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Himalayan-Bengal Model for Flysch Dispersal in the Appalachian-Ouachita System

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ABSTRACT

The relation of the modern Bengal subsea fan to the Cenozoic Himalayan suture belt and the analogous relation of the Carboniferous Ouachita flysch to a presumed Paleozoic Appalachian suture belt suggest a guiding principle of synorogenic sedimentation. 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.

India first encountered a southern Eurasian subduction zone near the end of Paleocene time. Northward movement of India since Oligocene time choked the subduction zone, stifled the associated magmatic arc, and created a suture complex of deformed Cretaceous flysch and younger Tertiary molasse. Strata derived from the resulting orogen include continental clastic wedges shed southward toward India and voluminous turbidites fed longitudinally through the Ganges-Brahmaputra Delta into the Bay of Bengal. The eastern flank of the Bengal subsea fan is being subducted now beneath the still-active eastern extension of the subduction zone.

The sequential, north-to-south welding of Europe and Africa to North America formed the complex Appalachian-Caledonide-Mauritanide suture belt, from which Taconic, Acadian, and Alleghanian clastic wedges were shed toward the North American craton. Turbidites of the Carboniferous Ouachita flysch were fed longitudinally, as sediment supplied through the Alleghanian clastic wedge, into a remnant ocean basin lying south of North America. The Ouachita system was then thrust northward across the continental edge during arc-continent collision that progressed from east to west. *Key words: historical geology, areal geology, tectonics, sedimentation, flysch, Cenozoic, Carboniferous, Himalayas, Bengal fan, Appalachians, Ouachitas.*

INTRODUCTION

The tectonic relations of the late Paleozoic Ouachita orogenic belt of the south-central United States are uncertain, both to the Appalachian belt along regional strike to the east and to the continental margin across regional strike to the south. Plate tectonic interpretations of the Appalachian and Cordilleran systems near the edges of the continent and of the Caribbean region south of it have not dealt with the Ouachita question. Recent interpretations of Ouachita tectonics rely upon a supposed analogy with active continental margins, marked by either a marginal magmatic arc of Andean type (Keller and Cebull, 1973) or a fringing island arc of Japanese type facing away from the continent (Morris, 1974a, 1974b). In our view, however, the general lack of appropriate volcano-plutonic complexes or their derivative sedimentary sequences within the Ouachita region encourages the consideration of alternative analogies.

We suggest a different interpretation by drawing a tectonic analogy between the Cenozoic Himalayas and the Bengal fan, on the one hand, and the Carboniferous Appalachians and the Ouachita flysch on the other. The Himalayas were formed by the collision of India with Eurasia, and the final development of the southern Appalachian system was marked by the collision of Africa with North America (Watkins, 1972). The immense Bengal fan of turbidites and associated oceanic strata is composed mainly of detritus fed into the remaining ocean basin east of India by the longitudinal drainage of the Himalayas and related highlands. We suggest that the Carboniferous clastic rocks, largely turbidites, of the Ouachitas are composed mainly of detritus fed into a remnant ocean basin south of North America by the longitudinal drainage of highlands formed along the trend of an Appalachian suture between Africa and North America. The flank of the Bengal fan adjacent to southeast Asia has been partly subducted by an arc-trench system that is still actively consuming oceanic lithosphere off Sumatra. We suggest that the Ouachita orogeny, during which the Ouachita system was thrust northward over the southern flank of the continent, was caused by the close approach of an analogous arc-trench system, whose remains lie buried now in the subsurface beneath Texas. Figure 1 shows the gross tectonic trends of the two regions at the same scale.

With available data, our interpretation is speculative but is compatible with current knowledge of the Ouachita and southern Appalachian regions. If valid, our view has implications concerning the tectonic framework of the Caribbean region just prior to the Mesozoic development of the Gulf of Mexico and the Antilles. From a broader standpoint, the concept that voluminous turbidites may be fed longitudinally into narrowing ocean basins that close sequentially between colliding continents may offer a general explanation for the common occurrence of synorogenic flysch derived from the same orogenic belt into which it is incorporated by later deformation (see also Dewey and Burke, 1974).

We first describe the Himalayan-Bengal events, discuss the Appalachian-Ouachita events by analogy, and then comment on the general implications of our interpretation. For each of the two regions, we summarize in order the general relations of (a) *oceanic closure* by plate consumption beneath continental margin arcs, (b) development of a *suture belt* along the orogen between colliding continents, (c) transverse dispersal of *clastic wedges* shed from the collision orogen across an adjacent continental block, (d) longitudinal dispersal of turbidites into a *remnant ocean basin* along tectonic strike from the collision orogen, and (e) deformation of the turbidite flysch by *continued subduction* along one flank of the remnant ocean. Our arguments are based upon the implications of the tectonic setting of the modern Bengal fan, as described by Curray and Moore (1971). However, we make no attempt to evaluate details they gathered subsequently on the

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evolution of the Bay of Bengal and adjacent parts of the Indian Ocean floor (Moore and others, 1975; Curray and Moore, 1974), although our thinking has benefited from the general thrust of their oral presentations (see below). Powell and Conaghan (1973) also presented a detailed analysis of the structural evolution of the Himalayan ranges, to which the reader is referred for additional structural and stratigraphic information. Our evaluation of published information pertaining to the Ouachita region is guided especially by the interpretations of Morris (1974a, 1974b), which we supplemented by reconnaissance visits only.

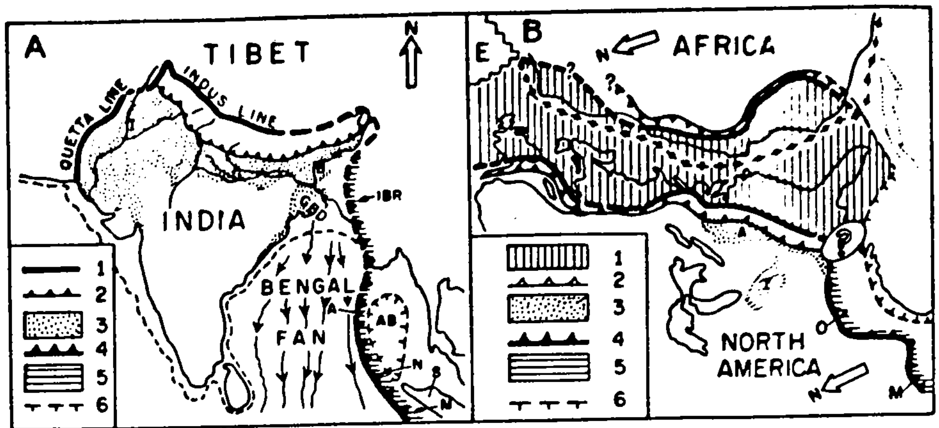

Figure 1. Main tectonic elements of Himalayan-Bengal (A) and Appalachian-Ouachita (B) regions. Symbols:
A. 1, Himalayan suture belt between India and Eurasia; 2, Main Boundary Fault (thrust) of Himalayan foothills; 3, Quaternary alluvium of Indus (I), Ganges (G), and Brahmaputra (B) river systems; arrows denote channel trends on subsea Bengal fan with head near Ganges-Brahmaputra Delta (GBD); 4, Indoburman-Sunda subduction zone, dashed where inactive; 5, mélanges and deformed flyschoid rocks of Indoburman ranges (IBR), Andaman (A)-Nicobar (N) insular ridge, and Mentawai (M) Islands off Sumatra (S); 6, schematic margin of extensional Andaman basin (AB); dashed line is 1,000-m isobath of continental slope off India.
B. 1, Complex belt of multiple Paleozoic suturing between North American and West African cratons shown joined (E is Europe) prior to Mesozoic opening of the Atlantic Ocean approximately along the stitched line; 2, fronts of foreland fold-thrust belts of Appalachian Valley and Ridge province (American side) and Mauritanides (African side); 3, coarse Carboniferous clastics, terrestrial and littoral, of Appalachian (A) and Illinois (I) basins; 4, frontal thrusts and folds of Ouachita system; 5, outcrop and known subcrop of Ouachita system; 6, schematic extensional margin of Mesozoic(?) Gulf of Mexico (present position of Cuba shown in dotted outline).

HIMALAYAN-BENGAL REGION

Indian Ocean Floor

The Indian Ocean basin has formed since the end of the Paleozoic Era, and mainly since the beginning of the Cretaceous, from the breakup of Gondwana and the dispersal of the continental fragments derived from it. Their initial configuration within intact Gondwana (Fig. 2A) has been inferred within close limits from the computer fitting of continental margins (Sproll and Dietz, 1969; Smith and Hallam, 1970; Dietz and Sproll, 1970). Paleomagnetic data compiled by McElhinny (1970) are compatible with the assembly shown, although Tarling (1972) used similar data to support a slightly different reconstruction. A recent analysis of marine magnetic anomalies in the Indian Ocean indicates that a reconstruction close to the assembly shown is probably correct (Sclater and Fisher, 1974). There is thus a prominent gap of about 1,000 km inferred between peninsular India and Australia on the original margin of Gondwana. In an oral discussion of the evolution of the Indian Ocean at the 1972 Conference on Geosynclinal Sedimentation at the University of Wisconsin, J. C. Curray and D. G. Moore suggested that the apparent gap in Gondwana's margin may have been filled by continental crust representing an extension of India beyond the present Himalayan border of peninsular India. McElhinny and Embleton (1974) recently made the same suggestion, reinforcing the paleomagnetic arguments in favor of the Smith-Hallam reconstruction shown in Figure 2A. Accordingly, we have depicted by dashed lines in Figure 2 an assumed extension of peninsular India that filled all or part of the apparent gap in the initial margin of Gondwana. A combination of subduction beneath Tibet and multiple thrusting in the Himalayas has obscured an indeterminate portion of the original continental margin of India (Crawford, 1971). Gansser (1966) estimated 500 km of crustal shortening in the Himalayas. The suggestion of Curray and Moore is also compatible with the information from JOIDES Legs 22, 26, and 27 that the sea floor of the Wharton Basin west of Australia is nowhere significantly older than Cretaceous (Scientific staff, 1972, 1973a, 1973b). Recently identified magnetic anomalies in the Wharton Basin suggest that western Australia initially separated from some appendage of the Indian continental block in Early Cretaceous time (Markl, 1974).

The evolution of the Indian Ocean basin during the past 75 m.y. has been outlined by McKenzie and Sclater (1971). By 75 m.y. B.P. (Fig. 2B), India, Africa, and Australia-Antarctica were already three discrete continental blocks separated by regions of oceanic crust through which various microcontinents were scattered. From 75 to 55 m.y. B.P. (Figs. 2B, 2C), India moved rapidly northward during an episode of fast spreading from an ancestral rise in the central Indian Ocean. At 55 m.y. B.P., near the end of the Paleocene Epoch, spreading slowed markedly or stopped. McKenzie and Sclater (1971) suggested that this behavior was probably a signal of the initial encounter between India and Eurasia. Presumably, the encounter involved peninsular India's appendage, now obscured by partial subduction and deformation beneath and within the Himalayan orogenic belt. Powell and Conaghan (1973) called special attention to previously unrecognized metamorphism of the edge of the Indian subcontinent as an additional factor that obscures detailed relations. It is unclear whether the crustal segment that first met Eurasia was firmly attached to India at the time or was a detached microcontinental fragment now caught up within the Himalayan orogenic belt.

The initial separation of Australia and Antarctica occurred about 45 m.y. B.P. (see Fig. 2C), and spreading resumed south of India — but in a new pattern — about 35 m.y. B.P. in Oligocene time. Since then, presumably, the driving of India against and beneath the Himalayan suture (Fig. 2D) has (1) choked the subduction zone at the southern margin of Eurasia where the once-intervening oceanic lithosphere was consumed, (2) stifled the associated magmatic arc along the edge of Eurasia in the region of continental collision, and (3) deformed the northern margin of India during partial subduction beneath Tibet and uplift in the Himalayas. Figure 3 depicts sequentially the geologic events related to the development of the suture belt as described below. Our discussion here is highly generalized; the reader is referred to Curray and Moore (1974) for interpretations that incorporate the most recent oceanographic information.

Himalayan Suture Belt

The crustal suture formed by continental collision between India and Eurasia lies along a complex tectonic zone, the Indus Line, marked by an elongate, semicontinu-

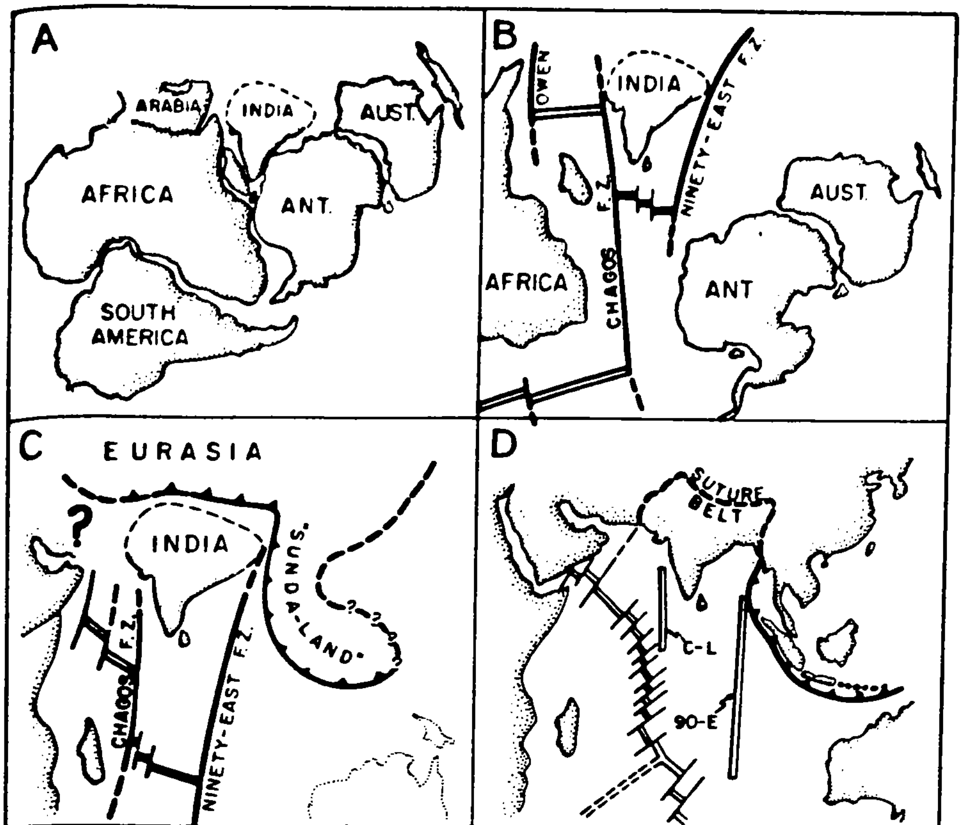

Figure 2. Diagrammatic evolution of the Indian Ocean (India shown as peninsular India with about 500-km width added arbitrarily to Himalayan border; see text). A. Smith-Hallam reconstruction of Gondwana prior to breakup. B. McKenzie-Sclater reconstruction of Indian Ocean region at 75 m.y. B.P. C. Early Cenozoic relations in Indian Ocean region. Rise crests and transforms shown were active mainly prior to 55 m.y. B.P.; Australia separated from Antarctica about 45 m.y. B.P. and reached position shown about 35 m.y. B.P.; latitudes of India and Australia shown for 35 m.y. B.P. from McKenzie and Sclater (1971); latitude of Eurasia for early Cenozoic estimated from paleomagnetic data (Irving, 1964); relative longitudes of India and Eurasia approximately as depicted by Le Pichon (1968); barbed line indicates subduction zone of arc-trench system along southern margin of Eurasia. D. Late Cenozoic relations in Indian Ocean region (present geography). Rise crests and transforms shown have been active since about 35 m.y. B.P.; open bars denote Chagos-Laccadive (C-L) and Ninety-East (90-E) Ridges along old fracture zone trends; barbed line denotes Indoburman-Sunda subduction zone trending into suture belt between India and Eurasia.

ous outcrop belt of deformed Upper Cretaceous Indus flysch and associated ophiolitic rocks (Gansser, 1966; Dewey and Bird, 1970). Where best known in the west-central Himalayas, the suture belt also contains younger molasse (Tewari, 1964). Within the Himalayas south of the Indus Line but still north of the Main Boundary Fault (Fig. 1), the Main Central Thrust forms the tectonic boundary between Paleozoic and Mesozoic strata with Gondwanic affinities to the south and Tethyan affinities to the north (Gansser, 1966). Although both of these sequences rest on broadly similar crystalline basements, the stratal contrast across the Main Central Thrust indicates severe structural telescoping and may even reflect the juxtaposition of two microcontinents, both once external to Eurasia. Powell and Conaghan (1973), who recently discussed the details of the structural evolution of the Himalayas in a time frame close to the one we outline here, regard the rocks between the Main Central Thrust and the suture belt along the Indus Line as a crustal sliver representing the distal northern edge of India itself, rather than as a separate entity.

The Indus flysch is composed of shale and sandstone of mainly Late Cretaceous age, and it is tectonically intermixed with (a) ultramafic rocks of the ophiolite suite; (b) sequences of red limestone, radiolarian chert, and slate — probably representing oceanic pelagites; and (c) exotic blocks of Permian to Jurassic rocks. We suppose this assemblage to be mélange. The southern boundary of the mélange belt is a south-directed thrust, and a north-directed counterthrust typically separates the Indus flysch from younger molasse lying to the north (Tewari, 1964). We interpret the mélange belt containing the Indus flysch and its lithologic associates as a subduction complex composed of materials accreted tectonically to the southern flank of Eurasia as the oceanic lithosphere that once lay between India and Eurasia was consumed by a trench along the southern margin of Eurasia. In this view, the thrust contact along the southern margin of the Indus flysch was formed at the time that crustal elements of India were lodged against the subduction zone and driven partly beneath Tibet. The counterthrust was formed when parts of the subduction mélange were pressed upward and outward from the suture belt during severe contractions that accompanied crustal collision (see also Dewey and Burke, 1973).

The molasse, at one time undifferentiated from the marine flysch, contains coal, marl, freshwater mollusks, plant fossils, and abundant current-bedding indicative of nonmarine deposition (Tewari, 1964). In Ladakh, where the molasse rests unconformably on the Ladakh granite, Tewari (1964) concluded that deposition was post-Eocene, and possibly post–middle Miocene. Near Kailas, a conglomerate of unknown age rests unconformably on the Kailas granite (Gansser, 1964) and is overlain along a north-directed thrust by the Indus flysch in a tectonic setting analogous to that of the sandstone and shale in the molasse of Ladakh. The younger molasse is nearly flat-lying and undeformed except near the thrust contact with the highly deformed Indus flysch. Similar nonmarine beds occurring in the Thakkhola region of Nepal have been interpreted as Miocene-Pliocene intramontane deposits (Bordet and others, 1967). We interpret the molasse of the suture belt as terrestrial deposits formed in local depressions when the initial stages of continental collision had closed the ocean between Eurasia and India. This occurred, however, before continued deformation had raised the Himalayan ranges as broad, unbroken highlands and masked the line of juncture between the two continental margins.

Mesozoic and Tertiary granitic plutons are abundant in the trans-Himalayan ranges north of the suture belt. Radiometric dates by the Rb/Sr method on 22 samples from the Hindu Kush in the western Himalayan region cluster in the following six approximate age ranges (Desio and others, 1964): 210 to 215, ~185, 130 to 135, 85 to 95, ~30, and 15 to 20 m.y. B.P. Farther east in the Karakorum Range, similar dates include two of about 50 m.y. B.P. for small isolated plutons and six from the main axial batholith ranging from nearly 25 m.y B.P. to less than 10 m.y. B.P., below which results with the Rb/Sr technique used are inconclusive. The Ladakh and Kailas granites lie south of the Miocene Karakorum axial batholith and were not dated by Desio and others (1964), but the Deosai pluton dated near 50 m.y. B.P. lies along the same structural trend. The

Ladakh granite has since been dated by the same method as 45 m.y. B.P. (Powell and Conaghan, 1973, p. 5). We interpret the granitic plutons of the Karakorum and related ranges as the exposed roots of the magmatic arc that lay along the southern margin of Eurasia while the nearby subduction zone was active.

The foregoing data are compatible with the following inferences: (1) During the rapid northward drift of India from 75 to 55 m.y. B.P., the flysch and associated rocks of the mélange belt were deformed in the subduction zone of an arc-trench system that had been active along the southern margin of Eurasia through much of Mesozoic time. (2) Following the supposed initial encounter of India with Eurasia, subduction stalled as spreading in the Indian Ocean slowed from 55 to 35 m.y. B.P. (3) With renewed spreading in the Indian Ocean after 35 m.y. B.P., further subduction is reflected by major plutons such as the Karakorum axial batholith, but the forced juxtaposition of continental margins later stifled subduction and fostered the crustal deformation that was recorded by overlapping of tectonic elements along thrusts and accompanied by uplift of the suture belt to its impressive late Cenozoic elevations. As roots of the magmatic arc, such as the Ladakh and Kailas granites, were exposed by erosion, terrestrial molasse was deposited in vestigial depressions along the suture belt and subsequently deformed.

The suture belt and related tectonic elements cannot be traced in detail on available geologic maps into the eastern Himalayas, but a structurally complex belt including exposures of ophiolitic rocks, granitic plutons, Mesozoic flysch, and Tertiary molasse is reported north of Mount Everest as far as Lhasa (Gansser, 1964). The suture presumably wraps around the eastern Himalayan syntaxis and trends into the Indoburman Ranges (Gansser, 1966; Krishnan, 1968). Similarly, the suture apparently wraps around the western Himalayan syntaxis and is expressed as the Quetta Line, an ophiolite-bearing zone of intense deformation separating the markedly contrasting facies belts of Sind and Baluchistan (Gansser, 1966; Krishnan, 1968).

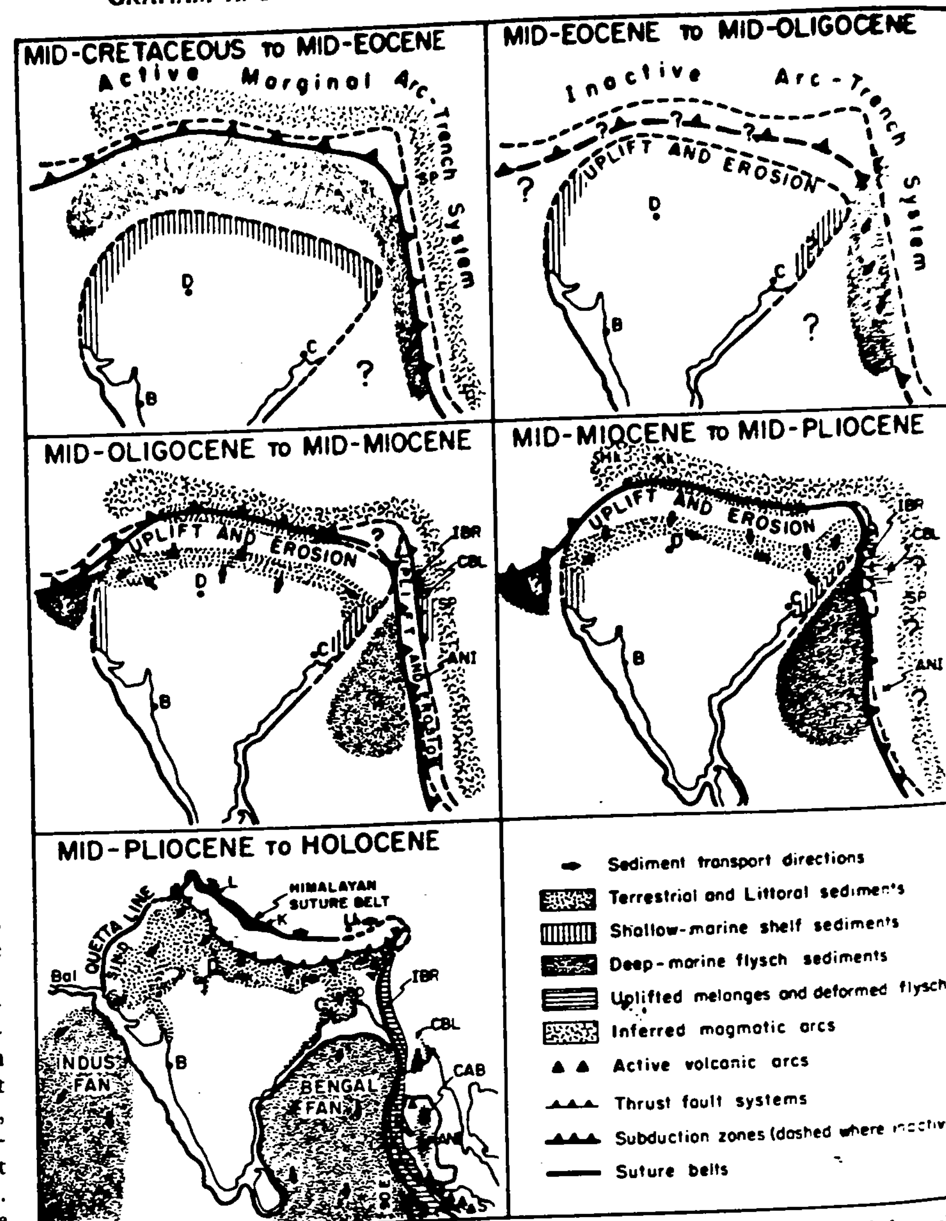

Figure 3. Sketches showing the inferred evolution of the Himalayan suture belt and the region along strike. Places shown include: ANI, Andaman-Nicobar islands; B, Bombay; Bal, Baluchistan; D, Delhi; BD, Bangladesh; C, Calcutta; CAB, central Andaman basin; CBL, central Burmese lowland; D, Delhi; HK, Hindu Kush; IBR, Indoburman ranges; K, Kailas; Kk, Karakorum; L, Ladakh; Lh, Lhasa; 90E, Ninety-East Ridge; S, Sumatra; SP, Shan Plateau. Line off coast of peninsular India is present 1,000-m isobath. Central Andaman basin began to open in middle Miocene time (see text), but early phases of opening are not shown here.

Himalayan Clastic Wedges

The Cretaceous to Holocene strata south of the Himalayas provide a sedimentary record of events along a continental margin involved in crustal collision. Cretaceous through lower Eocene strata in the sub-Himalaya or foothill belt are shallow-water marine sequences of shale, sandstone, and limestone (Gansser, 1964). They apparently represent shelf sedimentation near the northern edge of the Indian subcontinent prior to collision. Faunal and lithologic evidence suggests that an unrestricted seaway existed from Afghanistan to Burma in Cretaceous time, separating the Indian subcontinent from Eurasia; however, convergence apparently reduced considerably the width of the seaway during Paleocene time (Sahni and Kumar, 1974). Overlying middle Eocene strata (Kirthar Series), which include a regressive sequence of nonmarine deposits in the north and marine beds in the south (Sahni and Kumar, 1974), apparently signal the initial encounter of India with Eurasia. A nonmarine connection between India and Eurasia likely was established during early Eocene time because terrestrial vertebrates with Mongolian and North American affinities were well established in India by the middle Eocene Epoch (Sahni and Kumar, 1972).

Extensive sedimentation in the sub-Himalayan belt followed closely in time upon the middle Tertiary rejuvenation of major spreading in the Indian Ocean. The Oligocene to middle Miocene Murree Formation apparently represents thin deltaic deposits derived partly from a growing Himalayan source and partly from the Indian Shield to the south, but it is succeeded gradationally by the continental Siwaliks d dominantly from the north as a conmable sequence extending into the Pleisene. Three major divisions of the aliks (lower, middle, upper) probably resent time-transgressive facies rather n time-stratigraphic units (Glennie and gler, 1964). The upper Siwaliks are irse alluvial fan deposits close to the irce, whereas the middle and lower aliks contain more distal river-channel d flood-plain deposits. Tandon (1971) orded variable but generally southerly leocurrent directions in the upper aliks, southerly directions in the middle aliks, and southeasterly directions in the ver Siwaliks. More recently, Parkash and iers (1974) noted both east-west and rth-south dispersal trends in the upper aliks, southerly directions in the middle aliks, and southwesterly directions in lower Siwaliks. Siwalik physiography s thus similar to that of the present nges flood plain and its piedmont outaries, except that the depositional sysn has shifted southward with time as the malayan uplift has continued. Some fold- and faulting of the Siwaliks preceded erthrusting along the Main Boundary ult, although deposition probably also ntinued during thrusting (Johnson and ndra, 1972), and the warping of modern er terraces suggests continued activity ansser, 1964).

In the Sind area to the west, but still east the suture belt along the Quetta Line, rmian-Carboniferous through Eocene irine units are predominantly shelf limene (Krishnan, 1968). In adjacent outcrop ts farther to the east, Kirthar, Murree, d Siwalik sediments in the north pass ithward into a marine section that was dently deposited in the head of a shallow lf lying west of peninsular India. Marine iocene limestone, sandstone, and shale ari and Gaj Series) gradually filled this lf from the north and were followed by ntinental beds (Manchhar Series) roughly uivalent to the Siwaliks but with a unger base (Krishnan, 1968). The gulf ccession was later folded during Pliocene d younger deformation.

In the Assam-Bangladesh area to the east, t still west of the suture belt along the Inburman ranges, Cretaceous through cene strata (Disang Series and Jaintia oup) are predominantly shallow-marine d paralic sediments that thicken southrd toward the ancestral Bay of Bengal aju, 1968; Bhandari and others, 1973). lift during Oligocene time apparently ncided with resumption of spreading in Indian Ocean and formed a paleoslope m northeast to southwest, with regresn and erosion in the northeast near the loburman ranges (Raju, 1968). Highrgy coastal environments followed by ckish lagoonal conditions caused deposition of strandline sandstone and coal measures as the sea retreated (Bhandari and others, 1973). A regional unconformity at the Oligocene-Miocene boundary perhaps corresponds with the major uplift of the Himalayas to the north, and a regional paleoslope from north to south has persisted from early Miocene time to the present (Bhandari and others, 1973). Lower Miocene strata (Surma Group) reflect fluvial, deltaic, brackish, and marine environments in facies succession from north to south (Raju, 1968) and thicken toward the south and southeast (Bhandari and others, 1973). Middle Miocene through Pleistocene strata (Tipam and Moran Groups) are fluvial sandstone and claystone that reflect a geography similar to the present. The upper parts of this sequence resemble the Siwaliks (Krishnan, 1968), and they thicken toward the Himalayas to the northwest (Bhandari and others, 1973).

Bengal Fan Turbidites

The Bengal fan described by Curray and Moore (1971) is a huge deep-sea fan built into the Bay of Bengal from the north by turbidity currents that issue from the front of the Ganges-Brahmaputra Delta. Both the deltaic platform and the fan are apparently the largest depositional features of their kind in the world today. The detritus fed to them is derived by longitudinal drainage of the world's highest mountain ranges in the Himalayan region. The area of the fan is roughly 3×10^6 km^2, the maximum thickness of sediment is perhaps 15 km beneath the delta-fan interface (Moore and others, 1971), the average thickness is about 7.5 km (Naini and Leyden, 1973), and the total volume of sediment is thus of the order of 20×10^6 km^3. From correlations of magnetic anomalies and the results of deep-sea drilling, the oceanic crust beneath the fan is inferred to be Cretaceous to Paleocene in age (Sclater and Fisher, 1974). The layering within the sediment is complex, as discussed by Moore and others (1975), and we offer here no detailed interpretation of the age and composition of the sediment. The sea-floor morphology in the Bay of Bengal and the immense total bathymetric relief of at least 3 km on the sloping fan surface (Curray and Moore, 1971) suggest to us that the bulk of the sediment is fan sediment derived from the erosion of highlands raised along the Himalayan suture during deformation induced by crustal collision of India with Eurasia. The fan surface slopes southward for 1,000 km, confined on the west by the continental slope off India and Ceylon and confined on the east by the Andaman-Nicobar island chain. The east flank of the fan fills the northern extension of the Java trench, and deformed fan sediments there dip eastward into a subduction zone at the front of the Andaman-Nicobar insular ridge (Curray and Moore, 1971, 1974). On the west, sedimentary basins along the east coast of India contain late Mesozoic and Tertiary paralic and shallow-marine strata that record progressive downwarping of the continental margin, together with recurrent development of local grabens, beginning near the end of Jurassic time (Sastri and others, 1973). The continental margin presumably formed by rifting during the breakup of Gondwana, and the sedimentary basins along the coast and beneath the shelf offshore then evolved by subsidence concurrent with the formation of adjacent oceanic crust now beneath the Bengal fan.

Indoburman-Sunda Subduction

The outer sedimentary arc of the Sunda arc-trench system continues northward from the Mentawai Islands off Sumatra to pass first along the Nicobar and Andaman Islands and then into the Arakan Yoma highlands of the Indoburman ranges west of the central Burmese lowland or trough now occupied by the Irrawaddy River (Van Bemmelen, 1949; Weeks and others, 1967; Rodolfo, 1969b). South of Burma, the insular ridge that marks this tectonic feature stands east of the Java trench and its filled northern extension. However, it is west of the volcanic arc, which lies along the Barisan Range in Sumatra and then along the Weh-Barren-Narcondam trend of volcanic islands and seamounts in the Andaman Sea. General analysis of arc-trench tectonics suggests that such partly submerged insular ridges standing above the inner walls of trenches are composed of structurally complex rock masses uplifted following or during deformation in the subduction zones associated with the adjacent trenches (Dickinson, 1973). Descriptions of exposures (discussed below) along the Mentawai-Nicobar-Andaman trend and in the Indoburman ranges along strike include reference to complex thrusting, glide sheets, ophiolitic complexes, exotic tectonic inclusions, foliated claystone, and related features characteristic of mélanges and associated rock masses. Continuous subbottom profiling by seismic reflection indicates that the western foot of the Andaman-Nicobar insular ridge is thrust westward over strata of the sea floor in the Bay of Bengal (Weeks and others, 1967). Recent work by Hamilton (1974) emphasizes the structural continuity of the Mentawai-Nicobar-Andaman trend that links subduction complexes of the Sunda arc-trench system in Java and Sumatra with deformed rocks of the Indoburman ranges along tectonic strike to the northwest. Curray and Moore (1974) further noted that the folds and faults detected in the deformed Bengal fan sediments of the Andaman-Nicobar insular ridge are geometrically similar to structures exposed subaerially in the Indoburman ranges.

In the Andaman and Nicobar islands (Karunakaran and others, 1964, 1968), the oldest rocks exposed are upper Mesozoic radiolarian chert and metagraywacke associated with ophiolitic volcanic rocks and mafic to ultramafic intrusives including serpentinite. The structurally complex ophiolitic rocks are overlain by Eocene conglomerate and sandstone. These pass upward conformably into the Andaman flysch, which is composed of graded sandstone and shale beds of Eocene to Oligocene age. Paleocurrent directions in the flysch are consistently south-southwest. This sequence is cut by abundant shear zones suggestive of thrusting or gliding mainly toward the west. The Andaman flysch was strongly deformed prior to the deposition of shallow-marine Miocene strata that are less deformed but overlain unconformably by Quaternary deposits. We infer that the Andaman flysch was derived from southeast Asia and deposited by turbidity currents on the sea floor within and near an inactive trench during the time of slow spreading in the Indian Ocean. The underlying ophiolitic rocks we interpret as a subduction complex related to the consumption of oceanic crust during the earlier period of fast spreading, and we relate the pre-Miocene deformation of the Andaman flysch to the resumption of spreading in the Indian Ocean during Oligocene time. Deformation of the unconformably overlying Miocene strata was apparently concurrent with the extensional opening of the central Andaman basin in the Andaman Sea to the east since middle Miocene time (Rodolfo, 1969b).

The Indoburman ranges northeast of the Bay of Bengal (Fig. 3) are composed largely of an immense thickness of strongly deformed flysch, mainly lower Tertiary but partly Upper Cretaceous (Brunnschweiler, 1966). Ophiolitic basalt and serpentinite, together with chert and fine-grained limestone, are tectonically intermixed with Cretaceous phases of the flysch, but they occur only as clasts or exotic blocks among Eocene rocks. A folded unconformity between Eocene and Cretaceous strata is prominent near the coast and recalls similar relations along the Andaman-Nicobar insular ridge. Numerous mélange belts ("exotic facies") occur as dipping layers inclined steeply to the east subparallel to bedding and to the axial surfaces of isoclinal folds. During Oligocene time, when spreading resumed in the Indian Ocean, the Indoburman flysch was uplifted to form a persistent highland barrier that since has separated the sedimentary basins of Assam-Bangladesh and the Burmese lowland. We infer that the Indoburman flysch was deposited as turbidites in deep water at the head of an ancestral Bay of Bengal and that it was deformed as closure of that deep basin allowed the Assam-Bangladesh shelf to approach closer to Burma.

The deformed Indoburman flysch is bounded along the eastern flank of the Arakan Yoma highlands by an east-dipping thrust system along which the deformed flysch is tectonically overlain by an ophiolitic slab of serpentinite-peridotite and diorite-gabbro (Brunnschweiler, 1966). The ophiolitic rocks are structurally overlain by an orderly Tertiary sequence beginning with lower Eocene or Paleocene marine strata older than the upper Eocene flysch exposed immediately to the west of the thrust zone beneath the ophiolitic rocks. The orderly sequence forms the base of a thick Tertiary succession, complete except for local disconformities, beneath the central Burmese lowland (Tainsch, 1950). We infer that the Indoburman flysch of the Arakan Yoma was thrust by subduction beneath an arc-trench gap (the ancestral Burmese lowland) that lay at the flank of a marginal magmatic arc. Igneous activity in the arc is recorded by extensive Eocene-Oligocene plutonism and metamorphism along the Mogok belt at the edge of the Shan Plateau east of the Burmese lowland (Searle and Haq, 1964). Some plutons in the Mogok belt are apparently as young as Miocene (Maung Thien, 1973), and in the northern Indoburman ranges, the thick Miocene-Pliocene Chindwin molasse is locally thrust westward over deformed flysch (Brunnschweiler, 1966). Late Cenozoic arc–volcanic rocks lie west of the site of the earlier Cenozoic arc along a trend within the central Burmese lowland (Maung Thien, 1973) and in line with the volcanic islands of the Andaman Sea.

With the uplift of the Indoburman ranges, the Burmese lowland became a narrow trough that was fed a thick succession of Oligocene, Miocene, and Pliocene clastic strata from the north. A longitudinal facies change from continental and paralic deposits in the north to marine strata in the south is characteristic of all horizons (Tainsch, 1950; Maung Thien, 1973). The lowland is now occupied by the Irrawaddy River, which has built a rapidly growing delta into the northern end of the Andaman Sea (Rodolfo, 1969b). Turbidity currents carry sediment from the delta to the floor of the central Andaman basin (Rodolfo, 1969a), which lies east of the volcanic trend beside the Andaman-Nicobar insular ridge. The longitudinal drainage of orogenic highlands associated with the Himalayan suture is thus currently supplying sediment to deep-water basins lying in tectonic settings in two different positions with respect to the magmatic arc still active along tectonic strike from the suture belt. The Bay of Bengal and the Bengal fan lie in front of the arc trend, but the central Andaman basin lies behind the arc trend at a place where extension has detached the arc structure from a continental mainland.

APPALACHIAN-OUACHITA REGION

Paleozoic Ocean Closure

Wilson (1966) suggested, on the basis of contrasting lower Paleozoic faunal realms within the Appalachian orogenic belt, that a pseudo–Atlantic Ocean lay off present-day eastern North America prior to the late Paleozoic assembly of Laurasia. The idea is especially attractive in terms of plate tectonic models for orogeny, because it permits the inference that the Caledonides of Europe and the Mauritanides of Africa, as well as the Appalachian orogenic belt, were related to plate consumption by subduction during closure of the old ocean. The opening of the present Atlantic along the general trend of the resulting juncture, but not along precisely the same line as the old suture, is viewed as a later event. The early Paleozoic ocean has been termed Iapetus east of Greenland in the region where it separated Europe from North America (Harland and Gayer, 1972). Farther south, paleomagnetic data suggest that arms of the early Paleozoic ocean separated Africa, then a part of Gondwana, from both North America and Europe (Hailwood and Tarling, 1973).

The manner and timing of ocean closure is still uncertain in detail, but the sequence of orogenic events in space and time along the Appalachian-Caledonide belt indicates that the suturing of other continental masses to the eastern margin of North America progressed from north to south in present geographic terms (for example, P. E. Schenk in Zietz and Zen, 1973; see also Fig. 4). Respective times of inferred suturing have been estimated crudely by examining polar wandering paths for the circum-Atlantic continents as plotted on Bullard-type reconstructions of the Atlantic region for Permian-Triassic time (prior to the opening of the present Atlantic). Paleomagnetic data suggest that polar paths for the two regions containing the cratonic interiors of North America and the Baltic Shield converged in Silurian time; furthermore, the joint path of the two regions, here assumed to have been connected, then converged with the polar path for the British Isles in the Devonian Period (Briden and others, 1973). Finally, the North American and African or Gondwanan polar paths converged in Carboniferous time (Hailwood and Tarling, 1973). General times of Paleozoic suturing of continental masses in the Atlantic region, as inferred from

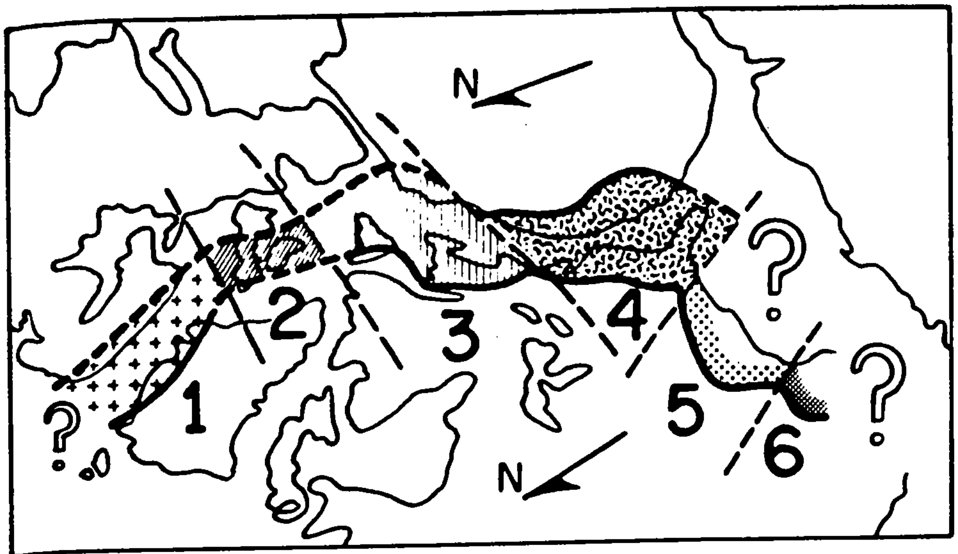

Figure 4. Bullard reconstruction of the closed North Atlantic with approximate Paleozoic margins of Appalachian-Caledonide-Mauritanide suture belt indicated. Patterns show schematically inferred times of final ocean closure along Appalachian belt (1–4) and of climactic deformation along Ouachita orogenic belt (5–6): 1, Middle to Late Silurian time in Scandinavia (Gale and Roberts, 1972); 2, Late Silurian to Early Devonian in Britain (Dewey, 1969); 3, Middle Devonian in the northern Appalachians (Bird and Dewey, 1970; Schenk, 1971); 4, Late Devonian to Carboniferous in the southern Appalachians (Hatcher, 1972); 5, Middle Carboniferous time in the Ouachita Mountains (Ham and Wilson, 1967); 6, Late Carboniferous to Early Permian in the Marathon region (Ham and Wilson, 1967).

paleomagnetic data, thus are broadly compatible with the specific timings suggested by various authors (indicated in Fig. 4) for segments of the Appalachian-Caledonide-Mauritanide belt on the basis of local geologic relations.

The positions of suture belts formed during Paleozoic ocean closure are difficult to specify because they involved interaction between North America and several different crustal entities. Most important were Europe (Eurasia) and Africa (Gondwana). Europe was sutured to North America along the northern Appalachian-Caledonide belt before Africa joined North America along the southern Appalachian-Mauritanide belt (McKerrow and Zeigler, 1972; see also Fig. 4). Also present along the Appalachian belt is an assemblage of rocks representing an elongate island-arc system that was cored by latest Precambrian and Cambrian volcanic and plutonic rocks (about 500 to 750 m.y. B.P.) and that probably stood within the ocean basin (Rodgers, 1971). This latter terrane is represented well in the Avalon belt of Newfoundland (Hughes and Bruckner, 1971), the Carolina slate belt (Fullagar, 1971), and basement beneath nearly flat-lying Paleozoic strata in Florida (Bass, 1969).

Various subduction zones and magmatic arcs associated with plate consumption may have stood on both sides of and within the ocean during a long history of closure spanning more than 100 m.y. Across any transect of the orogenic belt, arc-continent collisions may have preceded final continent-continent collision (Dewey and Bird, 1971; Bird and others, 1971). Tectonic overprints from successive episodes of deformation thus mark the orogenic terranes, and disruption of the orogenic belt by the later opening of the Atlantic has left large parts of it hidden beneath the modern continental shelves. In the south (see Fig. 1B), the region of possible sutures extends from the Brevard zone between the Blue Ridge and the Piedmont on the American side (Watkins, 1971) to an ophiolitic zone within the Mauritanides on the African side (Sougy, 1969). Estimates of the times of uplift of highlands formed by crustal collisions along the complex suture belt can be gained best from the ages of exogeosynclinal wedges and other clastic sediments shed westward from the orogenic belt toward the North American craton. Figure 5 depicts sequentially the late Paleozoic geologic events in North America related to the final development of the suture belt in the southern Appalachian-Mauritanide region.

Appalachian Clastic Wedges

The Cambrian and Early Ordovician continental shelf along the eastern edge of the North American craton is recorded by a carbonate-bank assemblage that extends from Newfoundland to Alabama and grades eastward to lutite deposited in deeper water (Rodgers, 1968). The top of this assemblage everywhere grades upward and eastward into Middle and Upper Ordovician shale and graywacke extending westward from uplands related to the Taconic orogeny. This orogenic event probably records either the collapse of a fringing island arc against the continental margin by closure of an intervening marginal sea (Dewey and Bird, 1971; Bird and others, 1971) or the collision of an intraoceanic island arc (Rodgers, 1971) with the continental margin following consumption of an indeterminate width of intervening oceanic lithosphere. Chapple (1973) has interpreted the sequence of diachronous events described as the Taconic orogeny (Rodgers, 1971) to be the result of abortive or partial subduction of the edge of North America beneath the flank of an approaching arc-trench system. For the southern Appalachians in particular, Odom and Fullagar (1973) inferred that an offshore island arc (Piedmont) was not a part of North America during early Paleozoic time but instead was facing North America and consuming intervening oceanic lithosphere by subduction beneath the near flank of the arc. They concluded that the arc collided with the continental edge in Middle Ordovician time to form an arc-continent suture along the Brevard zone. Whatever the details of the arc-continent interaction that produced the Taconic orogeny, the events did not mark terminal suturing within the orogenic belt (for example, Chapple, 1973; Odom and Fullagar, 1973).

The Ordovician clastics shed from sources uplifted or brought into position at the edge of the continent by the Taconic orogeny in the central Appalachians include turbidites (Martinsburg) that were deposited in a deep-water trough with longitudinal paleocurrents trending parallel to the continental margin (McBride, 1962). The depositional site may have been a structural furrow formed by depression of the continental edge as it was drawn down against the flank of a colliding arc-trench system (for example, Neill, 1975). Yeakel (1962) has shown that overlying clastics of Silurian (Tuscarora) age in the central Appalachians form a simple clastic wedge with fluvial paleocurrent trends leading transversely away from the orogenic belt toward the craton. To the southeast, the Tuscarora overlaps deformed older beds that were exposed along the eroded flanks of uplands raised by the Taconic orogeny; to the northwest, it thins and interfingers through deltaic facies with epicontinental marine shelf deposits of the continental interior.

A younger (Devonian) clastic wedge, which is not prominent south of the central Appalachians (Colton, 1970), reflects terminal suturing of the northern Appalachians associated with the Acadian orogeny (Bird and Dewey, 1970). Devonian continental deposits of the Old Red landmass formed by the joining of Europe to North America began to accumulate by Late Silurian to Early Devonian time in

Scandinavia and Britain, by Early to Middle Devonian time in maritime Canada, and by Middle to Late Devonian in New York and Pennsylvania (Friend, 1969). Pelletier (1958) has shown that the Mississippian (Pocono) upper part of the Acadian clastic wedge in the central Appalachians also displays fluvial paleocurrent trends leading from thick continental deposits near highlands on the southeast toward thinner shoreline facies gradational to marine deposits on the northwest.

Widespread marine deposition was not resumed in the northern Appalachians, although thick Carboniferous continental strata deposited in faulted basins and locally strongly deformed are prominent in maritime Canada (Belt, 1968) and New England (Quinn and Moore, 1968). In the southern Appalachians, however, Mississippian strata include shallow-marine sandstone, shale, and limestone with highly variable paleocurrent trends suggestive of shelf deposition (Schlee, 1963; Englund, 1968; Colton, 1970) in a part of the orogenic belt not yet fully sutured. A prominent Pennsylvanian clastic wedge (which does not extend north of the central Appalachians) records the Alleghanian orogeny during which the suture in the southern Appalachians was presumably completed. Mid-Carboniferous plutons (about 300 m.y. B.P.) that form a distinct trend in the southern Appalachian Piedmont (Fullagar, 1971) but are not prominent in the northern Appalachians may also reflect later closure of the orogenic belt in the south. At the northern extremity of Carboniferous marine-shelf deposition in the central Appalachians, alluvial-plain deposits (Mauch Chunk) form a continuous section linking the Mississippian (Pocono) upper part of the Acadian clastic wedge to the Pennsylvanian (Pottsville) lower part of the Alleghanian clastic wedge (Meckel, 1970).

In the southern Appalachians, regression associated with deposition of continental and littoral deposits in the Pottsville clastic wedge derived from the east began near the Mississippian-Pennsylvanian boundary. Complex intertonguing of marine and alluvial facies developed as seaward progradation advanced toward the west and northwest. Offshore shale and carbonate, beach-barrier sandstone, and delta-plain deposits with coal are encountered along traverses from northwest to southeast across an elongate depocenter, the Appalachian basin, that lay between the uplifted orogenic belt and the craton (Englund, 1968; Ferm, 1970; Smith and others, 1971; Ferm and others, 1972). Paleocurrent studies (Schlee, 1963; Chen and Goodell, 1964; Meckel, 1967; Jones, 1972) indicate that fluvial transport in the proximal facies was toward the northwest away from rising highlands in the orogenic belt and that the dispersal direction in the distal facies was toward the southwest subparallel to the orogenic belt (Fig. 5). From the nature of the sedimentary deposits, we infer important bypassing of sediment toward the still undeformed southern or Ouachita margin of the craton. We thus compare the Pottsville dispersal system geometrically to the Siwalik-Ganges dispersal of Himalayan sediment to the head of the Bengal fan.

At the northern end of the Appalachian basin, sediment was contributed from the craton to the north as well as from the orogenic belt (Meckel, 1967). Sediment from the northern Appalachians may also have reached the Ouachita region by a wide circuit through the Illinois basin in the interior of the craton (Potter and Pryor, 1961). In the Black Warrior basin at the southern end of the Appalachian basin, Pennsylvanian clastics thicken from north to south on the outcrop (Ferm and others, 1967; Colton, 1970) and from northeast to southwest in the subsurface (Thomas, 1972). These have been interpreted as fluvial and deltaic deposits spread northward from an orogenic highland along regional strike from the slightly younger Ouachita belt to the west (Ferm and others, 1967). In view of the uncertain structural relations near the tectonic juncture between the Appalachian and Ouachita systems (for example, Thomas, 1973), this

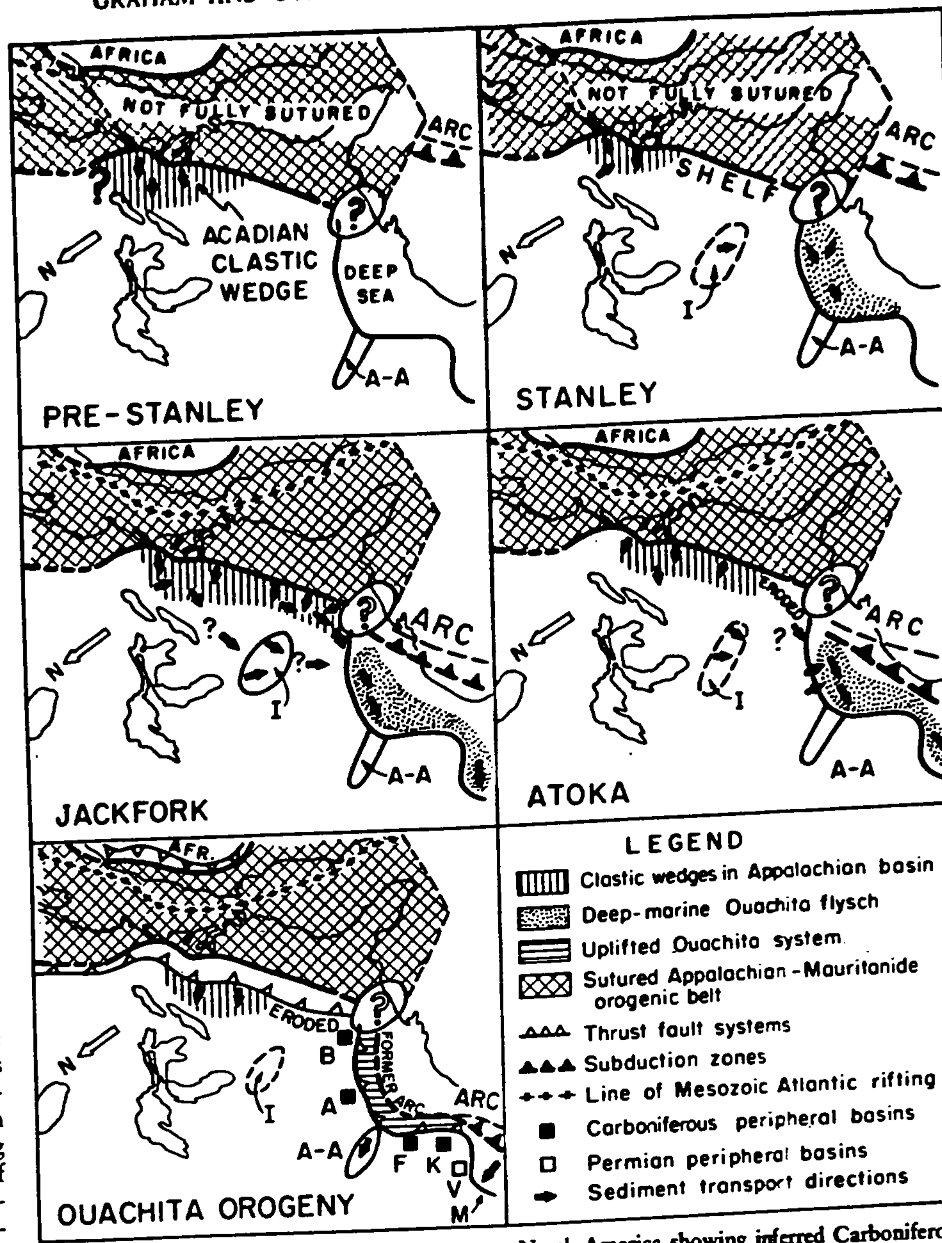

Figure 5. Interpretive sketches of southeastern North America showing inferred Carboniferous evolution of Ouachita flysch and Ouachita system. Time frame of each sketch identified by reference to subdivisions of Ouachita flysch (see text for ages). Features shown include: A-A, Anadarko-Ardmore basin; A, Arkoma basin; B, Black Warrior basin; F, Fort Worth basin; I, Illinois basin; K, Kerr basin; M, Marathon region; V, Val Verde basin.

suggestion is difficult to evaluate. Pottsville paleocurrents on the outcrop in the same area indicate generally westward movement of sediment (Schlee, 1963; Metzger, 1965). We infer tentatively that an orogenic syntaxis lying beyond the exposed end of the southern Appalachian belt blocked axial dispersal of sediment moving southward along the trend of the Appalachian basin and diverted sediment westward toward the Ouachita region. An analogous dextral diversion of sediment dispersal occurs now at the eastern end of the Himalayas, where the Ganges drainage along the foot of the mountains is blocked by the Indoburman ranges in Assam-Bangladesh and is turned toward the deltaic complex at the head of the Bay of Bengal.

Carboniferous Ouachita Flysch

Carboniferous turbidites, in aggregate 7.5 to 12.5 km thick (Cline, 1970; Haley and Stone, 1973; Morris, 1974b), are exposed in the Ouachita Mountains for 350 km along strike. The exposed rocks form but a small segment of a largely buried fold belt, the Ouachita system, which extends for 2,000 km from the southern Appalachians to the Mexican border (Flawn and others, 1961). The flyschoid nature of the Ouachita Carboniferous rocks was documented by Cline (1970), and we refer to the sequence here as the Ouachita flysch. The volume of the Ouachita flysch within the Ouachita Mountains is of the order of 3 $\times$ 10^6 km^3, or about one-seventh of the volume of the Bengal fan within about one-seventh of the length of the Ouachita system.

The Ouachita flysch was deposited conformably on a continuous sequence of Ordovician to Mississippian strata less than 2 km thick and exposed as the highly deformed core of the Ouachita Mountains. These strata are largely graptolitic shale, chert, and mainly thin-bedded sandstone interpreted generally as deep-marine deposits of a sediment-starved sea floor (Cline, 1970; Haley and Stone, 1973; Morris, 1974b). No depositional base to the sequence has been observed. Thrusts have carried the Ouachita core northward into near juxtaposition with age-equivalent but carbonate-rich shelf sections (Ham, 1959; Frezon and Glick, 1959). We infer that the lower and middle Paleozoic sequence of the Ouachita core was deposited in a deep oceanic basin that lay south of a shelf edge at the southern margin of Paleozoic North America. The Carboniferous Ouachita flysch was also deposited in deep water and contrasts lithologically with thinner shelf equivalents in the Arkoma basin north of the present Ouachita Mountains (Cline, 1970; Chamberlain, 1971). Morris (1971a, ... Carboniferous sequence as the clastic fill of an initially deep trough or elongate oceanic basin, and we accept this view. Morris (1974b) has provided summary descriptions of each of the four stratigraphic divisions of the Ouachita flysch:

1. The Stanley Group (~4 km) of Mississippian (Osagean to Chesterian) age is mainly dark shale with some intercalated silicic tuffs, but it includes turbidite sandstone beds that increase in number, thicken individually, and become more feldspathic toward the south in outcrop sections. Tuff and tuffaceous sandstone are also more prominent in southern outcrops. Paleocurrent studies (Johnson, 1968; Morris, 1974b) of turbidites indicate northwesterly flow in eastern outcrops and westerly flow in western outcrops, with westerly directions everywhere dominant in the upper Stanley. Locally, basal quartzose sandstone and slumped limestone blocks reflect minor sediment transport off the shelf to the north.

2. The Jackfork Group (~3 km) of Pennsylvanian (approximately early Morrowan) age consists of shale and quartzose turbidite sandstone with westerly paleocurrent trends (Briggs and Cline, 1967; Morris, 1974b). Presumed slump structures are again important locally along the northern fringe of exposures, and turbidite beds appear more proximal toward the southeast.

3. The Johns Valley Shale (<1 km) of Pennsylvanian (approximately late Morrowan) age contains only distal turbidite sandstone, but it is known for large exotic blocks of shallow-marine limestone and sandstone. Shideler (1970) successfully correlated individual exotic blocks with a number of specific pre-Pennsylvanian formations exposed in the Paleozoic shelf sequences of the Arbuckle and Ozark regions north of the Ouachita system in Oklahoma and Arkansas, respectively. The exotic blocks are generally supposed to have been derived from an unstable continental slope and to have been transported southward into deeper water as clasts within submarine mass flows.

4. The Atoka Formation (~5+ km) of Pennsylvanian (approximately Atokan) age records transition upward to shallow-water deposits. Atoka sandstones apparently contain more lithic grains than those of the Jackfork (Morris, 1974b). The lower Atoka is shale and turbidite sandstone beds with westerly paleocurrent trends, and proximal turbidites are more prominent on the outcrop toward the east (Briggs and Cline, 1967; Morris, 1974b). The lower Atoka thins northward and grades into more massive sandstone units of shelf and littoral environments (Stone, 1969). As the sequence accumulated, the shelf subsided by growth faulting (Koinm and Dickey, 1967) whose effects extended progressively northward with time. In the upper Atoka, deposition in shallow-marine waters and on adjacent coastal plains became dominant as deltaic and nearshore clastic accumulations prograded southward from the continental margin to the north, as reflected by southerly paleocurrent trends (Briggs and Cline, 1967; Morris, 1974b; Stone, 1969). The Atoka is overlain unconformably by mid-Pennsylvanian (approximately Desmoinesian) coal-bearing strata of the Krebs Group along the northern flank of the Ouachita Mountains.

The continental margin that lay north of the Paleozoic ocean floor upon which the Ouachita flysch was deposited probably formed by continental rifting near the Precambrian-Cambrian time boundary. The elongate Anadarko-Ardmore basin (Ham and Wilson, 1967) extends nearly 500 km into the continent from the northwesternmost bulge in the Ouachita system. The basin is apparently an aulacogen that formed as a gash in the continental margin along the failed arm of a triple junction associated with the continental separation (for example, Hoffman and others, 1974). Though less than 100 km wide, this basin is floored by 2 to 3 km of Cambrian volcanic rocks, which do not appear on the cratonic platforms to either side, and has a Paleozoic sedimentary fill 10 to 12 km thick. Except for their greater thickness, the sedimentary deposits are much like those of the adjacent cratonic cover. Contrasting sediment thicknesses within the Arbuckle Mountains sequence (Ham, 1969) suggest rapid thermal subsidence within the aborted rift soon after its initiation. In less than 100 m.y. from Late Cambrian to Late Ordovician time, 3 to 4 km of mainly shallow-water carbonate strata accumulated, yet less than a kilometer of beds were deposited in the succeeding 100 m.y. and more up to mid-Mississippian time. Evidence for progressive early Paleozoic subsidence of a gradually cooling new sea floor south of the continental margin may be preserved also in the lithology of the strata underlying the Ouachita flysch. Shale in strata older than Middle Ordovician age in the Ouachita core sequence are commonly calcareous, but pre-Stanley shale younger than Middle Ordovician age are associated with bedded cherts (Morris, 1974b); the contrast may reflect the time of subsidence through the carbonate compensation depth.

The sources for the Ouachita flysch suggested by Morris (1971a) are generally compatible with our inference of a subsea fan system derived mainly from the Appalachian suture belt to the east and southeast. Morris inferred a dispersal system for the more quartzose sand from the northern Appalachians and the Canadian Shield across the Illinois basin of the craton, and a dispersal system for the more lithic sand directly from the region around the southern end of the Appalachian basin. Thomas (1972) and Stone (1968) also suggested a link between the Ouachita flysch and clastics derived in part from south of the present Appalachians. The tuffs and feldspathic sands of the Stanley Group may reflect an additional source south of North America.

Ouachita Orogenic Events

The youngest beds of the Ouachita flysch are of Middle Pennsylvanian (Atokan) age. Shortly after deposition, the whole section was folded, together with the underlying core sequence, and carried northward at least 75 km in complex thrust sheets that rode across the coeval shelf section on the Paleozoic continental margin. Viele (1973) suggested that the thrusting began during flysch deposition, and he has documented the overall geometry of a complex stack of folded nappes in which both the Carboniferous flysch and the underlying core sequence are involved. Middle Pennsylvanian (Desmoinesian) strata in the Arkoma basin are folded in the subsurface in sympathy with Ouachita structures; equivalent and younger Pennsylvanian strata farther west contain chert-pebble conglomerate probably derived from rising Ouachita highlands (Ham and Wilson, 1967). Strong Pennsylvanian deformation of the southeastern end of the Anadarko-Ardmore aulacogen nearest the Ouachita orogenic belt was accompanied by deposition of approximately 5 km of clastics along the trend of that basin.

Most of the remainder of the Ouachita system is concealed by overlapping Cretaceous strata of the Gulf Coast lowlands. In the subsurface to the southeast (Thomas, 1972), slate and chert that apparently represent the Ouachita core sequence can be traced toward the southern end of the Appalachian belt. In the subsurface to the southwest (Flawn and others, 1961), a sinuous deformed belt of the Carboniferous Ouachita flysch and the older Paleozoic core sequence can be traced discontinuously in well cores to exposures in the Marathon region of West Texas. In the Texas subsurface, as on the outcrop, the flyschoid rocks of the Ouachita system are thrust northwestward over coeval shelf strata. Additional major thrusts, however, also bring metasedimentary rocks from the southeast across the Ouachita sedimentary units in the subsurface. Subcrops of post-orogenic(?) late Paleozoic strata apparently rest depositionally on rocks of the Ouachita system southeast of the main outcrop belt (Vernon, 1972; Meyerhoff, 1973; Woods and Addington, 1973). In the Marathon exposures farther west, the overthrust section of Carboniferous flysch and underlying Paleozoic strata is grossly similar to that in the Ouachita Mountains, although only roughly half as thick; the time of main deformation was also slightly later, near the Pennsylvanian-Permian boundary (Ham and Wilson, 1967). The extension or termination of the Ouachita system within Mexico remains unclear (López-Ramos, 1969).

In the Marathon region, pre-Carboniferous strata of the overthrust Ouachita system were once widely interpreted as mainly shallow-water deposits, but Thomson and McBride (1964) showed that the depositional features of most of the sequence are indicative of deep-water conditions. The Carboniferous flysch includes two packets of terrigenous turbidites with generally westerly paleocurrents separated by a sequence of carbonate turbidites (Dimple Limestone) with southeasterly paleocurrents. The carbonate flysch was derived from a shelf region along the edge of the adjacent continental platform (Thomson and Thomasson, 1969). The underlying terrigenous flysch (Tesnus Formation) contains quartz-rich turbidites with westerly and northwesterly paleocurrents, and the overlying terrigenous flysch (Haymond Formation) contains somewhat less quartzose sandstones with westerly and southwesterly paleocurrents (McBride, 1970). The turbidites grade upward into shallow-water molasse deposits at the top of the Marathon sequence.

Closely related to the Ouachita fold belt are a series of basins, distributed around the curving periphery of the Ouachita system, with late Paleozoic sections 2.5 to 5 km thick resting depositionally on undoubted continental basement or its cratonic cover (Ham and Wilson, 1967). Each of these peripheral basins is marked by rapid downwarping and filling that began shortly before or concurrently with climactic deformation in the adjacent segment of the Ouachita system. For example, in the Fort Worth basin of central Texas, the thickest fill is of Pennsylvanian age (Weaver, 1956; Turner, 1957), whereas the thickest fill is of Permian age in the Val Verde basin of West Texas (Vertrees and others, 1959). It seems clear that the major episodes of subsidence and deposition in the peripheral basins were related to the tectonic depression of an elongate region adjacent to the Ouachita orogenic belt. The string of peripheral basins thus defines a pericratonic foreland belt lying between the late Paleozoic orogen and the craton proper.

Our tectonic analogy suggests that the Ouachita orogeny reflects the crustal collision of a north-facing arc-trench system, with a south-dipping subduction zone, against the southern margin of North America. This conclusion is highly speculative, because the analogy with the Indian Ocean would involve some future collision between the Andaman-Nicobar segment of the Indoburman-Sunda subduction zone with the eastern continental margin of peninsular India and thus crumpling of the Bengal fan and thrusting of its sediment westward over shelf sequences now along the coastal zone.

Most published hypotheses for polarity of subduction in the northern Appalachians (Bird and Dewey, 1970), the southern Appalachians (Hatcher, 1972), and the Ouachitas (Keller and Cebull, 1973) all rely on arc-trench systems built near the edge of a Paleozoic North American plate and facing away from the continent. However, a recent analysis of relations in Newfoundland indicates that the opposite polarity, with an arc-trench system facing North America and having an east-dipping subduction zone, was the more likely configuration there prior to continental collision (Strong and others, 1974). In the Piedmont, many fundamental relations must yet be deciphered before any model can be defended in detail (Glover and Sinha, 1973), but Dewey and Burke (1973) inferred an east-dipping subduction zone there in Carboniferous time. This zone was presumably related to an arc-trench system along the edge of Africa south of the South Atlas fault. Paleozoic tectonic trends have yet to be traced from West Africa or the Piedmont into the circum-Caribbean region with any confidence.

In any case, we conclude that geologic relations within the Ouachita region favor the concept of orogeny owing to collision with a north-facing arc over the concept of orogeny associated with a south-facing arc built along or near the continental margin. Paleozoic shelf sequences north of the Ouachita Mountains betray no evidence of orogenic activity to the south until the sharp onset of the Ouachita orogeny in the Pennsylvanian Period, and it main effects were complete prior to Permian time. This style of orogenic development (Dickinson, 1971) is compatible with the concept of crustal collision, strictly limited in time, following consumption of offshore oceanic lithosphere beneath an arc-trench system facing the continent on the edge of a separate plate, but it is difficult to reconcile with long-continued arc-trench activity along the continental margin itself. In either case, the roots of the arc must now lie in the Texas subsurface as the buried former Paleozoic sediment source called Llanoria (Denison and others, 1969).

If the arc faced the continent, the main suture belt must lie somewhere within the Ouachita system. The only alpine serpentinite known is exposed as small tectonic slivers within the Ouachita core sequence (Sterling and Stone, 1961; Wicklein and Carpenter, 1973). If these are scraps of oceanic lithosphere, they could be fragments of the oceanic substratum that lay contiguous with the edge of the adjacent North American continent through the whole Paleozoic Era and need not mark a true suture. Because intact lateral gradations are commonly inferred between the Ouachita flysch and the shelf equivalents to the north, the most likely location of the suture belt probably now lies along or to the south of the metamorphic front in the subsurface of the Ouachita system, rather than anywhere on the outcrop. However, detailed facies relations are not clear in strata older than the Atoka Formation, and any palinspastic restoration of the Ouachita system is uncertain. Moreover, if the overthrust mass of the Ouachitas is viewed as a structurally telescoped subduction complex beneath which the edge of the continent was tucked, then the fundamental suture may be masked beneath the piles of nappes that form the exposed structural levels of the orogen. This possibility is enhanced by the presence within the Ouachita flysch of a variety of disturbed bedding types (Morris, 1971b) of the sort common elsewhere in subduction complexes. In addition, the exotic-bearing Johns Valley Shale, although regarded now as olistostromal, is similar in general character to lithologic belts that in other orogens are interpreted as mélanges indicative of differential tectonic movement within subduction zones.

Regardless of where the suture is located, the peripheral basins strung along the edge of the craton north and west of the Ouachita system are regarded here as indicating depression of the continental margin by partial subduction just before the colliding arc-trench system was stifled. As in other foreland belts, some proportion of the subsidence in these basins can be attributed to the downwarping of an isostatic moat in front of the tectonic load of nappes and overthrust sheets advancing northward along the flank of the adjacent orogen (Price and Mountjoy, 1970). Hamilton (1973) recently discussed similar downbowing of subducted slabs of lithosphere in response to the tectonic load imposed by a growing wedge of subduction complexes, and Coney (1973) has discussed general plate tectonic relations of foreland fold-thrust belts. The interplay between subduction tectonics related directly to plate consumption and gravity tectonics due to gravitational spreading of an overthickened subduction complex is poorly understood, but both must influence the overall development of foreland tectonic systems.

If the arc did face away from the continent as others have inferred, then the subduction zone would lie buried far to the south beyond Llanoria. The proper Himalayan-Bengal analog for the Ouachita flysch, thus deposited in a Ouachita trough (Morris, 1974a, 1974b) north of a fringing offshore arc, would then be a potential turbidite fill of the Andaman basin, rather than the Bengal fan as we suggest. Moreover, the analog for the Oklahoma-Arkansas shelf would be the Southeast Asian mainland rather than the east coast of peninsular India, and the geometric similarities implied by Figure 1 would dissolve. In our view, the only evidence that favors this interpretation is the presence of a few thin tuffs and feldspathic sandstones that are reportedly more prominent southward in the Stanley (Mississippian) division of the Ouachita flysch. Although this evidence is concrete, we do not regard it as conclusive and believe that it is outweighed by other considerations we have discussed.

LONGITUDINAL SEDIMENT DISPERSAL

Others have also emphasized the importance of longitudinal drainage of orogenic highlands (Eisbacher and others, 1974). We hold that the Himalayan-Bengal and Appalachian-Ouachita analogy illustrates an important principle of sedimentation in relation to tectonics. At any given time, existing oceans are the chief repository of sediment, and orogenic highlands are the main sediment sources. It follows that the ocean floors closest to the orogenic highlands have the best access to sediment. Ocean floors immediately adjacent to orogenic highlands include those that lie off active continental margins marked by arc-trench systems, but sediment deposited in this setting will be returned quickly to subduction zones at the continental margins and converted to mélanges or metamorphic terranes. The ocean floors best sited to receive thick continuous sections of clastic sediment are thus the remnant oceans that lie off the ends of sequentially developing suture belts between colliding continents. Because colliding continental margins cannot be aligned perfectly for simultaneous contact along their full length, sequential collisions, accompanied by repeated adjustments in plate motions, must be the general rule. Sequential collision thus affords a general model for the successive deposition of longitudinally transported flysch sequences, which in turn are successively deformed and incorporated into suture belts along the same orogenic belt that served as a sediment source (Fig. 6). An important corollary is that detritus shed from orogenic highlands that were developed along and beside a collision suture should not be sought only in clastic wedges shed transversely on the juxtaposed continents, but mainly along tectonic strike within the same orogenic belt. Conversely, sediment sources for synorogenic flysch within a sutured orogenic belt should be sought along tectonic strike within parts of the same

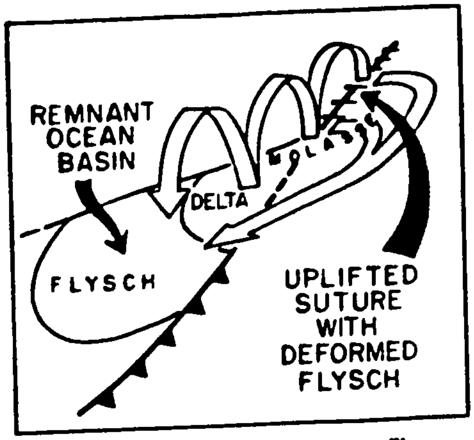

Figure 6. Conceptual diagram to illustrate progressive incorporation of synorogenic flysch within an orogenic suture belt by sequential closure of remnant ocean basin.

orogenic belt where each progressive stage of orogenic suturing occurred at a slightly earlier time. Although we do not pretend to have solved the puzzle of characteristically longitudinal paleocurrents for all synorogenic flysch sequences, we hope we have offered an approach that will lead to a satisfactory explanation of many common occurrences.

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