CENTENNIAL ARTICLE

Tectonics of sedimentary basins

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ABSTRACT

Simultaneous breakthroughs in our understanding of plate-tectonic processes, depositional systems, subsidence mechanisms, chronostratigraphy, and basin-exploration methods have resulted in rapidly improving actualistic models for sedimentary basins. Basin analysis has become a true science with the development of quantitatively testable models based on modern basins of known plate-tectonic setting. Major subdivisions of basin settings include divergent, convergent, transform, and hybrid; 23 basin categories occur within these settings. Basins are classified according to primary plate-tectonic controls on basin evolution: (1) type of substratum, (2) proximity to plate boundary, and (3) type of nearest plate boundary(s). Sedimentary basins subside primarily owing to (1) attenuation of crust as a result of stretching and erosion, (2) contraction of lithosphere during cooling, and (3) depression of lithosphere by sedimentary and tectonic loads. The first two processes dominate in most divergent settings, whereas the third process dominates in most convergent settings. Intraplate, transform, and hybrid settings experience complex combinations of processes. Several basin types have low preservation potential, as predicted by their susceptibilities to erosion and uplift during orogeny and as confirmed by their scarcity in the very ancient record.

Key references concerning actualistic plate-tectonic models for each type of basin form the basis for reviewing the present state of the science. The key references come from many sources, with diverse authorship, including several publications of the Geological Society of America. The further development and refinement of actualistic basin models will lead to improved testable paleotectonic reconstructions.

INTRODUCTION

The 1970s and 1980s have been a dynamic era in the science of basin analysis. Several breakthroughs in our understanding of modern and ancient processes, at large and small scales, have resulted in major advances in several disciplines central to basin analysis. Best known to geoscientists, of course, is the "revolution in the earth sciences" brought about by the widespread acceptance of the paradigm of plate tectonics (for example, Cox and Hart, 1986). Equally important to basin analysts has been the revolution in our understanding of modern depositional systems and the consequent major advances in sophistication of actualistic depositional models (for example, Davis, 1983; Walker, 1984; Reading, 1986). Actualistic petrologic models relating sediment composition, especially sand and sandstone, to plate-tectonic settings have been developed (for example, Dickinson and Suczek, 1979). Exploration techniques, especially seismic stratigraphy and detailed mapping of ocean floors, have provided new avenues for investigation of both ancient and modern basins. In addition, refinement in analytical methods, including chronostratigraphy, subsidence analysis, microanalytical investigation of thermal history, paleomagnetism, and paleoclimatology, to name just a few, has revolutionized basin analysis, as summarized by Miall (1984) in a landmark textbook. Miall pointed out that a "new stratigraphy" is upon us. The past 20 years have been the beginning of the "golden age of the new stratigraphy" owing to the concurrent revolutions affecting the earth sciences. In fact, stratigraphy has become a true science only with the development of testable actualistic models based on a combination of theory, observation, and experiment. These actualistic models are the key to basin analysis from the micro- to the megascale. Dott (1978) reviewed the development of pre–plate-tectonic (primarily nonactualistic) models for megascale basins (geosynclines).

The emphasis of this article is on actualistic plate-tectonic models for sedimentary basins. Models dealing solely with plate-tectonic processes, without application to sedimentary basins, have been excluded. Also excluded are models dealing with details of depositional systems, basin architecture, petrology, or other aspects of the basins themselves. Several excellent case studies of specific basins are not mentioned, not because they are deficient, but because they do not develop actualistic plate-tectonic models for a class of basins. To paraphrase Walker (1984, p. 6), a basin model should act as (1) a norm, for purposes of comparison; (2) a framework and guide for future observations; (3) a predictor in other situations; and (4) an integrated basis for interpretation of the class of basin it represents.

Selection of key papers for each type of basin ranged from obvious and easy to arbitrary and difficult; selection was limited to the English literature. Each paper discussed herein captures the essence of the important controls on basin formation and evolution for that specific plate-tectonic setting. Some of these papers are reviews that summarize years of both the authors' and others' work. Other papers are breakthrough insights concerning the development of some basin types. Each selected paper is not necessarily the first or the final statement on a particular type of basin; rather,

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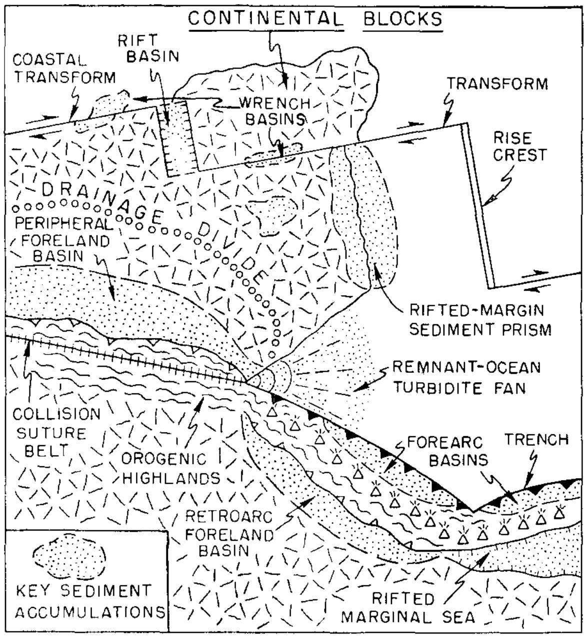

Figure 1. Sketch map showing sites of key sediment accumulations (basins) in relation to plate boundaries, continental margins, and associated sources of detrital sediment. Unpatterned areas represent oceanic crust. Continental margins and tectonic features are indicated by solid lines; other basin margins are indicated by dashed lines. Stipples indicate areas of sediment accumulation; "smoking" triangles represent magmatic arc. Solid barbs represent subduction zones; open barbs represent foreland fold-thrust belts. An intracratonic basin is shown between "drainage divide" (peripheral bulge) and "rifted-margin sediment prism" (continental rise and terrace). Modified from Dickinson (1980).

each paper represents (in my opinion) the best over-all statement and/or model at present. "Truth emerges more readily from error than from confusion" [Francis Bacon (Kuhn, 1970, p. 18)].

As is appropriate for a field as diverse as basin analysis, the authors of the papers range from sedimentologists to geophysicists. One of the most exciting aspects of this field of research is the integration of diverse disciplines.

One of the purposes of this Geological Society of America Centennial article is to highlight contributions of publications of the Society. The key papers selected come from diverse sources. As discussed below, however, publications of the Society have had a large impact on the science.

OVERVIEW AND BASIN CLASSIFICATION

Dickinson (1974, 1976) has provided the most comprehensive actualistic classification of basin types related to plate-tectonic processes. This classification is the basis for the organization of the remainder of this paper. Dickinson (1974) is a more formal publication, whereas Dickinson (1976) provides more detail. The true-scale diagrams of orogenic basins are an important addition to the latter publication.

To paraphrase Dickinson (1974, 1976), plate tectonics emphasizes horizontal movements of the lithosphere, which induce vertical movements due to changes in crustal thickness, thermal character, and isostatic adjustment. These vertical movements cause the formation of sedimentary basins, uplift of sediment source areas, and reorganization of dispersal paths. Primary controls on basin evolution (and the basis for classification) are (1) type of substratum, (2) proximity to plate boundary, and (3) type of nearest plate boundary(s). Types of substratum include continental crust, oceanic crust, transitional crust, and anomalous crust. Primary types of plate boundaries are divergent, convergent, and transform; hybrid boundaries also occur. Throughout this discussion, the distinction between continental margins and plate boundaries is important; they may or may not correspond. [For this reason, usage such as "collision of plates" should be avoided. Plates "converge"; only nonsubductable (buoyant) crustal components can "collide."] "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" (Dickinson, 1974, p. 1).

Subsidence of the crust to form sedimentary basins is induced by the following processes (Dickinson, 1974, 1976): (1) attenuation of crust due to stretching and erosion; (2) contraction of lithosphere during cooling; and (3) depression of both crust and lithosphere by sedimentary or tectonic loads, which are isostatically compensated either locally or regionally. The latter type of regional compensation of loads results in lithospheric flexure that can result in both subsidence and uplift far from applied loads. The first two processes dominate along most divergent plate boundaries, whereas the third process dominates along most convergent boundaries. Intraplate (including rifted continental margins), transform, and hybrid settings experience complex combinations of processes. Ancillary effects such as paleolatitude, paleogeography, and eustatic changes may provide modifying influences.

Tectonic controls on basin formation are diverse; therefore, the basin analyst must assimilate an enormous literature. Gross subdivision of basins into compressional and extensional types, or active versus passive margins, is even less useful than geosynclinal classification, which if nothing else, recognized the diversity and complexity of basin types (for example, Kay, 1951).

The first step in identifying essential components controlling basin development commonly is the construction of accurate maps and cross sections (preferably at true scale) of modern plate-tectonic systems. Figure 1 illustrates a courageous attempt to summarize almost every basin type in one diagram. Each of these basin types is discussed in terms of a key reference in the following sections.

Table 1 summarizes the types of basins discussed below. Dickinson (1974, 1976) discussed the over-all controls on basin development of most of these types. The key references discussed below provide either more complete reviews of controlling processes, further work detailing these processes, or more refined and quantitative models. Some of these basin models have been brought to high levels of quantitative sophistication, whereas others remain general and qualitative. Significant additional work is necessary before quantitatively testable models are available for all basin types.

Bally and Snelson (1980), Miall (1984), and Klein (1987), among others, provide useful reviews of additional aspects of basin classification, kinematics, and dynamics.

DIVERGENT SETTINGS

Sequential Rift Development and Continental Separation

Kinsman (1975) reviewed the sequential evolution of terrestrial rift valleys into juvenile ocean basins and ultimately, into mature rifted continental margins. Figure 2 illustrates the post-rift stages, using somewhat different nomenclature from that of Kinsman (1975). Kinsman's model is based on isostatic adjustment to lithospheric temperature and density distributions, and subsidence is analogous to the thermal subsidence of sea floor formed at spreading centers, with additional subsidence due to sedimentary loading. Early domal uplifts related to mantle plumes are postulated to precede crustal rifting, without consideration of the possibility (in fact, probability) that domal uplifts are responses to lithospheric extension rather than its cause (Sengor and Burke, 1978; Morgan and Baker, 1983). Kinsman also assumed "normal" continental crust prior to rifting, which is at odds with theoretical studies showing that thickened and heated crust is weaker and, thus, is more likely to be the locus of extension when tensile stresses are applied (Kusznir and Park, 1987; Lynch and Morgan, 1987). In any case, elevated rift shoulders (arch rims of Veevers, 1981) form on either side of the terrestrial rift and persist through the proto-oceanic phase (also, see Hellinger and Sclater, 1983) (Fig. 2A). Supracrustal thinning (erosion) occurs while these rims are elevated, resulting in subsidence below sea level during subsequent cooling as the continental edges move away from the spreading center. Kinsman suggested that the width of attenuated continental crust seaward of the rim is relatively narrow (60–80 km), although other workers have suggested that wider attenuated margins are common (for example, Cochran, 1983a; Lister and others, 1986). Rifting and continental separation commonly occur along alternating divergent and transform margins linking hot spots (Burke and Dewey, 1973), resulting in interplume, transcurrent, and plume margins, respectively. Kinsman's (1975) model emphasized simple isostatic compensation at a depth of 100 km, whereas regional lithospheric flexure (for example, Walcott, 1972) is a major control on subsidence, especially in terms of the seaward tilting of the shallow-marine (shelf) prism (for example, Pitman, 1978). Nonetheless, Kinsman's model successfully predicts the maximum stratigraphic thickness found in continental embankments (16–18 km) based on the loading capacity of oceanic lithosphere.

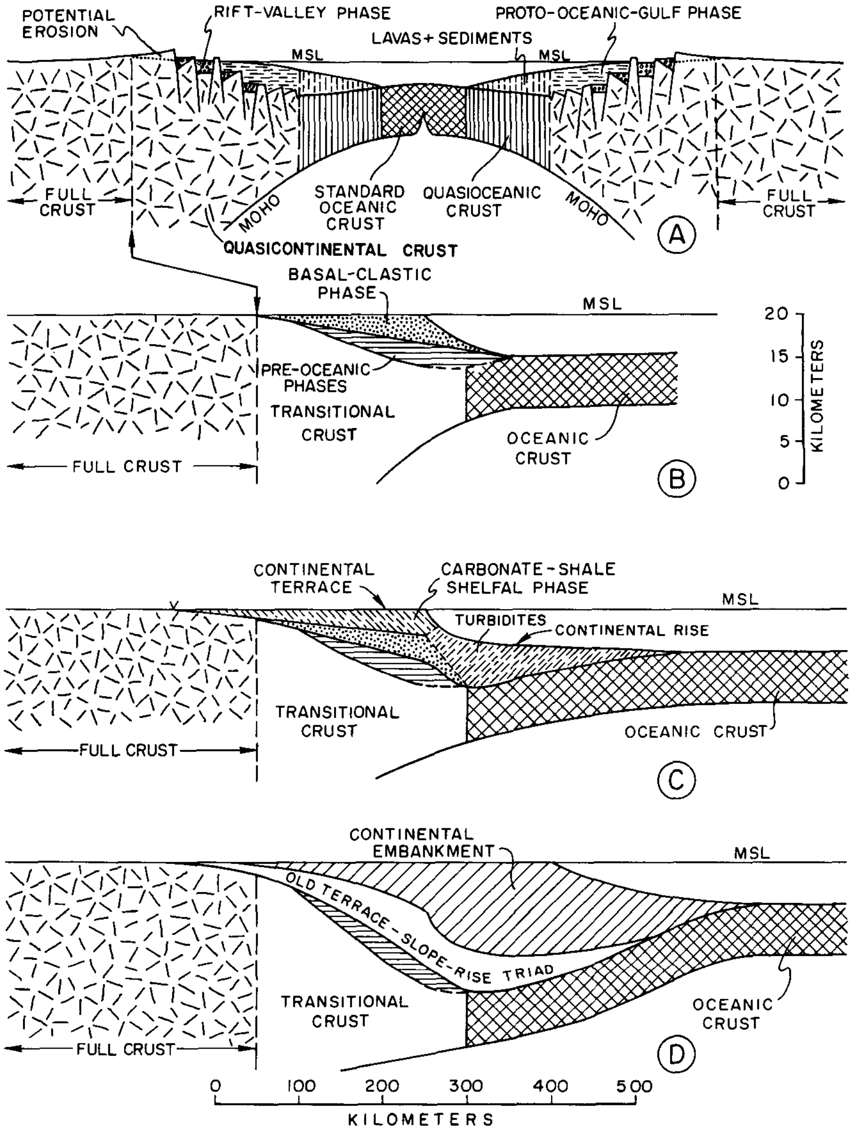

Figure 2. Schematic diagrams (vertical exaggeration 10×) to illustrate general evolution of rifted-margin prism along rifted continental margin: (A) proto-oceanic stage showing rift-valley depositional phases on top of attenuated continental (quasi-continental) crust, adjacent to thickened basaltic (quasi-oceanic) crust; (B) end of proto-oceanic stage when thermal subsidence is nearing completion; quasi-continental and quasi-oceanic crust are combined into transitional crust, underlying the subsiding continental margin; (C) continental terrace-slope-rise configuration during open-ocean stage, during which sediment loading is predominant subsidence mechanism; (D) continental-embankment stage, reached only where sediment delivery is voluminous enough to cause progradation of shoreline over oceanic crust (in areas of major deltas, usually at open ends of failed rifts). Modified from Dickinson (1976).

Terrestrial Rift Valleys

Surprisingly little work has been published on the tectonic development of terrestrial rift valleys in terms of basin models. Many volumes have been published on geophysics, geochemistry, volcanism, geomorphology, and other aspects of modern rifts, but few attempts have been made to synthesize the temporal evolution of rifts in terms of general models for basin development. Rifts form in many plate-tectonic settings: (1) craton interiors unrelated to orogeny (East Africa), (2) intracontinental zones related to continental collision (Rhine graben and Baikal rift), (3) transtensional rifts along transform faults (Dead Sea and Salton Sea), and (4) settings related to complex plate interactions at subduction zones and transform margins (Rio Grande rift). Orogenic activity immediately following rifting may deform rifts and convert them into complex basins. If no orogenic activity occurs following a finite amount of rifting, the rifts may "fail," and intracratonic basins may develop subsequently. In contrast, if rifting is "successful," then continental separation occurs and the rift basins are pulled apart to form new ocean basins. Thus, ancient rift basins commonly are buried under younger sediments, deposited either in intracratonic basins (overly-

TABLE 1. BASIN CLASSIFICATION

Divergent settings	
	Terrestrial rift valleys: rifts within continental crust, commonly associated with bimodal volcanism
	Proto-oceanic rift troughs: incipient oceanic basins floored by new oceanic crust and flanked by young rifted continental margins
	Continental rises and terraces: mature rifted continental margins in intraplate settings at continental-oceanic interfaces
	Continental embankments: progradational sediment piles constructed off edges of rifted continental margins
	Failed rifts and aulacogens: inactive terrestrial rift valleys, which may be reactivated during convergent tectonics and become aulacogens at high angles to orogenic belts
	Intracratonic basins: broad cratonic basins floored by failed rifts in axial zones
	Oceanic basins: basins floored by oceanic crust formed at divergent plate boundaries unrelated to arc-trench systems
	Oceanic islands, aseismic ridges, and plateaus: sedimentary aprons and platforms formed in intraoceanic settings other than magmatic arcs
Convergent settings	
	Trenches: deep troughs formed by subduction of oceanic lithosphere
	Trench-slope basins: local structural depressions developed on subduction complexes
	Forearc basins: basins developed between subduction complexes and magmatic arcs
	Intra-arc basins: local basins within magmatic arcs
	Interarc and backarc basins: oceanic basins between and behind intraoceanic magmatic arcs, and continental basins behind continental-margin magmatic arcs without foreland fold-thrust belts
	Retroarc foreland basins: foreland basins on continental sides of continental-margin arc-trench systems
	Remnant ocean basins: shrinking ocean basins caught between colliding continental margins and/or arc-trench systems, and ultimately subducted or deformed within suture belts
	Peripheral foreland basins: foreland basins above rifted continental margins that have been pulled into subduction zones during crustal collisions
	Piggyback basins: basins formed and carried atop moving thrust sheets
	Foreland intermontane basins: basins formed among basement-cored uplifts in foreland settings
Transform settings	
	Transtensional basins: basins formed by extension along strike-slip fault systems
	Transpressional basins: basins formed by compression along strike-slip fault systems
	Transrotational basins: basins formed by rotation of crustal blocks about vertical axes within strike-slip fault systems
Hybrid settings	
	Intracontinental wrench basins: diverse basins formed within and on continent crust due to distant collisional processes
	Successor basins: basins formed in intermontane settings following cessation of local orogenic activity

Note: table modified after Dickinson (1974, 1976).

ing failed rifts) or along rifted continental margins (overlying successful rifts and adjacent failed rifts).

Leeder and Gawthorpe (1987) have provided the latest synthesis of sedimentary models for extensional tilt-block/half-graben basins. They outlined the primary control that faulting exerts on dispersal paths, depositional systems, and basin architecture and presented four qualitative models relating surface morphology, fault activity, and facies. In all of the models, coarse-grained steep fans of small volume are derived from footwall uplands, whereas finer-grained broad cones are derived from hanging-wall dip slopes, except in the case of marine inundation. Lacustrine, fluvial, or deep-marine sediments may accumulate in areas immediately adjacent to fault scarps, depending upon over-all paleogeography. Tilt-block/half-graben basins are common in the ancient record, but it is not clear how many of them represent primarily high-angle faulting, listric faulting, or low-angle detachment faulting, as envisioned by Wernicke (1985) and Kusznir and others (1987). Development of refined depositional models based on studies of young extensional basins is needed to provide constraints for interpreting fault geometry and tectonics in ancient settings (for example, Frostick and Reid, 1987).

Proto-oceanic Rift Troughs

The Red Sea is the type proto-oceanic rift trough. Caution is needed, however, in constructing a general model for such settings because the Red Sea (including the Gulf of Aden) is the only modern proto-ocean on Earth. The Gulf of California is primarily a transtensional feature, although it shares many characteristics with the Red Sea. It is not surprising, therefore, that a summary of the history of the Red Sea is the basis for a proto-oceanic model.

Cochran (1983a) presented convincing geophysical evidence for active sea-floor spreading in the southern Red Sea, whereas north of 25°N, the center of the trough appears to be attenuated continental crust. Plate reconstructions dictate that significant divergence has occurred between Africa and Arabia since the end of the Oligocene. This necessitates the presence of quasicontinental and quasioceanic crust under most of the Red Sea (for example, Fig. 2A). Cochran (1983a) argued that the northern and southern Red Sea represent the earlier and later stages, respectively, of the transition from terrestrial rifting to sea-floor spreading. "An initial period of diffuse extension by rotational faulting and dike injection over an area perhaps 100 km wide is followed by concentration of extension at a single axis and the initiation of sea-floor spreading" (Cochran, 1983a, p. 41).

An extended phase of rifting and diffuse extension must be accounted for in thermal models for post-rifting subsidence (Cochran, 1983b). Specifically, horizontal heat flow causes additional cooling within the rift and uplift of the rift shoulders. Subsequent thermal subsidence of the continental margins is less than would be expected following "instantaneous" subsidence (for example, McKenzie, 1978).

Continental Rises and Terraces

Subsidence mechanisms at rifted continental margins include (1) thinning of crust due to stretching and erosion during doming; (2) thermal subsidence following rifting; (3) loading by sediment, resulting in both local isostatic and regional lithospheric flexure; and (4) lower-crustal and sub-crustal flow and densification following rifting.

Pitman (1978) elegantly demonstrated how transgressive/regressive sequences are formed on shelves along rifted continental margins. He showed that subsidence rates at shelf edges are normally greater than rates at which eustatic sea level changes. Shelves can be modeled as platforms rotating about a landward hinge line. Thus, shorelines seek locations that reflect a balance among sea-level change, subsidence rate, and sedimentation rate. The net result is that transgressive/regressive sequences reflect changes in rates of sea-level change, rather than changes in sea level. All modern rifted continental margins are younger than Paleozoic (post-breakup of Pangea), and most seismic "sea-level curves" (for example, Vail and others, 1977) are constructed primarily from Cretaceous and Cenozoic sequences along Atlantic margins. Sea level has been generally falling since its high point in the Cretaceous, owing to the combination of increase of the average age of oceanic crust (for example, Heller and Angevine, 1985) and the initiation of continental glaciation during the middle Cenozoic. The record of Cenozoic transgressive/regressive sequences along Atlantic margins is consistent with continually falling sea level, but at varying rates. Diverse tectonic processes provide mechanisms for second-order changes in sea level (for example, Cloetingh and others, 1985; Karner, 1986).

Pitman's model shows how the seaward-thickening wedge of the continental terrace can form (for example, Fig. 2C) in response to the combined effects of tilting about a landward hingeline, changes in sea level, and changes in sedimentation rates. The shoreline would be at the shelf edge only during times of unusually rapid sea-level fall (for example, glaciation or sudden flooding of a formerly isolated dry ocean basin as in the case of, for example, the Messinian Mediterranean) or in areas of very high sedimentation (see below).

Continental Embankments

Burke (1972) summarized depositional processes leading to the development of the Niger Delta (the surficial expression of a continental embankment), as schematically represented in Figure 2D. Prevailing southwest winds impinge symmetrically on the Niger Delta and neighboring coasts, thus setting up converging longshore drifts, which meet at the delta corners. Submarine canyons at these corners (and at the center of the delta during low sea level) feed voluminous sediment to submarine fans at the delta foot. The net result is a five-layer structure as the continental margin progrades seaward over oceanic crust and older pelagic sediments. From bottom to top, these layers are (1) deep-sea–fan sands; (2) transitional sand and mud; (3) slope mud, with abundant diapirs and slumps; (4) transitional shoreline and shelf sand and mud; (5) continental (primarily fluvial) sand. Continental embankments are gravitationally unstable owing to rapid burial of low-density, water-saturated sediments, and the common occurrence of deeply buried evaporite deposits inherited from the proto-oceanic stage. Therefore, salt diapirs, mud diapirs, and growth faulting are common.

Continental embankments such as the Niger Delta and the Mississippi Delta form in areas of unusually rapid sedimentation, primarily at the mouths of failed rifts (Burke and Dewey, 1973). Drainage from large cratons commonly is directed toward the mouths of failed rifts and

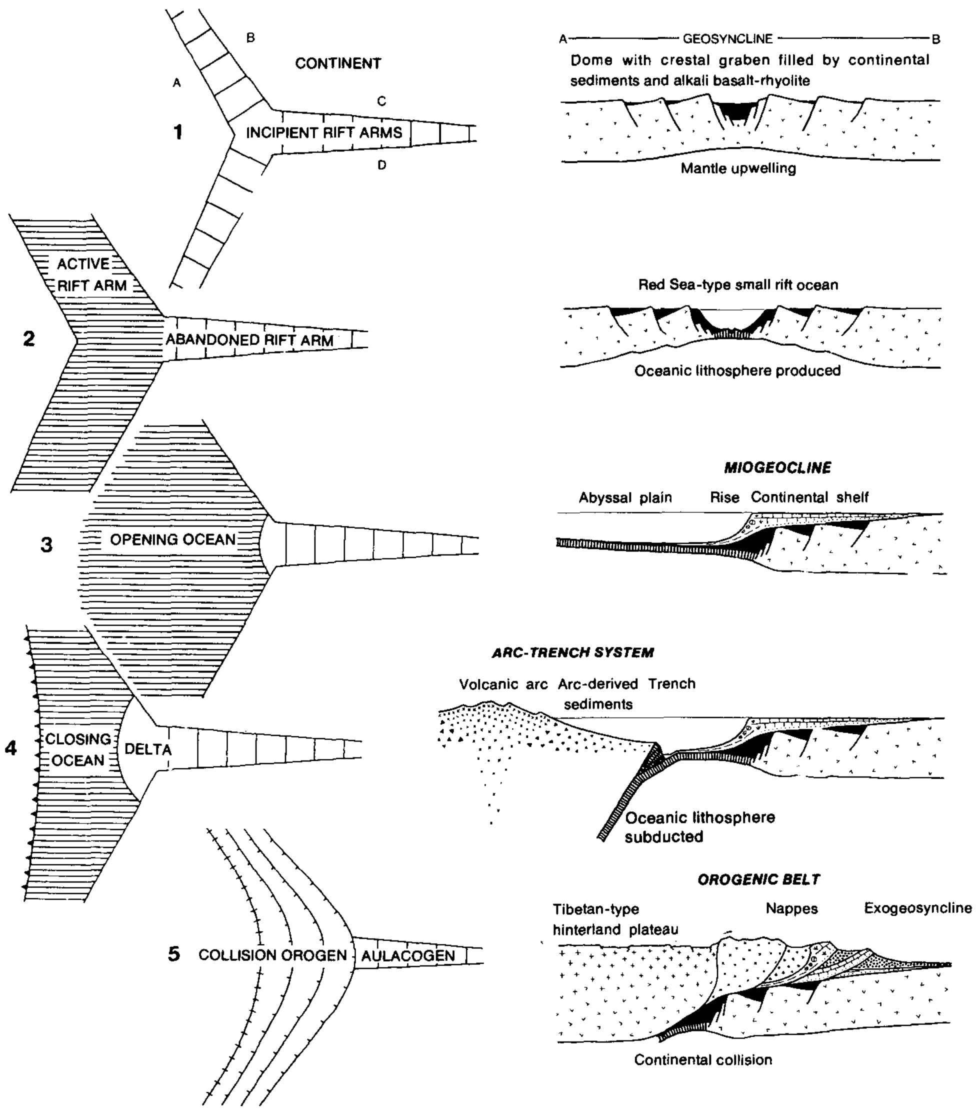

Figure 3. Model for evolution of an aulacogen and related orogenic belt. (1) A three-armed radial rift system accompanied by crustal doming; strike-slip component (transform fault) may be important along any of the rift arms. (2) Two rift arms spread to produce narrow ocean similar to Red Sea; rifting of third arm insufficient for continental separation. (3) Spreading of two active rift arms produces large ocean basin. Downwarping and terrace-rise sedimentation occur along new aseismic continental margins. Third arm fails and remains as transverse trough located at re-entrant on continental margin. The failed rift arm evolves from incipient rift to broad downwarp (intracratonic basin). (4) Ocean is closed by subduction along trench, producing adjacent magmatic arc. History of ocean closure may take many paths, only one of which is shown here. (5) Closing of ocean ultimately results in continental collision and development of collision orogen. Abandoned rift arm is preserved as an aulacogen located where orogen makes re-entrant into its foreland. Aulacogen is further loaded with peripheral-foreland sediments derived from advancing orogen, and its medial block founders, resulting in final stage of compressional deformation and faulting. Reproduced by permission of Society of Economic Paleontologists and Mineralogists (from Hoffman and others, 1974).

away from adjacent "normal" rifted continental margins. Similar massive outbuilding of continental margins due to deltaic progradation occurs in remnant ocean basins (for example, Bengal and Indus fans; see below); however, these latter deposits form in tectonically active settings related to continental suturing. Therefore, provenance, dispersal orientations, sediment types, related petrotectonic assemblages, and final deformational features contrast with continental embankments formed at failed rifts, even though gross structural cross sections of these progradational continental margins are similar. Kinsman (1975) predicted that the maximum possible stratal thickness of such margins is 16–18 km.

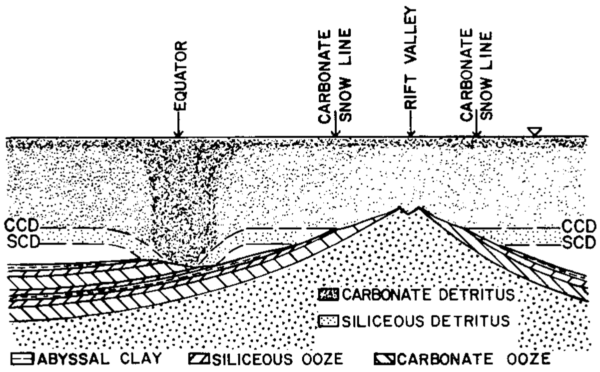

Figure 4. Model of axially accreting oceanic sedimentation (extreme vertical exaggeration, especially for stratal thicknesses). Deposition of carbonate ooze dominates on oceanic crust (solid-triangle pattern) shallower than the CCD (carbonate compensation depth); siliceous ooze accumulates between the CCD and SCD (silica compensation depth), and only abyssal clay accumulates below the SCD. Both the CCD and the SCD are depressed near equator and in other areas of high biologic productivity (zones of upwelling). Predictable stratigraphic sequences result as oceanic crust formed at spreading center (rift valley) cools and subsides as it moves away from spreading center (see text). After Heezen and others (1973).

Failed Rifts and Aulacogens

During continental rifting, three rifts commonly form at approximately 120°, probably because this is a least-work configuration (Burke and Dewey, 1973). Regardless of whether initiating processes are "active" or "passive" (Sengor and Burke, 1978; Morgan and Baker, 1983), in the majority of cases, two rift arms proceed through the stages of continental separation (outlined above), whereas one rift arm tends to fail (Fig. 3). Hoffman and others (1974) discussed the resulting sedimentary accumulations, with emphasis on a Proterozoic example. They outlined five stages in the development of the Athapuscow aulacogen, which with slight modification, provide a model applicable to most aulacogens (linear sedimentary troughs at high angles to orogens): (1) rift stage, (2) transitional stage, (3) downwarping stage, (4) reactivation stage, (5) post-orogenic stage.

Although the actualistic model developed by Hoffman and others (1974) may be applied to several failed rifts and aulacogens, Hoffman (1987) has recently questioned whether it is the best model for the Athapuscow "aulacogen." Also, Thomas (1983, 1985) has discussed the possibility that the southern Oklahoma "aulacogen" originated as a transform boundary rather than as a failed rift. Neither of these reinterpretations discredits Hoffman and others' (1974) original model; rather, both examples illustrate the difficulty in applying any model to the complexity of the real world. Progress results from the application and testing of models; the real world is always more complex.

Successful rifts evolve into shelf-slope-rise margins, with continental embankments at reentrants (for example, Niger Delta). Failed rifts grade from embankments at their mouths to terrestrial rifts within cratons. As lithospheric extension ceases (for any reason), the rift areas cool and flexurally subside to form intracratonic basins, especially where three failed rifts meet. Upon activation or collision of the continental margin, the rifted-margin sedimentary prisms are intensely deformed, especially at continental promontories (Dewey and Burke, 1974; Graham and others, 1975). As orogeny proceeds, failed rifts become aulacogens, which may experience compressional, extensional, or translational deformation.

Impactogens (Sengor and others, 1978) resemble aulacogens, but without the pre-orogenic stages (1–3) (see below).

Intracratonic Basins

With increased exploration of deeper levels of well-known intracratonic basins (for example, the Michigan basin), it has become clear that most intracratonic basins have formed above ancient failed rifts (see Klein and Hsui, 1987).

Derito and others (1983) developed a lithospheric-flexure model with a nonlinear Maxwell viscoelastic rheology that helps explain how intracratonic basins can have long histories of synchronous subsidence over broad areas of continents. They pointed out that predicted subsidence due to stretching and cooling following cessation of rifting (for example, McKenzie, 1978; Sclater and Christie, 1980) is insufficient to explain the magnitude or timing of subsidence for most intracratonic basins. Many intracratonic basins (for example, Michigan, Illinois, and Williston basins) have experienced renewed subsidence during times of orogeny in adjacent orogenic belts. These periods of reactivation may be due to the reduction of effective viscosity of the lithosphere resulting from applied stress (due to adjacent orogeny) (Derito and others, 1983). Due to flexural rigidity of the lithosphere, dense loads emplaced in the crust during rifting (for example, basaltic dikes) remain isostatically uncompensated for geologically long periods of time. Any stress applied to the lithosphere results in a geologically instantaneous relaxation and increased subsidence. Repeated changes of stress due to orogenic activity decrease the effective viscosity beneath the basin, and the basin subsides more rapidly than during nonorogenic periods. Thus, the nearby orogenic activity allows the basin (underlain by a failed rift) to approach isostatic compensation more rapidly than it would without orogenic activity. Cumulative subsidence over hundreds of millions of years approaches that predicted by simple thermal models (for example, McKenzie, 1978). The model of Derito and others (1983) presents an elegant explanation for continent-wide synchronism of depositional sequences due to orogenic activity along nearby continental margins.

Quinlan (1987) has provided a thorough review of mechanisms and models for intracratonic basins.

Oceanic Basins

Heezen and others (1973) presented a kinematic model to explain the distribution of Cretaceous and Cenozoic sediment in the modern Pacific Ocean. The model is based on the systematic increase in depth of oceanic crust with age, combined with the dependence of sediment type on depth of water (Fig. 4). Biogenic oozes are volumetrically predominant away from continental margins, with generally more rapid production of calcareous pelagic detritus than siliceous pelagic detritus. Virtually all of this biogenic material is produced within the photic zone (approximately the upper 150 m of water),

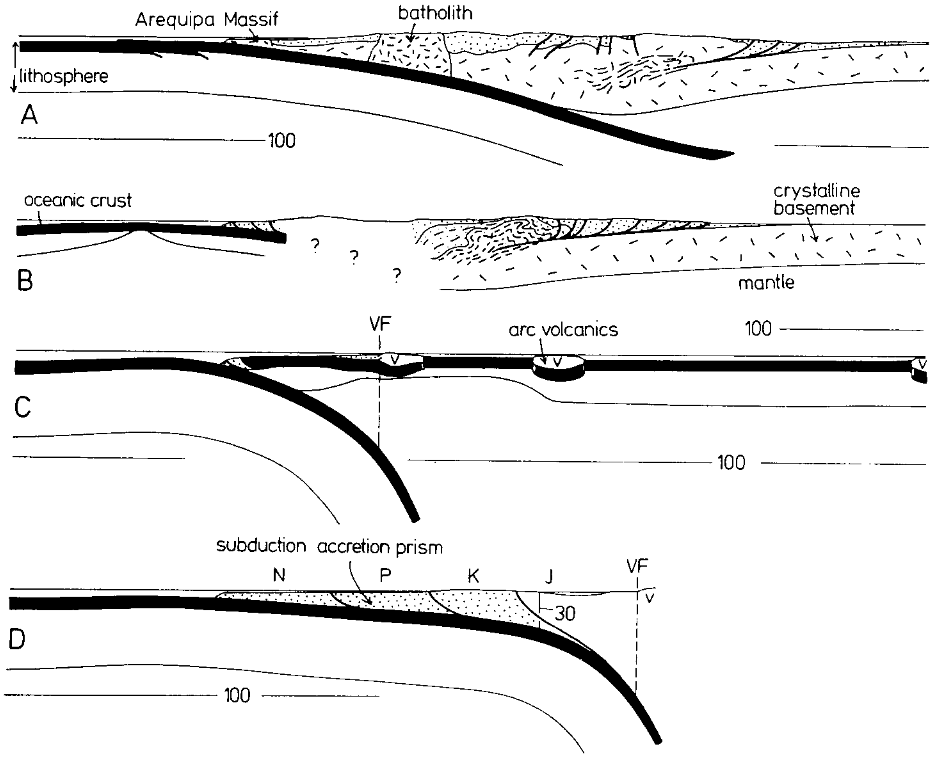

Figure 5. True-scale sections across Pacific plate margins (scale in kilometers). (A) Central Peru. (B) Western Canada. (C) Marianas. (D) Alaska (age of subduction-accretion prism: J = Jurassic, K = Cretaceous, P = Paleogene, N = Neogene). See text for discussion. Modified from Dewey (1980).

and upon the death of the secreting organisms, these tests settle slowly through the water column, dissolving as they fall. The depth below which no carbonate remains is the carbonate compensation depth (CCD). This depth is determined by the balance of biologic productivity in the photic zone and of the rates at which tests fall and dissolve in the water column (Berger and Winterer, 1974). Changes in any of these rates can change the CCD, which presently averages −3,700 m in the open ocean (Heezen and others, 1973) and −5,000 m near the equator (Berger, 1973). The CCD is depressed near the equator (and in other areas of upwelling) owing to higher biologic productivity. The silica compensation depth (SCD) is less well defined than is the CCD, but it is universally deeper than the CCD owing to the slower dissolution of silica tests. Below the SCD, only nonbiogenic clay accumulates (Fig. 4).

Oceanic crust formed at the East Pacific Rise south of the equator has the potential to subside below both the CCD and the SCD before crossing the equator; given the right depth (age) of crust and depth of equatorial CCD and SCD, an equatorial sequence of siliceous and calcareous oozes also may be deposited. Ancient oceanic crust with this history should, therefore, be overlain by the following vertical lithologic sequence: carbonate ooze, siliceous ooze, abyssal clay, siliceous ooze, carbonate ooze, siliceous ooze, abyssal clay. Heezen and others (1973) developed this model to illustrate possible combinations of ridge location and equatorial-transit history. The result is a predictive model that is in excellent agreement with the Cretaceous and Cenozoic stratigraphy of the Pacific plate. They discussed the formation of time-transgressive lithofacies (see the interesting discussion of oceanic stratigraphic principles by Cook, 1975) including pyroclastic material derived from western Pacific magmatic arcs and turbidite fans derived from the North American margin. Winterer (1973) has provided an additional example of paleotectonic reconstructions using Pacific plate stratigraphy.

Heezen and others' (1973) kinematic model is a potentially powerful tool for interpreting depositional histories of ophiolite sequences preserved in subduction complexes and suture zones. Caution is advised, however, because most ophiolites probably formed in ancient marginal basins rather than open-ocean settings, in which case, models for interarc and backarc basins are more useful for reconstructing paleotectonic settings. In addition, biologic controls on carbonate and siliceous sedimentation have changed markedly owing to evolutionary processes (Berger and Winterer, 1974); therefore, pre-Cenozoic oceanic models cannot be truly actualistic.

Oceanic Islands, Aseismic Ridges, and Plateaus

The model discussed above for the systematic subsidence of oceanic lithosphere as it travels away from divergent boundaries has important implications for intraoceanic sedimentation in general. All islands, aseismic ridges, and plateaus constructed of unusually thick basaltic material (for example, Hawaiian-Emperor chain) experience subsidence equal to the subsidence of oceanic lithosphere on which they ride. During growth due to active basaltic volcanism, only minor fringing sediments are likely to be deposited.

Clague (1981) divided the post-volcanic history of seamounts into three stages: subaerial, shallow water, and deep water or bathyal. Following cessation of volcanism, seamounts are likely to accumulate pelagic and shallow-marine carbonate sediments, with rate and type of sedimentation controlled largely by latitude (Clague, 1981). If carbonate sedimentation is equal to or greater than thermal subsidence, then fringing reefs and atolls will form atop the submerged islands and plateaus. Some of the depositional and migratory behavior of ancient islands, ridges, and plateaus accreted to continental margins at subduction zones may be reconstructed based on this model.

CONVERGENT SETTINGS

Arc-Trench Systems

Dewey (1980) classified arc-trench systems as extensional, neutral, or compressional, analogous to Dickinson and Seely's (1979) migratory-detached, noncontracted-stationary, and contracted categories (Fig. 5). Extensional arcs are intraoceanic owing to the formation of oceanic crust within and behind magmatic arcs as a result of trench rollback being faster than trenchward migration of the overriding plate. The western Pacific intraoceanic arcs are typical modern examples, with steep Benioff-Wadati zones dipping westward and subduction of old oceanic lithosphere (for example, Molnar and Atwater, 1978). Compressional arcs occur where a continental margin advances trenchward faster than trench rollback. The Andes typify these arcs, with shallow Benioff-Wadati zones dipping eastward and subduction of young oceanic lithosphere. Neutral arcs result where trench rollback is approximately equal to the trenchward advance of the overriding plate. The Aleutian and Indonesian arcs typify these arcs, with intermediate-dip Benioff-Wadati zones dipping northward and subduction of intermediate-age oceanic lithosphere.

Whether facing direction, age of subducted lithosphere, or a linked combination of both processes is the controlling factor in arc behavior is yet to be resolved (Dickinson, 1978). Dewey's (1980) kinematic model for arc behavior is useful, however, whatever the dynamic controls. Second-order effects include the probability that extensional arcs experience primarily basaltic magmatism and have low relief, thin sediments, and deep trenches. In contrast, compressional arcs experience silicic magmatism and have high relief, abundant sediments, and shallow trenches. Most arc-trench systems have intermediate characteristics, commonly including transform motion along arc trends; the superposition of several complex convergent zones results in great complexity (for example, Hamilton, 1979).

Further development of predictive models for the evolution of continental margins during the opening and closing of ocean basins (the Wilson cycle; see below) will come primarily from ancient orogenic belts, owing to the limited variety of young continental collisions. Dewey's (1980) kinematic model based on actualistic examples of magmatic arcs is a step in the direction of constraining the possibilities.

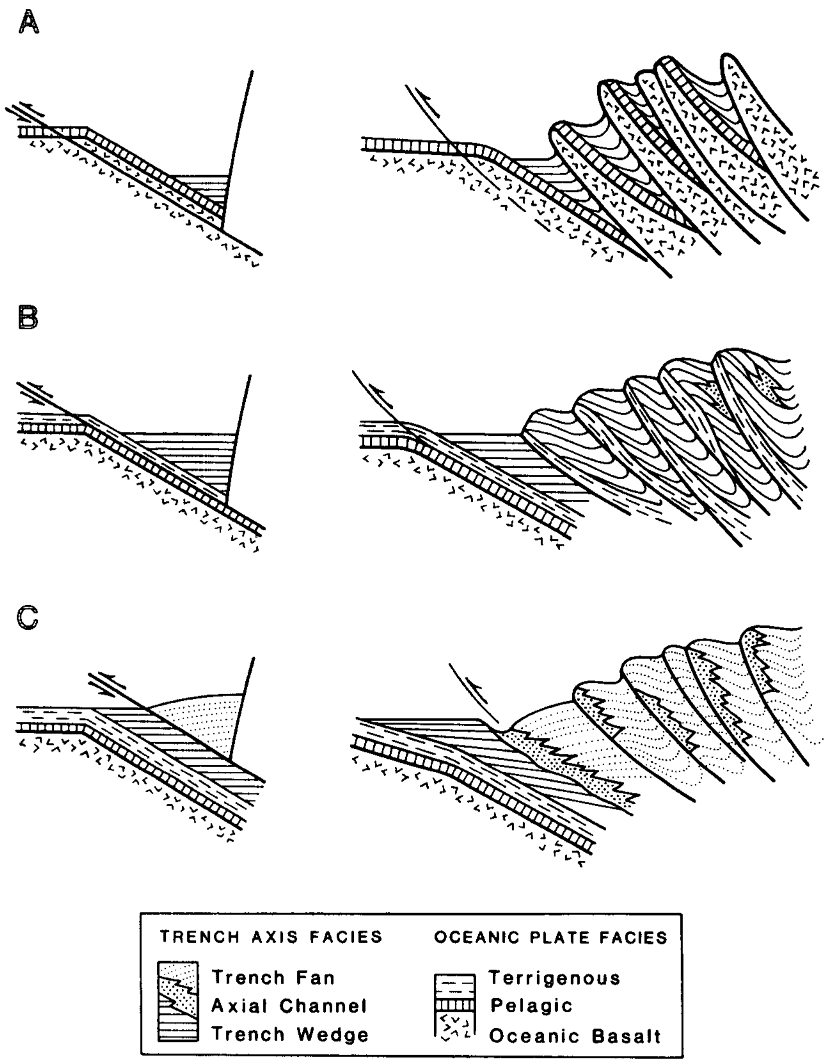

Figure 6. Accretion and preservation of trench deposits; stratigraphic control on position of décollement zone (thrust symbol) within the trench basin (Chile Trench). The lithofacies succession (left) determines the stratigraphic assemblage that is preferentially accreted to the adjacent continental slope (right). Sediment supply varies from meager (A) to abundant (C) along a margin, resulting in subduction complexes that include slivers of oceanic crust (ophiolite) with thin stratal packages (A), mixtures of trench and oceanic facies (B), or predominantly trench facies (C). Sediments that are not accreted at the trench presumably are underplated and/or melted at greater depths. From Thornburg and Kulm (1987).

Trenches

Thornburg and Kulm (1987) have provided one of the most detailed studies of sedimentation in a modern trench. The Chile Trench is especially interesting because it provides markedly different climatic and sedimentologic conditions along the studied length from 18° to 45° S latitude. Subducted crust is younger to the south, but subduction is orthogonal and rapid everywhere. Thus, plate-tectonic processes are relatively constant along the trench, and sedimentary processes can be isolated as causes and effects.

Karig and Sharman (1975) and Schweller and Kulm (1978), among others, discussed the dynamic nature of sedimentation and accretion at trenches. The sediment wedge of a trench is in dynamic equilibrium when subduction rate and angle, sediment thickness on oceanic plate, rate of sedimentation, and distribution of sediment within the trench are constant. Schweller and Kulm (1978) presented an empirical model relating convergence rate and sediment supply to types of trench deposits. Thornburg and Kulm (1987) provided documentation for the dynamic interaction of longitudinally transported material (trench wedge with axial channel) and transversely fed material (trench fan). With increasing transverse supply of sediment to the trench, the axial channel of the trench wedge is forced seaward, and the trench wedge widens. Contrasts in dynamic trench-fill geometries help determine not only trench bathymetry and depositional systems but also accretionary geometry (Fig. 6). This dynamic model may be useful for reconstruction of sedimentary and tectonic processes in trenches, as expressed in ancient subduction complexes.

Trench-Slope Basins

Moore and Karig (1976) developed a model for sedimentation in small ponded basins along inner trench walls. Deformation within and on the subduction complex results in irregular bathymetry, commonly including ridges subparallel to the trench. Turbidites are ponded behind these ridges, and trench-slope basins form. Average width, sediment thickness, and age of basins increase up slope due to the progressive

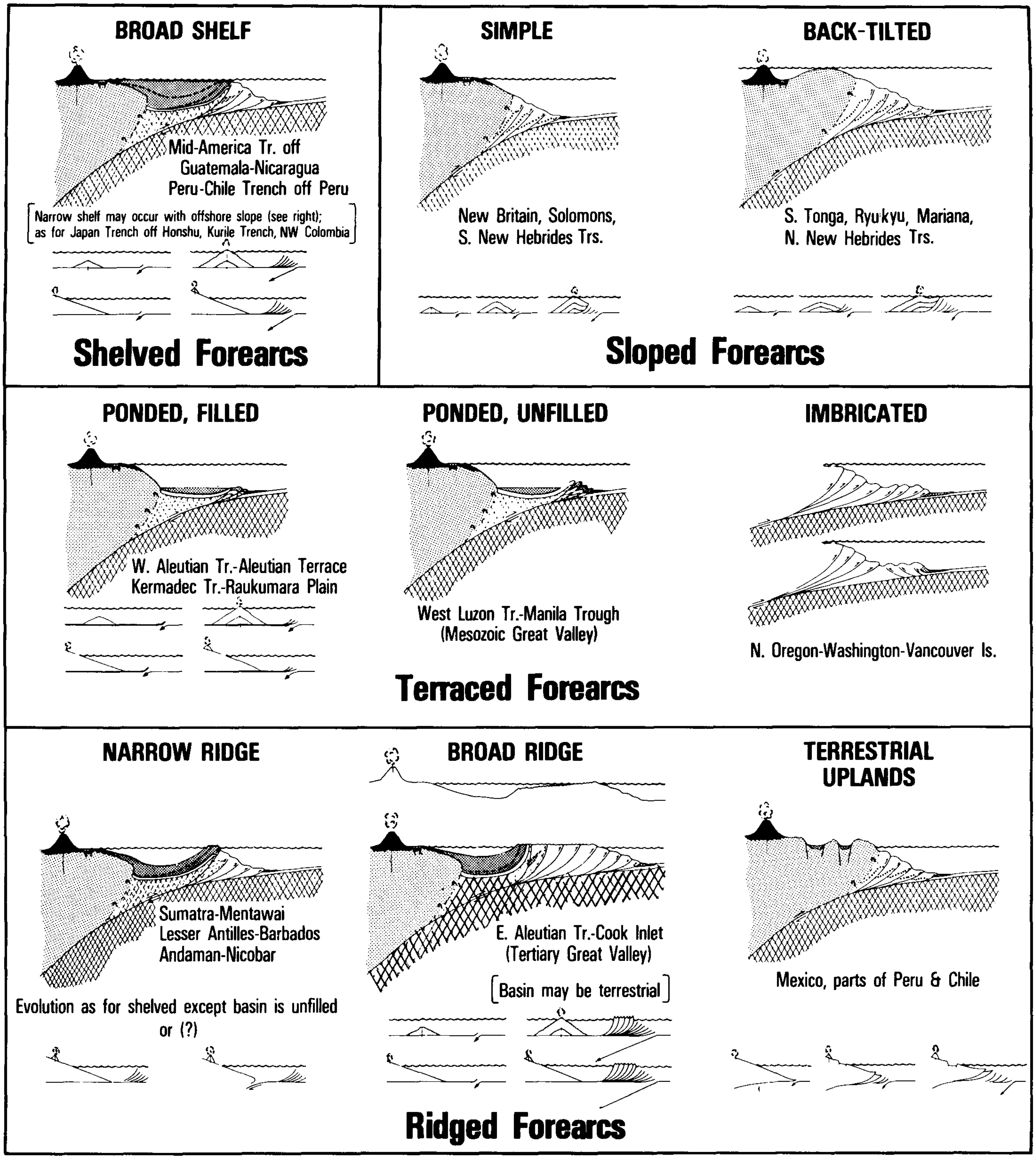

Figure 7. Configurations of modern forearc regions. Volumes of subduction complexes in modern forearcs are unknown in most areas (note question marks on diagrams), as are subsurface extent and nature of forearc igneous and metamorphic rocks. Where volumes of subduction complexes are small, arc massifs may be nearer trench than shown. Modified from Dickinson and Seely (1979).

uplift of deformed material and the widening of fault spacing during dewatering and deformation of offscraped sediment. In ancient subduction complexes, trench-slope basins are filled with relatively undeformed, locally derived turbidites surrounded by highly deformed accreted material of variable origin. Contacts between trench-slope basins and accreted material are both depositional and tectonic.

Moore and Karig's (1976) model was developed for Nias Island near Sumatra, an area of rapid accretion of thick sediments. Their model is less useful for sediment-starved forearc areas. Nonetheless, their general principles governing the development of sedimentary basins on the lower trench slope are fundamental to reconstructing ancient subduction settings.

Forearc Basins

Dickinson and Seely (1979) provided a classification of arc-trench systems, similar to Dewey's (1980), and outlined plate-tectonic controls governing subduction initiation and forearc development. Figure 7 illustrates the variability of forearc types. Factors controlling forearc geometry include (1) initial setting, (2) sediment thickness on subducting plate, (3) rate of sediment supply to trench, (4) rate of sediment supply to forearc area, (5) rate and orientation of subduction, and (6) time since initiation of subduction. Arc-trench gaps tend to widen through time (Dickinson, 1973) owing to prograde accretion at trenches and retrograde migration of magmatic arcs following subduction initiation. Prograde accretion is especially rapid where thick sequences of sediment are accreted. The net result of widening of the arc-trench gap is the general tendency for forearc basins to enlarge through time (for example, Great Valley forearc basin; Ingersoll, 1979, 1982).

Forearc basins include the following types (Dickinson and Seely, 1979): (1) intramassif (one type of intra-arc), (2) accretionary (trench-slope), (3) residual (lying on oceanic or transitional crust trapped behind the trench when subduction began), (4) constructed (lying across the boundary of arc massif and subduction complex), and (5) composite (combination of above settings). Residual and constructed basins tend to evolve into composite basins; commonly, this evolutionary trend is accompanied by filling and shallowing of the forearc basins.

Intra-arc Basins

Intra-arc basins are dominated by volcanic flows and volcaniclastic deposits, owing to their intimate association with active volcanoes. Depositional environments are varied but are determined, in large part, by whether the arc-trench system is extensional (oceanic), neutral (shelfal), or compressional (mountainous) (Dickinson, 1974, 1976; Dewey, 1980). Composition of magma, and therefore, style of eruption and volcanism, is determined, in large part, by type of underlying crust, oceanic (basaltic), transitional (andesitic), or continental (rhyodacitic) (Hamilton, 1979). Basins form primarily owing to tectonomagmatic collapse within eruptive centers, especially within silicic systems. Thus, "tectonic" processes are dominated by local extension above rising plutons, with little regard to regional stresses. Intra-arc basins may evolve into interarc basins during backarc spreading in extensional systems (see below), they may collapse as calderas and be engulfed in plutons in neutral systems, or they may be uplifted and destroyed in compressional systems. Thus, intra-arc basins have low preservation potential, and ancient examples are scarce.

Few studies of modern intra-arc basins have been published. This dearth of studies is due largely to the lack of integration of volcanology, sedimentology, and basin analysis. Publication of two recent textbooks (Fisher and Schmincke, 1984; Cas and Wright, 1987) suggests that this neglect is ending.

Busby-Spera (1984a, 1984b) provided a model to explain marine intra-arc sedimentation along the early Mesozoic continental margin of California. This basin model is not rigorously actualistic owing to the lack of modern studies. The sedimentologic and volcanologic aspects of the study, however, are actualistic. There is a risk in developing "actualistic" basin models based on ancient examples, especially in structurally complex, metamorphosed terranes such as Sierra Nevada "roof pendants" (for example, Christensen, 1963). Nonetheless, Busby-Spera (1984a, 1984b) outlined broad characteristics of intra-arc basins formed within neutral arc-trench systems and developed a model with predictive value in other studies.

Busby-Spera (1984a, 1984b) has demonstrated that complex facies changes are to be expected in the dynamic situation typified by silicic-andesitic volcanism near sea level along an open-ocean coast. Voluminous eruptions of rhyolite ash-flow tuff, construction of small andesite stratocones, and growth of submarine fans and debris aprons occur during times of volcanic activity; quiescence results in deposition of fine-grained epiclastic sediment and ash and/or progradational shelf sequences. All of the variability of modern shelves, shorelines, and nearby basins occurs within intra-arc basins near sea level. High sedimentation rates, complex facies interfingering, and complex interaction of magmatic, tectonic, eustatic, and sedimentary processes are expected. Development of more predictive actualistic models awaits study of modern systems.

Interarc and Backarc Basins

Carey and Sigurdsson (1984) have proposed a model for volcaniclastic sedimentation behind intraoceanic magmatic arcs that experience backarc spreading. Their model is a refinement of the model of Karig and Moore (1975), which in turn, is based on the tectonic model of Karig (1971). This model is applicable to intraoceanic arcs (for example, Marianas and Lesser Antilles), where trench rollback exceeds trenchward motion of the overriding plate (see Dewey, 1980); the magmatic arc is split to form a backarc basin, bounded by the still-active part of the magmatic arc and an inactive remnant arc (hence, this type of backarc basin also is called an "interarc basin"). The model needs modification in order to be applicable to backarc basins formed by the rifting of a continental-margin arc (for example, Japan Sea, where a rifted continental margin replaces the oldest remnant arc) or to backarc basins formed by the trapping of old oceanic crust (for example, Bering Sea, where no backarc spreading has occurred). Royden and others (1982) discussed an example of failed backarc spreading closely associated with continental collision. Also, Klein (1985) discussed modifications in sedimentation patterns in intraoceanic settings, resulting from climatic and oceanographic effects.

In Carey and Sigurdsson's (1984) model, the relative importance of sediment sources are (1) magmatic arc (primarily volcaniclastites, resulting from subaerial and subaqueous eruptions, and erosion of arc complex), (2) backarc spreading center (hyaloclastites and hydrothermally derived sediments), and (3) remnant arc (minor subaqueous gravity-flow deposits). Deposition of magmatic-arc–derived volcaniclastites results in the development of a sedimentary apron on the back side of the arc, with pronounced asymmetry of basin fill. Backarc basins experience four stages of evolution: (1) early rifting, with rapid influx of volcaniclastites; (2) basin widening, with active volcanism and spreading, and development of an asymmetric apron; (3) basin maturity, with waning volcaniclastic input and increased pelagic and hemipelagic deposition; and (4) basin inactivity, with cessation of spreading, and continued pelagic and hemipelagic deposition. A new cycle initiates with splitting of the arc and formation of a new backarc (interarc) basin.

Retroarc Foreland Basins

Compressional arc-trench systems commonly develop foreland basins behind the arcs owing to partial subduction of continental crust beneath the arc orogens (Figs. 5A and 5B). "Foreland basin" is a pre-plate-tectonic term used to describe a basin between an orogenic belt and a craton (Allen and others, 1986). Dickinson (1974) proposed that the term "retroarc" be used to describe foreland basins formed behind compressional arcs, in contrast to peripheral foreland basins formed during continental collisions (see below). Thus, although "backarc" and "retroarc" literally are synonymous, the former is used for extensional and neutral arcs, whereas the latter is used for compressional arcs.

Jordan (1981) presented an analysis of the asymmetric Cretaceous foreland basin associated with the Idaho-Wyoming thrust belt. She used a two-dimensional elastic model to show how thrust loading and sedimentary loading resulted in broad flexure of the lithosphere. The location of maximum flexure migrated eastward as thrusting migrated eastward. The area of subsidence was broadened owing to the erosional and depositional redistribution of part of the thrust load. Comparison of modeled basin and basement geometries with isopach maps provides tests of possible values of flexural rigidity of the lithosphere. The modern sub-Andean thrust belt and foreland basin have similar topography to that proposed for the Cretaceous of the Idaho-Wyoming system. Topography is controlled by thrust-fault geometry and isostatic subsidence.

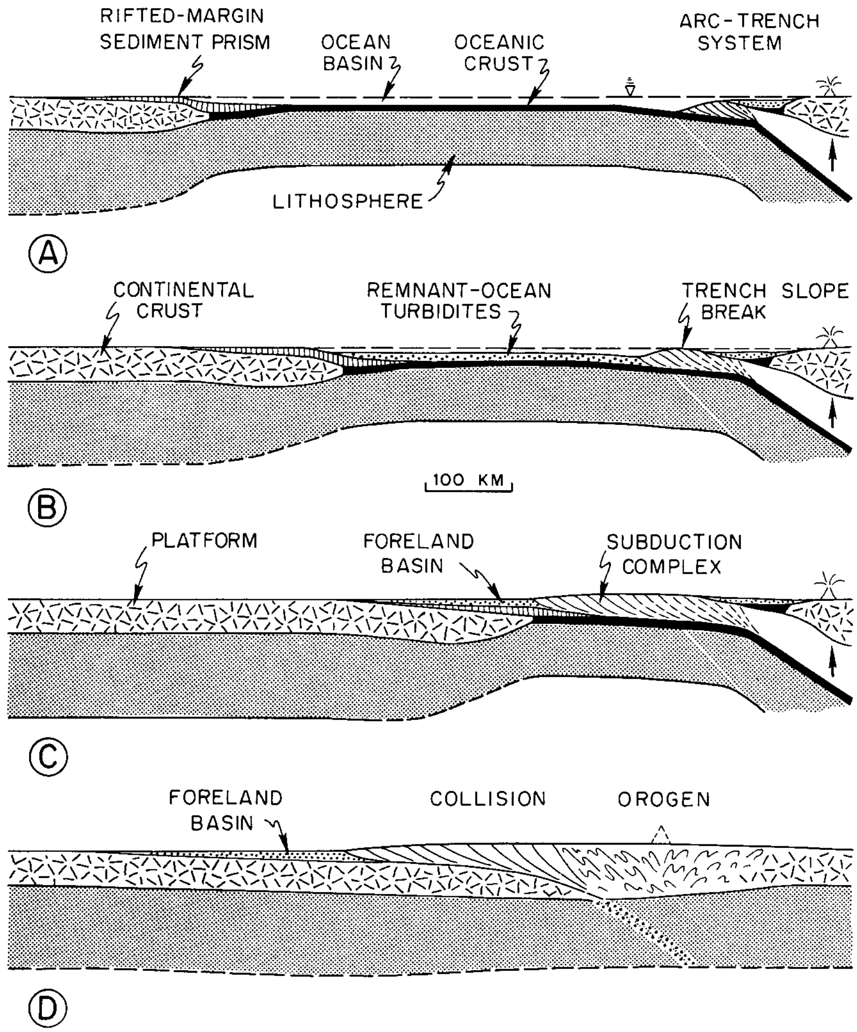

Figure 8. Idealized true-scale diagrams showing inferred evolution (A to D) of sedimentary basins associated with crustal collision to form cryptic intercontinental suture belt within collision orogen. Diagrams represent a sequence of events in time at one place along a developing collision orogen or coeval events at different places along a suture belt marked by diachronous closure. Hence, erosion in one segment (D), where suture has formed, provides sediment which is dispersed longitudinally past a migrating transition point (B to C), to feed subsea fans in a remnant ocean basin (B) along tectonic strike. Also, see Figure 9. Reproduced by permission of American Association of Petroleum Geologists (from Dickinson, 1976).

The model presented by Jordan (1981) is broadly applicable to other retroarc foreland basins. It demonstrates how tectonic activity in the foreland fold-thrust belt is the primary cause of subsidence in associated foreland basins. Sedimentary redistribution, autocyclic sedimentary processes, and eustatic sea-level changes are important modifying factors in terms of regressive-transgressive sequences, but compressional tectonics behind the arc-trench system is the driving force. The Cretaceous seaway of North America was largely the result of this compressional tectonic activity (combined with high eustatic sea level). Details concerning timing of thrusting and initial sedimentary response to thrusting within the Idaho-Wyoming thrust belt are debated (for example, Heller and others, 1986), but the essential role of compressional tectonics in creating retroarc foreland basins is clear.

Remnant Ocean Basins and Suture Belts

Intense deformation occurs in suture belts during the attempted subduction of nonsubductable, buoyant continental or magmatic-arc crust. Suture belts can involve rifted continental margins and continental-margin magmatic arcs (terminal closing of an ocean basin) or various combinations of arcs and continental margins. Figure 8 illustrates stages in the development of suture belts, either in time or in space. Colliding continents tend to be irregular, and great variability of timing, structural deformation, sediment dispersal patterns, and preservability occurs along strike (Dewey and Burke, 1974).

Graham and others (1975) used the Cenozoic development of the Himalayan-Bengal system as an analogue for the late Paleozoic development of the Appalachian-Ouachita system and proposed a general model for sediment dispersal related to sequentially suturing orogenic belts (Fig. 9). "Most sediment shed from orogenic highlands formed by continental collisions pours longitudinally through deltaic complexes into remnant ocean basins as turbidites that are subsequently deformed and incorporated into the orogenic belts as collision sutures lengthen" (Graham and others, 1975, p. 273). This model provides a general explanation for many synorogenic flysch and molasse deposits associated with suture belts (Fig. 8), although many units called "flysch" and "molasse" have different tectonic settings. For this reason, use of these geosynclinal terms is not recommended outside of their type areas in the Alps.

Peripheral Foreland Basins

As continental collision occurs between a rifted continental margin and the subduction zone of an arc-trench system, a tectonic load is placed on the rifted margin, first below sea level and later subaerially (Fig. 8). A peripheral foreland basin forms as the elastic lithosphere flexes under the encroaching dynamic load. The discrimination of ancient retroarc and peripheral forelands is difficult, but it may be possible based on the following characteristics: (1) polarity of magmatic arc, (2) presence of oceanic subduction complex associated with earliest phases of peripheral foreland, (3) asymmetry of suture belt (closer to peripheral foreland), (4) protracted development of retroarc (long-term arc evolution) versus discrete development of peripheral foreland (terminal ocean closure without precursor), and (5) possible volcaniclastic input to retroarc throughout history versus minimal volcaniclastic input to peripheral foreland.

Stockmal and others (1986) have provided a dynamic two-dimensional model for the development of peripheral foreland basins, following finite times of rifting. They modified the model of Speed and Sleep (1982) and demonstrated the effects of rifted-margin age and topography on lithospheric flexure and basin development. The primary effect of age shows up as a higher flexural forebulge and thicker trench fill during the earlier stages for an old (120 m.y.) margin. Subsequent development is relatively insensitive to margin age. Foreland-basin subsidence is strongly sensitive to overthrust load, with depths possibly exceeding 10 km. Crustal thickness may reach 70 km during the compressional phase (for example, Himalayas). Tens of kilometers of uplift and erosion, of both the allochthon and the foreland basin, are predicted during and after deformation. Most of this eroded material is deposited elsewhere owing to uplift within the foreland; longitudinal transport into remnant ocean basins results (Graham and others, 1975). Thick overthrusts with low topographic expression are to be expected where broad, attenuated rifted continental margins have been pulled into subduction zones.

Piggyback Basins

Ori and Friend (1984) defined "piggyback basins" as basins that have formed and been

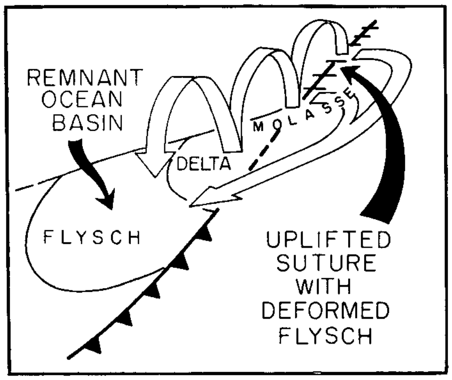

Figure 9. Conceptual diagram to illustrate progressive incorporation of synorogenic flysch within an orogenic suture belt by sequential closure of remnant ocean basin. Figure 8B illustrates the "flysch phase," Figure 8C illustrates the "delta phase" (transition point), and Figure 8D illustrates the "molasse phase." From Graham and others (1975).

filled while being carried on top of moving thrust sheets. They discussed examples from the Apennine and Pyrenean fold-thrust systems. Piggyback basins are dynamic settings for sediment accumulation; most sediment, if not all, is derived from associated fold-thrust belts. The fold-thrust belts can be in peripheral, retroarc, or transpressional settings. Piggyback basins share characteristics with foreland basins and trench-slope basins. They have low preservation potential owing to their formation on top of growing thrust belts; therefore, they are generally found only in young orogenic systems (for example, Burbank and Tahirkheli, 1985).

Foreland Intermontane Basins

Low-angle subduction beneath compressional arc-trench systems may result in basement-involved deformation within retroarc foreland basins (Dickinson and Snyder, 1978). The Rocky Mountain region of the western United States is the best-known example of this style of deformation, although similar modern provinces have been documented in the Andean foreland (Jordan and others, 1983). Overthrusting and wrench deformation, similar to processes related to intracontinental wrench basins, are likely.

Chapin and Cather (1981) discussed controls on Eocene sedimentation and basin formation of the Colorado Plateau and Rocky Mountain area. Types of associated uplifts include (1) Cordilleran thrust-belt uplifts, (2) basement-cored Rocky Mountain uplifts, and (3) monoclinal uplifts of the Colorado Plateau. Resulting basins can be classified into three types: (1) Green River type (large, equidimensional to elliptical, bounded on three or more sides by uplifts, and commonly containing lakes), (2) Denver type (elongate, open, asymmetric synclinal downwarps with uplift on one side), and (3) Echo Park type (narrow, highly elongate, fault-bounded, with through drainage, and strike-slip origin). Composite uplifts and basins also formed. Green River–type basins have quasi-concentric facies zonation, in contrast to wedge-shaped, unidirectional facies distribution in Denver-type basins. Echo Park–type basins have complex facies, with common sedimentary breccias and sheetwash deposits, associated with active faulting, erosion, and stream diversion typical of strike-slip basins. Chapin and Cather (1981) also discussed geomorphic and climatic effects on basin evolution and used the occurrence and timing of different basin types to constrain interpretations of the Laramide orogeny. They proposed a two-stage model for basin formation, which can be related to changes in North American plate interactions both in the Atlantic and the Pacific oceanic basins. Studies in similar modern settings (for example, Andean foothills; Jordan and Allmendinger, 1986) should improve these models.

TRANSFORM SETTINGS

Strike-Slip Systems

The complexity and variety of sedimentary basins associated with strike-slip faults are almost as great as for all other types of basins. Transform faults in oceanic lithosphere generally behave according to the plate-tectonic model, whereas strike-slip faults in continental lithosphere are extremely complex and difficult to fit into a model involving rigid plates. Simple mechanical models based on homogeneous media have little application to the heterogeneous media of continental crust.

The Reading cycle (for example, Reading, 1980) predicts that any strike-slip fault within continental crust is likely to experience alternating periods of extension and compression as slip directions adjust along major crustal faults. Thus, opening and closing of basins along strike-slip faults (Reading cycle) is analogous, at smaller scale, to the opening and closing of ocean basins (Wilson cycle).

Basins related to strike-slip faults can be classified into end-member types, although most basins are hybrids. Transtensional (pull-apart) basins form near releasing bends, and transpressional basins form at constraining bends (Crowell, 1974b). Basins associated with crustal rotations about vertical axes within the rotating

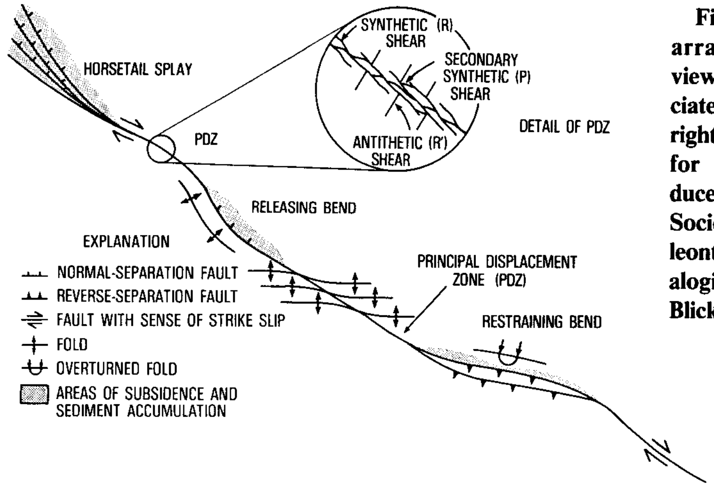

Figure 10. The spatial arrangement, in map view, of structures associated with an idealized right-slip fault. See text for discussion. Reproduced by permission of Society of Economic Paleontologists and Mineralogists (from Christie-Blick and Biddle, 1985).

blocks (herein termed "transrotational") may experience any combination of extension, compression, and strike slip.

Christie-Blick and Biddle (1985) have provided the most comprehensive summary of the structural and stratigraphic development of strike-slip basins, based, in large part, on the pioneering work of Crowell (1974a, 1974b). They illustrated the structural complexity likely along strike-slip faults (Fig. 10) and the implications for associated basins. Primary controls on structural patterns are (1) degree of convergence and divergence of adjacent blocks, (2) magnitude of displacement, (3) material properties of deformed rocks, and (4) pre-existing structures (Christie-Blick and Biddle, 1985, p. 1). Subsidence in sedimentary basins results from crustal attenuation, thermal subsidence during and following extension, flexural loading due to compression, and sedimentary loading. Thermal subsidence is less important than in elongate orthogonal rifts due to lateral heat conduction in narrow pull-apart basins. Distinctive aspects of sedimentary basins associated with strike-slip faults include (Christie-Blick and Biddle, 1985, p. 1) (1) mismatches across basin margins; (2) longitudinal and lateral basin asymmetry; (3) episodic rapid subsidence; (4) abrupt lateral facies changes and local unconformities; and (5) marked contrasts in stratigraphy, facies geometry, and unconformities among different basins in the same region.

Transtensional Basins

Pull-apart basins form at left-stepping sinistral fault junctures and at right-stepping dextral fault junctures. Mann and others (1983) proposed a model for such basins based on a comparative study of well-studied pull-apart basins at various stages of development. Pull-apart basins evolve through the following stages: (1) nucleation of extensional faulting at releasing bends of master faults; (2) formation of spindle-shaped basins defined and commonly bisected by oblique-slip faults; (3) further extension, producing "lazy-S" or "lazy-Z" basins; (4) development into rhombochasms, commonly with two or more subcircular deeps; and (5) continued extension, resulting in the formation of oceanic crust at short spreading centers offset by long transforms. Basaltic volcanism and intrusion may become important in the transition from stages 3 through 5 (for example, Crowell, 1974b). Most pull-apart basins have low length-to-width ratios, owing to their short histories in changing strike-slip regimes (Mann and others, 1983).

Transpressional Basins

Transpressional basins include two types: (1) severely deformed and overthrust margins along sharp restraining bends that result in flexural subsidence due to tectonic load (for example, south side of San Gabriel Mountains, southern California) and (2) fault-wedge basins at gentle restraining bends that result in uplift of one or two margins and downdropping of a basin as one block moves past the restraining bend (for example, Ridge basin, southern California) (Crowell, 1974b). A basin model for type 1 would involve flexural loading similar to that of the foreland models discussed above, although at smaller scale.

Ridge basin is one of the most elegantly exposed and carefully studied transpressional basins in the world, as summarized by Crowell and Link (1982). Crowell (1982) presented a dynamic model for the evolution of Ridge basin (12–5 Ma), a narrow crustal sliver caught between the San Gabriel fault to the southwest and a set of northwest-trending faults that became active sequentially in a northeast direction (Fig. 11). Ridge basin became inactive when motion was transferred completely to the modern San Andreas fault. As a result of movement on the San Gabriel fault, the southwest side of the basin was uplifted and the Violin Breccia was deposited along the basin margin. The depressed floor of the basin moved past this uplifted margin, as

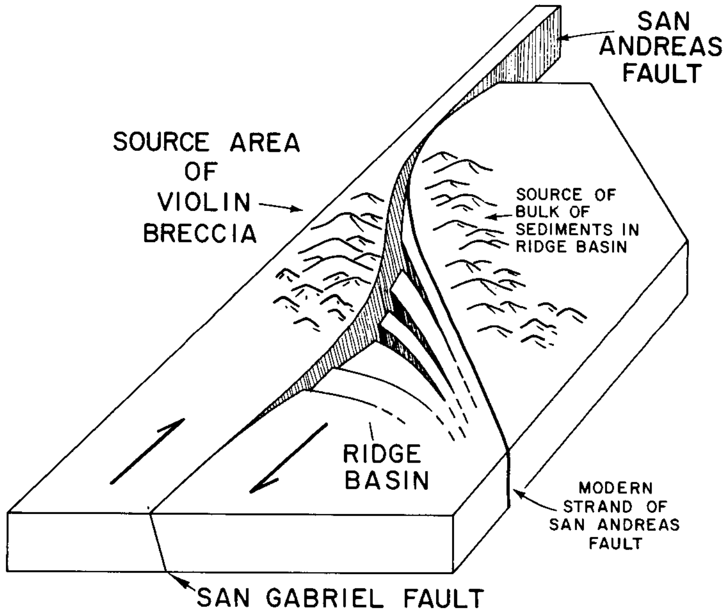

Figure 11. Block diagram (looking toward the northwest) illustrating origin of Ridge basin at a sigmoidal bend in the San Andreas fault. See text for discussion. Reproduced by permission of the Pacific Section, Society of Economic Paleontologists and Mineralogists (from Crowell, 1982).

TABLE 2. SUMMARY OF KEY REFERENCES AND PUBLISHERS

Topic	Key reference	Publisher
Overview and basin classification	Dickinson (1974, 1976)	SEPM and AAPG
Sequential rift development and continental separation	Kinsman (1975)	Princeton U. Press
Terrestrial rift valleys	Leeder and Gawthorpe (1987)	Geol. Soc. London
Proto-oceanic rift troughs	Cochran (1983a)	AAPG
Continental rises and terraces	Pitman (1978)	GSA
Continental embankments	Burke (1972)	AAPG
Failed rifts and aulacogens	Hoffman and others (1974)	SEPM
Intracratonic basins	Derito and others (1983)	Tectonophysics
Oceanic basins	Heezen and others (1973)	Nature
Oceanic islands, aseismic ridges, and plateaus	Clague (1981)	SEPM
Arc-trench systems	Dewey (1980)	Geol. Assoc. Canada
Trenches	Thornburg and Kulm (1987)	GSA
Trench-slope basins	Moore and Karig (1976)	GSA
Forearc basins	Dickinson and Seely (1979)	AAPG
Intra-arc basins	Busby-Spera (1984a, 1984b)	AGU and Pac. Sec. SEPM
Interarc and backarc basins	Carey and Sigurdsson (1984)	Geol. Soc. London
Retroarc foreland basins	Jordan (1981)	AAPG
Remnant ocean basins and suture belts	Graham and others (1975)	GSA
Peripheral foreland basins	Stockmal and others (1986)	AAPG
Piggyback basins	Ori and Friend (1984)	GSA
Foreland intermontane basins	Chapin and Cather (1981)	Arizona Geol. Soc.
Strike-slip systems	Christie-Blick and Biddle (1985)	SEPM
Transtensional basins	Mann and others (1983)	Journal of Geology
Transpressional basins	Crowell (1982)	Pac. Sec. SEPM
Transrotational basins	Luyendyk and Hornafius (1987)	Prentice-Hall
Intracontinental wrench basins	Sengor and others (1978)	Am. Jour. of Science
Successor basins	Eisbacher (1974)	SEPM

it received abundant sediment from the northeast. Previous depocenters moved southeastward past the restraining bend, after receiving sediment in conveyor-belt fashion, with uplift and tilting following deposition. The result is a stratigraphic thickness of more than 11 km in outcrop, although vertical thickness of the basin fill is approximately one-third of this. Many extraordinarily thick coarse clastic units in ancient, narrow fault-bounded basins likely were deposited in similar settings.

Transrotational Basins

Paleomagnetic data from southern California have documented extensive clockwise rotation of several crustal blocks, beginning in the Miocene and continuing today (Luyendyk and others, 1980; Hornafius and others, 1986). Luyendyk and Hornafius (1987) further developed their geometric model in order to make testable predictions concerning amount and direction of slip on faults bounding rotated and nonrotated blocks and concerning areas of gaps (basins) and overlap (overthrusts) among blocks. Within this setting, all of the complexity of transtensional and transpressional basins is likely. The unique aspect of Luyendyk and Hornafius' (1987) geometric model is that it successfully predicts the positions, shapes, areas, and ages of most southern California Neogene basins. Similar geometric models may be possible along other complex transform boundaries involving significant crustal rotations.

HYBRID SETTINGS

Intracontinental Wrench Basins

The collision of continents of varying shapes and sizes can lead to bewildering complexity in ancient orogenic belts (for example, Dewey and Burke, 1974; Graham and others, 1975; Molnar and Tapponnier, 1975; Sengor, 1976; Tapponnier and others, 1982). As Tapponnier and others (1982) demonstrated through the use of plasticine models, the collision of India and Asia has resulted in major intracontinental strike-slip faults, with associated transtensional, transpressional, and transrotational basins, including the formation of impactogens.

Sengor and others (1978) developed criteria for distinguishing failed rifts formed during the opening of oceans that are later closed (aulacogens) from intracontinental rifts formed owing to crustal collisions (impactogens). Both types of rift valleys trend at high angles to orogenic belts; however, aulacogens have a rifting history coincident with formation of a new ocean basin prior to collision, whereas impactogens have no pre-collisional history. A third category of rifts at high angles to orogenic belts is random rifts unrelated to either ocean formation or collision orogenesis. All of these rift basins are deformed during suturing in associated orogens. Tests for distinguishing them must come from the stratigraphic record because temporal correlation of initial rifting (or lack thereof) is the primary test for the geodynamic origin for these ancient rifts located at high angles to orogenic belts. One suggestive type of evidence for discrimination is that aulacogens tend to form at re-entrants along rifted continental margins (Dewey and Burke, 1974), whereas impactogens are more likely to form opposite coastal promontories, where deformation of colliding continents is more intense (Sengor, 1976). This criterion must be applied cautiously, however, due to the difficulty of definitively reconstructing pre-collision geometry (for example, Thomas, 1983, 1985).

Successor Basins

The original definition of successor basins (King, 1966) as "deeply subsiding troughs with limited volcanism associated with rather narrow uplifts, and overlying deformed and intruded eugeosynclines" (Kay, 1951, p. 107; Eisbacher, 1974) needs modification; "deeply subsiding" and "eugeosynclines" should be replaced by "intermontane" and "terranes," respectively. Within the context of plate tectonics, successor basins form primarily in intermontane settings on top of inactive fold-thrust belts, suture belts, transform belts, and noncratonal failed rifts. The presence of successor basins indicates the end of orogenic activity, and so their ages constrain interpretations of timing of suturing and deformation. Thus, they have special significance in "terrane analysis"; they represent overlap assemblages which provide minimum ages for terrane accretion (Howell and others, 1985).

Little work has been published on actualistic models for such basins; Eisbacher (1974) summarized models based on work in the Canadian Cordillera. This dearth of work may reflect the diversity of successor basins and their tectonic settings. In a sense, all basins are successor basins because they form following some orogenic event represented in the basement of the basin. In fact, one of Kay's (1951) examples of epieugeosynclines (successor basins) is the post-Nevadan basin of central California, which is now interpreted as a forearc basin, overprinted in the Tertiary by transform tectonics (Ingersoll, 1982; Ingersoll and Schweickert, 1986; Graham, 1987). Modern use of the term "successor basin" should be restricted to post-orogenic basins that do not fall into any other plate-tectonic framework during their development. Additional work on modern successor basins is needed before an actualistic model is available.

DISCUSSION

This review discusses actualistic models for basins in all plate-tectonic settings. It should be obvious that some of the basin types are common and volumetrically important, whereas others are rare and volumetrically minor. An important, but seldom discussed, factor in basin analysis and paleotectonic reconstruction is the preservability of tectonostratigraphic assemblages. Several basin types are rarely preserved in the very ancient record (for example, intra-arc, trench-slope, and successor basins); their absence in ancient orogenic belts is predicted by their locations and susceptibilities to erosion and deformation. Thus, their absence is not a valid test of plate-tectonic models. Veizer and Jansen (1979, 1985) have provided an empirical method of determining the "half life" of tectonostratigraphic elements. They estimated the half lives of active-margin basins at 30 m.y., oceanic sediments at 40 m.y., oceanic crust at 55 m.y., passive margins at 80 m.y., immature orogenic belts at 100 m.y., and mature orogenic belts and platforms at 380 m.y. The application of Veizer and Jansen's type of analysis to all of the basin types discussed herein will provide additional quantitative constraints on paleotectonic reconstructions.

Table 2 lists the key references and where they were published. Not surprisingly, more than half of these references were published by three societies with strong emphases on basin analysis and tectonics: American Association of Petroleum Geologists, Society of Economic Paleontologists and Mineralogists, and Geological Society of America. Table 2, however, also demonstrates the diversity of information sources and the interdisciplinary nature of the field; staying current in the field of sedimentary tectonics is a challenge.

It is hoped that this review will focus attention on important references, as well as suggest where work is needed most. Certain tectonic settings (for example, continental rises, continental terraces, and foreland basins) have received much attention in the last decade, so that actualistic models are becoming quantitatively sophisticated. In contrast, other settings (for example, successor basins, intra-arc basins, and continental embankments) have received relatively little attention. As all of these models improve, it is likely that more integration of successive orogenic processes into a continuum of models will provide strong predictive capabilities for paleotectonic reconstructions.

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