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PALEOGEOGRAPHIC INTERPRETATION: With an Example From the Mid-Cretaceous

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INTRODUCTION

Progress in global paleogeography in the 1970s was related to several key developments in the 1960s, among them the plate tectonics concept, the growth of the paleomagnetic data base, and the ability to determine relative water depths using fossil communities and sedimentary structures. As a result, generalized paleogeographic maps showing continental relationships and environmental interpretations are now available for all Phanerozoic periods (Ziegler et al 1979, 1982b). Progress in the 1980s will result from the recent recognition that continental margins can be stretched more than twofold during rifting (Montadert et al 1979, LePichon & Sibuet 1981, Dewey 1982), and that such stretched regions can be subsequently telescoped by hundreds of kilometers during collisions (Dewey 1982, Bally 1981). Indeed, the western margin of North America was stretched during the late Precambrian to earliest Cambrian, probably by a factor of 2 or more (Bond & Kominz 1984), compressed in the late Jurassic through early Tertiary Sevier through Laramide orogenies by a factor of 2 to 3 (Price 1981), and subsequently stretched in the Great Basin by at least a factor of 2 (Hamilton & Myers 1966, Coney 1980). Such mobility of the continental crust must be accounted for in palinspastic base maps before general paleogeographic relationships can be refined.

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In addition, much recent work within orogenic belts has focused attention on the widespread occurrence of relatively small crustal fragments that have been variously named exotic, allochthonous, or suspect terranes. These terranes record conspicuously different geologic histories from adjacent regions, including distinct lithologic, tectonic, magmatic, paleomagnetic, and biogeographic patterns. Terranes are separated from one another by sutures, marked by ophiolites and accretionary prisms, or by strike-slip faults, often of very large displacement. The Wrangellia terrane of western North America (Jones et al 1977) has travelled at least 30° and more likely 90° northward since the Triassic (Stone et al 1982), and during this transit it has passed from the coral reef zone into the temperate clastic zone (Monger et al 1982). It is at least locally separated from inner terranes by the Bridge River and Hozameen ophiolitic and accretionary prism assemblages of British Columbia and Washington State (Davis et al 1978). It is therefore important to establish the sequence and timing of terrane assembly and motion before more realistic global paleogeographic maps are reconstructed. Additional information on the motions of the oceanic plates can be gleaned from such studies (Page & Engebretson 1984).

This paper demonstrates how continent-scale palinspastic and paleogeographic maps are prepared, and how basin development and lithofacies patterns are related to extensional, compressive, and shear tectonic regimes. We illustrate this by a mid-Cretaceous (Cenomanian Stage) map centered on the United States and Mesoamerica, and by the geography and lithofacies of the Recent for comparison. The maps are segments of world-wide maps for 16 stages of the Mesozoic and Cenozoic eras being prepared by the Paleogeographic Atlas Project at the University of Chicago.

PALINSPASTIC RESTORATION

Most of the paleogeographic maps published every year do not employ tectonically restored base maps, and this practice can lead to a significant misrepresentation of relationships in areas that have experienced subsequent deformation. A map that shows regions restored to their undeformed state is called *palinspastic* [from Greek, meaning "stretched back" (Kay 1937)]. Although early paleogeographers (including C. Schuchert, E. Argand, and M. Kay, among many others) appreciated the problems of using unrestored bases (see Kay 1945), it was not until relatively recently that the severity of this problem was widely recognized. The realization of the extreme mobility of continental and oceanic lithosphere within the context of plate tectonics is due to the advent and increasing availability of various geophysical techniques, including multichannel seismic reflection profiles and cryogenically determined paleomagnetic poles, as well as to better geologic and structural maps of most parts of the world. For example, multichannel seismic reflection profiles of the southern Appalachians (Cook et al 1979) and eastern Great Basin (Allmendinger et al 1983) have suggested a minimum of 260 km of shortening and 30–60 km of extension, respectively, in these two regions alone. Paleomagnetic pole data have been particularly important in demonstrating large-scale latitudinal motions of various exotic terranes, such as the Wrangellia and Peninsular terranes, as well as large-scale rotations of terranes of the western Cordillera of North America. Structural techniques have also improved and now allow increasingly reliable determinations of rock strain and sense of shear along zones of high strain (Ramsay & Huber 1983, Simpson & Schmid 1983). These data taken together with improving global stratigraphic and geochronologic data provide much of the information necessary for making palinspastic restorations. In the following sections, we discuss the conceptual framework being used by the Paleogeographic Atlas Project to make first-order palinspastic restorations.

The primary information that is required to make palinspastic restorations is (*a*) the nature of deformations (i.e. do they reflect increases or decreases in surface area), (*b*) the amount of displacement associated with the various structures, and (*c*) the time interval during which displacements occurred. There is now a surprising amount of available information, albeit sometimes controversial, and thus first-order reconstructions are now possible. These reconstructions are required if we are to begin to fully appreciate the complexity of geological history and to recognize the next generation of important regional geological problems.

The primary goal of this paper is to outline how paleogeographic interpretations are made and depicted. As a part of this we present an interpretive map of Mesozoic and Cenozoic tectonic elements of the United States, Mesoamerica, and adjacent oceanic realms of the eastern Pacific, central Atlantic, and Gulf of Mexico (Figure 1; see color insert). This map provides much of the basic information required for making palinspastic restorations and has been used as the basis for the Cenomanian reconstruction depicted in Figure 6 (color insert). Before discussing our maps, we outline both the temporal and tectonic framework employed on them.

Tectonic Time Divisions of the Mesozoic and Cenozoic

Our approach to the temporal subdivisions of the Mesozoic and Cenozoic is based on the assumption that changes in global plate motions result in (or alternatively, result from) approximately synchronous, globally recognizable tectonic events. Such events include major continental collisions, periods of intracontinental rifting and/or ocean opening, and changes in plate velocities. This assumption follows from basic principles of plate kinematics on an Earth with a limited number of plates (some of which represent significant portions of the Earth's surface) that require that a change in relative motion between any two or more of the major plates must in turn result in changes of motion between all neighboring plates until a new quasi-steady state is reached, or until a new perturbing force operates on the system.

Five major changes can be detected during Mesozoic and Cenozoic times, which serve to define six intervals ranging in duration from 35 to 50 m.y. The interval lengths and boundaries have been established by choosing the beginnings or endings of major orogenic events and/or times of marked change in global plate motions. Thus, the sharp bend in the Hawaiian-Emperor and other Pacific hotspot tracks at about 43 m.y. is chosen as the beginning of the latest tectonic interval. This change in Pacific-hotspot relative motion in the middle Eocene corresponds quite well to the initiation of suturing of India with Asia, and since convergence continues today in this area, this interval is synonymous with the Himalayan Orogony. The opening of the Cayman Trough and the eastward motion of the Caribbean plate also occurred during this time (Pindell & Dewey 1982).

The time interval preceding this one corresponds temporally with the Laramide Orogeny, which extends from approximately the mid-late Cretaceous (Coniacian) to the middle Eocene in the type region. In North America, important "Laramide" effects include the reactivation of the Rocky Mountains, the closure of marine troughs in Mexico, and the collision of the Greater Antilles with the Bahamas platform (Pindell & Dewey 1982). Ocean-opening phases between India and Madagascar, Australia and Antarctica, New Zealand and Australia, New Zealand and Antarctica, and Greenland and North America, and northward Pacific-hotspot relative motion (Coney 1978, Henderson et al 1984) all occurred within this interval.

The opening of the South Atlantic near the beginning of the Cretaceous must represent another important change in plate motions, and it serves to define the start of an interval that corresponds roughly in time with the Sevier Orogeny of the western United States overthrust belt (Armstrong 1968). It should be mentioned that foreland thrusting along this belt continued for a long time and spanned the Nevadan, Sevier, and Laramide orogenic intervals. Subduction along the western margin of North America was also active during these intervals, but it was at about the mid-Cretaceous that suturing of Wrangellia occurred in the Pacific Northwest (Davis et al 1978).

Perhaps the most important upset in plate relationships happened at about the mid-Jurassic with the opening of the central Atlantic Ocean.

Extension along the eastern margin of North America coincided with the transformation of the western margin into an Andean system signaled by the onset of the Nevadan and Columbian orogenies of the United States and Canada, respectively. The transtensional opening of the Somali and Mozambique oceanic basins between east and west Gondwana was also an important event of this interval (Rabinowitz et al 1983).

Still earlier episodes of the Mesozoic must be defined on the basis of events far afield of North America. The collision of South and North China at about the mid-late Triassic (Norian) to begin the Indosinian Orogeny (Wang 1980) provides a convenient datum, and it coincides in time with the beginning of stretching along the Atlantic and Gulf margin of the United States. The Cape Orogeny of South Africa terminated at about the Norian (Dingle et al 1983) and may be used to define an interval that includes latest Paleozoic through mid-late Triassic times.

In conclusion, the Cape, Indosinian, Nevadan, Sevier, Laramide, and Himalayan orogenies in succession provide convenient terms for relatively restricted tectonic episodes of the Mesozoic and Cenozoic, and structures portrayed on our tectonic map (Figure 1) are correlated with these time intervals. An important point is that when intervals are defined in this way, the interval boundaries need not, and often do not, correspond with traditionally defined period or even stage boundaries.

In the following sections are discussed the major tectonic environments that we recognize: ocean floor, stretched continental crust, fold-thrust belts, subduction-accretion prisms, and strike-slip systems. The basic styles of deformation associated with each of these environments, the ways these deformations are treated palinspastically, and the methods useful in dating them are described. Specific examples of our treatment of these structures have been chosen from our Cenomanian palinspastic restoration (Figure 6).

Ocean Floor

Oceanic lithosphere is generated by seafloor spreading in a number of tectonic settings, including mid-ocean ridges, back-arc basins, and as the end product of pull-apart basin formation. The limits and age of ocean floor are for the most part readily determined from its typical geophysical signature (Emiliani 1983). Controversies generally arise over the location of continent-ocean boundaries, and over correlations of magnetic anomalies in some areas. The Gulf of Mexico is a good example of an oceanic realm for which both the extent of oceanic crust and the age of ocean floor have proved extremely controversial. The Gulf is characterized by low-amplitude magnetic anomalies that are difficult to correlate with a magnetic reversal sequence, by thick, extensive Jurassic evaporites that impede subsalt seismic reflections, and by thick sedimentary wedges along the continental margins. Comparison of maps depicting the limits and age of ocean floor within the Gulf (e.g. Buffler et al 1980, Cebull & Shurbet 1980, Martin 1978) shows that virtually all of them are different because they are based on different assumptions concerning the relationship of salt to crustal type, as well as on different prejudices concerning the opening history of the Gulf.

We depict the oceanic crust of the Gulf to be of middle-to-late Jurassic age, based on the regional constraints imposed by late Paleozoic stratigraphic and tectonic histories of circum-Gulf continental realms (Pindell & Dewey 1982), on the ages of the synrift Eagle Mills clastics and basaltic volcanics and syn- to postrift Louann salts, on the absence of correlatable magnetic anomalies due to seafloor spreading during the Jurassic quiet zone, and on the relative motion of Gondwana and Mesoamerica with respect to North America (Anderson & Schmidt 1983). Recent International Program of Ocean Drilling/Deep Sea Drilling Project (IPOD/DSDP) drillings in the Florida Straits bottomed in late Precambrian continental crust that is several tens of kilometers seaward of the 3000-m isobath (Schlager et al 1984); these results forcefully indicate the problems of interpreting the circum-Gulf region.

In the past the recognition of linear magnetic anomalies has been used to support arguments concerning the presence of oceanic basement in several areas [Labrador Sea (Srivastava 1978), South Africa (Rabinowitz & La Brecque 1979), Red Sea and Gulf of Aden (Cochran, 1981, 1983a)], but recent results derived primarily from multichannel seismic reflection profiling have shown, at least in the case of South Africa, that the innermost region of identified linear anomalies overlies stretched continental crust and not ocean floor (Austin & Uchupi 1982). With increasing appreciation of the magnitudes and frequent occurrences of stretched continental crust, there will no doubt be many more reinterpretations of this type along continental margins. As a final note, it should be pointed out that there are still important controversies concerning the correlation of linear magnetic anomalies with the reversal time scale. A good example can be seen in the west Philippine Sea, where Weissel (1981) and Shih (1980) have quite different interpretations of the anomaly correlations.

Stretched Continental Crust

Continental stretching involves an increase in surface area without the generation of new crust. The basic geometry is shown in Figure 2. In Figure 2*A* the ruled area of continental crust and subcrustal lithosphere is extended, leading to the development of rotated fault blocks at upper, brittle crustal levels and ductile stretching in the lower crust and subcrustal lithosphere (Figure 2*B*). The hallmark of stretching at surface levels is the

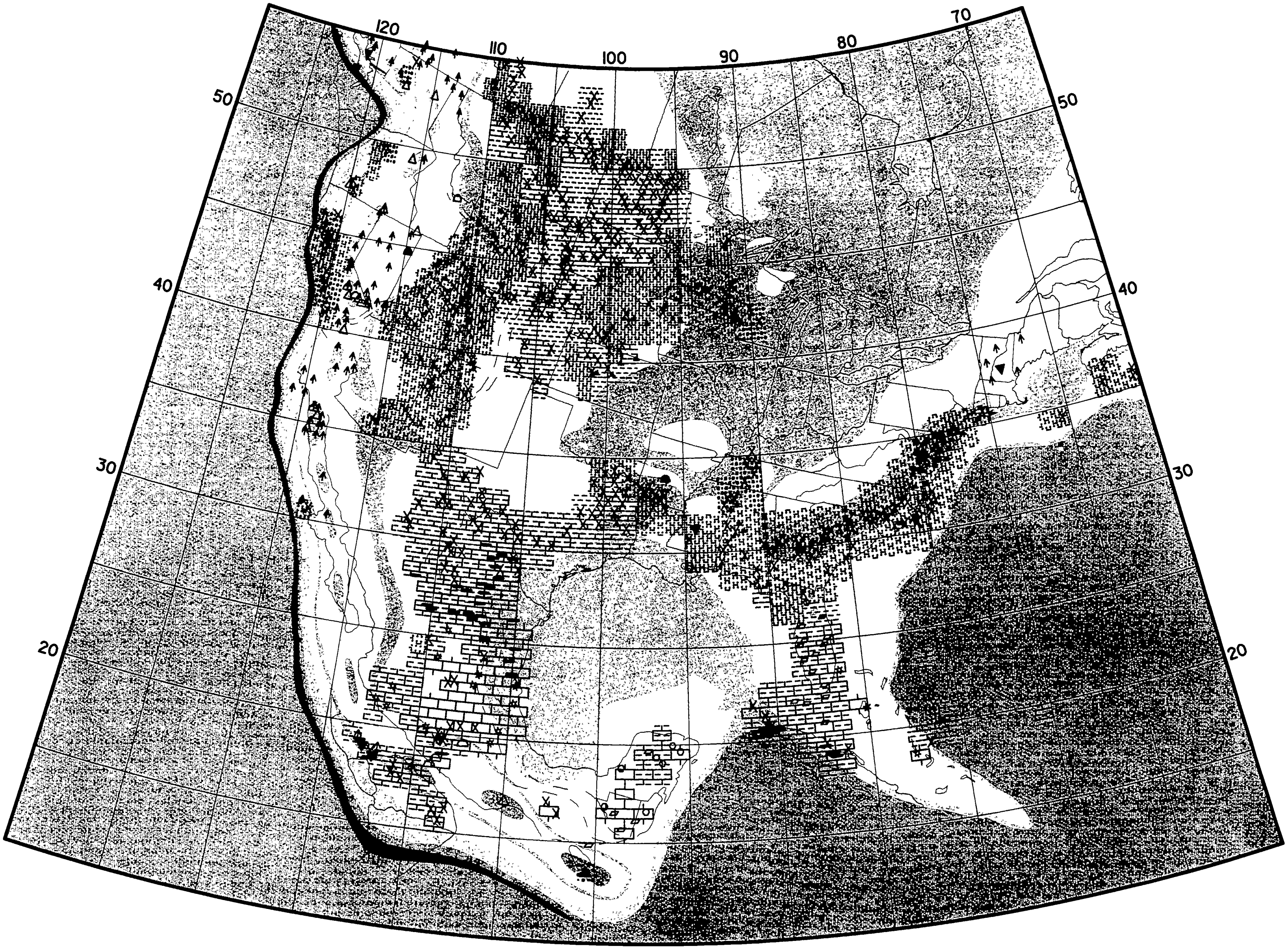


FIGURE 6 Cenomanian Paleogeography and Lithofacies

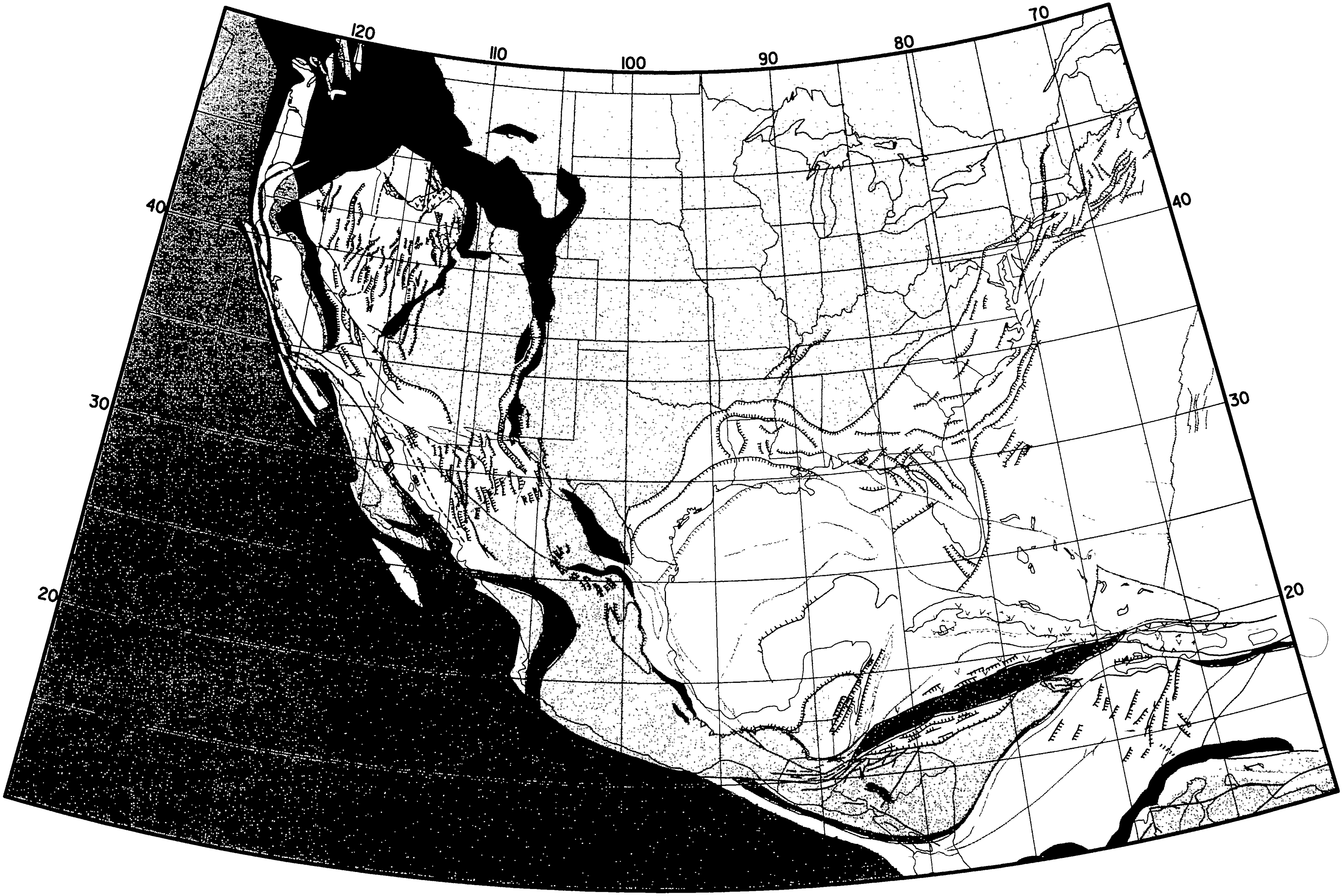
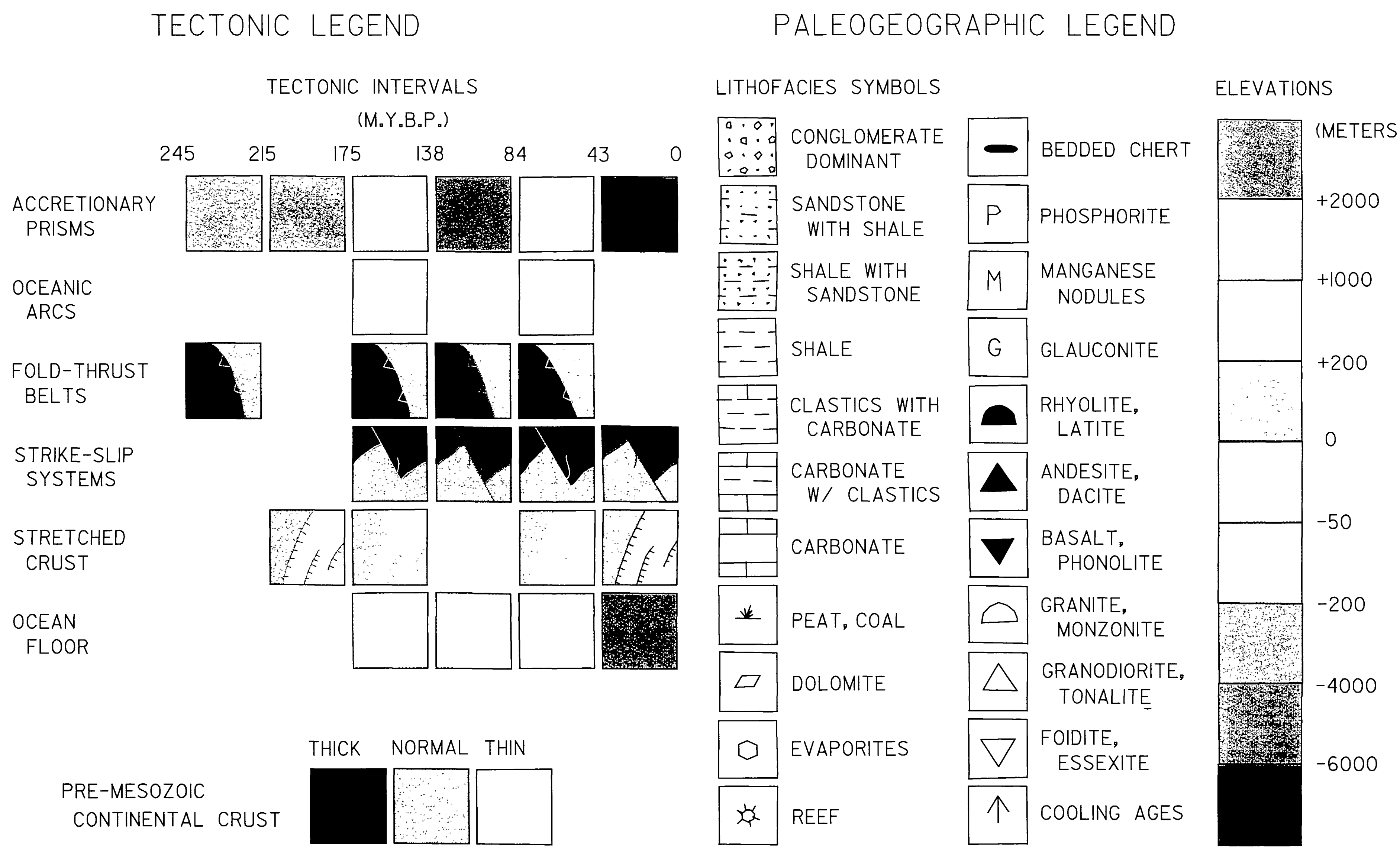


FIGURE 1 Mesozoic–Cenozoic Tectonic Map


TECTONIC LEGEND
TECTONIC INTERVALS
(M.Y.B.P.)
245
215
175
138
84
43
0
ACCRETIONARY PRISMS
OCEANIC ARCS
FOLD-THRUST BELTS
STRIKE-SLIP SYSTEMS
STRETCHED CRUST
OCEAN FLOOR
THICK
NORMAL
THIN
PRE-MESOZOIC CONTINENTAL CRUST
PALEOGEOGRAPHIC LEGEND
LITHOFACIES SYMBOLS
CONGLOMERATE DOMINANT
SANDSTONE WITH SHALE
SHALE WITH SANDSTONE
SHALE
CLASTICS WITH CARBONATE
CARBONATE W/ CLASTICS
CARBONATE
PEAT, COAL
DOLOMITE
EVAPORITES
REEF
BEDDED CHERT
P
PHOSPHORITE
M
MANGANESE NODULES
G
GLAUCONITE
RHYOLITE, LATITE
ANDESITE, DACITE
BASALT, PHONOLITE
GRANITE, MONZONITE
GRANODIORITE, TONALITE
FOIDITE, ESSEXITE
COOLING AGES
ELEVATIONS
(METERS)
+2000
+1000
+200
0
-50
-200
-4000
-6000

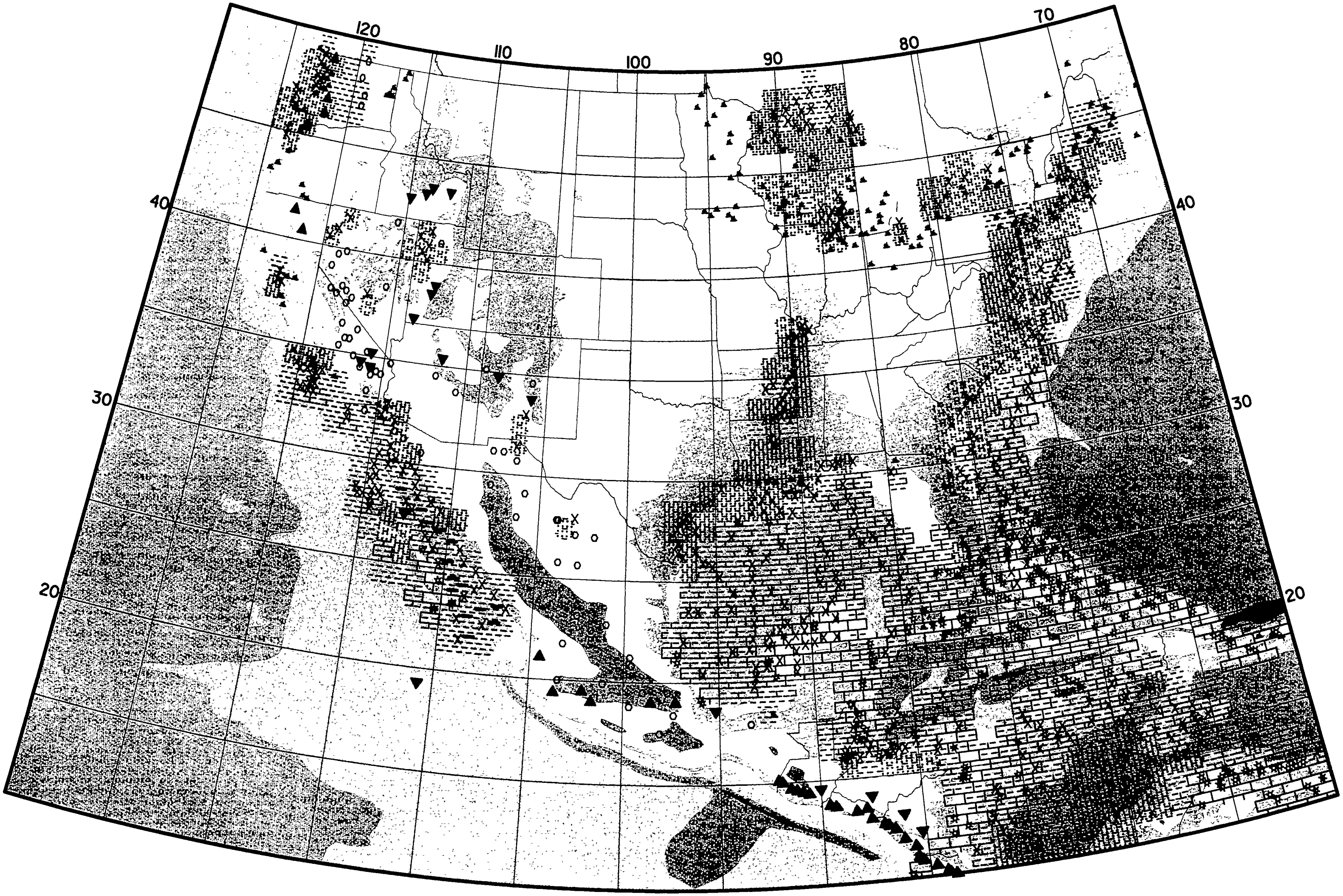


FIGURE 7 Recent Geography and Lithofacies

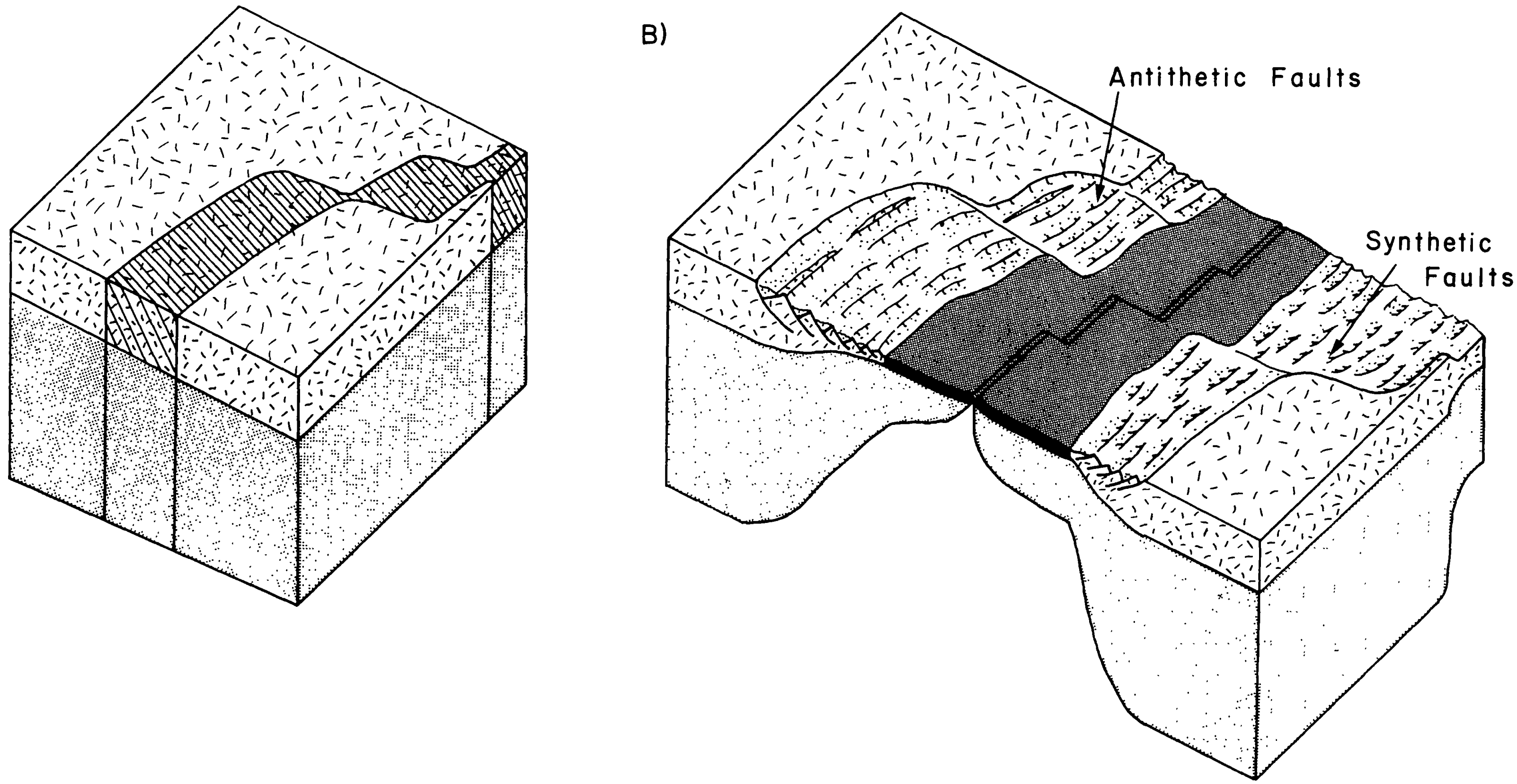


Figure 2 Lithospheric stretching and Atlantic-type margin development. *A* Initial condition of continental crust (random dashes) and subcrustal lithosphere (fine stipple) prior to stretching. Diagonal rules delimit the region to be stretched. *B* Continental crust thinned and stretched by an average factor of 2.5, resulting in pervasive normal faulting in the upper crust (with faults either synthetic or antithetic to the margin) and ductile flow in the lower crust and mantle. Crustal stretching ceases with initiation of seafloor spreading and generation of oceanic crust (black). The ridge axis need not be symmetrically situated within the stretched region. Reconstruction only of continent-ocean boundaries without palinspastically restoring the stretched continental crust to its initial state (diagonally ruled region in *A*) results in significant underfitting of the continents.

development of normal faults and associated basins that can occur as either synthetically or antithetically developed sets across the margin.

Until relatively recently, the geological and geophysical consequences of large-scale continental stretching were not fully appreciated. McKenzie (1978) presented a relatively simple model involving uniform, instantaneous stretching, which he showed to have the following consequences: (*a*) rapid synstretching subsidence where the initial ratio of crustal to lithospheric thickness (Cz/Lz) is greater than ~ 0.2, (*b*) surface uplift where Cz/Lz is less than ~ 0.2, (*c*) elevated geothermal gradients due to concomitant thinning of the thermal lithosphere, and (*d*) poststretching long-term thermal subsidence due to time cooling of the lithosphere. Subsequent work has examined the consequences of stretching over finite time periods (Jarvis & McKenzie 1980, Beaumont et al 1982, Cochran 1983a) and of nonuniform stretching (Royden & Keen 1980, Beaumont et al 1982), as well as the possible geologic and petrologic consequences of lithospheric stretching (Le Pichon & Sibuet 1981, Dewey 1982, Steckler & Watts 1981, Cochran 1981, 1983b, Vink 1982, Wernicke & Burchfiel 1982), but the basic relationships of McKenzie's initial model hold true.

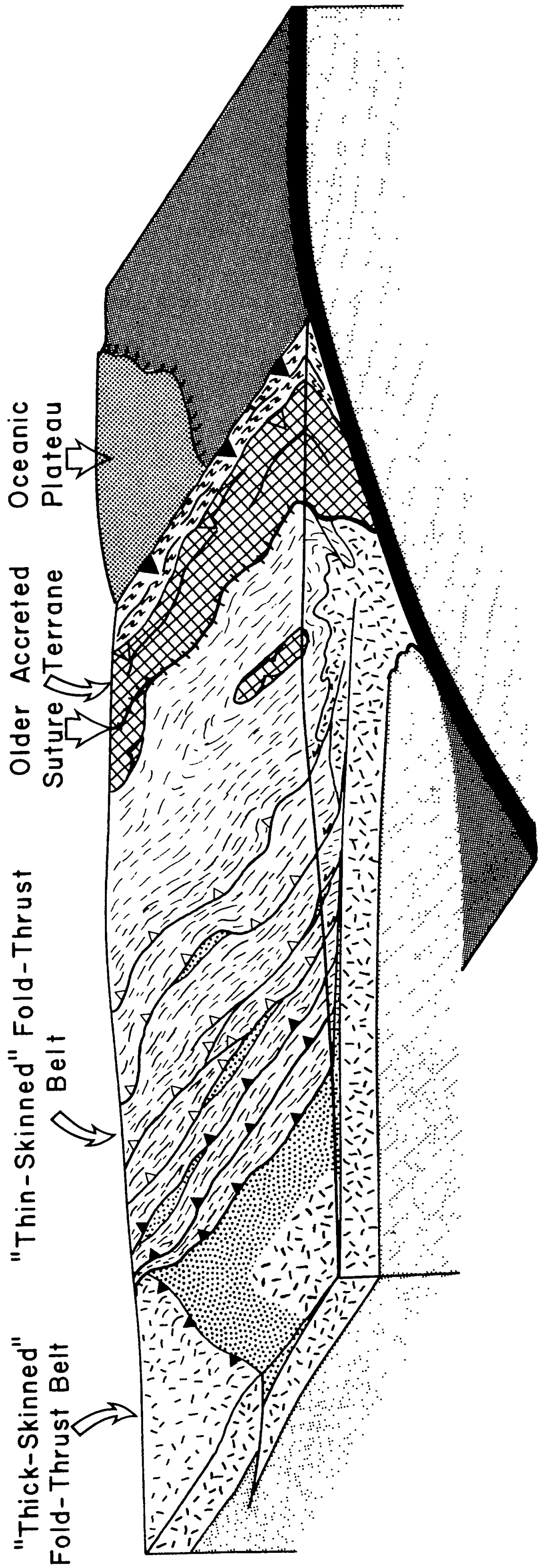
Crustal stretching has affected areas along the East Coast of the United States, the circum-Gulf of Mexico, and in the Basin and Range province (Figure 1). The amounts of stretching in these regions are difficult to determine precisely, but present estimates of average stretching factors range from $\beta = 2$ (100%) to $\beta = 3$ (150%) in all of them [East Coast (Keen 1981), Gulf Coast (J. L. Pindell, written communication), Basin and Range (Hamilton & Myers 1966, Hamilton 1978, Proffett 1977, Wernicke et al 1982)].

Palinspastically and paleogeographically, it is very important to delimit regions affected by crustal stretching. Along Atlantic-type margins, such as the East and Gulf coasts, stretching was succeeded by seafloor spreading. In the past, the basic premise of most continental reconstructions (Bullard et al 1965, Dietz & Holden 1970, Le Pichon & Fox 1971, Pitman & Talwani 1972) was that the best fit was one in which continental crust of conjugate margins did not overlap. Refitting of the continent-ocean boundaries shown on Figure 1 would not yield the original continental positions, because stretching has not been accounted for. Where regions affected by stretching are very wide (e.g. 100–300 km), this systematic underfitting can have a significant impact on reconstructions in other contiguous and noncontiguous regions. For example, misfitting North America, South America, and Africa in the Triassic has significant implications for the latitudinal position of Australia. In order to restore palinspastically regions that have been stretched, we defined the limits of stretched crust and compiled information on the thicknesses of the stretched and unstretched crust. The region that has been stretched can then be restored until its thickness equals that of the adjacent unstretched crust. The limits of stretched continental crust shown on the tectonic map define quite substantial areas that must be palinspastically restored. In regions with good multichannel seismic reflection data, the methods of Le Pichon & Sibuet (1981) and Wernicke & Burchfiel (1982) can be applied to determine stretching factors.

Basin and Range stretching postdates the Cenomanian, and therefore it has to be palinspastically restored for our Cenomanian reconstruction. Although no detailed fault-by-fault reconstructions of this region have yet been made, the general consensus is that about 100% of the essentially east-west extension has occurred in this region since the end of the Eocene (Hamilton & Myers 1966, Hamilton 1978, Wernicke et al 1982, Coney 1980). Although the thickness of the crust prior to Basin and Range extension is unknown, a twofold thinning of the crust seems likely (Wernicke et al 1982). The absence of significant rotation of the Sierra Nevada (Hannah & Verosub 1980) suggests that the amount of stretching in northern and southern parts of the Great Basin has been roughly the same. The extension directions derived from structural analysis of lineations developed in the regional detachment surfaces support an average east-west direction as well [e.g. ~S70°W in southwestern Arizona and California (Davis et al 1980), N70°W in the Snake Range (Miller et al 1983)]. We have therefore decreased the width of the present Basin and Range by approximately one half for our Cenomanian palinspastic restoration.

Fold-Thrust Belts

Fold-thrust belts as used in the tectonic map (Figure 1) refer specifically to those regions characterized by folding and/or thrusting that affect rocks derived from, or deposited on top of, a preexisting continental basement. Fold-thrust belts are thereby distinguished from other folded and thrust rocks that typify accretionary prisms and that were derived from, or deposited on top of, oceanic basement. Under this definition, fold-thrust belts generally form within collisional environments, such as the Himalayas, Appalachians, and Brooks Range, and within noncollisional back-arc regions of compressional ("Andean-type") arcs (Dewey 1980, Molnar & Atwater 1978, Dickinson 1979), including the sub-Andean ranges (Allmendinger et al 1983, Burchfiel 1980), and the Laramide and Sevier belts (Burchfiel & Davis 1975, Burchfiel 1980). Some fold-thrust belts, such as the Canadian Rockies (Monger et al 1982, Monger & Price 1979), result from a progression from collisional to back-arc environments. At present it is not clear what role the subduction of anomalously thick,


"Thick-Skinned" Fold-Thrust Belt
"Thin-Skinned" Fold-Thrust Belt
Older Suture
Older Accreted Terrane
Oceanic Plateau

A)

B)

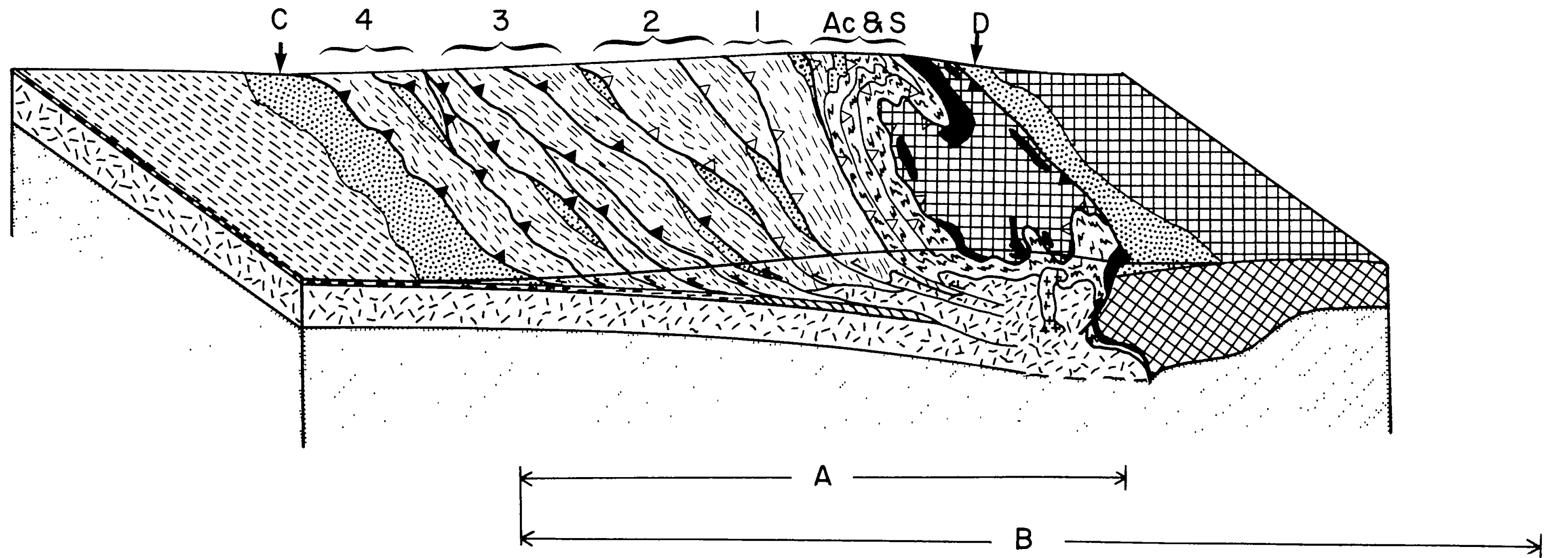


Figure 3 Fold-thrust belts. *A* Geometry of an Andean-type subduction zone with well-developed back-arc fold-thrust belts of "thin-skinned" and "thick-skinned" character. Shallow dip of Benioff Zone probably reflects subduction of either young oceanic lithosphere or an oceanic plateau. Thrust-loaded basins (coarse stipple) overlie depressed continental crust (random dashes). The continental crust is increasingly remobilized in the arcward direction. "Thin-skinned" thrust belt developed by progressive stepping out of thrust front. Previously accreted terrane (crosshatch) and accretionary prism also shown. Arc-related plutons and volcanics not shown. *B* Geometry of a collision-related fold-thrust belt. Collision of a continent (continental crust = random dashes, undeformed cover = dashed lines, deformed cover = wavy dashes) with an arc [arc and forearc basement = crosshatch, ophiolites = black, accretionary wedge = squiggly lines (Ac & S)]. Thrust-loaded basins (coarse stipple) develop in front of fold-thrust belt (C) and near the suture associated with back thrusting (retro-charriage) (D). Fold-thrust belt develops by outward stepping of thrust front from 1 to 4 with time. Sediments within fold-thrust belt change facies from onshore (4) to offshore (1). Crustal thickening within suture is associated with remobilization of basement and anatectic melting to produce syncollisional granites (pluses). Palinspastic restoration of continent side involves increase in width from present (A) to former width (B).

buoyant crust (such as oceanic plateaus and aseismic ridges) plays, but at least some reconstructions of the Laramide Orogeny relate it to subduction of the Farallon plate equivalent of the Hess Rise (Livacarri et al 1981, Henderson et al 1984).

The geometries within fold-thrust belts of different origins are comparable, but as yet only collisional fold-thrust belts have been shown to be associated with very large-scale (i.e. order of several hundred kilometers) crustal shortening (see, for example, Cook et al 1979, LeFort 1975).

The general geometry and structural development of fold-thrust belts of noncollisional and collisional origin are shown in Figures 3*A* and 3*B*, respectively. Fold-thrust belts are generally characterized by the following components in a hinterland direction: (*a*) A "foreland" or thrust-loaded basin, filled with terrestrial and/or marine sediments derived from the adjacent fold-thrust belt that results from flexure of the lithosphere adjacent to a thrust-related load. Evolution of these basins reflects the amount of loading, the flexural properties of the foreland lithosphere, and the rate of sedimentation versus sea-level rise or fall, as discussed by Beaumont (1981), Jordan (1981), and Dewey (1982), among others. (*b*) A frontal detachment in which previously deposited "foreland" basin deposits are commonly observed to be involved in the deformation. (*c*) The fold-thrust belt proper, which is characterized by (1) a temporal sequence of thrusts, such that structurally higher, more hinterland thrusts moved earlier and structurally lower, more foreland thrusts moved later; (2) a change in facies of the involved strata from generally platformal in the foreland to marginal in the hinterland; and (3) the increasing involvement and deformation of preexisting continental basement within the fold-thrust belt toward the hinterland. Laramide-style fold-thrust systems involve less shortening and generally bring basement to the surface immediately adjacent to the frontal belt. COCORP data suggest that the thrusts penetrate and offset the Moho and are not markedly listric (Smithson et al 1979). This contrasts quite strongly with typical "thin-skinned" fold-thrust belts.

The timing of thrust belt development is generally best constrained by ages of synorogenic flysch and molasse deposits. In addition, the ages of the youngest rocks involved in the deformation, as well as the isotopic ages of metamorphic minerals and syn- to posttectonic anatectic granites, also provide data on timing.

The methods generally used to restore palinspastically fold-thrust belts were outlined in classic papers by Dahlstrom (1969, 1970), in which the concept of "balanced cross sections" was introduced. The structural geometry of fold-thrust belts has been the focus of much recent study (Bally 1981, Boyer & Elliott 1982, Suppe 1983), in part owing to increasing interest in the hydrocarbon potential of overthrust environments. Figure 3*B* illustrates the importance of palinspastically restoring a fold-thrust belt. The present width of the fold-thrust belt is marked by A, whereas a width B is suggested for the predeformation state, based on approximate area balancing of the continental crust shown on the end section. Width A represents approximately 65% of the total shortening that occurred during two time intervals. The generation of intermediate palinspastic restorations is quite straightforward, since balancing involves undoing the deformation associated with each fold and thrust, starting in the foreland and progressing to the hinterland. Balanced (or retro-deformable; see Suppe 1983) cross sections of many North American orogenic belts have now been published, including the Canadian Rockies (Bally et al 1966, Price 1981), the Idaho-Wyoming fold-thrust belt (Dixon 1982), and the Valley and Ridge province of the Appalachians (Cook et al 1979). These, among others, provide critical tie-points for palinspastic restorations of these fold-thrust belts.

We have restored palinspastically the Laramide (i.e. Coniacian to middle Eocene) folding and thrusting from southern Mexico to the Canadian border, as well as a portion of the Sevier deformation in the same area. Laramide age deformation can be divided into two components: a generally eastern belt characterized by "thick-skinned" deformation, and a western "thin-skinned" deformation. The eastern belt of the Front Range Rockies, Black Hills, Big Horn Mountains, and Wind River Range appear to be associated with 10–30 km of east-west shortening and probably an equivalent amount of north-south-directed strike-slip motion (Hamilton 1978). We have moved regions affected by thick-skinned deformation back by about 40 km in a southwesterly direction. The western, thin-skinned belt is affected by apparently continuous activity from late Jurassic to middle Eocene times, and it is more difficult to determine what component reflects Cenomanian and younger deformation. Detailed studies in the Idaho-Wyoming fold-thrust belt (Wiltschko & Dorr 1983, Dixon 1982) suggest approximately 40 km of east-west motion. Comparable amounts of displacement appear to have occurred to the south, and therefore we have moved westward by approximately 40 km all regions to the west of the thin-skinned belt.

Other post-Cenomanian fold-thrust-related deformations are recorded in the Pacific Northwest, where imbrication along the Shuksan and Church Mountain thrusts record more than 60 km of east-west shortening (Misch 1966, Vance et al 1980, Monger 1977). In addition, approximately 50 km of northeast-southwest-directed shortening in South Florida and the Bahamas associated with the middle Eocene–Oligocene collision of Cuba and Hispaniola (Gealey 1980, Pindell & Dewey 1982) has been restored. This places the carbonate bank assemblages now observed on the northern margins of these islands back as a southward extension of the present Florida-Bahama platform.

Subduction-Accretion Prisms

Subduction of oceanic crust is often associated with the development of accretionary prisms (Karig 1974, Dickinson & Seely 1979), which, where well developed, represent substantial increases in the area of "continental" crust (Figure 4). Accretionary prisms are characterized by complex associations of trench-fill turbidites, upper and lower slope basin sediments, oceanic pelagic sediments usually of siliceous character, and various igneous and metamorphic rocks of ophiolitic affinities. These assemblages are primarily products of offscraping of rocks of trench or underthrust plate origin that accumulate along the leading edges of many subducting margins. Accretionary prisms are not ubiquitous elements of subduction zones, as was previously assumed, but they are now recognized as the result of interplay between several components, including the rate of sediment influx into the trench from the underriding and overriding plates, the presence or absence of seamounts or other topographic features on the underriding plate, and the age and relative motions of the overriding plate with respect to the subduction zone hinge line.

The rate of sedimentation within oceanic realms is controlled by biogenic productivity, which in turn relates to oceanic circulation patterns and/or proximity and access to continental sources, such as must have produced the Zodiak fan of the North Pacific. An important, but often ignored, factor that strongly controls the rate of sediment influx from the overriding plate is climate (in particular, rainfall). Ziegler et al (1981) pointed out that the absence of sediment fill within the Peru-Chile Trench closely parallels regions of the Andes characterized by low annual rainfall and thus by low net rate of sediment influx to the trench. Regions characterized by thick accretionary prisms form adjacent to river systems draining high mountainous terrain along those portions of continents that lie within areas of relatively high annual rainfall. This is certainly the case along the Oregon-Washington coast, the Chugach and Prince William terranes of southern Alaska, Barbados, Makhran, and Sumatra. Tectonic controls also play a role and have been discussed by Dewey (1980).

The ages of the accretionary prisms shown in Figure 1 indicate when the rocks were accreted; they generally do not reflect the ages of the rocks within the accretionary wedge. The dating of the time of accretion is often problematic. The best and most straightforward way is to date the trench-fill turbidites in each successive package, since the residence time of these sediments in the trench is generally quite short (< 3 m.y.). Unfortunately, trench-fill turbidites are often poorly fossiliferous, they can occur within

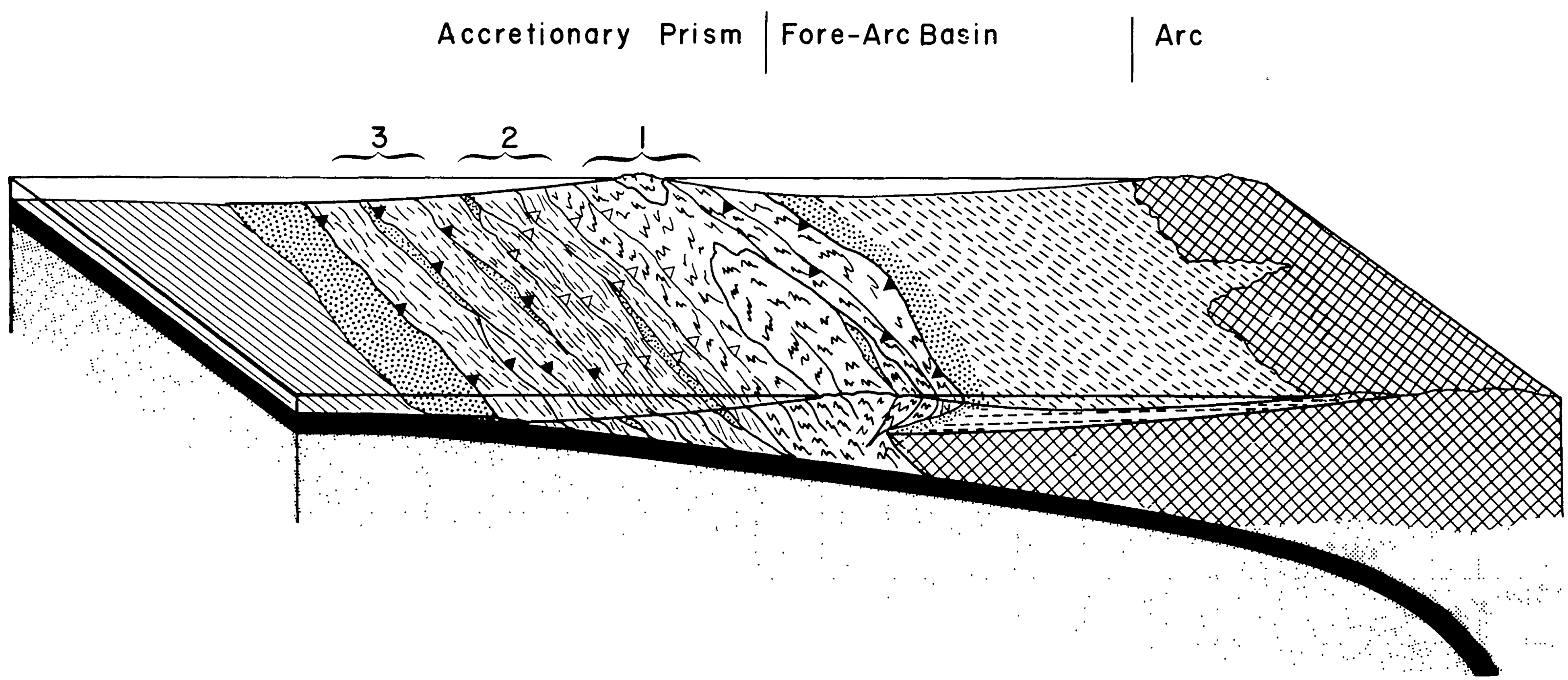


Figure 4 Accretionary prisms and Pacific-type margin development. Development of an accretionary prism (wavy and squiggly lines) by offscraping of trench-fill turbidites (coarse stipple), of oceanic pelagic deposits (ruled), and locally of fragments of ophiolite and seamount from the downgoing slab (oceanic crust = black, mantle = fine regular stipple). Growth of the accretionary prism occurs by progressive oceanward addition of wedge-shaped packages of offscraped material (numbers 1–3), schematically represented by oldest (1 = open triangles, squiggly lines), younger (2 = open triangles, wavy dashes), and youngest (3 = black triangles, wavy dashes). Upper and lower slope basins (fine irregular stipple) record continued motion on thrusts within the accretionary prism. In thick prisms, arcward thrusts (black triangles) emplace older accretionary prism rocks over ponded turbidites (stipple) of mixed forearc ridge and arc derivation and over forearc basin sediments (dashed lines); this process results in a raising of the forearc ridge. Crosshatch is arc and forearc basin basement.

completely disrupted mélange terranes where fossil ages are of uncertain meaning, and they cannot always be differentiated from turbidites within slope basins that postdate accretion. Where ages from the trench-fill turbidites are not available, alternative methods include (*a*) dating of the youngest oceanic pelagic sediments and/or sediments associated with accreted seamount complexes providing maximum ages, (*b*) isotopic dating of metamorphic rocks, such as blueschists, within the prism providing minimum ages, (*c*) isotopic age dating from near-trench igneous rocks providing minimum ages, and (*d*) extrapolation from ages of volcanism within adjacent arc terranes. None of these methods yield unequivocal ages, and whenever possible they have been used in conjunction to determine times of accretion.

Accretionary prisms are relatively easy to account for palinspastically, since the major components of large accretionary wedges are synaccretion trench-fill turbidites. In such cases the areas presently underlain by accretionary wedges that postdate the time of a given reconstruction are simply removed from the map. Thus, on the Cenomanian reconstruction we have removed the areas underlain by later Cretaceous and Tertiary accretion along the west coast of the United States, the Baja Peninsula, and the Middle America Trench. These sequences vary considerably along strike, ranging from regions characterized by thick basalts and associated reef carbonates, such as the Olympic Peninsula and farther south along the Oregon-Washington coast (Snavely et al 1968, Duncan 1982), to turbidite-dominated sequences of the Franciscan (McLaughlin et al 1982). DSDP/IPOD drilling results show that accretion along the Middle American Trench is quite limited (Moore et al 1982). In addition, the accretionary prisms exposed along the north coasts of Cuba and Hispaniola have also been removed. It is possible to make some inferences concerning the age of ocean floor being subducted at the time of the reconstruction if ages are available from offscraped oceanic pelagic sequences. This approach may be the only way to constrain ages of ocean floor in early Mesozoic and older times. More detailed palinspastic restorations are probably inappropriate at this time, since the nature of the subduction process generally leaves a sparse, but exceedingly complex, record of itself.

Strike-Slip Systems

Strictly speaking, strike-slip systems are associated with neither increase nor decrease of surface area. They are therefore the most straightforward tectonic elements to restore palinspastically. Some strike-slip systems, such as the San Andreas, define plate boundaries and generally parallel relative plate motion vectors. Other strike-slip faults define schollen (a crustal fragment bounded by faults) boundaries (Dewey & Sengör 1979) that accommodate various portions of relative motion, such as the faults associated with collision (McKenzie 1972, Dewey & Sengör 1979, Molnar & Tapponnier 1975), subduction (Fitch 1972, Dewey 1980), and transpressional strike-slip (Davis & Burchfiel 1973); these faults do not parallel relative plate motions.

Strike-slip faults commonly have segments that are oblique to the general fault trend. Such segments were described by Harland (1971) as transpressional and transtensional for compressional and extensional segments of strike-slip faults (Figure 5). Subsequently, Lowell (1972) and Harding & Lowell (1979) have shown that transpressional systems are commonly associated with upward-branching fault patterns that they term flower structures. Transtensional segments, on the other hand, are characterized by pull-apart basins, which are associated with crustal stretching (Mann et al 1983). When displacements are large, pull-apart basins can evolve into small ocean basins, such as the Gulf of California and the Andaman Sea.

Several important strike-slip systems are recognized within the area of Figure 1. These include the late Tertiary faults of the San Andreas system—the San Andreas (310 km), San Jacinto (24 km), and Elsinore (40 km) faults—and the Hosgri–San Gregorio (110 km) fault system of the western United States (Dickinson 1983). Earlier Tertiary and late Cretaceous faults include the proto–San Andreas (195 km; Dickinson 1983), as well as the Straight Creek–Fraser River system (150 km; Davis et al 1978) of the Pacific Northwest. The Ross Lake (160 km; Davis et al 1978) and Nacimiento (560 km; Dickinson 1983) faults represent earlier Cretaceous faults. Farther south, in Mexico and Central America, major strike-slip motion is associated with the Jurassic Mohave-Sonora megashear (~800 km)

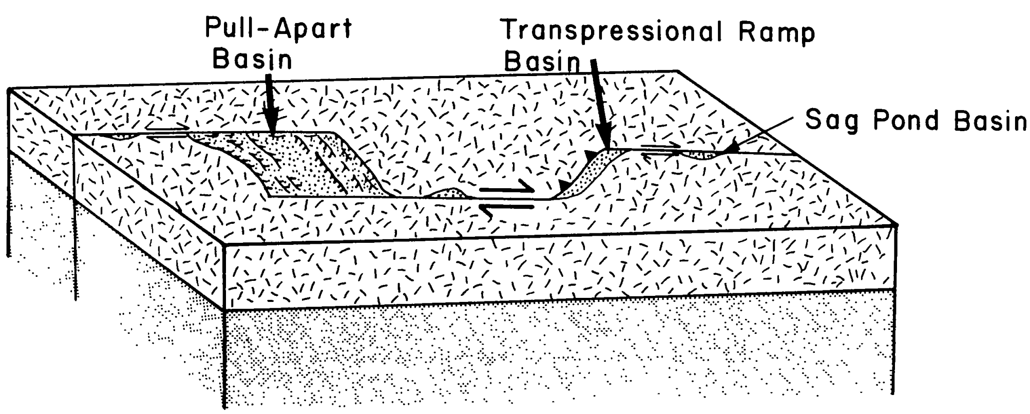


Figure 5 Strike-slip systems. Schematic representation of strike-slip system, with pull-apart basin, transpressional thrust and thrust-loaded basin, and local sag pond basins along the strike-slip fault.

trans-Mexican volcanic belt (~300 km), with latest Cretaceous(?) and younger motion along the Acapulco-Guatemala (~1300 km) fault (Anderson & Schmidt 1983) and its continuation in the Motagua system, and with offshore extension in the Swan transform (Pindell & Dewey 1982). Reconstruction of the Acapulco-Guatemala fault places the Chortis block back against southwestern Mexico and requires that the Middle America Trench evolve during migration of a trench-trench-transform system. Jurassic motion also occurred along the South Florida transform fault (Klitgord et al 1984). Motions on all of these faults, except the South Florida, Mohave-Sonora megashear, and Ross Lake faults, are reversed to produce the Cenomanian palinspastic restoration.

PALEOGEOGRAPHIC METHODOLOGY

Operationally, paleogeography consists basically of (*a*) dating rocks, (*b*) characterizing them environmentally, and (*c*) placing them in a paleolatitudinal and paleolongitudinal context. The first operation is concerned with paleontology, geochronology, and seismic and magnetic reversal stratigraphy; the second with paleoecology, sedimentology, and tectonics; and the third with paleomagnetism, paleoclimatology, and paleobiogeography. Operations (*a*) and (*c*), the "when" and "where" of paleogeography, are well known to most Earth scientists (if not, see Payton 1977, Harland et al 1982, McElhinny & Valencio 1981); thus, the emphasis in the following sections is on operation (*b*), the questions of "what" was deposited and "what" does it mean in terms of paleobathymetry and paleotopography. This discussion parallels and expands on that in Ziegler et al (1979, pp. 480–82).

Lithologic Data

There are 25 types of lithologies that are either common or significant in terms of depositional, climatic, or tectonic regimes and so are appropriate for regional paleogeographic studies (Table 1). For the Paleogeographic Atlas Project, the lithologies of approximately 38,000 localities world-wide have been assessed, and some 750 are displayed on our Cenomanian map (Figure 6) and 1100 on our Recent map (Figure 7). In our standard format, the five most common lithologies are recorded and listed in order of decreasing abundance. A computer printout of the data for Figures 6 and 7 is available from the authors on request and gives the country, state, latitude and longitude to the nearest tenth of a degree, lithologies, environment, formation, reliability of correlation at the stage level, and reference in terms of author, date, page, and locality number. A second printout is also available of the 234 references that were employed in compiling this data base.

The clastic-carbonate sediments (Table 1) are of course pervasive and are herein referred to as the "background lithologies." A rudimentary but

Table 1 Basic lithologic types appropriate for regional paleogeography[a]

Clastic-carbonate sediments	
(C)	Conglomerate
(S)	Sandstone
(M)	Mudstone, shale
(L)	Carbonate
Climatically significant sediments	
(T)	Tillite and glacio-marine beds
(P)	Peat, coal
(D)	Dolomite
(G)	Gypsum, anhydrite
(H)	Halite and bittern salts
(E)	Evaporites (G and H above)
(R)	Reefs
Oceanographically significant sediments	
(Q)	Bedded chert, radiolarite, diatomite
(V)	Phosphorite
(W)	Ferromanganese nodules and concretions
(X)	Limonite, goethite, or hematite
(Y)	Chamosite
(Z)	Glauconite
Soils	
(N)	Nonmarine, nondeposition
Acid-basic volcanic sequence	
(K)	Rhyolite, rhyodacite, trachyte, latite
(A)	Andesite, basaltic andesite, dacite
(B)	Basalt, phonolites, basanites, dolerite dikes
Acid-basic intrusive sequence	
(J)	Granite, monozonite, adamellite, alkali granite
(I)	Granodiorite, diorite, albitic granite, tonalite
(F)	Foidite, foyaite, essexite, theralite, etc
"Cooling ages" on intrusive and metamorphic rocks	
(U)	Uplift and unroofing

[a] Letters in parentheses are standard Paleogeographic Atlas Project codes.

effective clastic-carbonate index has been constructed from the relative abundance data of the background lithologies:

1. Conglomerate dominant
2. Sandstone with shale
3. Shale with sandstone
4. Shale
5. Clastics with some carbonate
6. Carbonate with some clastics
7. Pure carbonates.

It is this regrouping of the background lithologies that is portrayed on the maps in the form of computer-generated lithofacies patterns.

The computer has been programed to search each 1° latitude-longitude square for instances of the background lithology number code. When it finds a locality, it plots the appropriate lithology symbol, or averages the lithology if two or more localities are found in the same square. In either case, it centers the symbol on the 1° square, and in this way the symbols connect to form patterns where localities are abundant. The computer has also interpolated shading symbols into "empty" squares if there is data on background lithology in three or more of the eight surrounding squares.

The climatically significant accessory sediments are represented on the maps by discrete symbols, as are all the remaining lithologies. For simplicity, only the most abundant of these accessory lithologies is shown for each locality. Tillite is of course associated with mid- to high-latitude settings, but it is not shown in Figures 6 and 7 because there are no known Cretaceous glacial deposits, and because none of the 40 present glaciers in the United States and Mexico are greater than 5 km^2. It is doubtful that such deposits would survive in the geologic record, so they have not been plotted on our map. Coal and evaporites indicate opposite extremes of precipitation, or more accurately evapotranspiration (which combines precipitation and temperature). An analysis of these deposits through the Mesozoic and Cenozoic (Parrish et al 1982) shows that the coals consistently represent the equatorial and temperate rainy zones, and the evaporites the intermediate subtropical dry zones throughout this time span. Both are found in coastal and tectonically or glacially blocked drainage systems, and so they are useful as environmental and climatic indicators. The coals and peats in Figures 6 and 7 are north temperate coals, while the evaporites represent the north subtropical zone. Carbonates in general, and reefs in particular, were limited throughout the Mesozoic and Cenozoic to 40°N and 40°S as they are today; this zonation is thought to be due to the geometrical effect of light penetration to the seafloor as a function of latitude (Ziegler et al 1984). Deep-water reefs are known at present, but most reefs encountered in the geologic record appear to represent shallow water, based on the diversity of coral species and their association with algal-derived sediments, oolites, and other indicators of shallow conditions.

The oceanographically significant sediments, with the exception of chert, are classed as authigenic minerals and generally occur in trace amounts. It is important to note that the authigenic minerals are detectable mainly in areas of low background sedimentation (Emery 1969), particularly adjacent to desert belts, where there is low clastic influx, and in depths too great for carbonate production. Such conditions were enhanced during periods with rapid transgressions, such as the Cambrian and Cretaceous, when glauconite and phosphorite-rich sediments were generated over large regions. Phosphorite and chert are commonly linked with organic-rich sediments in outer-shelf settings and are likely to have been generated by oceanic upwelling systems of Ekman type (Parrish & Curtis 1982). The term "bioproductite" has been proposed for this compositionally diverse, but genetically linked, triumverate (Ziegler et al 1982a). In such settings it is usually reasonable to infer that water depths exceeded 50 m, because Ekman transport is limited to such conditions. However, dynamic upwelling, wherein deeper waters are forced over shoal areas by surface currents, can result in high surface productivity in shallow areas, so depth assignments based on the occurrence of bioproductites must be made with caution. Ferromanganese nodules occur over large tracts of the deep ocean today, characteristically in areas remote from terrigenous and biogenic source areas and also in lake deposits (Figure 7). Even so, they are rarely found in the rock record, largely because of its bias toward shelf settings. The iron minerals limonite, chamosite, and glauconite represent an onshore-offshore spectrum linked to the oxidation potential of these environments (Berner 1971). The presence of iron minerals, as with other authigenic minerals, is indicative of low background sedimentation, but it may also indicate complete chemical breakdown of soils in the source areas in specific climatic and topographic situations. The paleogeographic distribution of the iron minerals has yet to be analyzed in detail.

Soil types have been taken to indicate paleoclimatic conditions, but they are difficult to date precisely and so cannot be incorporated into paleogeographic data bases on a wide scale. Occasionally, soils can be dated by stratigraphic superposition and provide useful limitations on the position of the shoreline.

The igneous rocks are not as easy to classify as sedimentary rocks, because igneous processes are less efficient in compositional differentiation than surface processes. As a result of variation in nomenclatural usage in the world literature, we have grouped related igneous rock types into felsic, intermediate, and mafic classes (Table 1) based on the relative proportions of quartz, alkali feldspar, plagioclase, and feldspathoids (foids) (Streckeisen 1979). These groupings of the volcanic and intrusive rocks yield satisfying map patterns indicative of general plate tectonic environments. The intermediate types generally occur in linear belts representing the subduction-related igneous rocks. Landward, in the zone of crustal thickening and melting, there is generally a diffuse zone of felsic and some intermediate types. Basaltic compositions are found in most igneous environments, but they predominate, of course, in oceanic, hotspot, or continental rifting situations. The intrusive equivalents, the "foidites," are rare but do occur associated with continental rifts.

Large data bases of radiometric ages of igneous and metamorphic rocks are available in the published record, but for paleogeographic purposes most of these must be interpreted as "cooling ages." As York & Farquhar (1972) have pointed out, "When an area is involved in a prolonged cooling period, the various radioactive clocks will commence recording at widely separated times as the temperatures slowly fall from one critical blocking temperature to the next. Thus the Rb–Sr whole rock and U–Pb zircon clocks will begin to record the lapse of time before any other systems." We have used the Rb–Sr and U–Pb clocks to date the igneous intrusions, but we class the K–Ar ages as cooling ages, with the implication that this cooling takes place during uplift and unroofing of the orogen and therefore is indicative of mountainous terrains.

It is critical in paleogeography, as in other fields, to separate data from interpretation. In the foregoing paragraphs, lithologic types have been treated as the basic data of paleogeography, and comments on environmental interpretation have been restricted to lithologies that are confined to narrow environmental categories. Lithologies constitute data to the extent that the descriptions for a particular time interval are balanced and comprehensive (which, in the world literature, they often are not), and to the extent that the time limits of the interval are well established in the rock section (a more severe problem). Some rock types, such as sandstone and shale, are unrestrictive to environment, although sources may use terms such as greywacke, flysch, turbidite, or trench deposit that imply varying levels of environmental connotation. In this case, sandstone and shale constitute the "data," and the environment the "interpretation," as is explained below.

Environmental Interpretation

On our maps (Figures 6, 7) we employ a strict bathymetric/topographic contour representation of environments (Table 2). This is hazardous in view of the fact that the shoreline is the only contour directly interpretable from the geologic record, so it must be stressed at the outset that the other contours, as represented by the meter scale, should be treated as idealized guidelines that approximate the original contours. We feel that the contour system has the advantage of representational simplicity, and that the tectonic context is given by the paleotectonic maps and implied by the

Table 2 Elevation ranges of environments selected for regional paleogeographic maps

Code	Elevation (m)	Environments	Geologic recognition
	+10,000		
9		Collisional mountains	High-T, high-P metamorphics
	+4,000		
8		Andean-type peaks	Andesites/granodiorites in continental setting
	+2,000		
7		Island-arc peaks Rift shoulders	Andesites/granodiorites in marine setting Adjacent fanglomerates
	+1,000		
6		Inland plains Rift valleys Some forearc ridges	Between environments 5 and 7 Basalts, lake deposits in graben Tectonic mélanges
	+200		
5		Coastal plains Lower river systems Delta tops	Alluvial complexes Major floodplain complexes Swamps and channel sands
	0		
4		Inner shelves Reef-dammed shelves Delta fronts	Heterogeneous marine sediments Carbonates of Bahamian type Topset silts and sands
	−50		
3		Outer shelves Some epeiric basins Pro-deltas	Fine sediments, most "bioproductites" Fine clastics or carbonates Foreset silts and proximal turbidites
	−200		
2		Continental slope/rise Mid-ocean ridges Pro-delta fans	Slump/contourite facies Ocean crust less than 20 m.y. old Bottomset clays and distal turbidites
	−4,000		
1		Ocean floors	Pelagic sequences on ocean crust
	−6,000		
0		Ocean trenches	Turbidites on pelagic sequences
	−12,000		

contour configurations on the paleogeographic maps. A number of comprehensive texts relating sediments to environments are available (Davis 1983, Friedman & Sanders 1978, Reading 1978, Selley 1970), and thus the following discussion is restricted to the problems of environmental interpretation encountered in regional paleogeographic work.

The general elevation ranges of the environments selected for inclusion in Table 2 were determined simply by perusing a geographic map of the world. The ocean-floor categories (codes 0, 1, and mid-ocean ridges of 2) are reasonably straightforward. The well-known subsidence-time relationship of oceanic crust allows a depth assignment of 4 km for the ocean floor 20 m.y. after it was generated. Since the age of the ocean floor is now relatively well known (Sclater et al 1981), the widths of former ocean ridges can be inferred (Thiede 1982) in areas where the crust has not yet been subducted. Present-day trenches are reasonably well defined by the 6-km contour, although some older portions of the Pacific and South Atlantic ocean floor lie beneath this contour. Along subducting margins, the occurrence of turbidites can be used to define the trench contour, although it must be mentioned that turbidites cover large areas of shallower ocean floor adjacent to passive margins (North Atlantic) or to shear margins (western Canada). Oceanic trenches typically confine turbidity currents to movement parallel to the continental margin, so current-direction studies can be used to test the presence of a trench in the rock record. Locally, however, where the trench is filled with sediments (as along present-day coastal Oregon and Washington), this relationship may not hold true. By contrast, in areas of low sediment yield, such as Peru and Chile, the trench may be devoid of sediments. In summary, the best guide to the presence of a trench is the regional perspective of a subducting margin, rather than the presence or absence of turbidite deposits, which might in any case be subducted or overridden in the event of a continent-continent collision.

The continental slope and rise category is a distinctive but areally restricted and tectonically vulnerable environment. Cretaceous examples are known in the form of extensive reef talus deposits in Mexico (Enos 1983), clastic slump facies in Texas (Siemers 1978), and reef talus and mixed hemipelagic and carbonate turbidites in northern Cuba (Pardo 1975). The delta subdivision is shown in codes 2 through 5, but it can, of course, be restricted to the very shallow categories depending on basin morphology and crustal substructure.

The shelf environments (codes 3 and 4) provide the most abundant data sets for paleogeographic studies, and the depth order (and even the depth range) of environments can be determined from paleoecological and sedimentological studies. [See Kauffman (1967) for Cretaceous examples.] Indeed, the subdivision of shelf environments now possible is too fine to be portrayed adequately on regional scale maps. In general, inner-shelf environments are subject to wave and current activity, light penetration, and temperature and salinity variations, and these combine to yield a diversity of sedimentary rock types and biologic assemblages that contrasts sharply with the uniformity of the outer-shelf environments.

Epeiric basins constitute a special problem. These are created in a variety of ways, including (*a*) sea-level rise rapid enough to drown carbonate environments or create estuaries that trap clastic influx; (*b*) foreland thrusting that locally depresses the continental crust, as in the case of the west margin of the Western Interior Basin (Jordan 1981); (*c*) continental stretching and subsequent thermal subsidence of the lithosphere, seen today in the Gulf of Thailand; and (*d*) glacial scouring, as in the case of the present Hudson Bay. Depth interpretations of ancient epeiric basins are difficult because most environmental models are based on the marginal shelf seas characteristic of the present, and because the diversity of oceanographic parameters operating at these depths would vary in ways difficult to predict in enclosed basins. The thicknesses of prograding delta wedges (Asquith 1970), if corrected for stratigraphic and structural loading as well as compaction, could be used in the case of the Western Interior Seaway. Unfortunately, the loading corrections have yet to be applied, and published estimates of 600 m for the late Cretaceous of Wyoming are in our view excessively high.

The reconstruction of elevations in the terrestrial regions creates the biggest problem for the paleogeographer. The coastal plains and lower river valleys of the past do have a rich legacy in the form of swamp and alluvial plain complexes (Figure 6), and the fact that they have been transgressed from time to time bears testimony to their low elevations. The main problem comes with areas from 200 to 1000 m above sea level, an interval that today constitutes about 65% of the land surface (Cogley 1984, Figure 10). Much of this could be classified as inland plains, and such areas are generally not capable of leaving a geologic record of any sort. Many rift valleys, in their early stages, also lie at this elevation, as do the higher portions of forearc ridges. The Jurassic and Cretaceous Franciscan complex of coastal California provides an excellent example of a forearc ridge. It can be dated by K–Ar techniques on blueschist metamorphics (Armstrong & Suppe 1973), but its elevation, which increased through time, has been determined by its effects on sedimentation patterns in the forearc basin (Ingersoll 1979).

The mountainous categories (codes 7, 8, and 9) are recognized chiefly on the basis of igneous and metamorphic associations. In the case of the Andean and island-arc regions, the elevations refer to the andesite peaks and not to the associated mountain terrains that would generally be in the next lower elevation categories. The tectonic and sedimentary context of the andesites can be used to distinguish the Andean- from the island-arc-type mountains. The 4000-m contour was chosen to divide category 8 from 9 simply because most recent collision zones lie above this height. An ancient example would be the late Paleozoic Appalachian Mountains chain, which was characterized by high-temperature/high-pressure metamorphics.

MID-CRETACEOUS PALEOGEOGRAPHY

Paleogeography involves more than plotting localities with lithofacies data and environmental determinations on a palinspastically restored base map. Stratigraphic patterns and the prior geologic history of a region offer valuable clues in tracing environments across areas where the record has been obscured by erosion, deep burial, or metamorphism. Accordingly, we depart from the style of the earlier portions of this paper (in which we rigorously codified tectonic elements, rock types, and topographic intervals) and present an example of the Cenomanian paleogeography of the United States and Mesoamerica (Figure 6), with a Recent map also displayed for purposes of comparison (Figure 7). It is perhaps inherent in geology that while broad generalities can be established, each area is idiosyncratic, and thus perhaps as much can be learned from the local variations as from the areas that conform to expectations.

The Cenomanian sedimentary and igneous rocks of the United States and Mesoamerica are exceptionally well known and widely distributed, and they record a western active margin, an eastern passive margin, and span the north temperate and subtropical climatic zones. The same is true of the Recent of this region, although, of course, it contrasts sharply in its much lower sea level. Despite the wealth of data available on the Cenomanian of North and Central America, a detailed paleogeographic map has never been attempted for this region. An atlas of lithofacies maps for the whole continent is available (Cook & Bally 1975), as is an integrated lithofacies/paleogeographic atlas of the Rocky Mountain states (Mallory 1972). Other general sources include symposium volumes on the Mesozoic paleogeography of both the western (Howell & McDougall 1978) and west-central states (Reynolds & Dolly 1983), as well as one on mid-Cretaceous events for many parts of the world (Reyment & Thomel 1976). These sources, together with others mentioned in the regional sections that follow, were used in compiling our map. Before describing the regional aspects of the map, however, we must first discuss correlation problems and sea-level considerations.

Correlation Problems

Accurate correlations are essential in paleogeography, and despite the fact that Cretaceous biostratigraphy is exceptionally well refined, some problems remain. In the Western Interior Seaway, these involve the time at which the sea reached various areas during the Cenomanian transgression. The *Geologic Atlas of the Rocky Mountain Region* (McGookey et al 1972, Figure 26) shows marine conditions at the beginning of the Cenomanian as restricted to central Colorado and northern states, while subsequent work indicates that the seaway was established in western Oklahoma (Kauffman et al 1977) by this time. A similar problem exists with the terrestrial and marginal marine sequences of the Atlantic seaboard. Here, early correlations based on nonmarine ostracods have been improved on by palynological studies, and we accept Valentine's (1982, p. 26) correlation of the widespread subsurface unit F with the Cenomanian, rather than with the Albian and early Cenomanian as previously thought. As supporting evidence, we note that the markedly transgressive nature of unit F has a decidedly Cenomanian signature (see discussion below).

The correlation of the paleontological and radiometric time scales also presents a problem. We accept an 8-m.y. span from 94–102 m.y. for the Cenomanian, which was derived by applying corrections for the currently accepted K/Ar decay constants (Steiger & Jager 1978) to Van Hinte's (1978) time scale. Specialists on the Cretaceous (Kauffman 1977, pp. 85–87) assume a considerably narrower range based on bentonite age determinations from the Western Interior Seaway. We point out that such radiometric measurements are rarely accurate to 2% of the total age and are therefore not useful for establishing relative lengths of geologic stages. Strict adherence by Cretaceous workers to the bentonite ages has resulted in a time scale in which stages have extremely disproportionate lengths, and the Cenomanian has been one of the victims (down to 2.5 m.y. by some estimates). To judge by the number of fossil zones, by the general number and thickness of formations, and particularly by the magnitude of the sea-level rise (see below), the Cenomanian should be regarded as of at least average Cretaceous stage duration (6 m.y.).

Correlation problems even exist for the Recent time interval. We define the Recent by depositional processes that can be seen to be occurring today, or at least occurring in historic times. Because of the extremely rapid sea-level rise following the Pleistocene, sedimentation patterns in the marine realm are in the process of readjustment. Many shelf areas are not subject to sedimentation at the present (Emery 1968), so care has been taken to select information for our map that represents current sedimentation processes.

Sea-Level Considerations

The characteristics of the Cenomanian sea-level curve must be established for both paleogeographic and correlation purposes. Kauffman (1977, pp. 85, 89) has argued effectively that Cretaceous transgressive-regressive cycles of the Western Interior Seaway reflect eustatic fluctuations, and he shows a transgression extending from the beginning of the Cenomanian into the early part of the succeeding Turonian Stage. Indeed, in New Mexico and Arizona the shoreline, as represented by the Dakota Sandstone/Mancos Shale contact, advanced about 500 km, with three minor sandstone tongues interpreted to represent stillstands of the shoreline during this transgression (Molenaar 1983, p. 212).

A rough idea of the magnitude of the sea-level rise responsible for the transgression can be derived from stratal thickness and water depth estimates published on a locality in western Oklahoma (Kauffman et al 1977). This area is at least 500 km from the thrust loading of the western margin of the Western Interior Seaway and a similar distance from the thermal subsidence of the Gulf Coast. Cenomanian strata are approximately 50 m thick, and water depth estimates are zero at the beginning and 60 m at the end of the stage. If we assume an initial 50% porosity upon deposition (Sclater & Christie 1980, p. 3731) and a 10% porosity at present, the original thickness would have been 90 m. If we correct for sediment loading, the minimum sea-level rise required to deposit this amount of sediment would have been 38 m (assuming a mantle density of 3.3 g cm^{-3}, a sediment grain density of 2.7 g cm^{-3}, and isostatic equilibrium at all times). The depression of the seafloor due to water loading would be about 31% of the depth. Therefore, a water column of 60 m at the end of the Cenomanian would imply a sea-level rise of 42 m in addition to the 38 m referred to earlier, adding up to a total eustatic rise of 80 m during this stage. Pitman (1978, p. 1392) assumes that a maximum rate of sea-level change due to alteration of the geometry of the mid-oceanic ridge system would be about 10 m $m.y.^{-1}$, and our calculated sea-level rise is exactly this figure if an 8-m.y. duration of the Cenomanian is correct.

The absolute sea level at the end of the Cenomanian can be approximated from a locality in the Mesabi Range of Minnesota. This area, like the Oklahoma site, is well away from Cenomanian tectonic effects and, in addition, is remote from the late Tertiary Great Plains uplift. The value of this area for establishing Cretaceous sea-level heights has been recognized by others (Sleep 1976, Hancock & Kauffman 1979), although a recent review of the stratigraphy of the area (Merewether 1983) allows some refinement of earlier estimates. A mid-late Cenomanian "shallow-marine or brackish-water" fauna has been described by Merewether (1983, pp. 34, 44) from an horizon presently at 390 m above sea level. The "structural relief" on the top of the Cenomanian is only 90 m over a distance of 300 km across Minnesota, and this could be largely due to compaction of the thicker western sections as well as to the original bathymetric relief. As Sleep (1976, p. 53) has pointed out, the original surface has probably been raised about 50 m from its Cretaceous level as a result of erosional unloading of surrounding Precambrian uplands. A further correction appears necessary to adjust for glacial effects. The postglacial rebound model of Walcott (1970, 1972) shows that Minnesota is located just south of the pivot marking the northern extent of the peripheral bulge, an uplifted ring around a glacier due to the flexural rigidity of the lithosphere. Free-air gravity data support this model, as north-central Minnesota has a regionally averaged positive anomaly of about 10 mgal (Bowin et al 1982), which would suggest that the area still has about 70 m to subside before reaching isostatic equilibrium (Walcott 1970, Figure 4). For a contrasting view, see Officer & Drake (1981, Figure 8), who published a map of "probable vertical movements" suggesting that Minnesota is being uplifted at a rate of 1 cm yr^{-1}. Applying our corrections, a figure of 270 m is indicated for the late Cenomanian sea-level high stand. The only other North American site useful for testing this figure lies to the east of the Mississippi Embayment in central Tennessee, where Cenomanian rocks of purported coastal origin presently lie at elevations ranging from 275 to 305 m (Marcher & Stearns 1962).

In summary, the Cenomanian seas rose approximately 80 m from an initial elevation of 190 m to 270 m above present sea level during an 8-m.y. interval. Our map (Figure 6) shows the paleogeography at the end of the stage, while the lithofacies components are summarized over the whole stage. The 50-m bathymetric contour corresponds approximately to the shoreline position at the beginning of the Cenomanian.

Western Interior Seaway

Foreland folding and thrusting along the western margin of the continent was well developed by Cretaceous times, and it served to produce and pond sediments and to load tectonically the continental crust of the Western Interior. By the Albian, the sea level was sufficiently high to result in transgression, and vast areas of the Rocky Mountain and Great Plain states were flooded. East of the thrust belt, the relief must have been extremely low, as these regions had been both tectonically quiescent and subject to deposition since Pennsylvanian time. However, the Ancestral Rockies seem to have remained a source area as late as Albian times, to judge by current directions in the braided stream deposits of the Dakota Sandstone of New Mexico (Gilbert & Asquith 1976).

Renewed transgression in the Cenomanian extended the seaway eastward across portions of Minnesota, Iowa and Kansas. A relief of over 400 m existed in the Precambrian terranes of Minnesota where Cenomanian deposits are preserved in valleys flooded at this time. Our shoreline and 200-m contours in Minnesota are based on Sloan's (1964) contours on the Precambrian bedrock surface and his assumption that this area has not been significantly warped since the Cretaceous; we have also extended this method to adjacent states. Unfortunately, the deposits representing the late Cenomanian shoreline have been stripped by erosion in states farther south. The coastline here is unlikely to have been as irregular as that in Minnesota, as the whole region is underlain by relatively uniform late Paleozoic carbonates.

The active western margin of the seaway was more complex. In southwest Wyoming, steep gradients are evident (de Chadenedes 1975), as might be expected along a thrust margin. Immediately north, however, the great Sheridan delta (Goodell 1962) seems to have prograded across the axis of subsidence to areas of west-central Wyoming. By the late Cenomanian, the sea-level rise evidently exceeded the sedimentation rate, since marine deposits overlie paralic deposits in both areas. To the south, in Utah, Arizona, and New Mexico, the effects of thrusting disappear, and the Cenomanian is represented by a very extensive, uniformly thin ($\sim$100 m) transgressive sheet of Dakota Sandstone and Mancos Shale (Molenaar 1983).

The area of the greatest uncertainty is the nature of the connections of the seaway with the Gulf of Mexico. Cenomanian deposits have been stripped from a large area in central Texas, although Albian shallow-marine deposits do occur. We assume from the magnitude of the Cenomanian transgression that a connection across this area would have existed. The Coahila platform of northeast Mexico seems to have been relatively stable, but to the west there evidently was a deeper water connection of the Western Interior Seaway through the rapidly subsiding Chihuahua Trough (Greenwood et al 1977, Hayes 1970, Powell 1965).

Ancestral Gulf of Mexico

The Gulf had opened as a small ocean basin in the Jurassic, and by the end of the early Cretaceous, its shelf margin was well defined by a nearly continuous reef tract (Martin & Case 1975). Portions of this reef-dammed bank in Florida and the Bahamas were drowned during the Cenomanian, resulting in the rather more dissected banks that have survived to the present (Sheridan et al 1981). The same effect was probably responsible for the extinction of the reefs in many parts of the Gulf, although few detailed paleoecological studies are available for Cenomanian sediments overlying the reef tract. An exception is the study of Mancini (1977) on the Woodbine and Eagle Fork sediments of southern Texas, which are interpreted to have been deposited in inner to middle neritic conditions. Also, along the Florida/Georgia border, dark carbonaceous silty shales of the Lower Atkinson Formation yield pelagic foraminifera (Applin & Applin 1967) and probably indicate the inception of a channel along the Suwanee basin connecting the Gulf and the Atlantic that persisted as a bathymetric feature into Tertiary times (McKinney 1984). This channel was bounded on the south by the Ocala arch, which is incompletely covered by Cenomanian clastic strata and evidently was a shoal area with some islands at this time.

The northern margin of the Gulf was complicated by local tectonic and igneous activity that began about Cenomanian time. The Sabine uplift of eastern Texas was active and helped to define a small gulf to the west where deltas formed; this area later became the petroleum reservoir rocks of the Woodbine Formation (Oliver 1971). On the southern margin of the Sabine uplift, a narrow shelf with steep gradients resulted, along which slumping occurred (Siemers 1978). The nature of the uplift is unclear to us, although igneous activity is manifest in the Monroe uplift of northeast Louisiana and the Jackson dome of west-central Mississippi (Murray 1961, pp. 116, 340) and must have been contemporaneous, to judge by the amount of tuffaceous material in the Cenomanian sediments of the area. These rocks are part of a predominantly later Cretaceous alkalic igneous province that extends along the Mississippi Embayment and parallels the inner margin of the Gulf Coastal Plain through Arkansas, Texas, and northern Mexico (Baldwin & Adams 1971). However, the Mississippi Embayment postdates the Cenomanian and was the site of the Pascola arch, an extension of the Ozark dome, at this time. Marcher & Stearns (1962) reconstructed the Cenomanian topography of this arch and inferred a narrow arm of the sea in central Tennessee.

The Cretaceous paleogeography of Mexico has been mapped in detail (Enos 1983). Here, basinal carbonates as well as rudistid-dominated reef platforms were subsequently involved in Laramide convergence and are thus available for study in surface sections. Some palinspastic restoration is shown on our map. Enos points out that "mid-Cretaceous platforms differ from Lower Cretaceous ones in that vertical accretion greatly predominated over lateral progradation, resulting in steeper forereef slopes and much higher depositional relief in the mid-Cretaceous." He concludes that a relief of 1000 m was established in some areas because great aprons of shallow water debris extend 10 to 20 km from the reef tracts. The basinal deposits consist of "dark-gray lime mudstone and wackestone with pelagic microfossils," and bedded chert is present in some areas.

The basinal deposits of the northern Gulf have been prograded by thousands of meters of Tertiary sediments and have not been penetrated by drilling. West of the Florida platform, however, they have been drilled and are described as "laminated, banded and bioturbated limestone with coarse bioclastic layers" (Schlager et al 1984). Seismic profiles indicate that they were deposited in a depth range of 1200 to 1500 m. To the southeast, in northern Cuba, Cretaceous platform, slope or scarp, and pelagic facies were thrust over the Bahamas platform in early Tertiary times (Pardo 1975). Steep slopes are implied by reef-derived talus in the slope deposits, while bedded chert occurs in the carbonates of the pelagic facies.

Atlantic Seaboard

The Atlantic margin history was similar to that of the Gulf in that rifting began in the late Triassic, drifting began in the mid-Jurassic, and thermal subsidence was well underway by the mid-Cretaceous. Like the Gulf, the entire margin up to Nova Scotia was probably reef dammed through the early Cretaceous (Austin et al 1980), with the reef tract having developed at the outer limit of stretched crust (Hutchinson et al 1982). Inundation of this reef tract during the Cenomanian transgression was evidently complete north of the Bahamas.

The position of the late Cenomanian coastline can be traced from the subsurface of the central Atlantic states to New Jersey (Sohl et al 1976, Figure 8), where it trends across the present shore to the Scotian Shelf (Sherwin 1973, Figure 15). It is well bracketed by coal deposits representing the coastal plain and shallow marine shelf deposits, and it lies about 500 m below present sea level. Assuming our sea-level datum of 270 m for the Cenomanian to be correct, one might expect that the coast would transgress the unstretched portion of the crust, including the foothills of the Appalachian Mountains. The fact that this did not happen indicates that the Appalachians have subsided in excess of 70 m since this time. We assume that the Cenomanian 200-m topographic contour would approximate the inner limit of stretched crust, and if so it would generally coincide with the present 200-m contour, except in northern New England. This contour is slightly inland of the preserved coastal plain deposits.

Igneous activity affected New England during the Mesozoic, and some authors have related this activity to the passage of the hotspot that generated the New England Seamount chain (Vogt & Tucholke 1979). This hotspot cleared the shelf margin by Cenomanian times, but late Cretaceous fission track ages from northern New England (Zimmermann et al 1975) indicate that uplift continued through this time. Accordingly, we show some highlands on our Cenomanian map.

Pacific Margin

This is the area of the map most fragmented by strike-slip faulting. (See the section on "Palinspastic Restoration.") Essentially, a once-continuous

Andean subduction accretion complex has been transported northwest from northern Mexico and the United States as part of the Pacific plate, and southeast from southern Mexico as part of the Caribbean plate. The best-preserved and most studied area is in central California, where the arc, forearc basin, and trench deposits are preserved (Ingersoll 1979). A continuous Mesozoic record of convergence is preserved here, and by mid-Cretaceous times, a submarine forearc ridge had developed and served to pond the Great Valley forearc sediments.

The arc itself is the most pervasively preserved element along the various segments, but erosion has generally stripped the surface volcanics, and only the granodiorite intrusions are left to mark its general trend. Problems associated with interpreting the various types of radiometric ages on these rocks in California and the Baja Peninsula have been summarized by Krummenacher et al (1975). The trend in the Pacific Northwest (Armstrong et al 1977) is partly obscured by Tertiary volcanism and is distorted by Basin and Range extension. The trend in Central America is east-southeast (Weyl 1980, p. 299), i.e. subparallel to the present southeast-trending arc. Our radiometric data come from these papers and from many articles in the journal *Isochron/West.*

Forearc basin sediments of Cenomanian age have been identified with confidence only in California and Oregon (Dickinson & Thayer 1978), and the forearc ridge can be recognized by radiometric ages on blueschist rocks from California and the Baja Peninsula (Suppe & Armstrong 1972). True oceanic trench deposits of this age are not known with certainty.

The Methow-Tyaughton Trough of northern Washington and adjacent British Columbia was a forearc basin in Jurassic–early Cretaceous times, but as a result of the collision of outboard terranes about the mid-Cretaceous, it became, for a time, a narrow marine inlet (Tennyson & Cole 1978). This trough is the only detected reentrant in the late Cretaceous Andean arc.

Discussion and Conclusions

The changes in paleogeography over the past 100 m.y. (Figures 6 and 7) result from the counterclockwise rotation of the North American continent, the tectonic deformation and volcanic constructs of the western and Caribbean portions of the continent, the considerably lower sea levels observed today, and the effects of erosion and sedimentation. The rotation of the continent has meant that areas east of the Mississippi are slightly higher in latitude today, while the opposite is true in western areas. The net effect is not great, and the 35°N latitude line on both maps generally divides sediments of subtropical aspect from those of the temperate zone.

The entire western portion of the continent has been remolded by the intracontinental Laramide Orogeny, followed by Basin and Range extension, and finally by strike-slip fragmentation of the western margin. These three major effects combined to produce by late Tertiary times a much more mountainous and irregular margin than in the Cenomanian, when a relatively well-integrated Andean-type margin existed. The motion of many of the faults shown on our tectonic map (Figure 1) is small in scale, but their cumulative effect is considerable, and this serves to demonstrate the need in paleogeography for palinspastic restorations. In addition, this level of structural analysis is essential in understanding the character and evolution of sedimentary basins.

The effect of sea-level change is best observed in the central portion of North America. The hypsography of the continent is profoundly different as a result of this change, with the lowland environments having gained at the expense of the shallow sea since the Cenomanian. Modification of bathymetry by sedimentation is most profound along the northwestern coast of the Gulf of Mexico, where the shelf margin has prograded about 300 km since the Cenomanian (Winker 1982) as a result of the drainage of the central part of the continent by the Mississippi and other river systems. By contrast, shallow-water carbonate areas drowned by the Cenomanian transgression or moved northward out of the subtropical belt by the rotation of the continent are today deeper as the result of continued subsidence and low sedimentation rates. Finally, erosion has played an important but poorly understood role (except for glacial scouring, which created the Great Lakes, many peat bogs, and the irregular topography of both the New England and Washington State coastal and offshore regions).

The level of detail possible in paleogeographic reconstructions, as demonstrated in this review, will provide the framework for answering a number of important geologic questions. By tracing the elevations of shorelines of the past, we can examine the long-term warping of the relatively stable cratons. Can observed variations be explained by isostatic effects, or do they instead reflect deeper mantle processes? By linking tectonic and stratigraphic approaches, we can better understand the character and evolution of sedimentary basins. Did a basin result from thermal subsidence, tectonic loading, a transtensional system, or just a sea-level rise? By mapping the elevations of past mountain chains, we can better estimate their orographic effects on atmospheric circulation patterns. How does this translate into rainfall and its influence on peat formation, or into oceanic currents and their influence on upwelling, organic productivity, and oil source rock generation? Finally, by improved paleogeographies, we can understand the context in which organic evolution took place. Did a taxon evolve in response to environmental or climatic change, to isolation in a constantly changing world, or simply to selection pressures imposed by other biotic elements? These questions are just a sampler, and they indicate the range of problems that can be addressed when accurate paleogeographic maps are prepared.

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