the intervening oceanic lithosphere offshore (Dickinson, 1971b). By assuming that the sedimentation of a rifted-margin prism can be arrested at any stage in the growth of its successive depositional phases by several kinds of orogeny, a broad spectrum of individual geosynclinal developments can be accommodated within the same conceptual framework of plate tectonics. Important complications in the succession of depositional phases within rifted-margin prisms are introduced also by the presence locally of marginal offsets of the continental blocks involved and by the failed or aborted arms of triple junctions distributed along the trend of a rift belt. The marginal offsets may give rise to marginal fracture ridges and the triple junctions, to aulacogens.

The series of plate-tectonic settings that mark the successive stages of the evolution of rifted continental margins can be denoted loosely by the following five terms: pre-rift arch, rift valley, proto-oceanic gulf, narrow ocean, and open ocean (see also Schneider, 1972). The five gradational stages of structural evolution are associated with depositional phases whose strata are intercalated locally as contemporaneous facies. The successive phases of deposition may form markedly diachronous facies along any rifted-margin prism, for the geometry of plate tectonics requires most continental separations to proceed as wedge-like openings, rather than as instantaneous separations along the whole length of given rift belts (Dickinson, 1972).

Pre-rift arch.—During the thermal arching that precedes and accompanies incipient rifting, peralkaline volcanism is characteristic (fig. 6a). This activity is apparently not uniform along the rift belt but is concentrated near the crests of broad domal uplifts, from 250 to 1250 km across, that are spaced like beads with centers at intervals of roughly 1000 to 2000 km along the trend of the rift belt (LeBas, 1971). The balance between the rate of accumulation of such volcanics and the rate of erosion of the thermal arches that they crown is uncertain, but relations in Africa and South America adjacent to the South Atlantic suggest that erosion of the thermal arch is the dominant effect areally. In that region, uplifted terranes of Precambrian basement are prominent along both coasts between the ocean and extensive inland basins in which Paleozoic and Mesozoic strata are preserved on both continents (Burke and others, 1971).

Rift valley.—When sufficient crustal extension affects the arched region, rift valleys begin to form as grabens and half-grabens (fig. 6b). Probably these develop first within the domal uplifts, but later they extend as an essentially continuous branching network along the full trend of the rift belt. In the rift valleys, continental redbeds are intercalated with volcanics that continue to erupt through the growing system of crustal fractures (Scrutton, 1973). Broad regions to either side of the eventual zone of rupture between the separating continents apparently can be scarred by extensional faulting during this time. Large-scale extensional faulting has offset continental basement rocks across a belt 100 to 250 km wide west of the axial rift zone in the modern Red Sea (Hutchinson and Engels, 1972), and the Triassic basins of the Appalachian region lie as much as 250 to 500 km inland from the present continental slope, which can be taken as marking roughly the line of Jurassic continental separation.

Proto-oceanic gulf.—As continued crustal distension induces subsidence along the zone of incipient continental rupture despite continued thermal effects, the floors of the main rift valleys become partially or intermittently flooded to form proto-oceanic gulfs. Restricted conditions in these basins, which are probably still rimmed by uplifts that block delivery of clastic sediment, promote the deposition of evaporites in suitable climates (fig. 6c). Immense thicknesses, as much as 5 to 7.5 km, of evaporites are present in the subsurface beneath parts of the Red Sea (Lowell and Genik, 1972; Hutchinson and Engels, 1972). Buried salt layers that feed extensive diapir fields are present off many North Atlantic coasts (Pautot and others, 1970). Extensive evaporites are known also from coastal basins on both sides of the South Atlantic, where they are apparently correlative and represent dismembered portions of the same elongate and initially continuous evaporite basin (Reyment, 1972). The proto-oceanic evaporites are presumably deposited mainly on transitional crust, probably in most instances of the quasi-continental variety representing attenuated continental basement. Deposition on oceanic crust, or as part of the sedimentary component of quasioceanic crust, could conceivably occur if thermal uplift along the rift belt were sufficiently pronounced during and just after full continental rupture.

Narrow ocean.—Once new oceanic crust and lithosphere begin to form along the belt between two separated continental blocks, fully oceanic conditions are attained in the structural sense (fig. 7). In the geographic sense, however, a distinction can be drawn between narrow oceans and open oceans. In narrow oceans, sediment delivery to a given oceanic site from both bounding continental blocks could con-

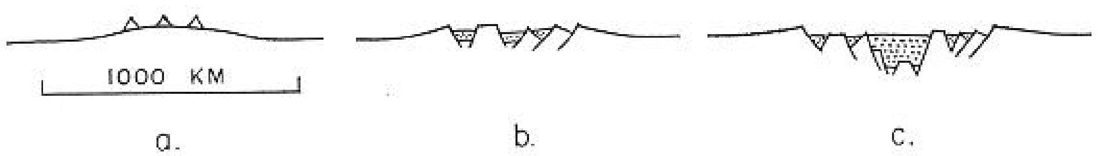

FIG. 6.—Sketches to illustrate successive pre-oceanic stages in evolution of rifted continental margins in cross-section (vertical exaggeration 25× except on dips of faults): *a*, pre-rift thermal arch shown about 750 km across with idealized volcanoes capping it; *b*, rift valley system with terrestrial sediments ponded locally across a belt about 500 km wide; *c*, proto-oceanic gulf with thick saline deposits shown within a belt about 250 km wide. Stipples indicate sediment.

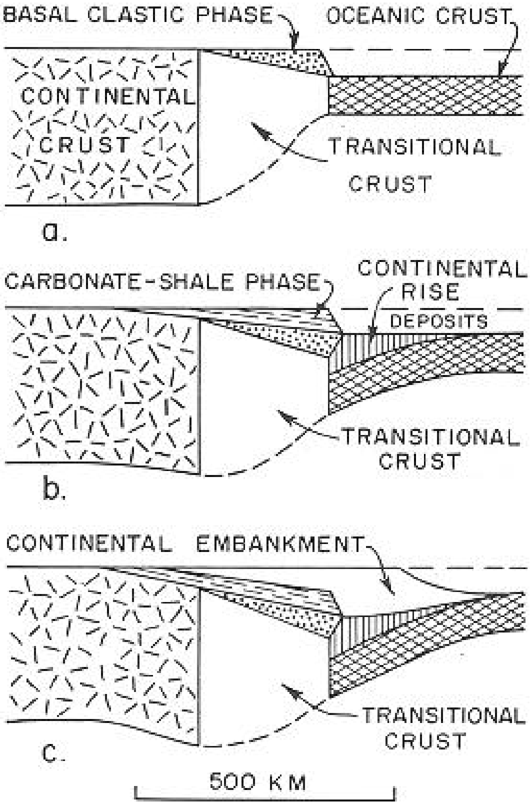

FIG. 7.—Idealized diagrams to illustrate successive depositional phases in evolution of rifted-margin prism along continent-ocean interface with sea level shown as dashed line (vertical exaggeration 10×): *a*, basal clastic phase of miogeocline deposited during thermal subsidence of transitional crust, within which earlier deposits of rift-valley redbeds, proto-oceanic evaporites, rift lavas, etc. are not differentiated; *b*, carbonate-shale phase of miogeocline deposited during shelf-slope-rise configuration of rifted continental margin; *c*, progradational continental embankment. See text for discussion.

ceivably occur, although the tendency of the spreading center to form an elevated midoceanic rise would tend to divide the oceanic basin into two halves with different sediment sources for beds deposited from bottom-hugging turbidity currents. If turbidity currents could reach or cross the actual spreading center, the net effect would be to continue forming transitional crust of the quasioceanic type.

More important in a narrow ocean is the fact that the transitional crust along and adjacent to the attenuated continental margins would continue to subside thermally as a closing stage of the plate interaction that formed the rifted continental margins. This period of subsidence probably persists along a rifted continental margin for perhaps 50 to 75 my, during which it proceeds independently of sedimentary loading. Its influence would tend to eliminate the uplifted belts that previously acted to bar sediment delivery to the rifted continental margin. This early thermal subsidence of the substratum beneath the belt of quasicontinental transitional crust is probably the factor that induces rather rapid accumulation of thick clastics as a basal phase of typical miogeoclines (fig. 7a). Such strata form a basal wedge, thickening seaward, as the oldest areally continuous deposits in the outer or oceanward parts of both the Appalachian and Cordilleran miogeoclines (King, 1969, p. 11-12; Stewart, 1972), in both of which the basal clastic phase is latest Precambrian and earlier Cambrian in age.

Open ocean.—When the strictly thermal subsidence of a rifted continental margin is complete, the margin is left in an intraplate setting and facing an open ocean. At this point, the drowned belt of transitional crust along the margin is already complex geologically. The attenuated continental basement rocks are faulted and covered locally by continental clastics, volcanics, and evaporites concentrated in varying degree within downfaulted blocks. Across this compound substratum, the basal clastic phase deposited during and closely following the main thermal subsidence is draped as a wedge of marine and paralic strata built upward to form an isostatically balanced continental terrace in the initial configuration of that feature. From this point onward, any further subsidence apparently is the result of sedimentary loading of crust offshore from the shelf break at the edge of the continental terrace.

The continued evolution of the rifted margin prism can be described using the terminology of Dietz (1963) for continental terrace, slope, rise, and embankment. The continental terrace, upon which shelf and paralic sedimentation dominates, extends to the slope break at the shelf edge, from which the continental slope leads down to deep water where the continental rise of turbidites accumulates along the edge of oceanic crust (fig. 7b). Bending of the lithosphere caused by the loading of the continental rise (Walcott, 1972) causes the continental terrace to tilt progressively seaward. This process enables the continental terrace to receive successive wedges of strata that thicken rather uniformly from a nearly common landward hinge zone toward the shelf edge (*e.g.*, Rona, 1973). The flexure at the hinge zone lies perhaps 100 to 250 km from the shelf edge. As subsidence of this kind is not linked directly to sedimentary loading of the continental terrace itself, erosional episodes on the shelf may produce disconformities within the shallow marine and shoreline deposits of the accumulating terrace wedge. These miogeoclinal strata, deposited more slowly than the underlying basal clastic phase, are probably represented by the succeeding carbonate-shale phase (*e.g.*, King, 1969) of the lower Paleozoic section in the Appalachian and Cordilleran miogeoclines. For the Appalachians, a carbonate platform marginal to the continent is recognized clearly for this interval (Rodgers, 1968), and similar strata appear in the Cordilleran case (Armstrong, 1968b).

The continental slope beyond the shelf break is largely a region of sediment bypass that serves for the transit of turbidity currents headed from shallower water toward the continental rise. Sediment thicknesses beneath the terrace-slope-rise association thus give an hourglass effect in section, with the pinched region of thin strata lying along the continental slope. Available data suggest that sediment thicknesses beneath the shelf break at the outer edge of the continental terrace, and also those beneath the continental rise, can reach at least 5 km.

If clastic sedimentation along a rifted continental margin is voluminous enough, upward construction of the continental rise and outward construction of the continental terrace lead to the development of a progradational continental embankment (fig. 7c). This type of feature is discussed here as a sequel to the stage of development represented by the terrace-slope-rise association, but appropriate relations among the details of structural development of transitional crust, the rate of thermal subsidence, and the timing and rate of sediment delivery could blur the distinction between the two stages of development in some instances. The continental slope on the front of a continental embankment becomes a constructional feature owing to wholesale progradation of the continental edge. Shelf break and slope thus advance seaward from the region of transitional crust until both reach a position above fully oceanic crust, as Dietz (1963) suggests for the Texas coast. The top of the embankment receives mainly fluvial and paralic sediments while the frontal slope and toe receive mainly turbidites. Immense thicknesses of sediment are possible for continental embankments; at least 12.5 km of sediment are present beneath the Texas coast and Walcott (1972) suggests that thicknesses of 17.5 km could be attained by simple isostatic subsidence of oceanic crust and lithosphere. By analogy with the deep-water Niger delta, the structure of the continental embankment in the region beyond the edge of the pre-existing continental terrace can be inferred to include three main depositional phases (Burke, 1972): a basal phase of sandy turbidites deposited near the toe of the embankment, a middle phase of mainly shaly rocks deposited on the advancing frontal slope of the embankment, and an upper phase of largely sandy paralic strata deposited along the prograding outer edge of the top of the embankment.

Marginal offsets.—On the floor of the modern Atlantic Ocean, the major transforms that offset the crest of the midoceanic rise are extensions of fracture zones whose extremities at the flanks of the ocean appear to coincide with abrupt offsets of the adjacent continental margins (Le Pichon and Hayes, 1971; Le Pichon and Fox, 1971). To some extent, therefore, the gross shape of the rectilinear trellis of rise segments and connecting transforms in the ocean is inherited from the shape of the initial rupture formed by continental separation. The marginal offsets were transform fault zones, rather than extensional rift zones, during continental separation. The edges of the continental blocks along the trends of the marginal offsets thus underwent a different early evolution than the edges that face toward the center of the ocean and are masked now by rifted-margin prisms of the type just discussed (fig. 8). Strike-slip along the marginal offsets during continental separation would disrupt and displace segments of the earlier phases of any sedimentary accumulations that might form along those parts of the continental edges. More important, however, is the fact that the structural character of the transitional crust along the marginal offsets is likely to be different in kind (Francheteau and Le Pichon, 1972).

During continental separation, continental margins along the marginal-offset transforms are swept by the butt ends of incipient midoceanic rise segments (see fig. 8). Although the full thermal and petrologic effects of this process are unclear, the lateral transit of the end of a rise segment along a marginal offset apparently leads to the formation of a distinctive feature termed a marginal fracture ridge (Le Pichon and Hayes, 1971; Le Pichon and Fox, 1971). Where fully developed, marginal fracture ridges extend along the marginal offset itself and at least that far again out to sea along the same trend. They are formed probably in part by crumpling and shearing along the slip zone, and in part by overthickened quasi-oceanic crust formed by exceptional leakage of igneous materials from the regions near the butt

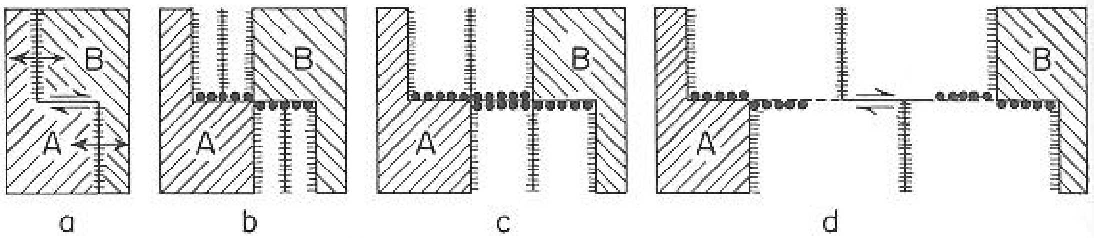

FIG. 8.—Diagram after LePichon and Hayes (1971) showing in plan view the development of marginal fracture ridges during continental separation of continental fragments A and B shown joined prior to separation in *a* and progressively farther apart in *b*, *c*, *d*. Barred lines are spreading centers linked by a transform and hachured lines show positions of normal rifted-margin prisms. Heavy dots are marginal fracture ridges along marginal offsets and offshore. Dashed lines in *d* are fracture zones along same trend.

ends of the migrating rise segments. Thick sedimentation along marginal offsets is probably delayed by prolonged thermal uplift, and clastic sedimentation may be inhibited locally by high-standing marginal fracture ridges. However, the marginal offsets are subject to prolonged thermal uplift and marginal fracture ridges may later actually promote abnormally thick sedimentation of biogenic sediments in the oceanic realm along offshore trends in line with marginal offsets. The Bahama platform, an elongate accumulation of 5 km of mainly shallow-water carbonates above perhaps 15 km of quasioceanic crust (Dietz and others, 1970) may reflect such a phenomenon, although other means of generating the quasioceanic crust beneath the carbonate platform have also been suggested.

Aulacogens.—The term aulacogen is applied here in the usage of Hoffman (and others, 1974) as adapted from the Russian literature, in which the term was devised for a class of features initiated mainly in the later Precambrian (Salop and Scheinmann, 1969). Aulacogens are elongate sedimentary basins that extend, as gradually narrowing wedges or pie slices in plan view, from the margins of cratons toward the interiors of cratons. The sedimentary sequences of aulacogens are mainly similar in general nature to facies equivalents in platformal sequences of the cratons adjacent on both sides, but are much thicker. Aulacogens are thought to evolve from semioceanic gashes formed at re-entrants in rifted continental margins during continental separations (fig. 9).

The overall geometry of RRR and RRF triple junctions (McKenzie and Morgan, 1969) is attractive as an explanation for the tectonic setting of aulacogens. Such a triple junction has two or three spreading centers as arms. The Afar region linking the Red Sea, Gulf of Aden, and East African rift systems is a modern example. Burke and others (1971) and Grant (1971) argue that the Benue Trough, which extends into Africa from the head of the Gulf of Guinea, was temporarily one spreading arm of a triple junction in the Cretaceous. When continuation of motion along the other two arms opened the South Atlantic, the Benue arm was aborted in an incipient stage of development. Cretaceous and younger sediments beneath the trough are more than 5 km thick for at least 500 km along its axis. Their accumulation was probably accommodated by the subsidence of transitional crust beneath the trough. Thermal subsidence following the failure of the Benue spreading arm to continue into a fully oceanic configuration probably served as a trigger to initiate sedimentation, which then forced further isostatic subsidence under sedimentary loading. The location of a long-lived Benue depression also apparently controlled the course of the Niger River and the position of its delta. The delta itself is evidently a local continental embankment containing sediment some 10 km thick built into deep water beyond the initial continental margin off the seaward end of the aulacogen.

In North America, the best example of an aulacogen containing Phanerozoic strata is the Anadarko-Ardmore basin, which extends inland nearly 500 km from the southern margin of the Paleozoic continent. Ham and Wilson (1967) describe the section in the elongate basin as 10 to 12 km of Paleozoic strata overlying at least 2 km of Cambrian volcanics. Late Paleozoic deformation and coarse clastic sedimentation within the basin was greatest toward its open end, and was probably related to the Ouachita orogeny along the nearby continental margin.

Arc-Trench Systems

Arc-trench systems are the characteristic geologic expression of convergent plate junctures (Dickinson, 1970). As recognized plainly by Kay (1951), the volcaniclastic rocks of volcanic island chains built along magmatic arcs are prominent within many eugeosynclinal terranes. Eugeosynclinal terranes also include the subduction complexes associated with trenches, where oceanic strata are mingled tectonically as they are detached from the tops of slabs of lithosphere descending beneath the flanks of arc-trench systems. Sedimentary sequences that accumulate on the flanks of magmatic arcs, which stand as positive topographic features during arc activity, receive a variety of geosynclinal designations locally, depending on details of their relationships to various types of substratum and also upon the nature of the strata themselves. A full discussion of the evolution of arc-trench systems requires an essential focus on magmatism and metamorphism, but the emphasis here is solely upon the facets of behavior that affect the associated sedimentary basins (Dickinson, 1974).

Arc-trench systems include the following five major morphotectonic elements (*e.g.*, Dickinson, 1973): (1) the *trench*, a bathymetric deep floored by oceanic crust; (2) the *subduction zone* beneath the inner wall of the trench and the trench slope break marking the top of the inner wall; (3) the arc-trench gap, a belt within which a *forearc basin* may occur between the trench slope break and the magmatic arc; (4) the magmatic arc, within which *intra-arc basins* may occur; and (5) the backarc area, within which may lie either an *interarc basin* floored by oceanic crust and separated from the rear of the arc by a normal fault system, or a *retroarc basin* floored by continental basement and separated from the rear of the arc by a thrust fault system.

Sedimentation in the various types of basins noted for the different elements of arc-trench systems is contemporaneous with both volcanism and plutonism along the magmatic arc, and with metamorphism both in the cool subduction zone and in the hot roots of the magmatic arc. Faulting and other deformation in the subduction zone, within the magmatic arc, and in the backarc area is also contemporaneous with sedimentation. Although sequential phases of sedimentation can doubtless be recognized for each of the kinds of sedimentary basins noted in the various morphotectonic settings, the areal contrast in facies among the various kinds of sedimentary sequences forms the most important and regular genetic pattern. This pattern of distinctive and parallel sedimentary terranes, coupled with their igneous and metamorphic associates, can be used as a means to identify the petrotectonic assemblages that form the geologic record of past arc-trench systems.

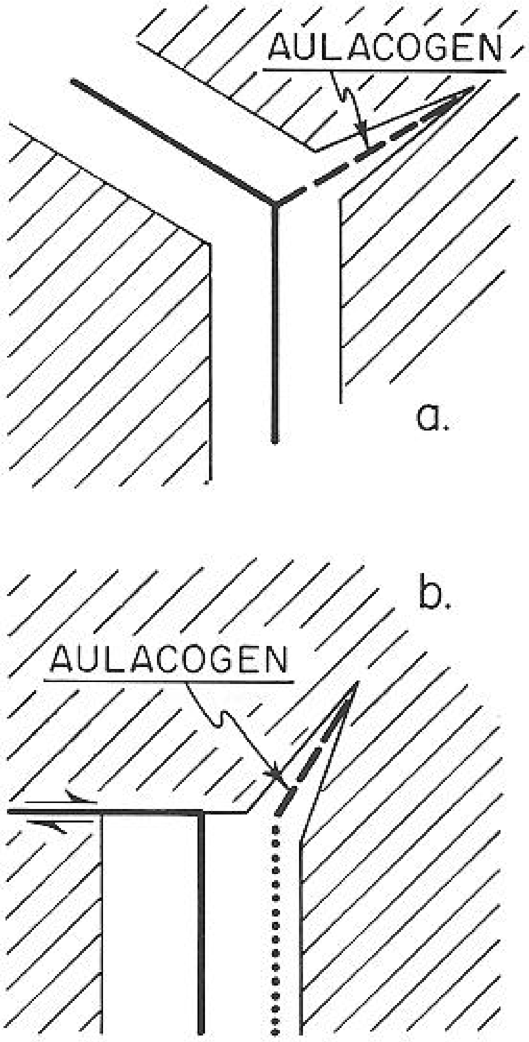

FIG. 9.—Diagrams showing in plan view alternate mechanisms for the development of aulacogens at re-entrants in rifted continental margins (continental blocks shaded, plate junctures that continue active shown as solid heavy lines, failed spreading centers along axis of aulacogen shown as dashed heavy line): *a*, aulacogen as failed arm of formerly stable RRR triple junction (spreading directions changed along the two arms that continued spreading when motion stopped along the failed arm); *b*, aulacogen as failed arm of inherently unstable RRF triple junction (after Grant, 1971).

Thick sedimentation associated with arc-trench systems is best discussed, therefore, in relation to subduction zones, forearc basins, intra-arc basins, interarc basins, and retroarc basins (fig. 10). The progressive development of these features presumably will continue until the plate consumption that fosters an arc-trench

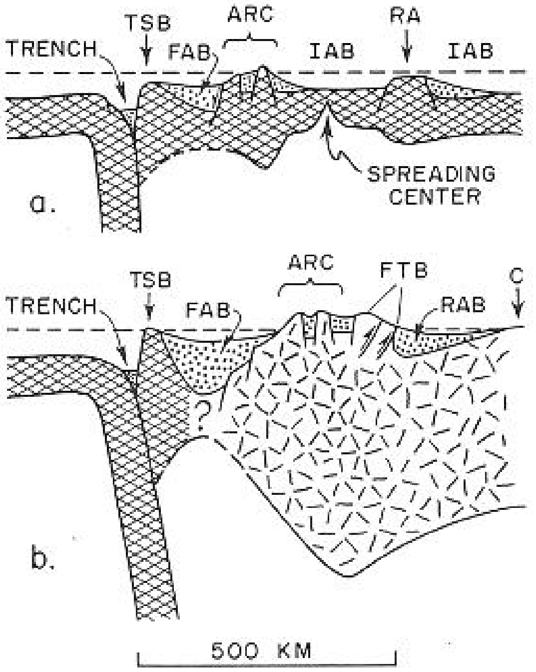

FIG. 10.—Idealized diagrams to illustrate tectonic settings of sedimentary basins (stippled) associated with arc-trench systems. Dashed line is sea level. Vertical exaggeration is 10×; note apparent steep angles of descent of plates beneath trench although true angles depicted are 60 degrees (*a*) and 30 degrees (*b*). Oceanic and paraoceanic crust is crosshatched; continental and paracontinental crust is jackstrawed. Trench slope breaks (TSB) lie above subduction zone complexes at thresholds of arc-trench gaps within which forearc basins (FAB) are shown. For intraoceanic arc (*a*), active or frontal island arc (ARC) is shown with a marine intra-arc basin, and remnant arc (RA) stands between two interarc basins (IAB), one active (left) and one inactive (right) but both with volcaniclastic wedges along one flank. For continental margin arc (*b*), volcanic highlands (ARC) are shown with a terrestrial intra-arc basin, a flanking intermontane lowland, and a foreland fold-thrust belt (FTB) above zone of partial crustal subduction lies between magmatic arc and retroarc basin (RAB) on depressed crust of pericratonic foreland adjacent to margin of interior craton (C).

system is terminated, commonly by crustal collision to form a suture belt.

Subduction zones.—Seaward from the trench in typical arc-trench systems is a broad upwarp of the ocean floor marking the flexure of the lithosphere as it bends to descend beneath the arc-trench system. The inner slope of this outer arch is the gentle outer slope of the trench, and is scarred in places by normal faults reflecting local extensional deformation of the ophiolite sequence represented by the igneous oceanic crust and its sediment cover. The trench is a bathymetric deep immediately adjacent to the tectonic front of the subduction zone, which begins at the base of the steep inner wall of the trench. On the trench floor, variable thicknesses of turbidites are ponded above the sediment layers rafted tectonically into the trench from the open ocean floor. Transport by turbidity currents within a trench is mainly longitudinal along the trench axis (von Huene, 1972; Marlow and others, 1973), although the initial entry of sediment into the trench may occur at intervals along the inner wall as well as from the ends of the trench (Ross, 1971; J. C. Moore, 1973). Where the rate of sediment delivery to a trench is high enough in relation to the rate of plate consumption, the trench may be filled with sediment and the trench site covered by subsea fans that mask the position of the tectonic front of the subduction zone (Silver, 1969). It is fair to infer that the volume of locally deposited turbidite sediment incorporated within the nearby subduction complex is thus in some measure inversely proportional to the bathymetric depth of the associated trench. An empty trench leaves little evidence in the geologic record.

Recent data leave little doubt that the steep inner wall of the trench is underlain almost directly by deformed and uplifted oceanic strata with only a local cover of undeformed sediment (Karig, in press). This material is interpreted here as a subduction complex of mélanges and crumpled beds sliced by thrusts and including ophiolitic scraps. The tectonic top of the subduction complex is assumed to lie at or just beneath the sea bottom at the trench slope break, a bathymetric transition point located at the top of the inner wall of the trench. The mass of the subduction complex is inferred to grow by the accretion of successive increments of oceanic crustal materials that either are jammed against its seaward flank at the trench axis or scraped into its basal levels from the top of the slab of lithosphere that descends beneath the subduction zone. It is important to note that these materials thus added to the subduction complex include not only indigenous trench turbidites deposited nearby, but potentially also include samplings of all the turbidites deposited over extensive areas of the ocean floor from sources wholly exotic to the arc-trench system into whose flank they are incorporated.

As oceanic materials are stacked tectonically within subduction zones, net uplift of the subduction complex must occur even while subduction continues, and should be dramatic when subduction ceases for any reason. The condition of a subduction complex where exposed to view on land is thus never the initial condition. Always there is the overprint of deformation during uplift, which must amount to a minimum of 5 to 10 km if the depths of modern trenches are representative. Mass movement of material off the steep inner wall of the trench may in time recycle some materials back through the process of subduction. We are still at a loss to understand fully the complex structures of mélanges and thrust-bounded slabs in subduction complexes (Suppe, 1972). It seems clear, however, that their great apparent thicknesses are tectonic, rather than stratigraphic.

Forearc basins.—The topographic and bathymetric configuration within the arc-trench gap between the trench slope break and the volcanic front is highly varied. The elevation of the threshold at the trench slope break is evidently controlled by the elevation of the top of the subduction complex, which may be emergent as islands or may lie at depths as great as 2 to 3 km. Different arc-trench gaps contain, singly or in combination, such diverse geographic elements as mountainous uplifts, longitudinal troughs, transverse submarine slopes, shallow shelves, deep-marine terraces or plains, and terrestrial plains or valleys. In a number of modern arc-trench gaps, thick sequences of largely undeformed sediments attest to progressive subsidence to develop forearc basins as the term is used here, and sequences interpreted as forearc basins in the geologic record attain thicknesses of 5 to 12 km. Subsidence may be related to the descent of a dense slab of lithosphere beneath the arc-trench gap.

The sedimentary sequences of forearc basins rest on a substratum of variable and partly uncertain character. On the arc flank of the arc-trench gap the substratum may include eroded igneous rocks, both plutonic and volcanic, of the magmatic arc. On the trench flank of the arc-trench gap the substratum may include parts of the subduction complex. Beneath the center of a forearc basin the substratum may be paraoceanic crust made of previously accreted elements of a subduction complex that broadens by growing seaward with time, or may be oceanic or transitional crust that existed before the arc-trench system was activated (*e.g.*, Grow, 1973). Thus, in general, forearc basins are commonly successor basins (King, 1969) in the sense that they overlie older, deformed elements or orogenic belts.

Forearc basins receive sediment mainly from the extensive nearby arc structures, where not only volcanic rocks but also plutonic and metamorphic rocks exposed by uplift and erosion may serve as sources. Sources may also include local uplands along the trench slope break or within the arc-trench gap itself. Facies gradations may presumably occur between strata of forearc basins and volcaniclastic beds of the volcanic arcs, but prominent normal fault zones commonly bound the basins on the arc side (Karig, in press). On the trench side, tectonic gradation into the disrupted strata of the subduction zone presumably occurs locally. In several instances, however, nearly intact ophiolite sequences that underlie continuous sedimentary sequences of inferred forearc basins are in sharp fault contact with the adjacent subduction complexes. This circumstance implies little or no transfer of material into the subduction zone from the part of the forearc basin now preserved. Instead, the forearc basins appear to have wholly overridden the subduction zones. This relation holds for the late Mesozoic Great Valley sequence of California where faulted against the coeval Franciscan complex (Bailey and others, 1971), for the late Paleozoic and early Mesozoic western marginal facies of New Zealand where faulted against the coeval eastern axial facies or Torlesse Group (Landis and Bishop, 1972; Blake and Landis, 1973), and for the early Tertiary succession of the central Burmese lowland where faulted against the Indoburman flysch terrane (Brunnschweiler, 1966).

By inference from the bathymetry of modern forearc basins, and from the sedimentology of older sequences inferred to have been deposited in similar settings, forearc basins may contain a variety of facies. Shelf and deltaic or terrestrial sediments, as well as turbidites with either transverse or longitudinal paleocurrents, may occur in different examples. The local bathymetry is presumably controlled by the elevation of the trench slope break, the rate of sediment delivery to the forearc basin, and the rate of basin subsidence acting in combination. Various facies patterns and successive depositional phases probably can occur in different cases.

Intra-arc basins.—Magmatic arcs include both intraoceanic and continental margin types (Dickinson, 1974). Intraoceanic arcs include those with only paraoceanic crust built by magmatic additions to oceanic crust and those underlain at depth by a sliver of continental crust detached as part of a migratory arc structure. Continental margin arcs include island arcs backed by shallow epicontinental seas as well as those standing along the edges of landmasses. Fault-bounded extensional basins that occur within many magmatic arcs may be related to local volcano-tectonic subsidence, or to arching that accompanies uplift of paracontinental crust, or to the development of an incipient interarc basin. Volcaniclastic strata are characteristic, but may include a range of types from terrestrial redbeds to turbidites, as well as various intermediary facies. Local sources of sediment within the arc structure are typical, but low-standing island arcs located near continental blocks may accumulate clastic strata from external sources as well; these have been termed basinal arcs (Berg and others, 1972).

Backarc areas.—The distinction made here between interarc and retroarc basins in the backarc area reflects the existence of two distinct variants of arc-trench systems (see fig. 10). Both types of basins are related indirectly to the convergent plate junctures to which the trenches and their associated subduction zones are related directly. The contrast between the tectonic settings of interarc and retroarc basins seemingly stems, therefore, from influences other than simple plate interactions at convergent plate junctures. The key control is apparently the relative motion of the plate of lithosphere in the backarc area with respect to the underlying asthenosphere (Coney, 1971; Dickinson, 1972; Hyndman, 1972; Wilson and Burke, 1972; Wilson, 1973).

The lithosphere is apparently not wholly intact across the region beneath the magmatic arc owing to thermal softening from the high heat flux. The narrow belt of lithosphere beneath the arc-trench gap can thus be viewed as a separate narrow plate. Where the lithosphere behind the arc has a component of motion, relative to underlying asthenosphere, away from the magmatic arc, then the arc structure may split. An interarc basin underlain by newly formed oceanic crust built by a backarc spreading center then opens between the active or frontal arc and a remnant arc (Karig, 1972), which may be of either intraoceanic or continental margin type. This mode of behavior is characteristic of eastward-facing island arcs in the western Pacific region (Karig, 1970, 1971a). Where the lithosphere behind the arc has a component of motion, relative to underlying asthenosphere, toward the magmatic arc, then partial subduction of continental lithosphere beneath the rear of the arc structure is assumed here to occur (e.g., Coney, 1972). A fold-thrust belt thus develops in the backarc area as cover rocks are stripped off descending basement. The resulting highlands shed debris into a downbowed retroarc basin along a belt that can be termed pericratonic between the continental margin arc and the craton. This mode of behavior is characteristic of the westward-facing Andean arc, which is flanked on the east by the Subandean fold-thrust belt, beyond which are the Subandean sedimentary basins that lie between the Andes and the craton (Ham and Herrerra, 1963; Sonnenberg, 1963).

Conceivably, the contrasting behavior of eastward-facing and westward-facing arc-trench systems may reflect the different tectonic regimes, respectively extensional and contractional, induced in barkarc areas by the postulated net westward drift of lithosphere with respect to asthenosphere as a result of tidal influences (G. W. Moore, 1973). By implication, arc-trench systems with a roughly east-west orientation might experience no marked deformation in backarc areas, and hence might display neither interarc nor retroarc basins.

Interarc basins.—The sedimentary record of interarc basins is not well documented, but their global abundance at present suggests that eugeosynclinal terranes of the past probably contain numerous examples. It must be inferred that some ophiolitic sequences of orogenic belts represent oceanic crust formed as the floors of interarc basins, rather than in open oceans. If there are significant differences between the igneous rocks of the two kinds of ophiolitic sequences, the distinction is not yet established.

The sedimentary strata in modern interarc basins include distinctive turbidite aprons of volcaniclastic beds shed backward from the rifted rear sides of migratory frontal arcs (Karig, 1970, 1971a, 1972). These turbidite wedges appear to rest almost directly on the igneous oceanic crust with little or no intervening pelagites present. Beyond the interarc spreading centers sedimentation varies markedly. Where a given interarc basin is bounded on the side away from the arc-trench system by a submerged remnant arc, no effective source of clastic sediment is present and oceanic pelagites accumulate. Where successive remnant arcs with paraoceanic crust are calved in succession from migratory frontal arcs, a broad oceanic region is formed in which the only thick sedimentary accumulations are turbidite wedges stranded behind each submerged remnant arc.

On the other hand, where an interarc basin forms by disruption of a continental margin arc, one side of the interarc basin is a form of rifted continental margin along which some variant of a rifted-margin prism can be formed (Mitchell and Reading, 1969) beside a marginal sea (Karig, 1971b; Packham and Falvey, 1971; Moberly, 1972). It may be argued that the pattern of parallel facies belts associated with such a continental margin fringed by migratory intraoceanic arcs lying offshore faithfully reproduces the classic miogeosyncline-eugeosyncline couple. If so, extreme horizontal motions of lithosphere may be unnecessary assumptions to explain the juxtaposition of diverse terranes within orogenic belts. The rifted continental margin on the inner side of the interarc basin is interpreted then as the miogeosynclinal belt, whereas the adjacent interarc basin, the offshore island arc, and the open ocean beyond together represent the complex tectonic elements of the eugeosynclinal belt. Although attractive, the analogy harbors a potential fallacy. Only if the substratum beneath the supposed miogeoclinal wedge includes igneous rocks representing part of the geologic record of the earlier stages of arc evolution prior to arc migration can the analogy be defended in detail. As most miogeoclinal wedges appear to rest on truncated continental basement considerably older than the base of the miogeoclinal wedge, this logical requirement of the analogy does not appear to be met in typical orogenic belts.

Retroarc basins.—The sedimentary record of retroarc basins includes fluvial, deltaic, and marine strata as much as 5 km thick deposited in terrestrial lowlands and epicontinental seas along elongate pericratonic belts between continental margin arcs and cratons. Where the magmatic arcs stand along continental margins that have grown seaward by tectonic accretion, some retroarc basins may be successor basins in the sense of resting upon previously deformed terranes. Sediment dispersal into and across retroarc basins is mainly tranverse, in a gross sense, from highlands on the side toward the magmatic arc, although contributions from the craton are also present. The deposits of retroarc basins are thus exogeosynclinal in the sense that debris is shed toward the craton from sources within marginal orogenic belts.

The sources of sediment may include the magmatic arc itself, but commonly the principal sources are uplifted strata in the fold-thrust belt formed by partial subduction behind the arc. Such was the case for the Cretaceous retroarc basin of the interior and Rocky Mountain region of North America (Weimer, 1970). The main highland sources were uplands of folded and faulted pre-Mesozoic strata lying just west of the retroarc basin, but still east of the batholith belt that marks the position of the associated magmatic arc (Hamilton, 1969). Part of the subsidence in retroarc basins is probably in response to flexure of the lithosphere or other isostatic adjustments induced by the tectonic load of thrust sheets in the adjacent foreland fold-thrust belt (Price and Mountjoy, 1971). As the retroarc basin evolved, contractional deformation disrupted piedmont facies along the highland flank of the basin, and ultimately crumpled the flank of the basin fill within the fold-thrust belt (Armstrong, 1968a). Where the main sources of sediment are thus in the fold-thrust belt behind the arc, rather than within the magmatic arc, the nature of the sources depends upon the previous history of the continental margin. Where the magmatic arc arises following the initiation of plate consumption along a previously inactive continental margin draped with a rifted-margin prism, the sources are apt to be uplifted miogeoclinal strata.

The fold-thrust belts that parallel the orogenic margin of retroarc basins thus may be described commonly as foreland thrust belts (Coney, 1973). In this sense, the foreland is simply the cratonal or platformal interior of the continent, and the foreland basin is a retroarc basin. However, foreland basins in this same setting with respect to the continental interior may form as a result of crustal collisions in which a rifted continental margin with its rifted-margin prism encounters the main subduction zone associated with the trench of an arc-trench system. The designation of the foreland can thus be ambiguous with respect to the polarity of the arc-trench system responsible for the orogenic belt. So long as parts of a rifted-margin prism are thrust back toward the continental interior, and a pericratonic fringe of continental basement is drawn down by partial subduction to form an elongate basin parallel to the fold-thrust belt, the concept of a foreland to the orogenic belt is appropriate. Foreland basins formed by partial subduction of continental margins during crustal collisions are here termed peripheral basins as discussed in the next section. Designation of a given foreland basin as either a retroarc basin or a peripheral basin thus depends upon a knowledge of the sequence and timing of tectonic events in the adjacent orogen.

Suture Belts

The term suture belt is used here for the complexly deformed joins along which crustal blocks are welded together by the crustal collisions that occur when lithosphere bearing thick crustal blocks reaches a subduction zone along a convergent plate juncture where oceanic lithosphere was previously being consumed (fig. 11). Crustal collisions include a variety of types involving both intraoceanic and continental margin arc-trench systems (Dickinson, 1971c). In all cases, crustal collision involves juxtaposition of the tectonic elements of an arc-trench system, together with its variety of sedimentary basins, against other crustal blocks across the

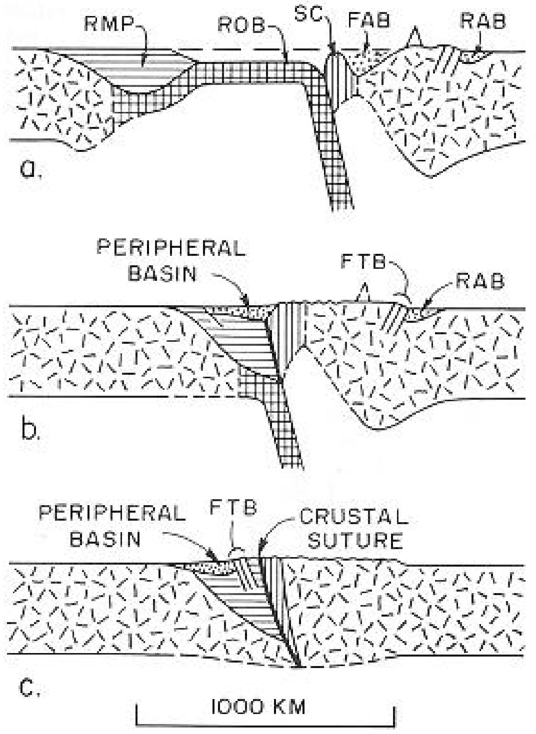

FIG. 11.—Idealized diagrams to illustrate hypothetical sequence of events, and associated sedimentary basins (stippled), during crustal collision between rifted continental margin (see fig. 7) on left and continental margin arc (see fig. 10b) on right: *a*, prior to collision; *b*, initial collision; *c*, final suturing (vertical exaggeration is 10×). Oceanic and quasioceanic crust is crosshatched; continental, quasicontinental, and paracontinental substratum is jackstrawed. Symbols: RMP, rifted-margin prism (horizontal rules); ROB, remnant ocean basin (sea level as dashed line); SC, subduction complex (vertical rules); FAB, forearc basin; RAB, retroarc basin; FTB, foreland fold-thrust belts.

suture belt. Along a sutured join, deformed sedimentary sequences that were deposited on the ophiolitic basement of an open oceanic basin or an interarc basin can be caught between the sutured crustal blocks. Such sequences commonly appear to view after suturing as tectonically scrambled mélanges of ophiolitic scraps and oceanic facies. Such suture-belt mélanges may not be visible within a suture belt if the extent of subduction during crustal collision was sufficient to hide them beneath rocks of the overriding plate. Clear evidence of the sutured join may also be absent if contractional deformation during collision is sufficient to squeeze the suture-belt mélanges upward to tectonic levels that are later removed by erosion. These two cases of obscured suture belts can be described as hidden sutures where telescoping thrust sheets cover the suture and as cryptic sutures where the materials caught in the suture are pressed out tectonically and lost by later erosion (e.g., Dewey and Burke, 1973).

Suture belts contain deformed examples of all the various types of sedimentary sequences discussed in connection with oceanic basins, rifted-margin prisms, and arc-trench systems. In addition, sedimentary basins of a unique type here termed *peripheral basins* with strata as thick as 5 km are also formed by processes related to collision (see fig. 11). As a continental crustal block is drawn toward a subduction zone just prior to crustal collision, bending of the lithosphere probably first causes extensional faulting analogous to that seen on the oceanic outer arch seaward of arc-trench systems. These faults might offset the strata of a rifted-margin prism in a sense similar to that of the earlier faults that were associated with continental rifting or with growth faulting during deposition of the prism. Later in the progress of crustal collision, the edge of the continental block is depressed by partial subduction to form a pericratonic or foreland basin peripheral to the suture belt on the plate being partly consumed. As the process of subduction is braked by crustal collision, the subsidence in this peripheral basin may be succeeded by marked uplift. As the peripheral basin is drained, a phase of evaporite deposition could conceivably ensue. The well-known sabhka deposition in the Persian Gulf may be an example of such deposition in a restricted seaway remaining along a belt parallel to the suture belt of the Zagros Crush Zone in Iran (Wells, 1969).

Perhaps the most characteristic deposits of peripheral basins are exogeosynclinal clastic wedges spread toward the craton as fluvial and deltaic strata shed from a suture belt involving the continental margin (Graham and others, in press). If the peripheral basin is deep enough, however, these deposits may be preceded by turbidites deposited on depressed continental or transitional crust, rather than oceanic crust (W. M. Neill, personal commun., 1973). Paleocurrent trends in the clastic wedges may be dominantly transverse to the orogenic trend, whereas those in the turbidites may be dominantly longitudinal to the orogenic trend. Clastic wedges of peripheral basins, as well as any clastic wedges shed toward the other side of highlands along the suture belt, may thus be termed molasse in many cases. The turbidites of peripheral basins, as well as the turbidites of oceanic basins or forearc basins caught within the suture belt, may thus be termed flysch in many cases.

The evolution of suture belts forms an attractive explanation, though not the only one, for the tectonic relations of flysch and molasse (Graham and others, in press). In general, any completed suture belt will represent the end result of a sequential closure of a remnant ocean basin (Dickinson, 1972). Only if the shapes of colliding continental margins are mirror images of one another, and the vector of the relative plate motion causing crustal collision is exactly as required, can crustal collisions be synchronous along their whole length. In the general case, extensive suture belts must be diachronous in development as successive adjustments in plate motions and boundaries allow progressive welding of crustal blocks to proceed. A tectonic transition point between the segment already sutured (fig. 11c) and the segment yet to be sutured (fig. 11a) will migrate along the developing suture belt with time. Behind the transition point, orogenic highlands, clastic wedges and filled peripheral basins are characteristic. Ahead of the transition point, remnant ocean floor and incipient peripheral basins are present. As the drainage of orogenic highlands is commonly longitudinal, much of the sediment derived from the collision orogen will not be shed transversely as clastic wedges, but will be shed longitudinally into the remnant oceanic basin and deepening peripheral basins along tectonic strike. Many of the deposits that reflect erosion of the collision orogen thus will be incorporated later into the same orogenic belt as the tectonic transition point migrates along the growing suture belt. In this fashion, synorogenic flysch of turbidites with mainly longitudinal paleocurrents and postorogenic molasse of clastic wedges with largely transverse paleocurrents may be seen as the natural result of crustal collision to form suture belts.

Intracontinental Basins

Intracontinental basins are the most difficult type to treat constructively in terms of plate tectonics if basins related to apparently intracontinental orogenic belts like the Urals are excluded. Provided such orogenic terranes are interpreted as suture belts (Hamilton, 1970), the associated basins can be interpreted variously in terms of former oceanic basins, rifted continental margins, arc-trench systems, and collision orogens. Basins related to these kinds of features include foreland basins of both retroarc and peripheral types where the basin fill is supracontinental in the sense of resting on continental crust or on older rifted-margin prisms.

Basins bounded on all sides by anorogenic terranes forming a basement that is uniformly older than the basin fill are the intracontinental ones difficult to explain using principles of plate tectonics. For infracontinental basins (see above) the basement does not extend unmodified beneath the floor. Partial attenuation of continental basement along an aborted rift that never advanced beyond an incipient stage could lead to conditions permitting marked crustal subsidence locally, especially under sedimentary loading. Unfortunately, detection of transitional crust hidden beneath an infracontinental basin depends upon geophysical observations, for the basin fill permanently masks the substratum.

Presumably, infracontinental basins would tend to be elongate in many cases, but not necessarily in all. If antecedent or contemporaneous domal uplifts were distributed at intervals along the belt of partial crustal attenuation, as appears to be the pattern for early stages of continental rifting, then crustal thinning by stretching and erosion might be concentrated within relatively equant areas. As a result, the infracontinental basins that developed after thermal tumescence gave way to thermal decay might appear as apparently isolated and more or less round features distributed apparently at random across a continental block. The only clue to their essentially common origin might be a rough contemporaneity of development. The initial stages of a major continental separation may well involve extensive gashing of the continental block, while still joined, over a broad region that would lapse eventually into quiescence except where the rifting was fully established along a single trend. Dispersed infracontinental basins might then remain as a record of the widespread extent of incipient rifting.

Alternatively, a fundamentally elongate belt of incipient continental separation might be marked by a chain of isolated infracontinental basins linked only by intracontinental transforms. If the transform segments of the integrated tectonic system were masked by cover rocks or later deformation, the fundamental pattern might be difficult to detect by any means. There seems an especially strong possibility that successor basins might form along recently completed suture belts in this fashion as residual plate motions were resolved into translation along transforms roughly parallel to the suture belt.

None of these speculations touch upon the possibility of long-lived supracontinental basins with underpinnings of normal continental crust. The motion of a plate of lithosphere over a bumpy asthenosphere accounts well only for reversible epeirogenic warping and temporary subsidence. Note that this effect might affect local areas distributed in unpredictable fashion over a continental block, but that any local and temporary subsidence would occur as part of a wave of shbsidence. The passage of the lithosphere over a bump or depression on top of the asthenosphere thus might leave a sort of subtle track in the stratigraphic record of any epicontinental seas covering a continental block. Mechanisms for permanent subsidence of supracontinental basins on a large scale in truly intraplate settings are not apparent from plate-tectonic theory.

Summary

The preceding tentative classification of sedimentary basins in a plate-tectonic framework indicates that satisfactory alternatives to the geosynclinal terminology can be devised, and that points of correspondence between the two schemes of nomenclature can be appreciated. The discussion also indicates that direct equivalency between individual terms in the two sets cannot be expected. For example, eugeosynclines apparently contain the strata of oceanic basins, intra-arc basins, and interarc basins as modified by deformation in subduction complexes, magmatic arcs, and suture belts. On the other hand, rifted-margin prisms may include the superimposed strata of taphrogeosynclines, miogeosynclines, and paraliageosynclines. Exogeosynclinal foreland basins may be either retroarc basins or peripheral basins in the terminology suggested. Forearc basins and aulacogens have been described by some as epieugeosynclines and zeugogeosynclines, respectively, but others have applied different terms to analogous features and the same terms to different kinds of features. Such discordances in terminology are to be expected, given the dramatic change in frame of reference. Whatever terminology is used, progress in applying plate-tectonic theory to problems of sedimentation can come easily only if sedimentary basins are classified and discussed in a manner that is congruent with concepts of plate tectonics. In this paper, I have tried simply to find phraseology that would convey meaning now, without prejudice to either past or future usage.

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