Plate-tectonic evolution of the Gulf of Mexico and Caribbean region

James Lawrence Pindell

FRONTISPIECE

"The number of attempts to synthesize the tectonic framework of the Caribbean are infinite, as are the number of different frameworks that have been suggested...In terms of the plate tectonic 'revolution' in the earth sciences, it would be very much preferable if the Caribbean area and the Bahamas did not exist."

—F. Nagle [1970, pp. 413-414]

Westward looking view of Pico Isabella des Torres, 2 km south of Puerto Plata, northern Dominican Republic. Curious silver clouds cap the 800 meter peak at the close of nearly every afternoon. On his first voyage to the new world, in early January of 1493, Christopher Colombus dubbed the small embayment below the peak as Puerto Plata, or Silver Port, referring to the mystical clouds above. Much has changed in Hispaniola since that time, but the silver clouds, and the mountain beneath them, remain the same.

PLATE-TECTONIC EVOLUTION OF THE GULF OF MEXICO

AND CARIBBEAN REGION

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A thesis submitted for the degree of

Doctor of Philosophy

at the University of Durham

Faculty of Science

Department of Geological Sciences

May, 1985

PLATE-TECTONIC EVOLUTION OF THE GULF OF MEXICO AND CARIBBEAN REGION

-abstract-

James Lawrence Pindell

A geologic-kinematic model for the evolution of the Gulf of Mexico and Caribbean region is built within a framework provided by a detailed Late Paleozoic (Alleghenian) plate reconstruction and a revised North American (NOAM) and South American (SOAM) relative motion history. From the Middle Jurassic to the Campanian, SOAM migrated east-southeast from NOAM. From the Campanian to the Eocene, little or no NOAM-SOAM relative motion occurred, although minor sinistral transpression is suggested. Since the Eocene, minor west-northwest convergence between NOAM and SOAM has occurred along pre-existing fracture zones. Three stages of evolution are recognized which correlate with these phases of relative motion. Stage 1: mainly carbonate shelves fringed the Gulf of Mexico and "Proto-Caribbean" passive rifted margins, during plate separation. Stage 2: the Caribbean Plate (CARIB) progressively entered the NOAM-SOAM gap from the Pacific by subduction of Proto-Caribbean crust beneath the Greater Antilles. Stage 3: CARIB migrated east by 1200 km, subducting Proto-Caribbean crust and forming the Lesser Antilles Arc. Transform faults have dissected the original Greater Antilles Arc, and nappes in the Venezuelan Andes have been emplaced southeastwards onto the northern SOAM margin, diachronously from west to east. Field work done in Dominican Republic, both near Puerto Plata and in the southwest sector, indicates that 1) Cuba and northern and central Hispaniola are parts of one original Greater Antilles arc, 2) this arc collided with the Bahamas in the Late Paleocene-Mid Eocene, and 3) Hispaniola has been assembled by strike-slip juxtaposition of terranes from the west.

ACKNOWLEDGMENTS:

I wish to express my deepest gratitude to my thesis supervisor, Professor John F. Dewey. John has granted me an "endless leash" concerning the nature of this thesis, entrusting me to study any and all of those geological aspects that help to constrain models of Gulf and Caribbean evolution, and assisting me so much with them. He has helped to support me at critical times, both personally and financially. And above all, he has given me the fortunate chance, at least (!), to obtain a firm understanding of the way in which the earth works.

Sincere appreciation and thanks are due also to Professor Walter Pitman of Lamont-Doherty Geological Observatory of Columbia University. Walter provided access to Lamont's magnetic data files, showed me how to "pick" seafloor-spreading magnetic anomalies, and taught me a lot about the world.

Field-related portions of this thesis are strengthened by the help of Dr. Edward Robinson and Dr. Pavel Cepek, who dated and provided faunal lists for many of the rock samples collected across Hispaniola.

Kevin Burke, Calvin Cooper, Jim Dolan, Grenville Draper, Gene Lane, Sharon Poissant and Steve Tanski have spent various amounts of time with me in the field throughout Hispaniola, keeping me "honest," helping me to see the not-so-obvious, and giving me often-needed companionship. In addition, Dr. and Mrs. Abraham Risk of Barahona kindly provided friendship and a comfortable place to stay at various times, and Jose De Coux of Sosua did the same and was invaluable for his endless supply of "Los Cuba Libres, Cervezas, y 100% beefburgers."

I acknowledge the kind support of the Mineria of the Dominican Government. Dr. Victor Beras and associates provided access to maps and other logistical support. Also, I am very grateful to the people of the Dominican Republic, particularly around Los Patos in the southwest and near Sosua and Puerto Plata in the north. I truly respected their kindness, genuine interest, and cheerful help during the field seasons.

I thank Michelle Kominz and Maxine Chute for initially teaching me how to use Lamont-Doherty's computer facilities.

Completion of this thesis has been greatly facilitated by the help of several friends and staff at the University of Durham. Particularly, I thank Sylvia Hutton, Lucy Mines, Carol Blair, Jerry Dresser, John Neilson, George Randall, and Andy Reid.

I am indebted to all of the following, from whom I have learned so much about general geology, the Gulf of Mexico and the Caribbean: Steve Barrett, Dwight Bradley, Kevin "Brownian motion" Brown, Dick Buffler, Kevin Burke, John Cagle, Steve Cande, Jim Case, Calvin "Klein" Cooper, James Dolan, Nick Donnelly, Grenville "Hombre Peligroso del Norte" Draper, Robert Erlich, Charles Evans, Stuart Hall, Donny "little red man" Hutton, Kent Johnson, James "Rico" Joyce, Garry "G. D." Karner, Bill Kidd, Kim Klitgord, John Ladd, Stuart "lack" Lake, John LaBrecque, David "stoke" Larue, Paul Mann, Florentine Maurrasse, Chad McCabe, Winthrop Means, Akio Miyashiro, Clyde Moore, Tim "Mundane" Munday, Frederick "of Hollywood" Nagle, George Putman, Eric Rosencrantz, David Rowley, Charlotte Schreiber, Dietmar Schumacher, Celal "Tectonic Turk" Sengor, Robert Spoelhof, Fred Taylor, Carol Tessier, Graham Westbrook, and Stan White.

Support for the project came from the following sources, to which I am truly grateful: Pennzoil Exploration Company; Amoco Production Company, International; British Petroleum Company; Phillips Petroleum Company; Robertson Research, Inc.; and an Overseas Research Student Award from British Universities.

Finally, but certainly not least, I thank my parents and my brother and sister and their families for standing by me in moments of indecision about the future. Also, I am very grateful to my parents for granting open credit to the PBA (Parent's Bank of America), whose interest-free loans have helped to make life a lot easier!

TABLE OF CONTENTS

TABLE OF FIGURES

CHAPTER 1

CHAPTER 2

CHAPTER 3

CHAPTER 4

CHAPTER 5

CHAPTER 6

CHAPTER 7

CHAPTER 8

TABLE OF TABLES

INTRODUCTION:

Purpose:

The ultimate goal of this thesis is to develop a plate-tectonic model for the evolution of the Gulf of Mexico and the Caribbean region. At the time of writing, several first-order, large-scale problems plagued geologists attempting to decipher bits and pieces of Gulf and Caribbean evolution. This is a framework study, designed to provide a plate-kinematic basis on which can be built more detailed geological analyses of any sort, in any area within the Gulf and Caribbean. Also, it provides a model that can be tested, modified, and revised by future workers. The model itself helps to identify problems on which more work is needed in order to gain a full or better understanding of the geologic history of local areas.

The evolutionary model developed within this thesis was derived largely from <u>deductive reasoning</u>. The initial premises are of the largest order, and successive attempts to build and refine the model have incorporated increasing amounts of geologic information. Refinement of a model can only go so far; at some point it becomes necessary to concentrate on problem areas, using <u>inductive reasoning</u>, to provide critical bits of information which can solve points of contention during the deductive phase of reasoning. Two such problems were identified early on and addressed in the field, in northern and southwestern Dominican Republic (Chapters 4 and 5). Of course, the ultimate goal of this process is for the two ends to meet in the middle, so that the model suggests no further problems, and additional detailed work does not disprove or alter the model. Fortunately for the

Caribbean geologist, this point lies far into the future in the case of the Caribbean!

The author would like to think that refinement of the model, on a regional scale at least, has reached the point where significant improvement cannot be achieved without new studies in problem areas identified by the model, or at least reinterpretations of some existing studies.

Methodology:

Permo-Triassic reconstructions [Bullard et al., 1965], except Pangea "B" [Irving, 1977], demonstrate that South America occupied the entire Caribbean and most of the Gulf of Mexico area prior to the dispersal of Pangea. Thus, the plate-tectonic development of both the Gulf and the Caribbean must be directly related to the paleopositions and relative motions of the larger encompassing plates. The existence of Jurassic seafloor crust in the Gulf of Mexico, and the fact that the Yucatan block is presently a part of the North American Plate, provides a measure of the potential complexities associated with that evolution. Another example is that Cuba, an integral part of the Caribbean collection of arcs, is not (any longer) a part of the Caribbean Plate as we know it today from the distribution of seismicity [Sykes and Ewing, 1965; Molnar and Sykes, 1969]. Thus, the present plate boundary system in the Gulf/Caribbean region bears very little relationship to former configurations, and the region may be considered historically as a complex "interplate" realm.

Early attempts to build evolutionary scenarios of interplate regions by relating geology to a framework of relative plate motions include Atwater [1970] for western North America, and Smith [1971] and Dewey et al. [1973] for the Alpine and Mediterranean region. A geologic-kinematic synthesis of this kind was first done for the Caribbean by Ladd [1974; 1976], who related Caribbean geology to the relative motions of North and South America since the Jurassic.

Derivation of evolutionary models for interplate realms must follow three basic steps. First, an attempt must be made to reconstruct the relative positions of the major and minor plates which existed at the onset of the geological development of the region. All pre-Mesozoic crust involved must be identified; all syn-rift and post-rift crustal attenuation, shortening and shear must then be restored to pre-rift limits; finally, the continents and blocks, in their pre-rift shapes, must be reassembled into a reconstruction which satisfies pre-rift deformation patterns, paleogeographic relationships, and depositional patterns. Secondly, beginning with the initial reconstruction, the relative motion history of the plates encompassing the interplate region must be defined in order to establish the framework on which to base the evolutionary model. Both paleopositions and the relative motion vector are important. The paleopositions define the size and shape of the region at various times, and the relative motion vector defines the direction and rate of change in the size and shape of the region. Finally, the geology of the interplate region must be integrated with, and related to, the framework provided by the relative motions of the encompassing plates.

An intimate relationship must exist between relative plate motions and the geology of an interplate region; iterative integration of the two derives plate boundary schemes through time [Dewey et al. 1973; Pindell and Dewey, 1982], and each provides a test for the other. Stable, transgressive carbonate platforms or passive margins may be associated with drift phases following continental breakup, whereas folding, arc volcanism, metamorphism and obduction of ophiolitic rocks may be associated with convergent phases of ocean closure. In addition, linear belts of uplifted basement rocks in flower structure often represent zones of strike-slip motion, which, in turn may relate directly to portions of relative motion vectors.

On a finer scale, motions of the smaller intervening microplates of the interplate region may be inferred by paleomagnetic measurements, deformation styles on microplate borders, vergence of thrusting in ophiolitic belts, polarity indicators in arc terranes, orientations of paired metamorphic belts, provenance and facies considerations in sedimentary rocks, linear magnetic anomalies in trapped oceanic basins, major strike-slip faults with measurable offsets, and a variety of other indicators which may be unique to the region in question. But all of the above must ultimately be constrained by the paleopositions and relative motions of the encompassing major plates. The importance of that framework cannot be overemphasized.

In the case of the Gulf of Mexico and Caribbean region, the starting point is the reconstruction of western Pangea (North and South America, Africa, and the pre-Mesozoic continental blocks of Yucatan, Central America and the Bahamas). The pertinent relative motion histories are North America/South America and, to a lesser extent, North America/Farallon. The region whose geology must be integrated into that framework includes the Caribbean itself, the Gulf of Mexico, Yucatan, Mexico, Central America, and northern South America.

Organization:

This thesis is organized by the logical progression of the methodology outlined above, which coincides roughly with a forward progression in the geological history of the Gulf and Caribbean region. In the first chapter, a detailed reconstruction of western Pangea (the Alleghenides) is derived.

In the second chapter, starting with the Late Paleozoic Alleghenian reconstruction, the kinematic framework of relative plate motions since the Triassic is developed and defined. This is followed by a chapter on the geology and evolution of the Gulf of Mexico region, which is theoretical in nature. Chapters 4 and 5 summarize the results of field work done in northern Dominican Republic (Puerto Plate area) and in southwestern Dominican Republic (eastern Sierra Baharuco, San Juan Valley), respectively. These results provide critical input to the Caribbean evolutionary model presented later. Chapter 6 is a review of the primary geological features of the Caribbean, in which first-order plate-tectonic elements and their significance are defined, such as arcs, suture zones, and back-arc basins. Chapter 7 is another theoretical analysis, on the neotectonic deformation and evolution of the northern South American borderlands. Finally, chapter 8 integrates all of the above into a geologic-kinematic plate-tectonic model for the evolution of the Caribbean region as a whole. The model is presented in a series of 11 plate boundary maps, with vector triangles defining the direction and rate of relative plate motions. In a final "chapter," conclusions derived from, and problems remaining in, the model are discussed, and areas requiring further work from an inductive approach are defined.

CHAPTER 1

THE ALLEGHENIAN RECONSTRUCTION

THE ALLEGHENIAN RECONSTRUCTION:

Introduction:

The Upper Paleozoic of eastern, central, and southern North America is underscored by two important facts. First, the Late Paleozoic belt of deformation (Appalachians, Ouachitas, Marathon, Huastecans) and its associated foreland basins in the U.S. interior are continuous from at least Nova Scotia to Mexico. Second, there is no evidence for the existence of marine conditions during the Upper Permian to Middle Triassic, throughout southern, central and eastern North America, or the circum-Gulf region. These two important facts suggest that the circum-Atlantic continents and the continental blocks within the Gulf/Caribbean realm must have achieved complete Paleozoic ocean closure.

Because of repeated Paleozoic orogeny in the Appalachians, and because the terms Appalachian, Ouachita, Marathon and Huastecan describe geographic portions of a single continuous, but diachronous, belt of Late Paleozoic deformation, the term "Alleghenides" is used here to describe collectively those Late Mississippian-Middle Permian tectonic features and metamorphism that were created by the diachronous ocean closure between Gondwana and American portions of Laurussia [Graham et al., 1975; Kluth and Coney, 1981; Pindell and Dewey, 1982; Dewey, 1982; Bradley, 1982].

The Alleghenian reconstruction must provide an internally consistent rearrangement of interpreted tectonic provinces and metamorphic zones from the various continental blocks. The rearrangement of provinces is made difficult by the fact that Jurassic continental breakup has followed approximately the lines of Late Paleozoic continental suturing; only rarely can pre-Permian geology be tied across Alleghenian sutures in the reconstruction because the opposing continents had little in common prior to collision, and the majority of Triassic deposition upon each consists of red beds that are lithologically similar and difficult to correlate biostratigraphically.

Before a reconstruction of Alleghenian tectonic features may be attempted, simple geometric constraints must first be considered. These include (1) identification of all pre-Mesozoic continental crust involved in Alleghenian orogenesis, (2) restoration to original crustal thickness the attenuation undergone by each margin and block during Late Triassic and Jurassic rifting, and (3) retraction, where possible, of post-Permian offsets on intracontinental fault systems such as those in Mexico [Anderson and Schmidt, 1983] and in northwest South America (see chapter 7). The most critical constraint, however, is the proper realignment of the circum-Atlantic continents by closure of the Central, Equatorial, and the South Atlantic Oceans. Various ocean-closure models differ greatly and have drastically different implications for Alleghenian reconstructions.

Three differently derived ocean-closure models (Figure 1.1a,b,c) stand out as important: least-error fitting of the 1000 fathom (2000 meter) isobaths [Bullard et al., 1965]; realignment of marginal offsets and fracture zones from opposing continental margins [LePichon and Fox, 1971]; and least-error fitting of paleomagnetic poles of Permian age for each of the three continents [Van der Voo et al., 1976]. The least-error fitting of present margins does not recognize the blanketing effect of post-breakup marginal sedimentation on the true marginal configuration of pre-Mesozoic basement. The realignment of marginal offsets and fracture zones is a requirement at the commencement of continental breakup. The fit suggested

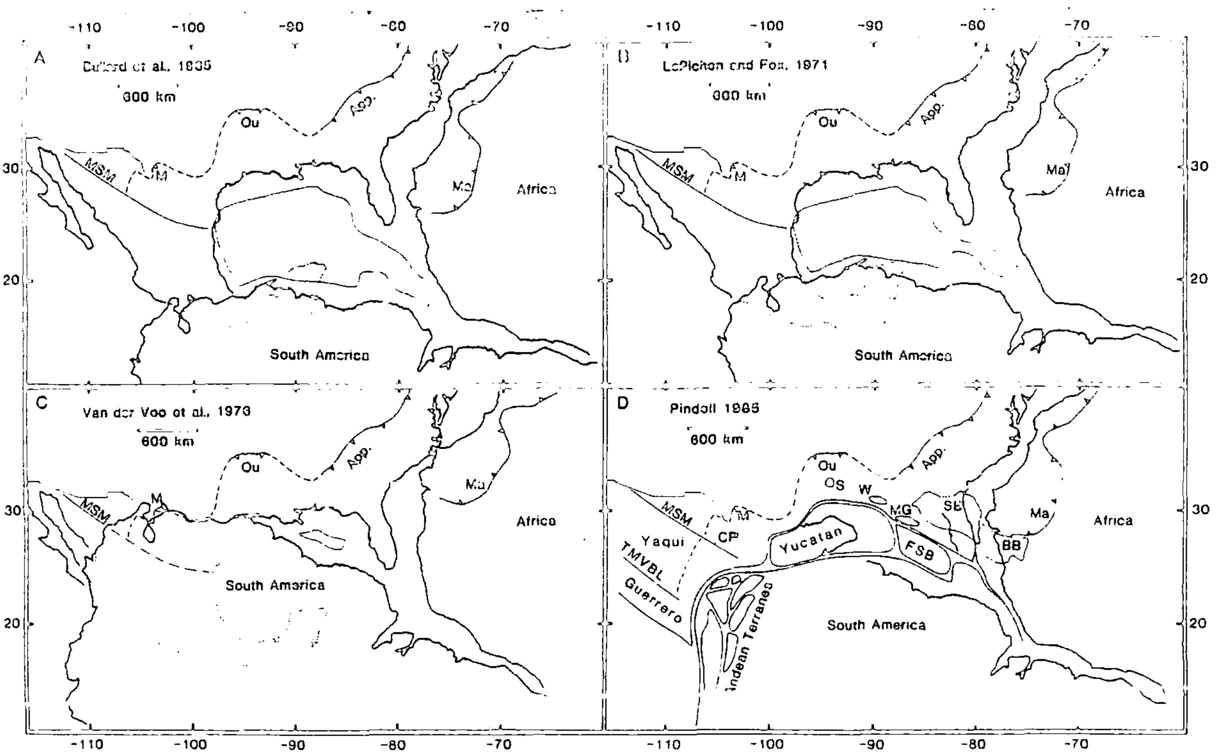


Figure 1.1. Four initial reconstructions of the major circum-Atlantic continents (Mercator Projection in a present-day North American reference frame) commonly used as starting points in models of Gulf/Caribbean evolution. Shaded areas indicate non-continental gaps within the reconstructions. MSM, Mojave-Sonora Megashear; App, Appalachians; Ou, Ouachitas; M, Marathons; Ma, Mauritanides; TMVBL, Trans-Mexican Volcanic Belt Lineament; FSB, Florida Straits Block; CP, Coahuila Peninsula; S, Sabine Uplift; W, Wiggins Arch; MG, Florida Middle Grounds Arch; SB, Suwannee Basin; BB, Bove Basin.

by paleomagnetism overlaps continental crust in Florida and northernmost Mexico, and leaves no room for the insertion of additional continental blocks between North and South America. As is, the paleomagnetic fit does not explain the geometry of the Alleghenian features of North America. All of these fits use the South Atlantic fit of Bullard et al. [1965], which contains significant error because the 1000 fathom isobath along the Amazon portion of the Brazilian shelf does not coincide with the continental limit [Pindell and Dewey, 1982; Klitgord et al., 1984].

Figure 1.1d schematically shows a fourth reconstruction, which is developed in this chapter. This reconstuction, among the other considerations mentioned above, incorporates a tighter Guinea-Brazil juxtaposition across the Equatorial Atlantic. This difference results in 1) North America and Africa remaining approximately in their syn-rift relationship as defined by marginal offsets of fracture zones, and 2) Late Paleozoic paleomagnetic poles for South America aligning with those for North America. Hence, the strong points of Figures 1.1b and 1.1c are preserved.

The following discussion on various aspects of the derivation of the Alleghenian reconstruction refers heavily to Figure 1.2, which outlines primary Gulf-region geological features. The Alleghenian paleoreconstruction is presented in Figure 1.3. Table 1.1 defines abbreviations, and interprets geological features, portrayed in Figure 1.3.

Louann and Campeche Salt Alignment:

One very simple but important key to Alleghenian paleogeography which arises from the Mesozoic rifting history, rather than the Late Paleozoic closure history, of the Orogen is the symmetrically disposed Louann and

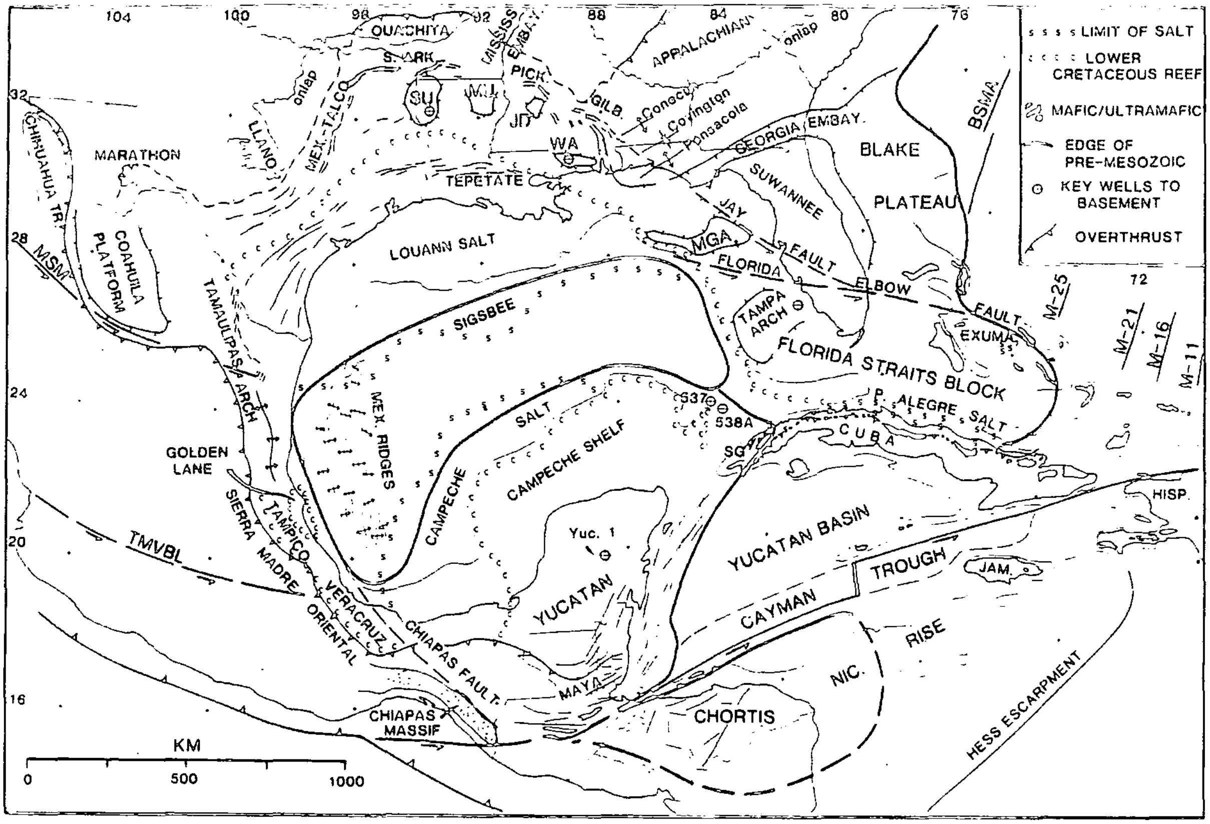


Figure 1.2. General tectonic and locality map of the Gulf region. MSM, Mojave-Sonora Megashear; TMVBL, Trans-Mexican Volcanic Belt Lineament; SU, Sabine Uplift; MU, Monroe Uplift; JD, Jackson Dome; WA, Wiggins Arch; MGA, Florida Middle Grounds Arch; SG, Sierra Guaniguanico, presently of western Cuba but was northeastern corner of Yucatan block prior to Paleogene; BSMA, Blake Spur Magnetic Anomaly; M-25, M-21, M-16, M-11 are Central North Atlantic magnetic anomalies.

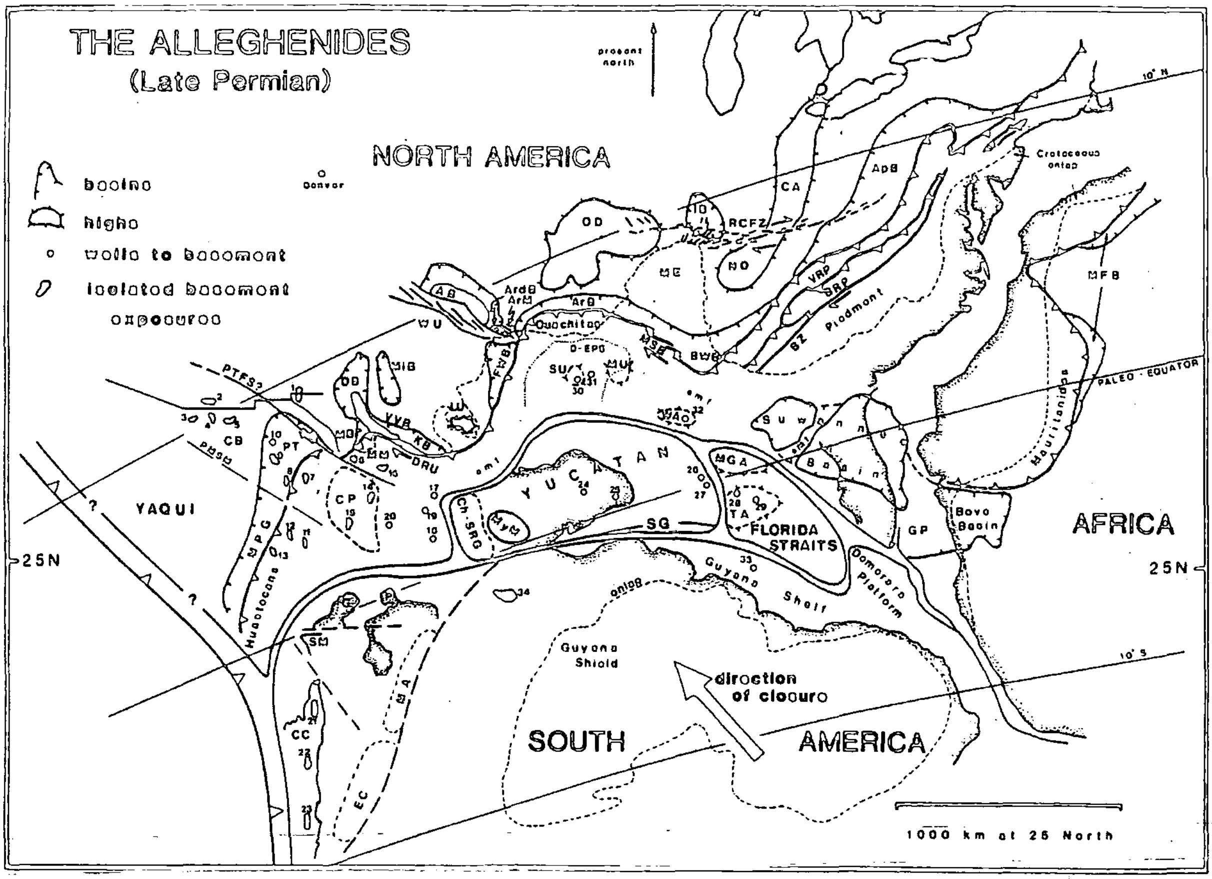


Figure 1.3. Late Permian Alleghenides reconstruction derived in text, abbreviations and numeric labels defined in Table 1.1. Total rotation parameters for closure defined in Tables 2.3, 2.4 and 2.5 (chapter 2). Chortis block and Mexico south of the Trans-Mexican Volcanic Belt not included because their Permian positions are unknown.

TABLE 1.1 Key and Interpretive Glossary to Figure 1.3

Abbreviation	Explanation
AB	Anadarko Basin. Carboniferous extensional and right-lateral strike-slip basin, compressed in Late Pennsylvanian and/or Early Permian.
ApB	Appalachian Basin. Devonian to Carboniferous foreland sedimentary basin resulting from load of Appalachian thrust sheets.
ArB	Arkoma Basin. Carboniferous foreland basin resulting from load of Ouachita thrust sheets to the south.
ArdB	Ardmore Basin. Subbasin of Anadarko Basin, sits amidst dextral strike-slip zone.
ArM	Arbuckle Mountains. Dextral shear zone in North American foreland, probably related to Wichita Uplift. Active during Pennsylvanian.
BB	Bove Basin. Platform sequence (terrigenous) of Paleozoic age, rocks correlatable to Florida's Suwannee Basin.
BRP	Blue Ridge Province. Metamorphic thrusted nappes in central Appalachians, thrusting caused by collision of Africa with North America.
BWB	Black Warrior Basin. Carboniferous foreland basin resulting from load of Appalachian thrust sheets.
BZ	Brevard Zone. High-angle reverse and/or strike-slip fault in southern Appalachians, possible zone of lateral escape during Alleghenian compression.
CA	Cincinnati Arch. Foreland bulge caused by lithospheric flexure from load of Appalachians.
CB	Chihuahua Block. Zone north of Mojave-Sonora Megashear but south of Rio Grande which underwent basement reactivation during Jurassic (sinistral shear) and Cretaceous (compression). In Figure 1.3, a homogeneous simple shear totalling 100 km across the block has been restored to the northwest.
CP	Coahuila Platform. Block accreted to North America during Alleghenian assembly, possibly originally part of Gondwana but stranded during rifting, may represent Late Paleozoic arc created by closure between North America and Gondwana.
CC	Cordillera Central, Colombia. Presumably formed northwest margin of South America throughout Paleozoic to Late Cretaceous. Age and chemistry of Paleozoic and Mesozoic plutonic rocks questionable, but may have been Late Paleozoic arc prior to ocean closure. Mesozoic-Cenozoic arc volcanism has occurred here as well.
Ch-SRG	Chuacus-Santa Rosa Group. Chuacus Group are metamorphosed (in early Carboniferous) igneous and sedimentary rocks of Guatemala belonging to Yucatan block. Metamorphism relates to continental collision. The Upper Pennsylvanian to Lower Permian Santa Rosa Group unconformably overlies the Chuacus, and was deposited in a shallow sea behind the main Alleghenian thrust belts.
DB	Delaware Basin. Extensional, right-lateral strike-slip basin of Early Permian age. Relates to the escape of the Mexican peninsula during continental collision.

TABLE 1.1. (continued)

Abbreviation	Explanation
D-EPS	Desmoinesian-Lower Permian Sediments. Postorogenic shallow-water molasse deposited behind Ouachita thrust zone. Correlated with Santa Rosa Group (Yucatan Block) and Permian of Colombia.
DRU	Devil's River Uplift. Right-lateral shear zone created by northwestward migration of Marathon region along southern side of Llano Uplift during ocean closure.
EC	Eastern Cordillera. Uplifted since Late Miocene due to compression and dextral strike-slip motion caused by Panama-Colombia collision. Was shallow-water sea and received Permian molasse from adjacent uplifted Cordillera Central during collision.
emf	Eagle Mills Formation. Predominantly Upper Triassic continental red beds filling grabens and obscuring basement.
FWB	Fort Worth Basin. Pennsylvanian foreland basin related to loading of Ouachita structural belt.
G	Guajira Block. A northern extension of Cordillera Central (see Cordillera Central).
GP	Guinea Plateau. Extended and submerged pre-Mesozoic continental crust which must be included in Alleghenian reconstruction.
IB	Illinois Basin. Intracontinental basin initiated during Early Paleozoic, reactivated in Alleghenian orogeny with the intersection of Mississippi Embayment and Rough Creek fault trends.
KB	Kerr Basin. Early Permian foreland basin, related to loading by Marathon thrust sheets (Ouachita structural belt).
LU	Llano Uplift. Stable buttress of North America during collision, foreland bulge to the Ouachita structural belt.
MA	Merida Andes. Uplifted since Late Miocene during Andean Orogeny. Was shallow-water sea and received Permian molasse from nearby uplifted Cordillera Central (Santa Marta and Guajira).
MB	Marfa Basin. Early Permian foreland basin related to loading of Marathon thrust sheets.
ME	Mississippi Embayment. Structural low, floored by Precambrian rift, only slightly reactivated during Alleghenian collision and Mesozoic rifting, but underwent extension and volcanism in Cretaceous.
MFB	Mauritanides Foreland Basins. Devonian to Permian foreland basins related to loading by Mauritanides thrust sheets during collision.
MiB	Midland Basin. Shallow, cratonic basin adjacent to Delaware Basin, cause uncertain.
MGA	Middle Grounds Arch of Northwest Florida Platform. Structural basement high, or horst, defines edge of pre-Mesozoic continental crust in northeast Gulf.
MM	Marathon Mountains. Metamorphosed Paleozoic continental rise and slope sediments and orogenic flysch thrust up onto North American shelf sequence during Late Pennsylvanian to Early Permian time.

TABLE 1.1 (continued)

Abbreviation	Explanation
MPG	Mexican Paleozoic Geosyncline. Foreland basin related to emplacement of Huastecan thrust sheets during Early to Middle Permian time.
MSB	Mississippi Slate Belt. Dextral shear zone created by migration of Ouachita region and Yucatan along southern end of Appalachians during Carboniferous ocean closure.
MU	Monroe Uplift. Basement high, possibly stranded continental remnant of Gondwana.
MyM	Maya Mountains. Deformed Upper Paleozoic igneous and sedimentary rocks. Shales and sands (Maya Series) correlated to Santa Rosa of Guatemala, Desmoinesian-Early Permian sediments of U.S. Gulf Coast, and Lower Permian of Colombia.
ND	Nashville Dome. Southern portion of Cincinnati Arch (see Cincinnati Arch).
OD	Ozark Dome. Structural basement high, foreland bulge related to loading of Ouachita thrust belt.
P	Paraguana Block. Northward extension of Cordillera Central (see Cordillera Central).
PMSM	Proto-Mojave-Sonora Megashear. Possible? zone of dextral continental escape during collision.
PT	Pedregosa Trough. Foreland basin related to emplacement of Huastecan thrusts in Permian time. May also be site of dextral transform motion allowing continental escape of Mexican peninsula.
PTFS	Pedregosa Trough Fault System. Postulated zone of dextral slip enhancing development of Pedregosa Trough and allowing continental escape of Mexican peninsula.
RCFZ	Rough Creek Fault Zone. Zone of intracontinental dextral shear, accommodating shortening in southern Appalachians during Carboniferous.
SG	Sierra Guaniguanico, Cuba. Prior to Paleogene Cuba-Bahamas collision, this continental crust belonged to northeast margin of Yucatan.
SM	Santa Marta Block. Northern extension of Cordillera Central (see Cordillera Central).
TA	Tampa Arch. Pre-Mesozoic basement high off southwest Florida Platform, forms western portion of Florida Straits Block.
VRP	Valley and Ridge Province. Folded platform cover in front of primary Appalachian thrust sheets.
VVB	Val Verde Basin. Early Permian foreland basin related to loading by Marathon thrust sheets.
WA	Wiggins Arch. Metamorphic pre-Mesozoic basement high, originally part of Gondwana.
WU	Wichita Uplift. Dextral shear zone related to minor Pennsylvanian northwestward displacement of Llano block relative to midcontinent.
1	Franklin Mountains. Upper Paleozoic carbonate shelf of North America [Bridges, 1970].

TABLE 1.1 (continued)

Abbreviation	Explanation
2	Bisbee, Arizona. Upper Paleozoic carbonate shelf of North America [Bridges, 1970].
3	Cananea. Upper Paleozoic carbonate shelf of North America [Bridges, 1970].
4	Sierra Los Ajos. Upper Paleozoic carbonate shelf of North America [Bridges, 1970].
5	El Tigre. Upper Paleozoic carbonate shelf of North America [Bridges, 1970].
6	Solitario. Surface exposure of southern continuation of Marathon belt [Bridges, 1970].
7	Mina Plomosas-Placer de Guadalupe. Marathon equivalent, Ordovician to Devonian limestone, dolostone, chert, Carboniferous shale and carbonate, Permian reefs and shale [King, 1975].
8	Sierra del Cuervo. Wolfcampian highly deformed but little metamorphosed shale and sand [King, 1975].
9	Sierra de la Mojina. Lower Cretaceous conglomerate with metasedimentary clasts with ages of 330-350 Ma [King, 1975].
10	Villa Ahumada Borehole. 1600 m of shale and silt, upper part Wolfcampian [Bridges, 1970].
11	Ciudad Victoria (west of). Thrust slices of Precambrian gneiss, schists, Silurian to Devonian black shale, cleaved Carboniferous-Permian flysch, ultramafics [King, 1975; De Cserna, 1976; Salas, 1970].
12	Aramberri. Pre-Mesozoic phyllites, schists, metavolcanics [King, 1975].
13	Catorce. Ultramafic (ophiolite?) obducted with the Huastecans [de Cserna, 1976].
14	Potrero de la Mula. Granodiorite plutons dated at 206 Ma, dating is questionable [King, 1975].
15	Las Delicias. Permian volcaniclastic shale and greywacke, reefy carbonates, highly folded, and intruded by Triassic (cooling age) granite [King, 1975].
16	Sierra del Carmen. Phyllite, marble, quartzite, greenschists, 263-275 Ma on metamorphics [King, 1975].
17	Borehole. Bottomed in Paleozoic schists [King, 1975].
18	Borehole. Bottomed in Paleozoic slate and quartzite [King, 1975].
19	Borehole. Bottomed in granite-gneiss, 358 Ma [King, 1975; Flawn et al., 1961].
20	Borehole. Bottomed in Paleozoic granite [King, 1975].
21	Cajamarca Group. Exposures of orthogneiss, tuffs, basic lavas, metatonalites. Metamorphic cooling ages: biotite, 239 Ma; muscovite, 214 Ma [Irving, 1971, 1975; Shagam, 1975].
22	Granitic schists, metamorphism 215 Ma [Irving, 1975].
23	Granitic schists, metamorphism 220 Ma [Irving, 1975].
24	Borehole. Yucatan no. 1. Meta-andesite, 330 Ma and 290 Ma [Marshall, 1974].
25	Borehole. Yucatan no. 4. Metaquartzite [Marshall, 1974].

TABLE 1.1 (continued)

Abbreviation	Explanation
26	Borehole. DSDP 537. Phyllite, 449 Ma and 456 Ma [Schlager et al. 1984].
27	Borehole. DSDP 538A. Amphibolite gneiss, 496 Ma and 348 Ma. Also diabase, 190 Ma and 165 Ma on different sills [Schlager et al. 1984].
28	Borehole. Pre-Jurassic(?) granodiorite R. N. Erlich, personal communication, 1984).
29	Borehole. Pre-Jurassic(?) granite (R. N. Erlich, personal communication, 1984).
30	Borehole. Mississippian rhyolitic tuffs [Nicholas and Waddell, 1982].
31	Borehole. Desmoinesian-Early Permian shallow-water, undeformed clastics and carbonates [Nicholas and Waddell, 1982].
32	Borehole. Granite and phyllite, metamorphic ages of 270 Ma to 325 Ma [Cagle and Khan, 1983; J. Cagle, personal communication, 1984].
33	Borehole. Precambrian at 8,877 feet [Case and Holcombe, 1980].
34	El Baul. Paleozoic sediments metamorphosed in Late Paleozoic, granite intrusion 287 Ma [Feo-Codecido et al., 1984].

Campeche salt deposits in the Gulf of Mexico (Figure 1.2). Models of plate separation (and hence of reconstruction) must recognize that seafloor spreading separated a once-single Gulf wide salt pan into its component parts [White, 1980]. Thus, it is probable that the Yucatan block, but not the Chortis block, fit between the U.S. Gulf Coast and Venezuela prior to salt deposition. This positioning of Yucatan provides the necessary geometry of opposing margins for a tight fit between Gondwana and Laurussia, and explains the Ouachita structural embayment of the Alleghenian Orogen.

The Florida Straits Block

Several lines of evidence suggest that southern Florida and much of the Bahamas Platform are underlain by pre-Mesozoic continental crust, a block (or blocks) referred to here as the Florida Straits block (Figure 1.2).

The nature and age of basement beneath the Bahamas has been debated for decades, as geophysical and well data have not yet provided a conclusive answer. However, gravimetric, magnetic, and seismic data [Uchupi et al., 1971] suggest that presumably pre-Mesozoic, attenuated continental crust underlies the western half of the platform, whereas the eastern half is oceanic. Basement of the oceanic portion must have been raised to the photic zone early on to initiate carbonate bank development, presumably by plate motions along transform faults, or by volcanism, during the separation of North America and Gondwana. Recently acquired seismic data along the northern Bahamian margin (J. W. Ladd, personal communication, 1984) indicate that the western Bahamas are underlain by attenuated continental crust as far east as Tongue of the Ocean. A similar eastward extent is suggested by the Punta Alegre evaporites of northern Cuba [Pardo, 1975] and by the possible salt occurrence [Lidz, 1973] beneath Exuma Sound (Figure 1.2), if it is assumed that evaporite sequences pertaining to continental breakup are formed primarily in shallow water, often on attenuated continental crust due to initial subsidence during rifting. The southern margin of the Bahamas was imbricated in thrusts as Cuba collided with the Bahamas in the Paleogene [Gealey, 1980; Dickinson and Coney, 1980], providing surface exposures in northern Cuba of early Bahamian stratigraphy. The eastward extent of the Punta Alegre closely matches the extent of continental crust suggested by Uchupi et al. [1971].

In the subsurface of the southern Florida Shelf, the recent drilling of granitic basement rocks of the Tampa Arch (Figure 1.2) at depths of 3 to 3.7 kilometres (R. N. Erlich, personal communication, 1984) also demonstrates the existence of attenuated continental crust. Further, seismic sections presented in Schlager et al. [1984] show a block-faulted basement beneath the sedimentary section of the northern wall of the Florida Straits Channel. Hence, the Jurassic rhyolites and basalts of the South Florida Volcanic Province of Klitgord et al. [1984] intrude and cover this stretched continental crust, and reflect intracontinental volcanism during Mesozoic rifting between Africa and southeastern North America.

It appears that the Florida Straits block (or blocks) of attenuated pre-Mesozoic continental crust underlies southern Florida, the southern Florida Shelf, and the western half of the Bahamas (Figure 1.2), and must be included in Alleghenian reconstructions. The fact that much of this crust overlaps Africa when the Atlantic Ocean is closed indicates that crust of the Florida Straits block has migrated some distance into its present position, either by internal extension during rifting or by translation from the site of the present eastern Gulf of Mexico along plate boundaries, as in Pilger [1978]. It will be seen that both mechanisms probably occurred.

Stretching Analysis of Rifted Margins:

Areas presently underlain by pre-Mesozoic continental crust that must be integrated into the Alleghenian assembly are shown in Figure 1.4, excluding western Africa, and include the U.S. Gulf Coast margin and eastern Mexico, northern South America, the Yucatan block, Florida, the Blake Plateau, and the Florida Straits block. The Antilles and the Panama-Costa Rican isthmus are post-Alleghenian, Cretaceous and Cenozoic island arc complexes built on Late Mesozoic ocean floor. The continental crust of the Chortis block (Figure 1.2) entered the Caribbean region from the Pacific realm during the Late Cretaceous-Paleogene [Dickinson and Coney, 1980; Wadge and Burke, 1983] and, hence, does not need to be considered in the reconstruction. The margins of the pre-Mesozoic blocks were considerably attenuated during Mesozoic rifting, so that their present geographic dimensions greatly exceed their former, prerift dimensions, and this extension must be restored in proper reconstructions of western Pangea.

The cross-sectional geometry of continental crust along each margin may be estimated by isostatically balancing various thicknesses of sediment, crust, upper mantle, and water, at several locations across the margin, with an equivalent column of "normal" continental crust. The depth to the sediment-crust interface can be determined seismically (reflection and refraction), but identification of the crust-upper mantle boundary is more difficult. Where possible, a component of subsidence due to flexure at the margin should be accounted for. Post-rift sedimentary onlap makes most rifted margins appear as though stretching has occurred farther inland than it actually has. This is matched by a corresponding sedimentary section in the stretched zone where more stretching has occurred than is indicated by Airy isostatic balancing, and the two effects compete with each other [Watts, 1981]. However, within the stretched zone itself, the effects of

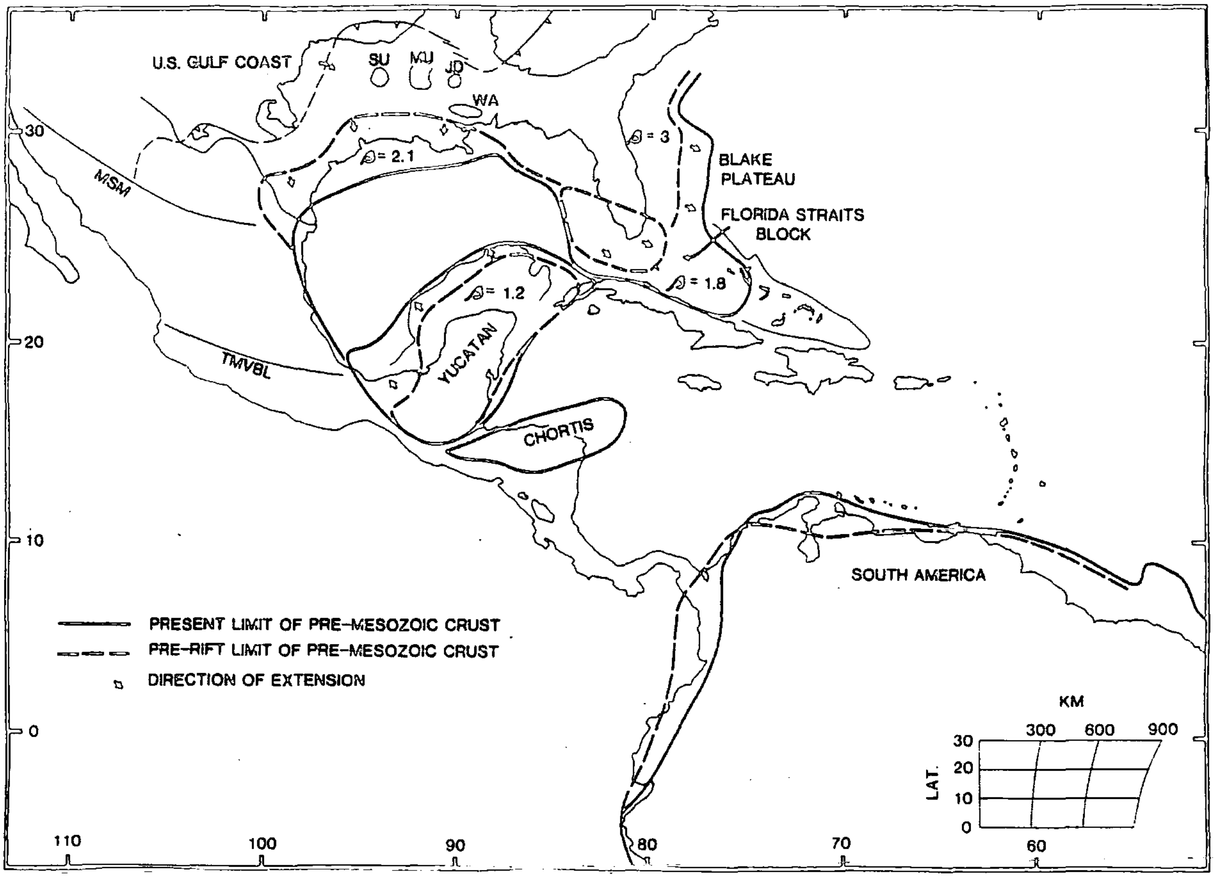


Figure 1.4. Areas of pre-Mesozoic continental crust in the Gulf-Caribbean region and amount of synrift extension and/or subsequent change in shape. Pre-rift limits were determined by restoring Jurassic lithospheric attenuation of the rifted margins and blocks, and by restoring Neogene offsets of faults within northern South America (see chapter 7), which collectively places the northern Andean terranes to the southwest in the present Panama Basin area. Beta values, or amounts of stretching, for various margins and blocks indicate average crustal attenuation within those margins. Prerift shapes are used in the Alleghenian reconstruction of Figure 1.3. Pre-Mesozoic crustal extent and extension of Chortis block are uncertain. MSM, Mojave-Sonora Megashear after Anderson and Schmidt [1983]; TMVBL, Trans-Mexican Volcanic Belt Lineament; SU, Sabine Uplift; MU, Monroe Uplift; JD, Jackson Dome; WA, Wiggins Arch.

flexure are not well understood because the crust undergoes complex changes in flexural rigidity during rifting. Therefore, an Airy model is used here in highly attenuated portions of margins. From lithospheric cross sections, total crustal extension may be estimated, assuming plane strain and conservation of continental crust, and discounting subsidence in the wings due to flexure, by defining a line that divides equally the cross-sectional area of continental crust on its oceanward side and the cross-sectional area of sediment and raised mantle on its landward side (Figure 1.5). By estimating the position of this line intermittently along each margin, the restored, prerift geometries may be approximated.

U.S. Gulf Coast and eastern Mexico. Figure 1.5 illustrates the restoration of attenuation in the case of the U.S. Gulf Coast, where crustal attenuation across a 760-km-wide zone was extreme. Sediment thicknesses of 13.4 to 14.6 km beneath 0 to 1 km of water in the Gulf Coast basin [Antoine et al., 1974] suggest beta values approaching 4. In northern Louisiana, southern Mississippi and east Texas, as far north as the Talco-South Arkansas-Pickens-Gilbertown fault system, shallow listric faults and rift structures indicate that basement attenuation occurred, but was relatively minor compared to southern Louisiana and offshore. Subsidence has been due largely to flexure caused by sediment loading in the Gulf, although significant rifting occurred in the East Texas, North Louisiana and Central Mississippi salt basins. Nunn [1982] estimated beta values in these small basins between 1.5 and 2, but these are perhaps overestimates because they are upper crustal extensions not involving the entire lithosphere. Integration across the 760-km-wide Gulf Coast Plain and margin yields a net north-south extension due to rifting of about 400 km, or an average beta factor of 2.1 (Figure 1.5). Thus, the prerift limit of pre-Mesozoic continental crust through central Louisiana lies at about 30.5 N,

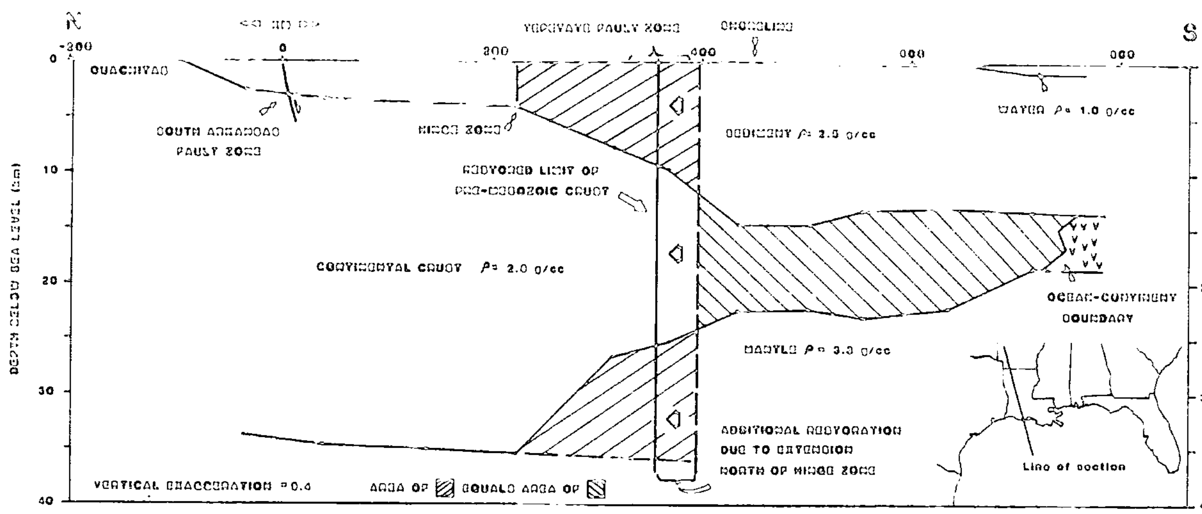


Figure 1.5. Theoretical cross section of the U.S. Gulf Coast margin through and offshore Louisiana, crossing the Sabine Uplift. Estimated depth to basement after Antoine et al. [1974] and Woods and Addington [1973], ocean-continent boundary after Buffler et al. [1981]. South of the apparent hinge zone, crust, mantle, sediment and water are balanced with an equivalent column of normal crust using an isostatic Airy model ignoring flexure, which is an adequate approximation in this area as the effective rigidity of the lithosphere was thermally reset (reduced) during rifting. Subsidence over the 220 km of the Gulf Coast to the north of the hinge zone is largely due to flexure, but extensional basement faulting did occur (South Arkansas fault zone, Gulf Coast salt basins, for example). An average extension of 20% (beta=1.2) is estimated for this zone, considering that little or no rifting occurred within Sabine Uplift and that beta estimations of 1.5 to 2 for stretching in the salt basins which assume homogeneous stretching of the entire lithoshere [Nunn, 1982] are probably overestimates. Hence, only an additional 44 km are removed from the restored limit determined by Airy balancing south of the hinge zone.

approximately along the Tepetate fault system (Figure 1.2). Similar analyses at other locations along the margin lead to the prerift limit of pre-Mesozoic crust shown in Figure 1.4.

Along eastern Mexico, basement drops off quickly along the eastern side of the Tamaulipas Arch, the Golden Lane high, and offshore along Veracruz (Figure 1.2). As will be discussed later, this curvilinear line defines a margin viewed here as the remains of a transform fault between Mexico and Yucatan which formed as the latter migrated away from the U.S. Gulf Coast. Hence, crustal attenuation due to rifting is believed to be negligible.

The Yucatan block. Basement beneath the Yucatan Peninsula lies beneath a 4-km-thick predominantly Late Jurassic to Cretaceous carbonate-evaporate section [Lopez Ramos, 1975], except in the southern, orogenically active portion and along the eastern basement high. The thickness of this section suggests a 20% extension of crust, but the absence of a well developed rift-related clastic section may indicate that much of this deposition is due to flexure-induced subsidence relating to the cooling and loading of the surrounding oceanic crust. Greater but uncertain sedimentary thicknesses occur beneath the Campeche Shelf, where rifting probably was more severe. Northwest of the Campeche Escarpment, an 80-km-wide rifted margin is overlain by 7 to 8 km of sediment and about 3 km of water [Buffler et al., 1980], suggesting a beta factor of about 5 and 64 km of extension within this narrow belt. Sediment thicknesses beneath Tabasco and westernmost Campeche (6 km) and offshore beneath Campeche Bay (8.5 km, under 2 km of water) [Buffler et al., 1984] indicate beta values of 1.5 to 3.7 over a 400-km-long, northwest trending, cross-section. In addition to stretching, Cenozoic faulting also has altered the original shape of Yucatan. First, continental, terrigenous shelf sediments and crust originally belonging to the northeastern Yucatan block but presently in Sierra Guaniguanico of western Cuba (Figure 1.2) was caught up in the Paleogene collision of Cuba with the Bahamas. Second, sinistral displacement (about 130 km) of Yucatan basement has occurred along the Polochic Fault of Guatemala during the Neogene [Burkart, 1978, 1983; Deaton and Burkart, 1984]. Both offsets must be restored to original geometry.

Following these considerations, and assuming that the Yucatan Platform's 4 to 5 km thick sedimentary section is half due to subsidence arising from stretching and half from flexure, the prerift shape of the Yucatan block (Figure 1.4) was approximately 20% smaller than it is today. The direction of extension is northwest (present coordinates), perpendicular to the Campeche Escarpment.

Synrift continental extension in Yucatan combined with that in the U.S. Gulf coast, then, is about 500 to 520 km, in close agreement with the estimate of 490 km by Sawyer [1984].

Blake Plateau and Florida Straits block. Crustal attenuation in the Blake Plateau and Florida Straits block appears to be genetically related to the separation of Africa from North America. The sedimentary section lying upon block-faulted basement beneath the Blake Plateau is about 11 km thick [Grow and Sheridan, 1981] and water depths average 1 km. These values indicate beta values of about 3. The extension appears to be in a southeast direction, parallel to the divergence of Africa from North America. Thus, in the Alleghenian reconstruction the Blake Plateau must be restored to about 33% of its present size in a northwest direction.

Beneath the northern half of the Florida Straits block, the postrifting sedimentary section is 7 to 8 km thick (J. W. Ladd, personal communication, 1984), but in the South Florida Basin (south of Tampa Arch, Figure 1.2) and north of Cuba thicknesses may reach 12 km [Meyerhoff and Hatten, 1974]. In the area of Cay Sal Bank, the 12 km sediment thickness is partly due to thrust imbrication pertaining to the Paleogene Cuba-Bahamas collision [Dickinson and Coney, 1980; Gealey, 1980], whereas the 9 km thick section in the Andros area ahead of the thrusts [Meyerhoff and Hatten, 1974] is largely due to increased subsidence and sedimentation rates caused by foreland loading. In the Andros well, Paleocene-Middle Eocene deposition occurred 5 times faster than before and after this foredeep development [Paulus, 1972], leading to an accumulation of about 1.5 km greater than would have been allowed by subsidence arising from thermal decay and sedimentary loading; therefore, only about 7.5 km of the total deposition was driven by stretching and thermally induced sedimentary loading. Offshore at Tampa Arch and its flanking Florida Elbow and South Florida Basins, depth to basement ranges from 3 km to over 6 km (R. N. Erlich, personal communication, 1984), suggesting beta values of 1.2 to 1.8.

Because of the comlexity of the Florida Straits region, flexure is only considered as it pertains to the Paleogene collision with Cuba. By integrating the above sediment thicknesses over the geographic area for which they exist and discounting the effects of the collision, a rifting-induced sedimentary sequence averaging about 8 km is indicated for the Florida Straits block as a whole, which suggests a beta factor of about 1.8 that must be restored in the Alleghenian reconstruction.

Northern South America. Because the northern South American rifted margin has been so highly obscured by compressive deformation related to Caribbean evolution, crustal attenuation due to rifting cannot be estimated. However, significant changes in the shape of northern South America due to accretion and to internal strike-slip motions may be quantified, as outlined in the next section.

Change in Shape of Northern South America:

Three episodes of tectonism have significantly altered the Jurassic shape of northern South America since the breakup of Pangea. The first was the accretion of Cretaceous aged basement of the Western Cordillera of Colombia and Ecuador in Late Cretaceous time [Barrero, 1979; Mooney, 1980]. The second was the accretion of the Netherlands-Venezuelan Antilles volcanic arc (Aruba, Curacao, Bonaire, Aves, Roques, and Orchila) against the northern margin of Venezuela in Late Cretaceous to Paleogene time [Maresch, 1974; Gealey, 1980; Beets et al., 1984]. The third has been internal deformation, mainly by motion along several strike-slip faults, of northern South America over the last 9 Ma amounting essentially to 290 km of northeastward migration of the Andean Cordilleran terranes of Ecuador, Colombia and northwestern Venezuela, relative to cratonic South America (see chapter 7). The effects of each episode on northern South American geography are approximated below.

Quantifying the late Neogene deformation of northern South America to determine its paleogeography at 9 Ma is achieved by restoring known and inferred post-nine Ma offsets upon faults and plate boundaries separating terranes and plates of the southern Caribbean realm (chapter 7). Late Neogene motions between the terranes and plates can be described by a vector-triangle diagram (Figure 1.6), the restoration of which produces the 9 Ma paleogeography of Figure 1.7. The cause of this uplift and deformation, and drastic paleogeographic change, is, most likely, the progressive collision of the Panama arc with western Colombia [Pindell and Dewey, 1982; Wadge and Burke, 1983], and the subduction of buoyant, young crust produced at the Galapagos spreading center, which includes the Carnegie and Cocos aseismic ridges, beneath the Cordilleran terrane of Colombia and Ecuador.

The islands offshore northern Venezuela (Aruba to Orchila, Figure 1.7, inset) are composed largely of Upper Cretaceous plutonic and volcanic rocks

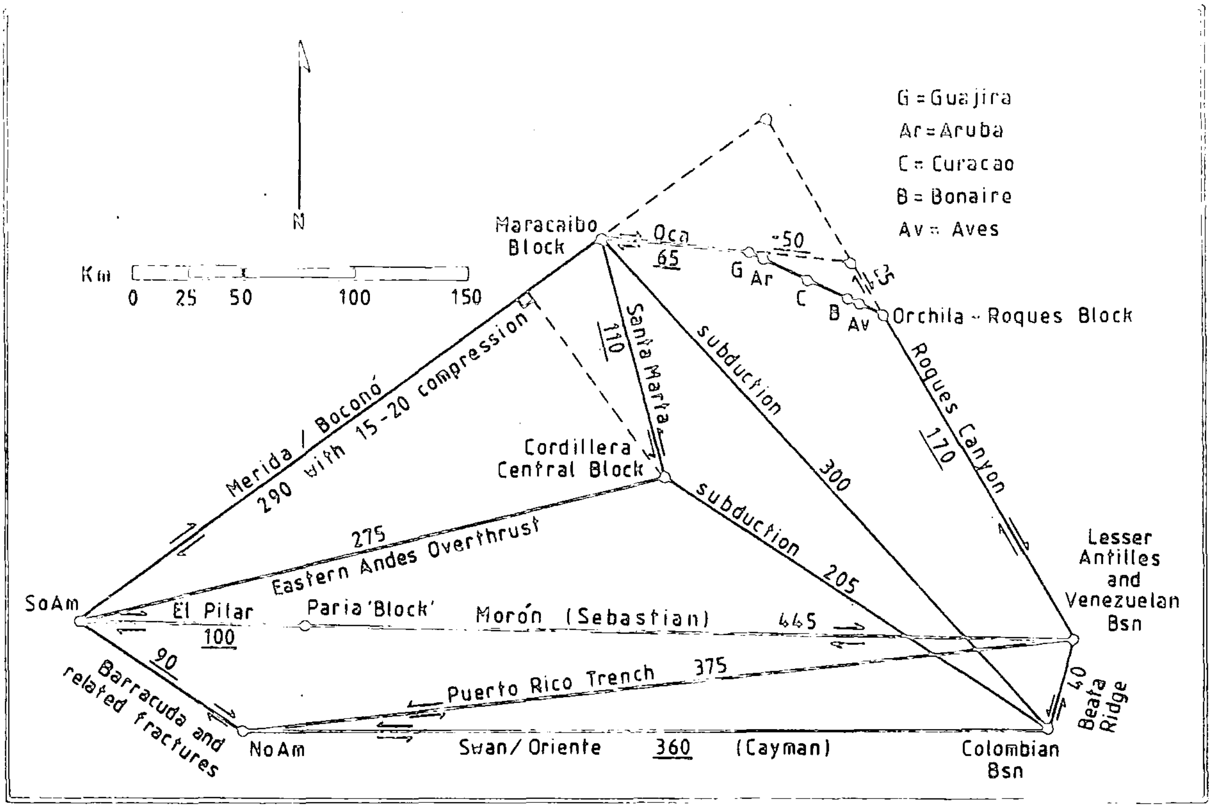


Figure 1.6. Vector-triangle diagram showing trend and magnitude of Neogene offsets between blocks and plates of the circum-Caribbean region (derived and referenced in Figure 7.1 of chapter 7). Restoration of offsets derives pre-Late Miocene shape of northern South America, shown in Figure 1.7.

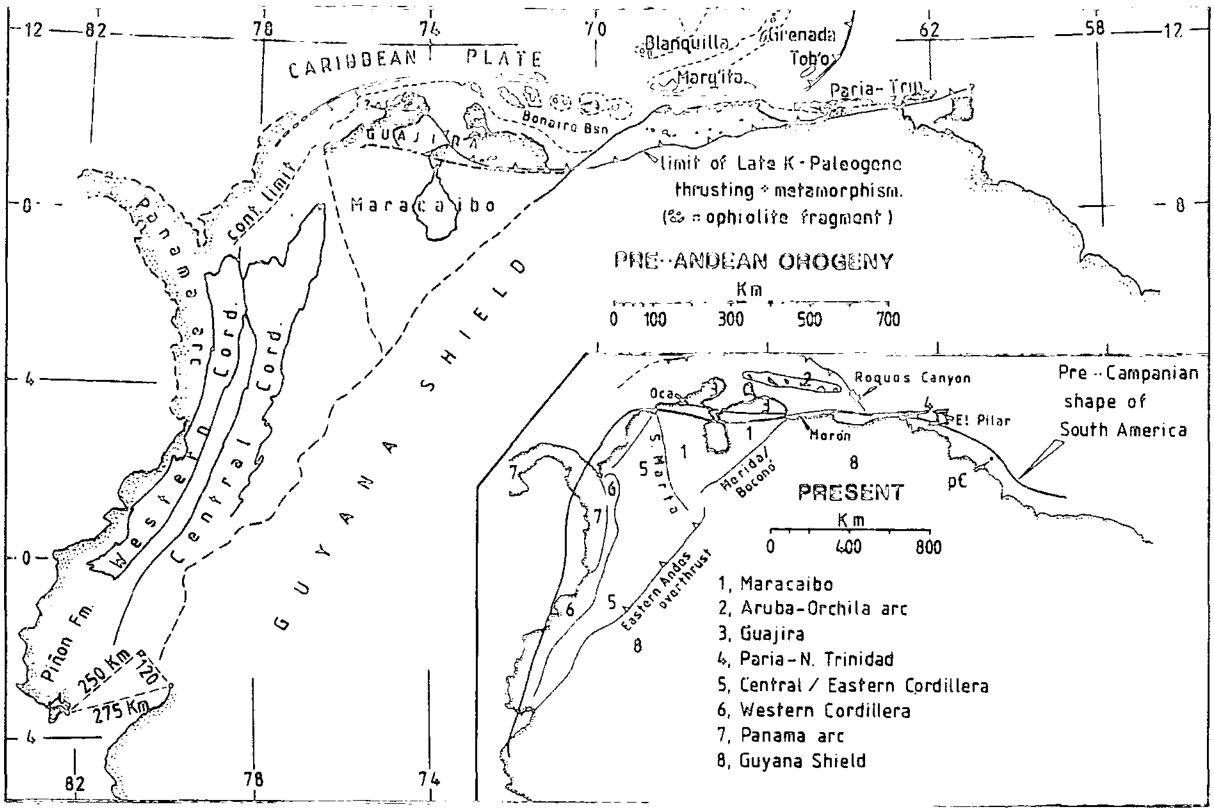


Figure 1.7. Pre-Late Miocene paleogeography of blocks and plates of circum-Caribbean realm, derived by restoring offsets defined in Figure 1.6. Inset: Blocks of northern South America which have undergone relative motion, on the faults shown, over the last 9 Ma. VC, thrust front of Villa de Cura and other nappes. Heavy line is pre-Campanian shape of northern South America used in Alleghenian reconstruction, derived by removing Aruba-Orchila and Western Cordillera accreted terranes from pre-Late Miocene paleogeographic reconstruction. Modified after Figure 7.2 of chapter 7.

intruding and overlying a Lower Cretaceous mafic series which probably represents oceanic crust [Maresch, 1974; Beets et al., 1984]. Hence, the islands comprise an intraoceanic arc which developed and collided with Venezuela subsequent to the breakup of Pangea. La Blanquilla (southern portion of Aves Ridge), Margarita (origin uncertain) and Tobago (accreted to leading edge of Caribbean Plate at an unknown time) also are composed entirely of post-Paleozoic rocks. In Venezuela, the ultramafic bearing Villa de Cura Klippe and associated nappes have been interpreted as the oceanic forearc [but see Beets et al., 1984] of this Netherlands-Venezuelan Antilles arc, which was obducted onto northern Venezuela during Late Cretaceous to Paleogene arc-continent collision [Maresch, 1974; Gealey, 1980]. Because pre-Mesozoic basement occurs near the coast to the north of the Villa de Cura, the precollision edge of northern South America apparently lies along the coast or offshore within the Bonaire Basin, but the offshore islands must be excluded from Pangean reconstructions.

In the Western Cordillera of Colombia and Ecuador (Figure 1.7, inset), pre-Cretaceous rocks are absent. Cretaceous basic igneous rocks (Pinon, Basic Igneous Complex, Diabase Group) overlain by deep-water abyssal sediments may represent oceanic crust accreted to the Cordillera Central in the Late Cretaceous [Mooney, 1980; Irving, 1975; Shagam, 1975; Feininger and Bristow, 1980; Barrero, 1979; Goossens and Rose, 1973; Duque-Caro, 1979]. Rocks typical of island-arc volcanism are present [Henderson, 1979], but these probably postdate the accretion of the basement complex to South America and, hence, represent arc volcanism along the newly accreted South American margin rather than volcanism at a preexisting arc of Pacific provenance that collided with Cordillera Central. The cause of accretion may have been due to the young age and consequent buoyancy of the oceanic crust; accretion occurred only a short time after formation. The basement complex may, in fact, be an occurrence of oceanic crust that was affected by the Middle Cretaceous B" extrusion event of the Caribbean Plate [Burke et al., 1978] that, rather than entering the Caribbean, was accreted to western South America. In addition to the Western Cordillera, the Neogene accreted Panama arc [Wadge and Burke, 1983] also must be removed from South America.

The removal of post-Paleozoic accreted terranes and restoration of Neogene deformation produces the pre-Campanian shape of northern South America shown in Figure 1.7, inset. This shape is believed to represent the Jurassic shape of northern South America, although Sierra de Santa Marta and the Guajira Peninsula may have migrated northeastward during the Late Cretaceous-Paleogene(?) from the area of the Lower Magdelena Valley [Duque-Caro, 1979]. Also, deformation prior to the Campanian is possible, but the stable shelf conditions that prevailed during the Early Cretaceous over much of northern South America [Maresch, 1974; Irving, 1975] suggests relative tectonic quiescence at that time.

The Equatorial Atlantic (Brazil-Guinea) Assembly:

The opening of the Equatorial Atlantic between northeast Brazil and the Guinea margin of Africa postdated the opening of the Gulf of Mexico. Because Africa's relative motion with respect to North America during the opening of the Gulf can be traced by magnetic anomalies in the Central Atlantic, the pre-rift reconstruction of South America with northern Africa defines, through the three-plate circuit, the relative relationship between North and South America during Gulf evolution.

Pindell and Dewey [1982] and Klitgord et al. [1984] suggested that the mid-Cretaceous South America-Africa fit of Rabinowitz and LaBrecque [1979], which is tighter than the South Atlantic fit of Bullard et al. [1965], existed also during the Jurassic, and should be used when modeling the Gulf of Mexico. However, this tighter fit overlaps Precambrian crust between Liberia and northernmost Brazil by about 30 km, demonstrating the need for minor improvement. A revised Equatorial Atlantic reconstruction is shown in Figure 1.8, which aligns the prerift limits of continental crust along the northern Brazil and Guinea margins. These limits were determined by removing marginal sediments and restoring the inferred syn-rift extension, as in Figure 1.5 for the Gulf margin. This fit describes the interplate relationship between northern Africa and South America prior to Aptian plate divergence in the Equatorial Atlantic and throughout the opening of the Gulf of Mexico.

The Yucatan Block:

The Yucatan block played an important role in the evolution of the Gulf of Mexico. Its rotated prerift position against Texas and Louisiana as portrayed by Pindell and Dewey [1982] accounts for such problems as the Ouachita salient and the apparent absence of Late Permian to Triassic marine rocks in the Gulf region by allowing complete continental closure by the Permian. Further, magnetic anomaly lineations in the central Gulf of Mexico are oriented nearly east-west and diverge to the west [Shepherd et al., 1982; S. Hall, personal communication, 1984]. This trend is clearly distinct from the Central North Atlantic trend and is consistent with an origin of the Gulf of Mexico by counterclockwise rotation of Yucatan, as suggested by Pindell and Dewey [1982].

Rock types also constrain this initial position of Yucatan. The Chuacus Group in Central Guatemala is a metamorphosed (greenschist to amphibolite) collection of marbles, quartzites, greenstones, schists, serpentinites and

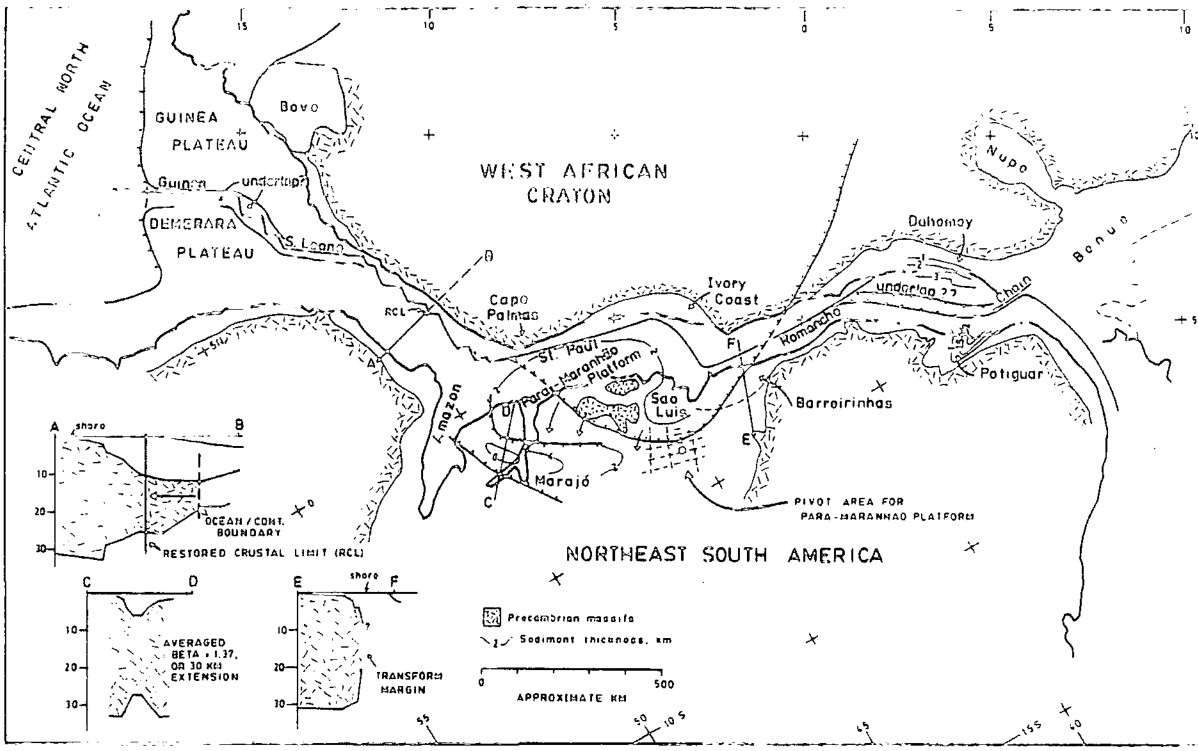


Figure 1.8. Prerift Equatorial Atlantic reconstruction, matching restored limits of pre-Mesozoic continental crust of the continental margins, methodology for which shown in Figure 1.5. Following "Hypothesis II" of Delteil et al. [1976], St. Paul fracture zone juxtaposes Cape Palmas with the mouth of the Amazon, the Romanche juxtaposes Cape of Three Points with the Berreirinhas Basin and the Chain juxtaposes the Ilesha Spur of southwest Nigeria with the Potiguar Basin. Guinea fracture zone margins well defined by Behrendt et al. [1974], Delteil et al. [1976], McMaster and Ashraf [1973], and Sibuet and Mascle [1978]. Isopachous data from Asmus and Ponte [1975], Milliman [1979], and Whiteman [1982]. Restoration of prerift continental limit beneath Niger Delta from Delteil et al. [1976]. Cross sections AB and CD constructed from figure 5 of Milliman [1979], and section EF from Ojeda [1982] and Asmus and Ponte [1975]. Removal of sediment at Amazon mouth, and restoration of attenuation in the Amazon and Marajo Basins, allows this tighter fit. Clockwise rotation of the Para-Maranhao Platform during plate separation produced the westward deepening Marajo Basin. The suggested pivot point for closure of Marajo Basin (arrows denote closure direction) is amidst area of basement fractures, as shown. Northern edge of Para-Maranhao Platform is a fracture zone. Note good alignment of Sao Luis and West African Cratons (heavy line with adjacent dots). Also, gap between Demerara and Guinea Plateaus seen in the Bullard et al. [1965] fit is avoided, which explains the absence of Jurassic marine sediment east of the plateaus. The adverse effect of this Equatorial fit on the southern South Atlantic fit can be reconciled by restoring extensional deformation in central Africa [Wright, 1968], and in Argentina. See Table 2.2 of chapter 2 for total rotation parameters.

volcanic rocks, with granodioritic intrusions, whose age is known only to be pre-Permian on the basis that it is unconformably overlain by the Permian Santa Rosa Group [Roper, 1978; Van den Boom, 1972; Anderson et al., 1973]. Preserved within a section apparently of the Chuacus Group that was only moderately metamorphosed, however, are Carboniferous (probably Lower Carboniferous) fossils [Van den Boom, 1972]. Thus, regardless of the age-range of the Chuacus stratigraphy, the age of metamorphism apparently is Carboniferous. Isotopic evidence [Gomberg et al., 1968] on the Rabinal Granite, a probable synmetamorphic melt from the granitization of a Chuacus arkose [Van den Boom, 1972], produces possible ages of 1075 Ma plus or minus 25 and 345 Ma plus or minus 20. The latter is compatible with a Lower Carboniferous age. It appears, therefore, that the deformation and metamorphism exhibited by the Chuacus Group can be attributed to Alleghenian orogenesis. Alleghenian orogenesis as modeled by Dewey [1982] would suggest a prolonged period of deformation and metamorphism for the rocks of Yucatan, as they occupied a shear zone between North and South America from latest Devonian to earliest Permian time. Further support for proximity of northwestern South America and North America by Late Devonian time is based on the presence of Appalachian basin marine invertebrate fauna of that age in Colombia and Venezuela [Boucot, 1975; Barrett, 1983]. Sediments of the unconformably overlying Santa Rosa Group (Permian) probably were shed from exposed portions of the Chuacus, and other structural highs in the orogenic zone, into shallow marine molasse basins during the final stages of orogenesis, as they are far less deformed and relatively unmetamorphosed.

In the Maya Mountains of Belize (Figure 1.2), a variably deformed, argillaceous to conglomeratic sequence of Pennsylvanian-Middle Permian age overlies older (?)gneissic basement with granitic intrusions [Dawe, 1984; Nelson, 1984; Bateson and Hall, 1977; Dixon, 1957]. At least the upper part of the sedimentary section consists of shallow-water, fossiliferous molassic sands, shales and conglomerates, and is lithologically and biostratigraphically correlative with the Santa Rosa Group of Central Guatemala [Anderson et al., 1973]. These deposits are very similar in lithology and age to the Desmoinesian to Lower Permian late to post-orogenic molasse deposits in the subsurface of the U.S. Gulf Coast [Woods and Addington, 1973] and to the Lower Permian strata of northwest South America [Shagam, 1975]. These localities align in a continuous belt behind the main Alleghenian thrusted zone to the north, providing support for the reconstruction from a sedimentological perspective.

In the subsurface of northern Yucatan, meta-andesite/dacite was drilled (Yucatan No. 1, see Figure 1.2) and isotopically dated as 330 Ma and 290 Ma [Marshall, 1974]. The andesite may represent an occurrence of the Alleghenian arc that theoretically should have existed; the isotopic age is consistent with a thermal resetting during the period of closure between Gondwana and Laurussia. At Catoche and a neighboring knoll of the Yucatan block, DSDP holes 538A and 537 (Figure 1.2) reached Paleozoic basement as well [Schlager et al., 1984]. Hole 538A sampled amphibolite gneiss with isotopic ages of 496 and 348 Ma, and hole 537 obtained phyllite with ages of 449 and 456 Ma. The 348 Ma age can be attributed to thermal resetting during the Alleghenian orogeny [Schlager et al., 1984], whereas the older ages are similar to many of those recovered from the basement of Florida [Smith, 1982]. That Florida belonged to Gondwana prior to Alleghenian collision is clear from its pre-Carboniferous Afro-South American fauna [Cramer, 1971; Pojeta et al., 1976]. Similarities between Yucatan and Florida would imply the same for Yucatan. Lithologies and Upper Paleozoic isotopic ages of 270 to 325 Ma on rocks (phyllite, granite) from a well penetrating the Wiggins Arch (Figure 1.2) [Cagle and Khan, 1983; J. Cagle, personal communication, 1984] are similar to those recovered from the Yucatan block and may suggest that the Wiggins Arch was part of Gondwana as well. All lie to the south of the proposed Alleghenian suture.

Synthesis of the Alleghenian Reconstruction:

The above considerations provide geological and quantitative geometrical constraints on reconstuctions of western Pangea. Although Middle to Late Paleozoic convergence between Laurussia and Gondwana is poorly understood, patterns of Alleghenian orogenesis and Late Paleozoic sedimentation appear to be best explained by complete ocean closure during continent-continent collision, within which the Yucatan block filled the gap between Venezuela and the U.S. Gulf Coast margin (Figure 1.3). The probable suture zone, from east to west, lies between the Appalachians and the Mauritanides of western Africa; crosses Georgia between the Suwannee Basin and the Southern Appalachians; continues north of the Wiggins Arch and Sabine Uplift, following approximately the trend of the Gilbertown-South Arkansas-Mexia graben system; is offset by the Texas Lineament along the Devil's River Uplift (essentially a tear fault); swings to the south of the Marathons but again is offset, possibly by a proto-Pedregosa Trough fault system, and passes to the northwest of the Coahuila Platform and then south between the Huastecan belt of Mexico and northwest South America. Whether the Chortis block was part of Mexico during Late Paleozoic time or whether it arrived later after migration along the Cordillera is unknown and, hence, Chortis is excluded from the reconstruction.

Accretionary complexes that were developed during ocean closure include the Ouachitas, the Marathons and the Huastecans. All of these accretionary complexes overthrust Paleozoic shelf rocks of North America, suggesting that final convergence was achieved by southeast dipping subduction with North America as the downgoing plate. The case for Alleghenian accretionary prisms in the southern Appalachians is questionable, but these may lie to the east beneath the coastal plain.

Alleghenian thrust-loaded foreland basins which developed on North America include the Appalachian, Black Warrior, Arkoma, Fort Worth, Ker, Val Verde, Marfa, Pedregosa basins and the Mexican Paleozoic "Geosyncline". Distal, foreland bulges include the Cincinnati Arch, Nashville Dome, Ozark Dome, and Llano Uplift.

Possible remnants of the expected arc between Gondwana and North America include the Upper Paleozoic granodiorites, andesites or volcanics presently found in the Central Cordillera of Colombia [Irving, 1975], the Chuacus Group of Guatemala, Yucatan No. 1 well of Yucatan, the Coahuila Platform, the Sabine Uplift, and the basement of Florida. Zones of Late Paleozoic strike-slip motion leading into the North American foreland which may have allowed escape of continental fragments include the Rough Creek Fault Zone, the Anadarko Basin/Wichita Uplift, the Delaware Basin [Dewey, 1982] and, possibly, the Pedregosa Trough. Closure was achieved by east to west oblique collision from the latest Devonian through the middle Permian, with probable migrations of various blocks within the suture zone.

CHAPTER 2

PLATE-KINEMATIC FRAMEWORK FOR GULF/CARIBBEAN EVOLUTION

PLATE KINEMATIC FRAMEWORK FOR CARIBBEAN EVOLUTION:

Introduction:

In cases where two continents are separated by an Atlantic-type ocean, relative paleopositions of the continents can be determined by finite plate rotations which realign continental margins and magnetic anomaly pairs of equal age, along flow lines defined by fracture zone traces. The relative motion history of the continental pair may then be determined by tracing the trajectory of one continent, relative to the other, through the successive relative paleopositions, throughout the spreading history of the intervening ocean [Pitman and Talwani, 1972].

In the case of the relative motion history of North and South America, however, a simple two-plate analysis like the above cannot be employed. This is because oceanic crust of the Caribbean Plate is allochthonous with respect to the Americas, such that it probably is not related to the initial separation of North and South America. Therefore, paleopositions of South America relative to North America are best determined by finite difference solutions at various times for the three-plate system of South America, Africa, and North America [Ladd, 1976]. The relative motion histories of both the Central and South Atlantic Oceans can be documented, and, from these, the paleopositions and the relative motion history of North and South America can be computed. As with the two-plate system, the relative-motion history of North and South America using the three-plate system may be portrayed by trajectories, or flow-lines, through the relative paleopositions through time.

The relative paleopositions of North and South America constrain Gulf/Caribbean evolutionary models by defining the size and shape of the Gulf/Caribbean intra-plate realm at intervals through time. The relative motion history vector, however, defines the direction and rate of change in the size and shape of the interplate realm. Thus, the relative motion vector directly constrains the nature of the North America-South America plate boundary through time. For example, periods of plate divergence, as shown by the relative motion vector, would suggest seafloor spreading at a particular rate at an intervening mid-ocean ridge system, whose transforms parallel the separation direction. Likewise, a period of no relative motion would suggest the absense of a North America-South America plate boundary for that period. These deductions are not empirical, however, in that motions upon combinations of plate boundaries may sum to equal the net motion indicated by the relative motion vector. Such complications must be inferred from the rock record.

Other plates whose motions, relative to North America, are important to the evolution of the Gulf/Caribbean region are the oceanic plates of the Pacific realm. The direction and rate of convergence of these plates may have controlled the style of subduction beneath, and possible microplate migrations along, the Pacific subduction systems of the Americas and the Caribbean Plate. These include the Farallon for the Cretaceous and Paleogene, and the Cocos for the Miocene to Recent. The Cocos is partly a remnant of the Farallon, the two having been separated by subduction of a portion of the Farallon/Pacific ridge and transform system in the middle Tertiary [Atwater, 1970]. This, and the Early Miocene initiation of seafloor spreading at the Galapagos Ridge [Klitgord and Mammerickx, 1982; Mammerickx and Klitgord, 1982], have resulted in differing motions of the Farallon and Cocos Plates, relative to North America, since the Early Miocene. Other

Pacific oceanic plates may have interacted with the Americas during the Late Triassic through earliest Cretaceous, but these cannot be identified with existing knowledge of Pacific plate-motion history.

In this chapter, 15 paleopositions and the relative motion history are defined for North America/Africa, and 10 paleopositions and the relative motion history are defined for South America/Africa. From these, 15 paleopositions and the relative motion history for South America/North America are computed. The difference in the number of paleopositions arises from the fact the initial rifting in the Equatorial Atlantic between northern Africa and northeast Brazil (Aptian) postdates that between Africa and North America (early Middle Jurassic). Hence, for the Middle Jurassic to Aptian interval, the spreading history of the Atlantic can be described as a two-plate system, involving relative motion of North America and Gondwana. In the southern South Atlantic Ocean, rifting began in the latest Jurassic and propagated northwards throughout the Early Cretaceous [Rabinowitz and LaBrecque, 1979], entering Benue Trough rather than the Equatorial Atlantic [Pindell and Dewey, 1982; Klitgord et al., 1984]. These factors do not affect the Gulf/Caribbean region, however, and are not treated here in any detail.

The time intervals and magnetic anomalies for which paleopositions and relative motions are defined are: early Middle Jurassic, Bathonian (Blake Spur), Late Callovian, Tithonian (M-21), Early Berriasian (M-16), Early Aptian (M-0), Late Albian, Early Campanian (34), Late Campanian (32), Late Paleocene (25), Middle Eocene (21), Early Oligocene (13), Early Miocene (6), Late Miocene (5), Present.

<u>Farallon and Cocos with respect to North America:</u>

Engebretson [1982] has analyzed Farallon/North America relative motion history since the Early Cretaceous. Unfortunately, estimates of pre-Campanian relative motion may be inaccurate, due to the absence of magnetic polarity reversals during the Cretaceous Magnetic Quiet Period, and to the possibility of the existence of unknown plates and plate-boundaries in the Pacific prior to that time. Engebretson's poles are in part derived from the assumption that a hot-spot reference frame exists within the mantle, such that hot-spot traces in various ocean basins are fixed in space and may be used to determine relative motions between pairs of plates, each of which possesses hot-spot tracks. This may or may not be true, but close agreement with Engebretson's analysis of Farallon/North America relative motion comes from integration of relative motions between members of a global circuit of plates only, using an east and a west Antarctica [Jurdy, 1984]. Further, the predicted relative motion history is iteratively supported by cause-and-effect relationships between plate motions and the geological development of the North America Cordillera [Page and Engebretson, 1984]. Despite the potential problems, and in lieu of a better analysis, Engebretson's Farallon/North America poles are assumed to be satisfactory for the entire interval Cretaceous to Oligocene, after which time relative motion of the Cocos Plate becomes more important.

Determining relative motion between the Cocos and North American Plates is highly conjectural, involving uncertain assumptions. First, an age must be estimated for the time when the Cocos began behaving independently of the Farallon. This appears to have occurred at about 22 to 25 Ma, as marked by the initiation of seafloor spreading at the Galapagos Ridge [Mammerickx and Klitgord, 1982]. Therefore, in this analysis, the Pacific plate of importance to vector triangle completion with North and South America for the times of anomalies 6 and 5 (21 Ma and 9 Ma, respectively) is the Cocos

Plate, rather than the Farallon.

The three-plate system of Cocos-Pacific-North America was used to determine Cocos/North America relative motion for the Neogene. Engebretson's Pacific/North America poles were used as a known, and this is perhaps one source of error, as they are the result of finite difference solutions themselves. The Cocos/Pacific poles for 21 Ma and 9 Ma are extrapolated from the instantaneous pole and rotation rate of Klitgord and Mammerickx [1982]. This extrapolation is another source of error, but, despite local ridge jumps, the spreading rate and pole position for Cocos/Pacific relative motion seems to have been quite constant throughout the Neogene [Mammerickx and Klitgord, 1982; Klitgord and Mammerickx, 1982]. Using these sources, rotation parameters describing Cocos/North America relative motion for the Neogene were computed.

In Table 2.1, Engebretson's Farallon/North America pole data have been interpolated or summed to define the total finite and stage pole parameters at the specific time intervals used in this study. In this way, Farallon/North America relative motion can be easily related to the calculated relative motions of North and South America. For the Neogene, Cocos/North America poles are listed in Table 2.1, as defined above. The Cretaceous-Paleogene Farallon/North America and Neogene Cocos/North America relative motions for the various time intervals are shown in Figure 2.1. The relative motion vectors for the intervals are calculated for latitude/longitude pairs at 10N, 90W and 5N, 80W, respectively, which lie in the Pacific-Caribbean interface. Hence, the vectors are useful for constructing velocity triangles involving the North American, the South American, the Caribbean, and the Farallon and Cocos Plates for the various time intervals used in this study.

Some implications of Figure 2.1 for Caribbean geological evolution are

TABLE 2.1. ROTATION PARAMETERS, FARALLON/COCOS WITH RESPECT TO NORTH AMERICA

	total finite parameters				stage pole parameters		
Anomaly	Age (Ma)	lat.	long.	angle	lat.	long.	angle
M-16	141	-27.0	295.0	104.0			
					14.5	54.2	18.4
M-0	119	-35.0	303.0	95.1			
					60.7	148.1	11.7
--	100	-31.0	305.0	85.2			
					12.1	112.9	15.7
34	84	-36.0	304.0	70.8			
					8.4	106.8	14.5
32	72	-43.7	303.7	58.9			
					30.9	100.2	19.2
25	59	-52.0	313.0	41.8			
					53.7	137.1	11.7
21	49	-51.0	312.0	30.1			
					61.7	169.8	16.1
13	36	-31.0	298.0	16.5			
					47.3	149.2	9.8
6	21	-5.6	279.3	9.2			
					24.1	-112.8	19.8
5	9	23.0	-117.2	13.8			
					23.0	-117.2	13.8
0	0	-.-	-.-	-.-			

1. Positive numbers are degrees north or east, or counterclockwise rotations looking down at pole.
2. Rotation angles for total finite rotations go back in time to the indicated anomaly, and rotation angles for stage pole rotations are coming forward in time, for the indicated interval.
3. Farallon parameters are after, or interpolated from, Engebretson [1982].
4. For anomalies 6 and 5, poles describe Cocos/North America motion rather than Farallon/North America (see text).

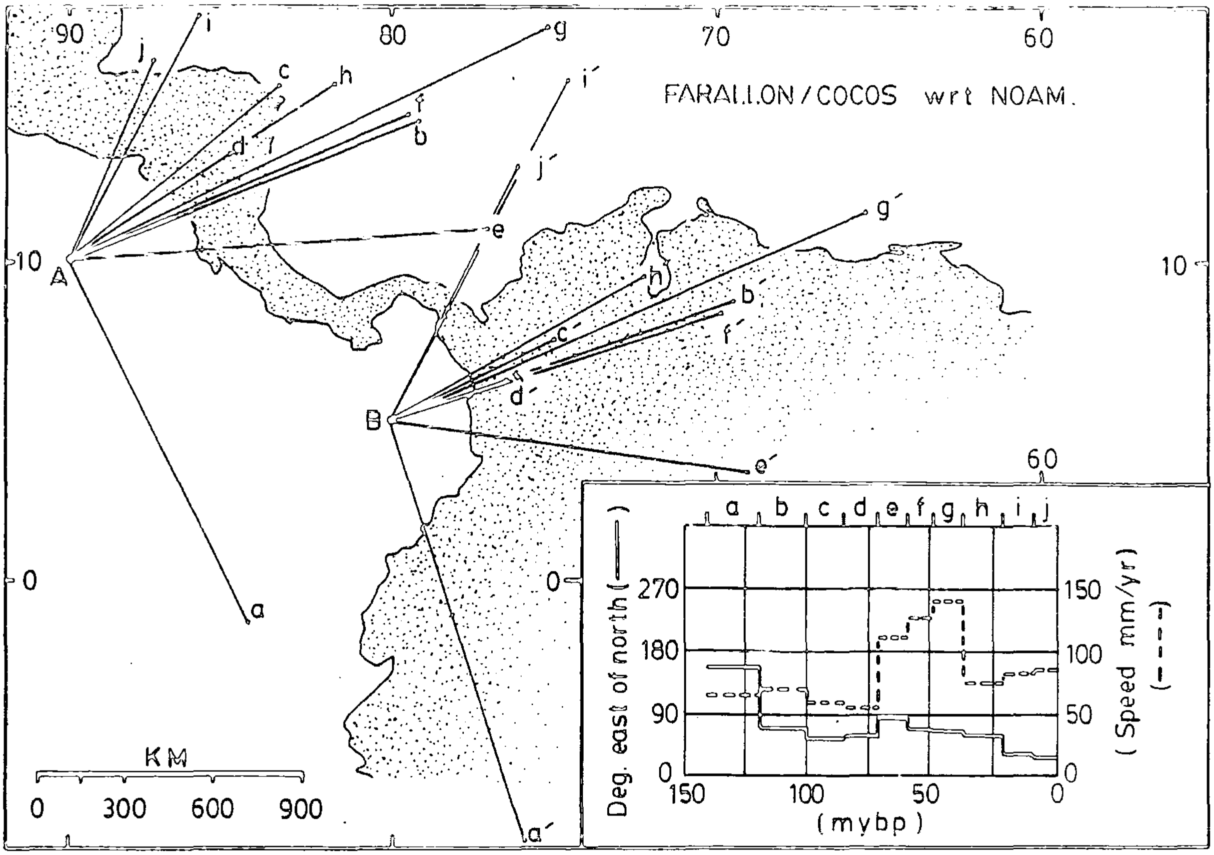


Figure 2.1. Stage motion vectors of the Farallon/Cocos Plate, relative to North America, determined for two points (10N, 90W) and (5N, 80W), from stage pole data of Table 2.1. Farallon Plate became Cocos Plate in Late Oligocene or earliest Miocene (see text). Inset: Graph of direction and velocity of Farallon/Cocos Plate measured at Point A (10N, 90W). Time intervals labeled with letters a through j correspond to the vectors of the figure. Interpretation of vectors given in text.

as follows. A strong component of sinistral shear along the early Cordilleran trench is indicated for the Early Cretaceous (Berriasian to Aptian). This shear may have been responsible for the southeastward migration of blocks of Cordilleran Mexico into South America's "overlap position" seen in most reconstructions of western Pangea (e.g., Bullard et al. [1965]). Beginning in the Aptian and continuing through the Campanian, the motion of Farallon with respect to North America was northeast directed at rates of 50 to 70 mm/yr. This period is associated with the early plutonic and volcanic development of the Greater Antilles arc; the initiation of head-on convergence between the Farallon Plate and the Proto-Caribbean may have been responsible for the formation of a subduction zone between the two. Fast rates of convergence (110 to 150 mm/yr) are indicated for the Late Campanian to Late Eocene, which is directly related to Laramide Orogenesis in and around the Caribbean and to the insertion of the Caribbean Plate into the gap between North and South America. Finally, Oligocene-Recent northeastward convergence has occurred at rates of 70 to 80 mm/yr. The sharply reduced convergence rate may be associated with an Oligocene westward shift in the axis of volcanism (theoretical steepening of Benioff Zone) from the Cordillera Central to the Western Cordillera of Colombia and Ecuador (see chapter 6).

A final note worth mentioning regards the extrusion of medial Cretaceous basalts (seismic horizon B") onto the Caribbean Plate. It has been suggested that one cause of the intra-plate volcanism and anomalously high topography of the African continent relative to other continents is that it rests nearly in a mantle reference frame, and that heat has built up beneath the continent as a result [Burke and Dewey, 1973]. If Africa is approximately in the mantle reference frame, the Americas have been migrating westward over the mantle during the Atlantic's opening.

Subduction of the Farallon Plate beneath the Americas at rates that are similar to the spreading in the Atlantic would imply that it, like Africa, may have had, at times, little motion relative to the mantle. This may be true for the medial Cretaceous; for the period 120 Ma to 100 Ma, Engebretson [1982] shows that the Farallon Plate's relative motion with the hypothetical hot spot reference frame is slower than 10 mm/yr. The extrusion of the B" horizon basalts onto the crust of the Caribbean occurred from the Cenomanian? to the Campanian, shortly following this interval of little relative motion. A possible explanation for the basaltic extrusion is that excess heat was accumulated beneath the Caribbean Plate during the Aptian-Albian, which caused partial melting and off-axis, intra-plate volcanism over the subsequent 20 million years.

South America with respect to Africa:

In chapter 1, a continental reconstruction between northern Brazil and the Guinea margin of Africa is advanced that is tighter than the fit proposed by Bullard et al. [1965]. Rotation parameters for this fit are given in Table 2.2. The beginning of the rift-related stage of basin formation and initiation of Brazil/Guinea relative motion across the Equatorial Atlantic is Aptian [Asmus and Ponte, 1973; Rabinowitz and LaBrecque, 1979], although formation of these basins was significantly preceded by the emplacement of dikes along the western Guinea margin. The first marine sediment in any of the Equatorial Atlantic's marginal basins is Albian [Asmus and Ponte, 1973]. It is assumed, for lack of conflicting evidence, that northern Africa and northern South America remained a single continent until the Aptian. Therefore, it is also assumed that the early evolution of the southern South Atlantic, which began in the earliest

TABLE 2.2. ROTATION PARAMETERS, SOUTH AMERICA WITH RESPECT TO AFRICA

		total finite parameters				stage pole parameters		
Anomaly	Age (Ma)	lat.	long.	angle		lat.	long.	angle
M-0	119	52.08	-34.03	51.39	P			
						35.0	-22.4	-9.7
--	100	56.5	-34.7	42.3	*			
						35.0	-22.4	-9.7
34	84	63.0	-36.0	33.6	$			
						63.0	-36.0	-6.3
32	72	63.0	-36.0	27.3	$			
						63.0	-36.0	-4.7
25	59	63.0	-36.0	22.6	$			
						63.0	-36.0	-3.4
21	49	63.0	-36.0	19.2	$			
						73.4	-46.4	-6.1
13	36	58.0	-35.0	13.3	$			
						53.6	-34.1	-5.8
6	21	61.4	-35.0	7.5	$			
						53.7	-34.2	-4.0
5	9	70.0	-35.0	3.6	$			
						70.0	-35.0	-3.6
0	0	-.-	-.-	-.-				

1. Positive numbers are degrees north or east, or counterclockwise rotations looking down at pole.
2. Rotation angles for total finite rotations go back in time to the indicated anomaly, and rotation angles for stage pole rotations are coming forward in time, for the indicated interval.
3. P is derived in chapter 1.
4. $ are modified after Ladd [1976], as discussed in text.
5. * is interpolated between paramters for M-0 and 34.

Cretaceous [Norton and Sclater, 1979], bears no relationship to the evolution of the Gulf and Caribbean region.

Figure 2.2 outlines a possible model for the Early Cretaceous opening of the southern South Atlantic. Southern Africa drifted from southern South America from anomaly M10 time to about anomaly M-0 time (Aptian). Prior to the Cretaceous, the shape of Africa was not that of today. Northward rift propagation which terminated within central Africa via the Benue Trough opened the southern South Atlantic and gave Africa the shape that it has today. Toward the end of this early phase of opening, the Rio Grande-Walvis Ridge hot-spot trace was breached by southern sea waters, depositing the Aptian salts in the narrow trough to the north. In the Aptian, rifting jumped from central Africa to the future site of the Equatorial Atlantic, and the shape of Africa has remained to the present essentially as it was at the end of the Aptian. Therefore, marine magnetic anomaly identifications in the southern South Atlantic since the Aptian can be used to determine South America/Africa paleopositions and relative motion history, for use in determining South America/North America relative motions.

Ladd [1974] analyzed the marine magnetic anomalies in the South Atlantic Ocean. At the time of Ladd's study, it was convention to pick the anomalies at the peak of the signal. More recently, accuracy in the method has become such that significant differences arise from picking different parts of the signals of some anomalies, and it is present convention to pick the inner (younger) sides of the signals. South Atlantic magnetic data were re-examined by the author, and it was found that Ladd's poles of rotation were quite accurate. However, the rotation angles were altered slightly to account for the differences between picking the anomalies at their peaks and at their inner sides, and to be consistent with the method used for picking the anomalies in the Central Atlantic. The poles of rotation of Ladd [1974;

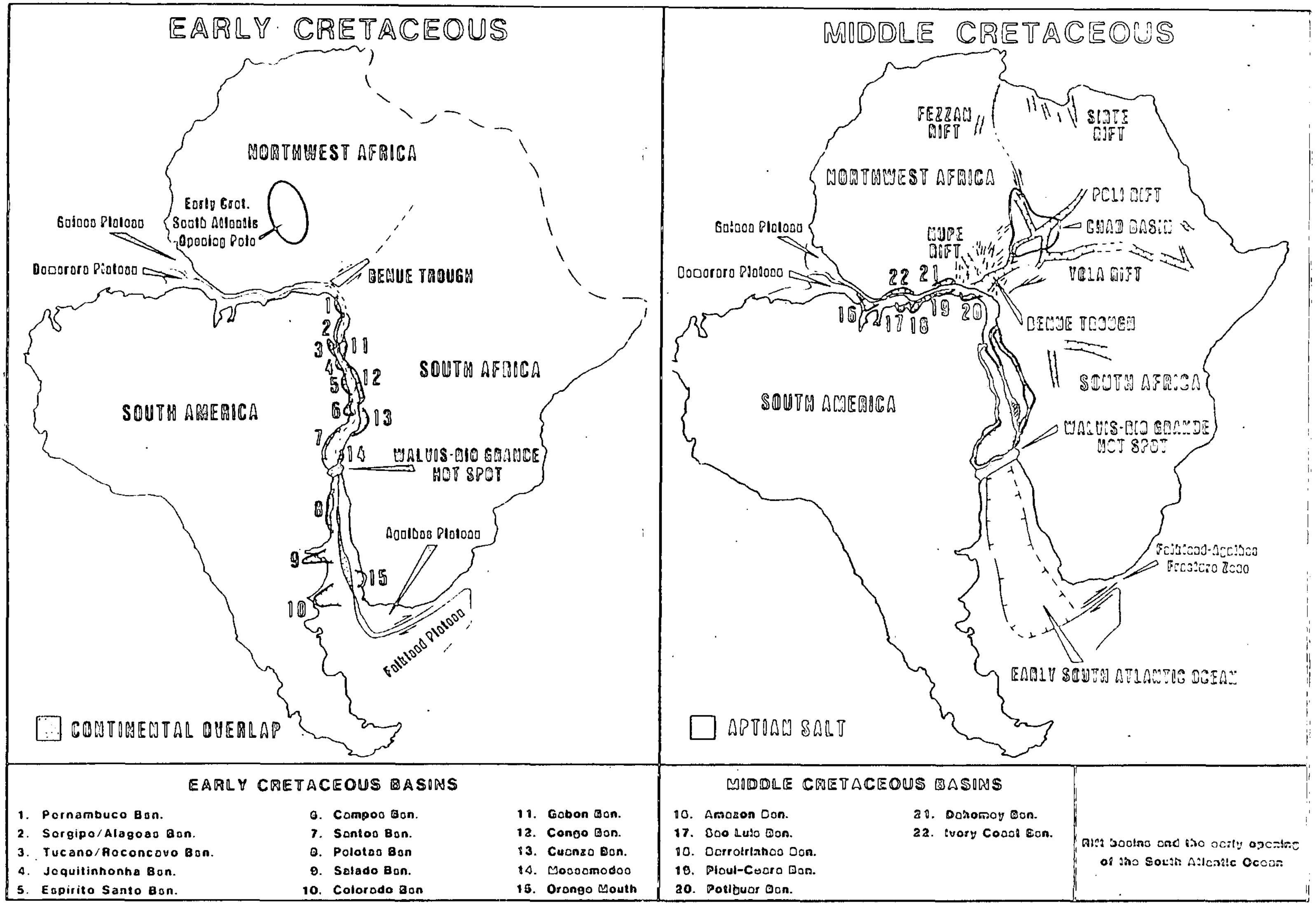


Figure 2.2. Simple model for the early opening of the southern South Atlantic Ocean. Early Cretaceous rifting propogated northwards and entered central Africa, rather than the Equatorial Atlantic. Africa was deformed and assumed approximately its present shape by anomaly M-C time, at which time rifting broke through the Equatorial Atlantic. The Aptian Salts of the South Atlantic resulted from evaporation of spills across the early Rio Grande-Walvis Ridge from the south [Burke, 1975].

1976] and the ammended rotation angles are shown in Table 2.2. Figures of the ammended picks are not included herein.

Africa with respect to North America:

At the time of this study, Keathley Sequence rotation parameters (Late Jurassic through Aptian) recently determined by K. Klitgord and H. Schouten were available to the author (personal communication). However, parameters for the time following the Cretaceous Quiet Period (Campanian through Present) had not been computed since the work of Pitman and Talwani [1972]. Bathymetric data collected through the seventies (GEBCO map of the Central North Atlantic Ocean) showed very accurately the trace of fracture zones across the Central Atlantic. A substantial "hitch" in the flow lines beginning at anomaly 34 (Campanian) and continuing to anomaly 21 (Eocene) was clearly indicated. As this hitch was unknown at the time of Pitman and Talwani's study, it was evident to the author that significant improvement could be made upon rotation parameters defining the Central Atlantic's spreading history.

The least structurally disturbed portion of the Central Atlantic Ocean between Africa and North America lies between the Atlantis and Carolina fracture zones, which includes the Kane fracture zone. The mid-ocean ridge system in this area is remarkably straight, with few offsets, for a distance of about 1500 km, making anomaly identification relatively easy as compared to areas where the ridge system is transected by closely-spaced transforms. Anomalies 34, 32, 25, 21, and 13 were picked over a 1500 km wide swath across the Central Atlantic (Figure 2.3). The GEBCO bathymetric maps provided the traces of the Atlantis and Kane fracture zones, which were used to control the trajectories of the plate motion paths. Updated rotation

Figure 2.3. Map of a portion of the Central North Atlantic, showing the picks used to compute the poles of rotation for anomalies 34, 32, 25, 21, and 13 of Table 2.3. The Atlantis and Kane fracture zones were used as flow lines along which Africa and North America were separated.

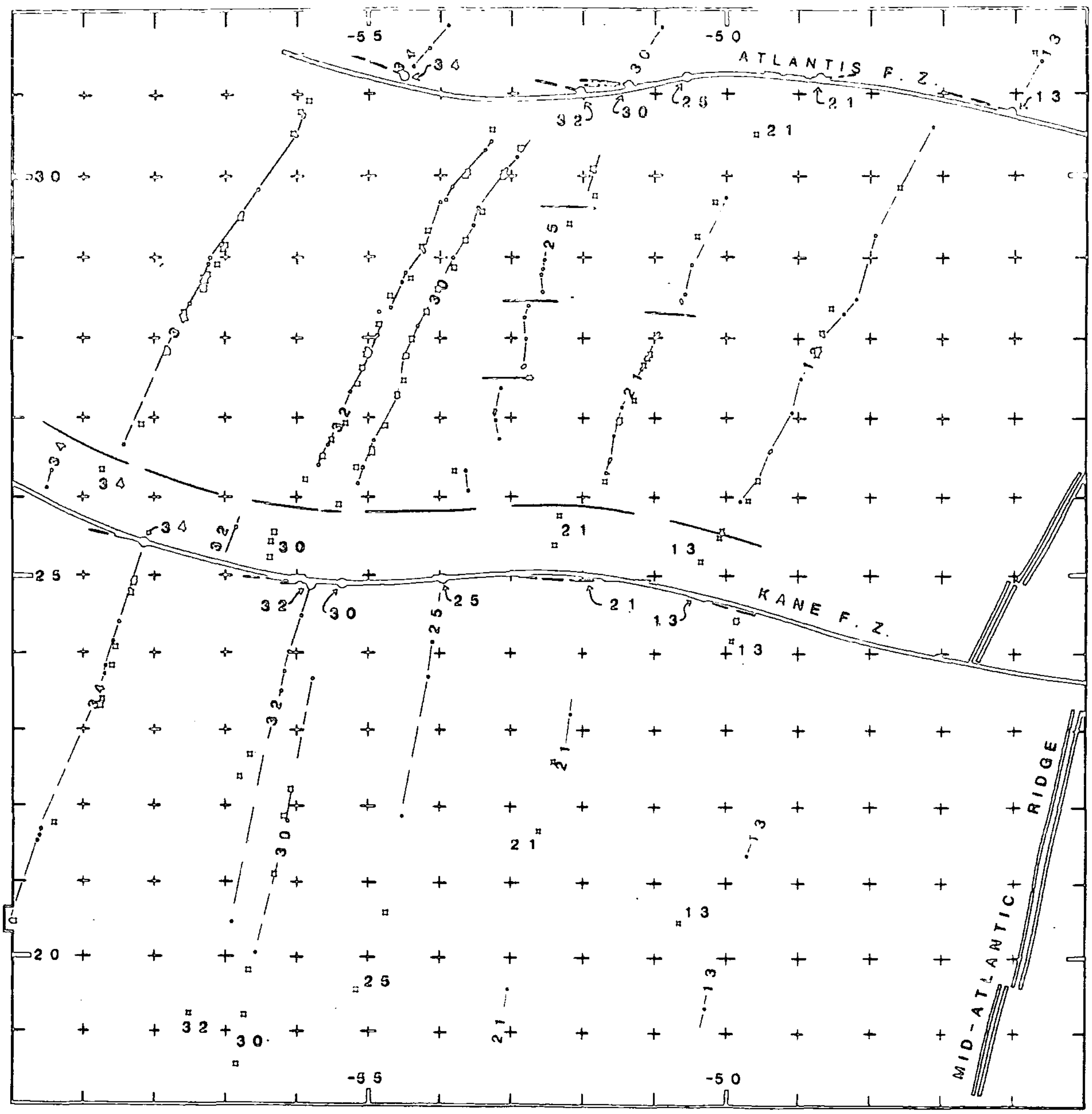


Figure 2.4. Map of western Central-North Atlantic showing the magnetic anomaly picks of Figure 2.3. Small solid circles are original picks on the western half of the ocean, crosses are the eastern picks rotated about the total poles of rotation determined for each anomaly pair (Table 2.3). Segments of the eastern Atlantic's fracture zones are also rotated and shown (heavy dashed lines). Congruence of fracture zones is imperfect because the fracture zones record flow lines about instantaneous stage poles during seafloor spreading, not the total finite poles. However, the points along the eastern Atlantic's fracture zone traces that were adjacent to points along the western Atlantic's fracture zone traces during the formation of the corresponding magnetic anomaly are shown in large solid dots and large open circles, respectively. The accuracy of the poles of rotation is shown by the congruency of the juxtapositions of the anomaly picks and fracture zone offsets, for each anomaly pair. For clarity, anomaly labels (such as 34, 25, etc.) that are horizontal correspond to rotated eastern Atlantic picks.

parameters for anomalies 6 and 5 were provided by K. Klitgord and H. Schouten (personal communication) during the course of the study.

The fracture zone traces and the picks for anomalies 34, 32, 25, 21, and 13 from the eastern Central Atlantic were rotated to those for the western Central Atlantic, and best fitting poles and rotation angles were determined. The rotated eastern picks and fracture zone traces and the non-rotated western picks and fracture zone traces are plotted in Figure 2.4. This figure shows the degree of coincidence of the two sets for each anomaly examined. Assuming the area studied (central portion of the Central Atlantic) is indicative of the entire North America-Africa plate boundary, the magnitude of possible error of the indicated relative paleopositions is less than 40 km. The close spacing of fracture zones makes other portions of the Central Atlantic less attractive than the central portion for magnetic anomaly identification.

Table 2.3 lists the rotation parameters, including those provided by K. Klitgord and H. Schouten, defining the spreading history of the Central Atlantic Ocean.

South America with respect to North America:

By finite difference solution [Dewey et al., 1973], rotation parameters defining the paleopositions of South America relative to North America were computed from the data in Tables 2.2 and 2.3. The results are given in Table 2.4. From this data, the relative motion history vector of South America with respect to North America may be plotted.

Figure 2.5 shows relative motion vectors of South America with respect to North America for three data sets, including the above. The vector of Pindell and Dewey [1982] is not shown because it is an amalgamation of data

TABLE 2.3. ROTATION PARAMETERS, AFRICA WITH RESPECT TO NORTH AMERICA

		total finite parameters				stage pole parameters		
Anomaly	Age (Ma)	lat.	long.	angle		lat.	long.	angle
Permian	~250	65.55	-13.69	-77.15	P			
						26.7	-63.4	2.5
M. Jur.	~180	66.95	-12.02	-75.55	K			
						60.0	0.0	3.5
BSMA	~170	67.02	-13.17	-72.10	K			
						60.0	0.0	6.3
Callovian	~160	67.13	-15.49	-65.93	*			
						61.1	22.0	4.2
M-21	~150	66.50	-18.10	-61.92	K			
						72.0	9.4	2.2
M-16	~140	66.10	-18.40	-59.79	K			
						61.3	-10.2	5.6
M-0	119	66.30	-19.90	-54.25	K			
						55.02	-29.90	12.95
--	100	70.16	-19.73	-41.68	*			
						55.02	-29.90	12.95
34	84	77.10	-19.20	-29.42	K,P			
						65.64	-36.12	6.47
32	72	80.35	-12.45	-23.15	P			
						75.19	-61.50	5.38
25	59	80.50	7.70	-17.90	P			
						85.10	-68.63	3.92
21	49	78.30	16.00	-14.05	P			
						81.12	49.01	4.39
13	36	76.00	8.00	-9.70	P			
						69.68	-11.18	4.50
6	21	79.57	37.84	-5.29	K			
						76.82	11.26	2.92
5	9	79.08	77.95	-2.41	K			
						79.08	77.95	2.41
0	0	-.-	-.-	-.-				

1. Positive numbers are degrees north or east, or counterclockwise rotations looking down at pole.
2. Rotation angles for total finite rotations go back in time to the indicated anomaly, and rotation angles for stage pole rotations are coming forward in time, for the indicated interval.
3. K are after Klitgord and Schouten [1982; or personal communication].
4. P are parameters determined in this study.
5. * are interpolated between adjacent rotations.

TABLE 2.4. ROTATION PARAMETERS, SOUTH AMERICA WITH RESPECT TO NORTH AMERICA

		total finite parameters			stage pole parameters		
Anomaly	Age (Ma)	lat.	long.	angle	lat.	long.	angle
Permian	~250	54.92	40.81	-31.22			
					26.7	-63.4	2.5
M. Jur.	~180	53.78	48.44	-30.71			
					60.0	0.0	3.5
BSMA	~170	52.68	53.59	-27.63			
					60.0	0.0	6.3
Callovian	~160	49.12	64.97	-22.41			
					61.1	22.0	4.2
M-21	~150	45.84	72.09	-18.74			
					72.0	9.4	2.2
M-16	~140	42.06	75.65	-17.10			
					61.3	-10.2	5.6
M-0	119	28.61	90.18	-14.36			
					57.68	51.32	5.29
--	100	13.99	101.84	-10.82			
					57.68	51.32	5.29
34	84	10.30	-61.79	9.16			
					12.49	42.49	0.76
32	72	11.13	-57.05	9.04			
					54.52	151.17	1.15
25	59	18.10	-58.67	8.71			
					36.16	137.81	1.38
21	49	26.16	-60.51	8.03			
					40.94	-75.57	-2.34
13	36	19.68	-56.20	5.87			
					14.57	-60.98	-2.20
6	21	22.50	-53.07	3.70			
					11.42	-52.84	-1.99
5	9	34.83	-52.89	1.79			
					34.83	-52.89	-1.79
0	0	-.-	-.-	-.-			

1. Positive numbers are degrees north or east, or counterclockwise rotations looking down at pole.
2. Rotation angles for total finite rotations go back in time to the indicated anomaly, and rotation angles for stage pole rotations are coming forward in time, for the indicated interval.
3. All parameters deduced by finite difference method using data in Tables 2.2 and 2.3.

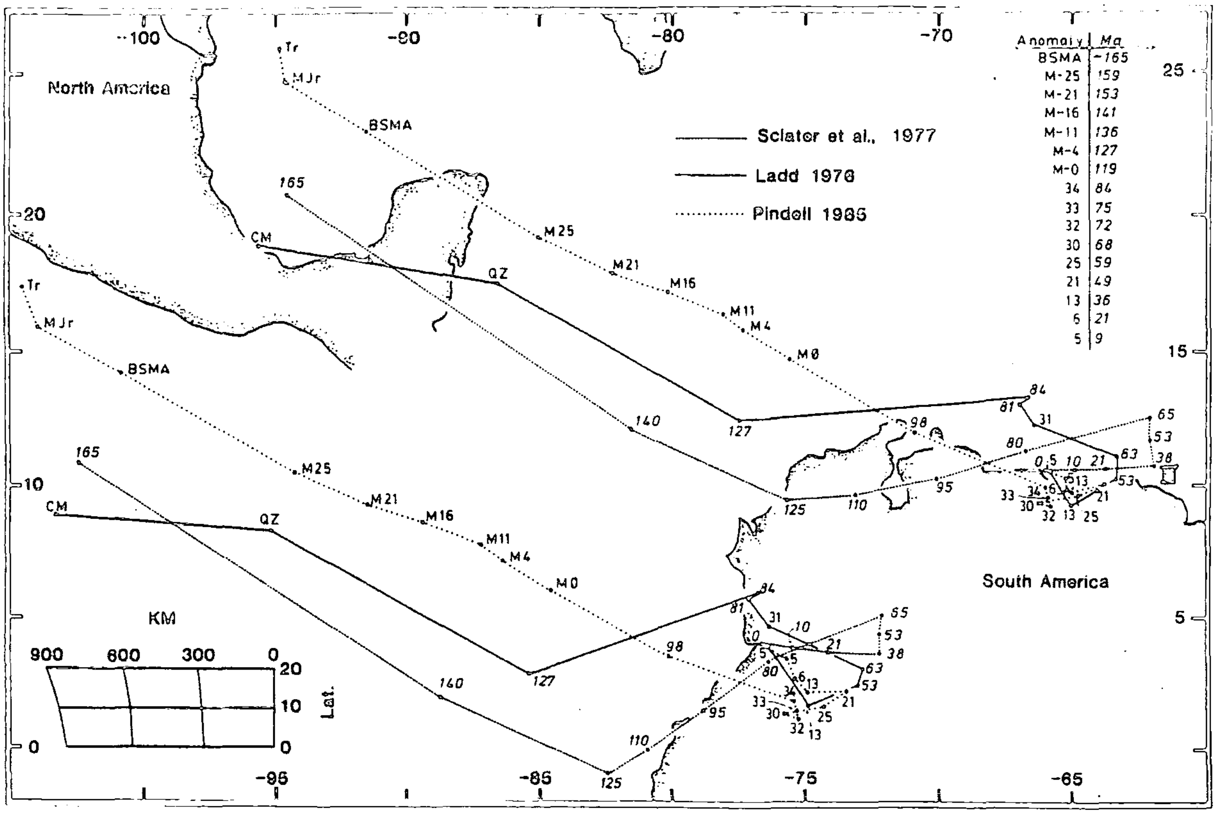


Figure 2.5. Three proposed post-Triassic relative motion paths, or trajectories, of two points on northern South America with respect to a fixed present North American reference frame, Mercator Projection. Positions identified by upright numerals were computed from triangle completion of Central and South Atlantic rotation parameters that were defined by magnetic anomaly identification. Inclined numerals indicate age in millions of years, and were interpolated between positions picked by anomaly identification. Differences in motion histories reflect improvements in magnetic anomaly data sets, and different initial reconstructions of South America with respect to Africa and North America. Trajectory of Pindell (this study) is derived from data of Table 2.4.

of the vectors shown. The three vectors of Figure 2.5 trace the positions through time of two points presently lying on the South American coastline. The differences in the vectors arise from differing initial reconstructions between North America, South America, and Africa, and from varied accuracies of magnetic anomaly identification and subsequent determination of total finite rotation parameters for the Central and South Atlantic Oceans. The vector of Pindell (this study) starts with the reconstuction defined in chapter 1, and is derived from the relative motion data of Table 2.4. This vector includes only one interpolated data point, at 100 Ma (Cretaceous Quiet Period).

The vector of Pindell (this study) is interpreted as follows. During the Late Triassic, a small amount of approximately north-south divergence is predicted due to the lithospheric extension associated with deposition of the Eagle Mills red beds, and to dike orientations along the eastern United States that suggest a dextral shear component accompanying the extension associated with the Newark graben series. This relative motion precedes the opening of the Central Atlantic.

By the early Middle Jurassic, southeastward divergence between North America and Gondwana began and is responsible for the opening of the Central Atlantic. A single ridge system could have accounted for this separation between North and South America. However, Jurassic seafloor crust in the Gulf of Mexico, whose anomaly trends differ from Central Atlantic trends [S. Hall, personal communication, 1984; Shepherd et al., 1982], suggests that two ridge systems operated during the Jurassic, one in the Gulf and one in the "Proto-Caribbean," the individual motions upon which summed to the total indicated relative motion. When the Gulf ridge system terminated (Berriasian), and Yucatan became a part of North America, Early Cretaceous to Campanian divergence could have been accomplished at a single ridge in the Proto-Caribbean. This is especially likely if the Caribbean Plate was allochthonous with respect to the margins of the Proto-Caribbean prior to the Campanian. A slight kink in the vector from anomaly M-21 to M-11 may have caused compression at any long left-lateral transforms in the developing ridge system, and extension at any long right-lateral transforms. Because of the geometry of the early Central Atlantic and Proto-Caribbean oceans, the eastern Bahamas probably was the site of long left-lateral transforms, the compression at which may have uplifted basement there to the photic zone, leading to initial carbonate deposition.

Divergence continued through the Early Campanian (anomaly 34). The Aptian opening of the Equatorial Atlantic Ocean, which created a three-plate system for the first time, apparently had little or no effect on the direction of relative motion between North and South America, although the rate could have changed. This cannot be determined because of the Cretaceous Quiet Period. Further, it is not clear whether relative motion ceased exactly at, or somewhat before, anomaly 34 time.

From anomaly 34, or just before, to anomaly 21 (Middle Eocene), relative motion seems to have nearly ceased. The minor oscillations in the vector from anomaly 34 to anomaly 21 are all within the magnitude of the error in the analysis. Thus, any prediction of significant relative motion at plate boundaries for this interval is unwarranted. However, minor motion (200 km) of South America, relative to North America, toward approximately 080 may be suggested for the Paleocene-Middle Eocene interval (20 million years). This divergence rate (10 mm/yr, total rate) is slower than spreading rates at known ridges, and the total offset is also small (200 km). It is suggested that this potential phase of relative motion was achieved by sinistral slip at the transforms and fracture zones of the Central Atlantic ridge system which at this time extended into the

Equatorial Atlantic, and not at a mid-ocean ridge system between North and South America. It is also possible, however, that spreading at a ridge and subduction at a trench, both located between North and South America, occurred synchronously, and at approximately the same rate, to produce the indicated relative motion.

Finally, from anomaly 21 (Middle Eocene) to the present, slow but continued convergence along west-northwest trending flow lines is suggested. This general trend nearly parallels, but is opposite to, the pre-Campanian divergent flow lines. In addition, the Barracuda, Tiburon, and other basement ridges east of the Lesser Antilles also roughly parallel the west-northwest trend. These ridges are probably ancient fracture zones associated with the the pre-Campanian seafloor spreading in the Central Atlantic, which have become reactivated in a dextrally convergent sense since the Eocene, in association of this last phase of North and South American relative motion. A comparison between the two flow lines for the Pindell data in Figure 2.5 indicates that in addition to minor convergence with North America since the Eocene, western South America is migrating faster toward North America than is northern South America. This suggests that the post-Eocene stage-poles describing the motion of South America with respect to North America are located relatively near, but to the north of, South America, perhaps within a few hundred km to the northeast of Puerto Rico. Stage-pole positions for the intervals between anomalies 21, 13, 6, 5 and the Present are all located in this general area (see Table 2.4), but each may contain considerable error owing to the fact that they are located so close to the plate whose rotation they define (South America).

The primary difference between the Pindell vector and the vectors of Ladd [1976] and Sclater et al. [1977] is that the latter vectors predict a significant Cretaceous phase of North America/South America compression.

The difference stems directly from the choice of the Equatorial Atlantic fit used in the analysis: that of Bullard et al. [1965], or the fit presented herein. This fit is also responsible for the relatively northerly placement of the Jurassic to Campanian portions of the vectors. Such is the importance of the Equatorial Atlantic fit. Cretaceous compression is well known from northern Venezuela, but the rocks exhibiting these compressional structures are allochthonous with respect to the Venezuelan margin, and probably were not obducted onto the margin until the Cenozoic. All of the compressional structures may have originated in a pre-obduction "subduction phase," involving the Caribbean plate, rather than the North American plate. No definite North American/South American convergence can be documented for the Cretaceous; in fact, gravity and magnetic studies of the ocean floor to the east of the southern Lesser Antilles arc indicate that the North America/South America plate boundary has always been that of strike-slip (G. Westbrook, personal communication, 1984).

In summary, South America's motion relative relative to North America has been very simple: first it migrated out (Jurassic to Campanian), then essentially stopped (Campanian to Eocene), then returned a very small amount (post-Eocene).

Despite the improvements in the understanding of the North America/South America relative motion vector, some nagging problems remain to be solved. First, the existence of the Jurassic and Cretaceous Quiet Periods (? to 160 Ma and 119 to 84 Ma, respectively) precludes magnetic anomaly identification in crust of those ages. Spreading rates cannot directly be measured for these these periods with much accuracy. Second, the timing of initial motion across the Equatorial Atlantic shear zones is poorly known because of the area's general pre-Albian emergence and the fact that linear magnetic anomalies with northerly strikes do not develop well in equatorial regions. Third, the non-rigid behavior of Africa during the Early Cretaceous [Burke and Dewey, 1974] is poorly quantified, and the amount of Santonian shortening in Benue Trough, and its effect on the South Atlantic, is unknown.

The Yucatan Block with respect to North America:

In chapter 3 (Geology and Evolution of the Gulf of Mexico), a model is proposed in which the motions of the Yucatan Block are constrained by the presence and motions of North and South America. For consistency, the rotation parameters describing the motion of the Yucatan Block relative to North America are included in this chapter (Table 2.5), rather than in chapter 3.

TABLE 2.5. ROTATION PARAMETERS, YUCATAN BLOCK WITH RESPECT TO NORTH AMERICA

Anomaly/age	Latitude	Longitude	Angle
Triassic	29.74N	80.36W	-43.41
Mid-Jurassic	29.50N	81.40W	-41.00
Blake Spur time	29.50N	81.40W	-34.00
Late Callovian	29.50N	81.40W	-20.00
Anomaly 21 time	29.50N	81.40W	-9.00
Anomaly 16 time	-.--	-.--	-.--

1. All parameters defined by modelling in chapter 3.
2. Paramters are measured in Mercator projection and contain inherent error.

CHAPTER 3

GEOLOGY AND EVOLUTION OF THE GULF OF MEXICO REGION

GEOLOGY AND EVOLUTION OF THE GULF OF MEXICO REGION:

Introduction:

In this chapter, geological features of the Gulf of Mexico region are reviewed and interpreted. The important features are shown on the generalized tectonic map of the Gulf of Mexico region (Figure 3.1). The analysis assumes the initial Alleghenian reconstruction derived in chapter 1, and is built within the plate-kinematic framework developed in chapter 2. Rotation parameters describing the motion of the Yucatan Block relative to North America are also given in chapter 2 (Table 2.5).

Interpretive Review of Primary Gulf-Region Geological Features:

Basement

Three types of crust exist in the Gulf region: normal-thickness continental crust, variably-attenuated continental crust, and oceanic crust. Prior to seafloor spreading in the Gulf, relative plate motions were accommodated by continental attenuation (as much as 500 km between the southern U.S. and Yucatan, see above). This attenuation must be incorporated when modeling Gulf evolution. Attenuated basement around the northern Gulf includes a collection of highs (Sabine, Monroe, Wiggins) separated by intervening basins (Gulf Coast Salt Basins). The Sabine and Wiggins basement highs consist of Paleozoic metamorphics, volcanics and

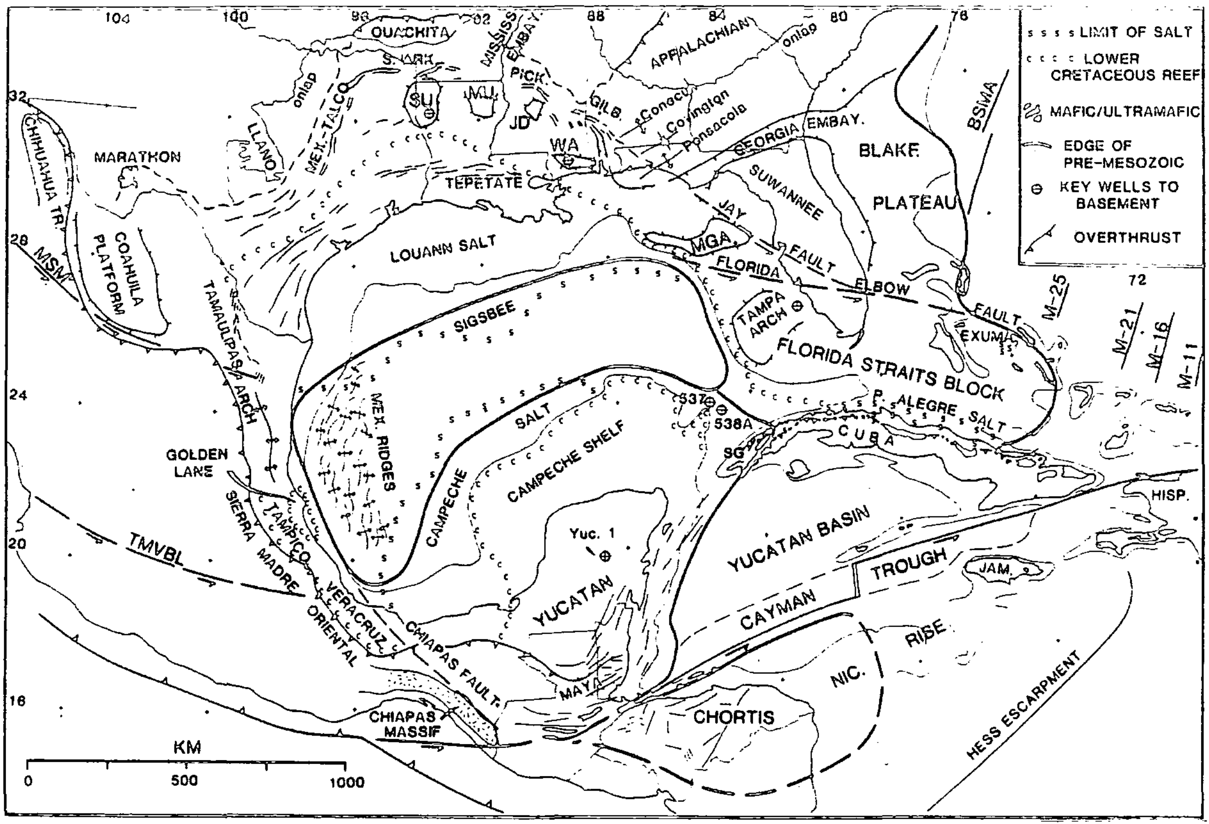


Figure 3.1. General tectonic and locality map of the Gulf region, features described in text. MSM, Mojave-Sonora Megashear; TMVBL, Trans-Mexican Volcanic Belt Lineament; SU, Sabine Uplift; MU, Monroe Uplift; JD, Jackson Dome; WA, Wiggins Arch; MGA, Florida Middle Grounds Arch; SG, Sierra Guaniguanico, presently of western Cuba but was northeastern corner of Yucatan block prior to Paleogene; BSMA, Blake Spur Magnetic Anomaly; M-25, M-21, M-16, M-11 are Central North Atlantic magnetic anomalies. This is same as Figure 1.2.

granite [Nicholas and Waddell, 1982; Cagle and Khan, 1983], which may be interpreted as remnants of the leading edge of Gondwana which were sutured to and overthrusted onto the margin of North America during the Alleghenian Orogeny and subsequently left behind during Late Triassic to Early Jurassic rifting [Pindell and Dewey, 1982; Smith et al., 1981]. The same applies for Florida, on the basis of Early Paleozoic fauna in Suwannee Basin [Cramer, 1971], and for the Coahuila Platform of northeastern Mexico, which possesses Late Paleozoic granodioritic intrusions [King, 1975] possibly representative of an arc. Basement in the attenuated margin of northwestern Yucatan block is unknown.

Extensive salt deposits overlie the zones of attenuated continental crust, and seaward limits of the Louann and Campeche salt provinces match reasonably the ocean-continent boundaries [Buffler et al., 1981], although oceanward halokinesis has occurred, particularly in the Sigsbee Escarpment [Lehner, 1969]. It seems likely that the onset of emplacement of oceanic crust by seafloor spreading in the Gulf at oceanic isostatic depths (2.6 km depth) allowed sufficiently open marine circulation to terminate salt deposition, and split the once continuous salt province into the distinct provinces known today. Unfortunately, the halokinesis on both sides of the Gulf prevents direct matching of opposing margins for paleogeographic reconstructions.

The oceanic portion of the Gulf of Mexico was created by primarily Late Jurassic seafloor spreading at a ridge system that must be included in plate-boundary reconstructions. Linear magnetic anomaly trends in the central Gulf fan slightly to the west and generally parallel the U.S. Gulf Coast margin and the Campeche Escarpment of Yucatan [Shepherd et al., 1982; S. Hall, personal communication, 1984]. This leads independently to the conclusion that Yucatan originated from the Texas-Louisiana margin and rotated counterclockwise to its present position, and further substantiates the geological arguments presented earlier for this initial position in Alleghenian paleogeography.

When the Equatorial Atlantic is closed (see above) and South America-Africa is then refitted to North America following Central North Atlantic flow lines [Klitgord and Schouten, 1982], overlap occurs between the Florida Straits block and the Guinea and Demerara Plateaus of Africa and South America, respectively. Also, the closure of Yucatan against the Texas-Louisiana margin leaves an oceanic hole in the northeastern Gulf with a southeasterly magnetic anomaly trend that is discordant with the trend in the central Gulf [Shepherd et al., 1982]. It appears that crust of the Florida Straits block migrated to the east-southeast, in addition to undergoing internal extension, during Gulf evolution. This motion relative to the remainder of Florida was left-lateral strike-slip, as discussed below, while motion relative to Yucatan was extensional, as defined by the eastern Gulf's magnetic anomalies.

Fault Systems of the Gulf Region

U.S. Gulf Coast. In the U.S. Gulf Coast margin, the Mexia-Talco-South Arkansas-Pickens-Gilbertown fault system defines the northern limit of significant normal faulting in basement. This system coincides closely with the probable Alleghenian suture. Assuming Yucatan originated from the Texas-Louisiana margin and rotated to its present position, the direction of extension in the U.S. Gulf Coast was north-south. Therefore, the Mexia-Talco fault zone owes its origin to differential subsidence and minor dextral shear between the basement of the stable Llano area and the basement of the Gulf Coast during attenuation. Similarly, motion in the Mobile Bay system may have been slightly sinistral.

Mexico. In most reconstructions of western Pangea, much of Mexico overlaps South America, if Mexico is kept in its present position with respect to North America. Understanding the emplacement of Mexico into this "overlap position" has been problematic because known evidence for large-scale motions upon one or more intracontinental transform faults across Mexico and the southwesternmost U.S. is meager. Evidence in favor of such large offset faults includes the interruption of northeasterly striking Precambrian tectonic belts, and a 700 to 800 km sinistral displacement of stratigraphic columns having "provocative similarities" [Anderson and Schmidt, 1983]. The theoretical structure responsible for these discontinuities has been termed the Mojave-Sonora Megashear (MSM) [Silver and Anderson, 1974]. The timing of motion along the postulated MSM is constrained between the first major motions of the breakup of Pangea (Late Triassic, first stretching, but early Mid-Jurassic is first significant motion) and deposition of the Oxfordian Zuloaga Group (Smackover equivalent), as the latter apparently masks the fault zone in northeast Mexico and is not offset. A 700 to 800 km offset during Bajocian-Callovian time (about 20-25 million years) indicates displacement rates between 2.8 to 4 cm/yr. It appears as though, in effect, blocks or slices of Cordilleran Mexico migrated more or less with South America during initial breakup, maintaining a land bridge from North to South America at least until Callovian time, when saline waters finally entered the Gulf of Mexico region and evaporated to form the once continuous Louann and Campeche salt deposits. From which ocean the saline waters entered, the Pacific or the Atlantic, is unknown. Another potential, but unproved, zone of major Mesozoic offset across Mexico is along the Trans-Mexican Volcanic belt. This area, during Late Cenozoic time, has been the site of significant calc-alkaline volcanism, possibly because it is a zone of weakened crust [Mooser, 1969; Anderson and Schmidt, 1983].

The number of major faults through Mexico and their offsets are poorly understood; proposed offsets are based largely on geometry alone. It is very possible that all of Mexico, since it resided on the western, convergent margin of Pangea during breakup, experienced severe internal shearing of the sort seen along the western North American margin today. If studies of North America/Farallon relative motion for Late Jurassic-Early Cretaceous time are accepted, Farallon subduction beneath southwestern North America possessed a strong left-lateral oblique component [Engebretson, 1982, but see Duncan and Hargraves, 1984]. This is in agreement with the postulation of a simple shear regime for Mexico during the Late Jurassic [Beck, 1983], and possibly Middle Jurassic as well.

Along the eastern Mexican margin, continental basement falls off abruptly along a linear trend defined by the eastern side of the Tamaulipas Arch, the Golden Lane high, and offshore at Veracruz. If extended to the south, this trend crosses into Chiapas, approximately along the northern flank of the Chiapas Massif. Sediment thickness adjacent to this flank exceeds 4500 metres [Viniegra, 1971], suggesting the existence of a major structural break. The linear trend as a whole is perpendicular to magnetic anomaly trends in the Gulf of Mexico [Shepherd et al., 1982; S. Hall, personal communication, 1984], and defines a small circle about a pole located in the vicinity of northern Florida. It is suggested that this trend, termed here the Tamaulipas-Golden Lane-Chiapas fault zone, was a right-lateral transform zone between Yucatan and eastern Mexico, that allowed migration of Yucatan away from the Texas-Louisiana margin. This is further supported by the conspicuous absence of salt diapirs marginal to the central portion of this trend, in the area of the Mexican Ridges [Buffler et al., 1979], in contrast to areas of the Gulf where attenuated continental basement exists with thick salt deposits and diapirs [Buffler et al., 1980, 1981]. A transform origin for the eastern Mexican margin which postdated salt deposition explains the absence of salt. In addition, the Tamaulipas Arch, during Late Jurassic time, was a linear basement high of Paleozoic rocks that supplied significant amounts of debris to proximal deposits of the Huizachal and Zuloaga Groups during initial rifting and platform subsidence [Sandstrom, 1982; Stabler, 1982; Meyer and Ward, 1982; Bracken, 1982]. The linearity and positive structural relief of the Arch during the Jurassic suggest strike-slip faulting as a possible cause of its uplift. In Chiapas, coarse, Upper Jurassic-Lower Cretaceous Todos Santos red beds fill northwest trending structural valleys and ridges whose relief reached 1000 to 2000 m at the time of deposition [Burkart and Clemons, 1971]. This scenario accords with models of strike-slip faulting equally as well as it does with rifting, and the northwest trend of the basins is perpendicular to the expected extension direction between North and South America. Finally, definition of a Mexico-Yucatan shear zone along the northeast side of the Chiapas Massif avoids the often invoked hypothesis that major displacement has occurred across the Isthmus of Tehuantepec; crystalline rocks cross the isthmus from southern Mexico into the Chiapas Massif with no obvious structural break [King, 1969; Case and Holcombe, 1980].

Florida region. Basement structure in Florida and the southeasternmost U.S. can only be inferred from gravimetric, magnetic and borehole information, but basement appears to be irregular and to consist of a collection of horstlike Paleozoic highs separated by grabens filled with red beds [Smith, 1983; Barnett, 1975; Klitgord et al., 1984]. Primary structural features include the Jay Fault, the Georgia Embayment "graben" system, and the Florida Elbow basin which separates the Middle Grounds and Tampa Arches

(Figure 3.1).

The northwest trending Jay fault [Smith, 1983] aligns with the Pickens-Gilbertown fault system and the Bahamas fracture zone of Klitgord et al. [1984], and defines a steep, down to the south drop-off in basement. Sinistral motion along this fault is speculative, but it clearly defines the northern limit of significant basement attenuation, as do the Pickens-Gilbertown fault system in the Gulf Coast and the Bahamas fracture zone across the Florida Peninsula. This attenuation probably occurred in the Middle Jurassic, as it did throughout the entire Gulf of Mexico. North of the Jay, relatively minor basement structures such as the Covington Embayment and Conecuh and Pensacola Arches formed due to general extension between North America and Gondwana.

Development of the northeast trending, Triassic Georgia Embayment system is older than the other fault zones, and is often referred to as an extensional rift on the basis of its Triassic red beds and basalts filling narrow troughs [Barnett, 1975]. If the system's formation were due to crustal extension, then a well-developed Jurassic sedimentary section related to basin subsidence could be expected. However, such a section is absent across southern Georgia and north-central Florida [Barnett, 1975; Smith, 1983] and, therefore, it is suggested that the Georgia Embayment system probably was formed by strike-slip shear, which caused local uplift and erosion, and deposition of Upper Triassic red beds into associated strike-slip basins. On the basis of geometry alone, the suggested sense of offset is right-lateral (discussed later).

Basement structure beneath southern Florida and the south Florida shelf is poorly known due to the thick post-Triassic sedimentary section. As mentioned earlier, the eastern part of Florida Straits block overlaps Africa when the continents are reassembled. This is true even after 80% internal extension is restored to the northwest (see above). In addition, the Yucatan block alone cannot fill the oceanic portion of the Gulf of Mexico; a small oceanic swath remains in the northeastern deep Gulf. Therefore, it is apparent that crust of the Florida Straits block migrated out of the eastern Gulf, in addition to undergoing severe internal extension. Hypothesized here is the existence of a fault (Florida Elbow Fault, Figure 3.1) along which the Florida Straits block, subsequent to crustal attenuation south of the Jay Fault, migrated 300 km east-southeast out of the eastern Gulf to its present position beneath the south Florida shelf and western half of the Bahamas. The Florida Elbow Fault runs from the southern escarpment of the Paleozoic Florida Middle Grounds Arch, continues through the Florida Elbow basin and crosses Florida near Lake Okeechobee, underlies the northwest Providence Channel, and defines the northern margin of Great Bahama Bank. Such a trend is readily seen on a magnetic anomaly contour map [figure 5 of Klitgord et al., 1984]. It is further suggested that sinistral motion along the fault system produced the Florida Middle Grounds escarpment by translating the Tampa Arch away from the Middle Grounds Arch.

Tectonism Following Gulf Formation

Since the Gulf of Mexico's formation, which was completed by or during the earliest Cretaceous, the Gulf region has experienced at least two periods of plate-tectonic deformation. The first was the eastward advance of thrust sheets of the Sierra Madre Oriental of Mexico (Figure 3.1) in Late Cretaceous to Eocene time [Dengo, 1975]. The cause of thrusting is still debated, but the load of the thrusts, which travelled across Jurassic salt, depressed the western portion of the eastern Mexican margin, reactivating the Tampico-Tuxpan and Veracruz sedimentary basins by flexural loading. The

Laramide orogenesis also elevated and reactivated the Tamaulipas Golden Lane highs, probably due to lithospheric flexure in response to loading by the Sierra Madre to the west, or by sedimentation offshore to the east, or both. The second period of tectonism was the early Paleogene arrival and collision of Cuba (Greater Antilles) with the Bahamas/southern Florida shelf [Dickinson and Coney, 1980; Gealey, 1980]. The emplacement of thrust sheets of the Cuban forearc tectonically loaded the southern Florida/Bahamian margin, depressing the carbonates of the latter to oil-maturation depths. Oil has since migrated upward into the fractured serpentinites that presently form small reservoirs in and offshore northern central Cuba [Wassal, 1956]. The causes of other possible "events" during the Cretaceous, including the Gulf-wide mid-Cretaceous unconformity [Buffler et al., 1980], volcanism and local plutonism in the Gulf Coast [Zartman, 1977; Smith et al., 1981], uplift of the Wiggins, Monroe and Sabine highs, and reactivation of the Mississippi Embayment [Ervin and McGinnis, 1975] are unclear. In addition to these plate-tectonic deformations, much of the northern and western Gulf has experienced severe sedimentary gravity sliding and halokinesis.

Evolutionary model of the Gulf of Mexico:

Late Triassic

In the Late Triassic, western Pangea began to rift along widespread, poorly defined zones of intracontinental block-faulting and dike emplacement

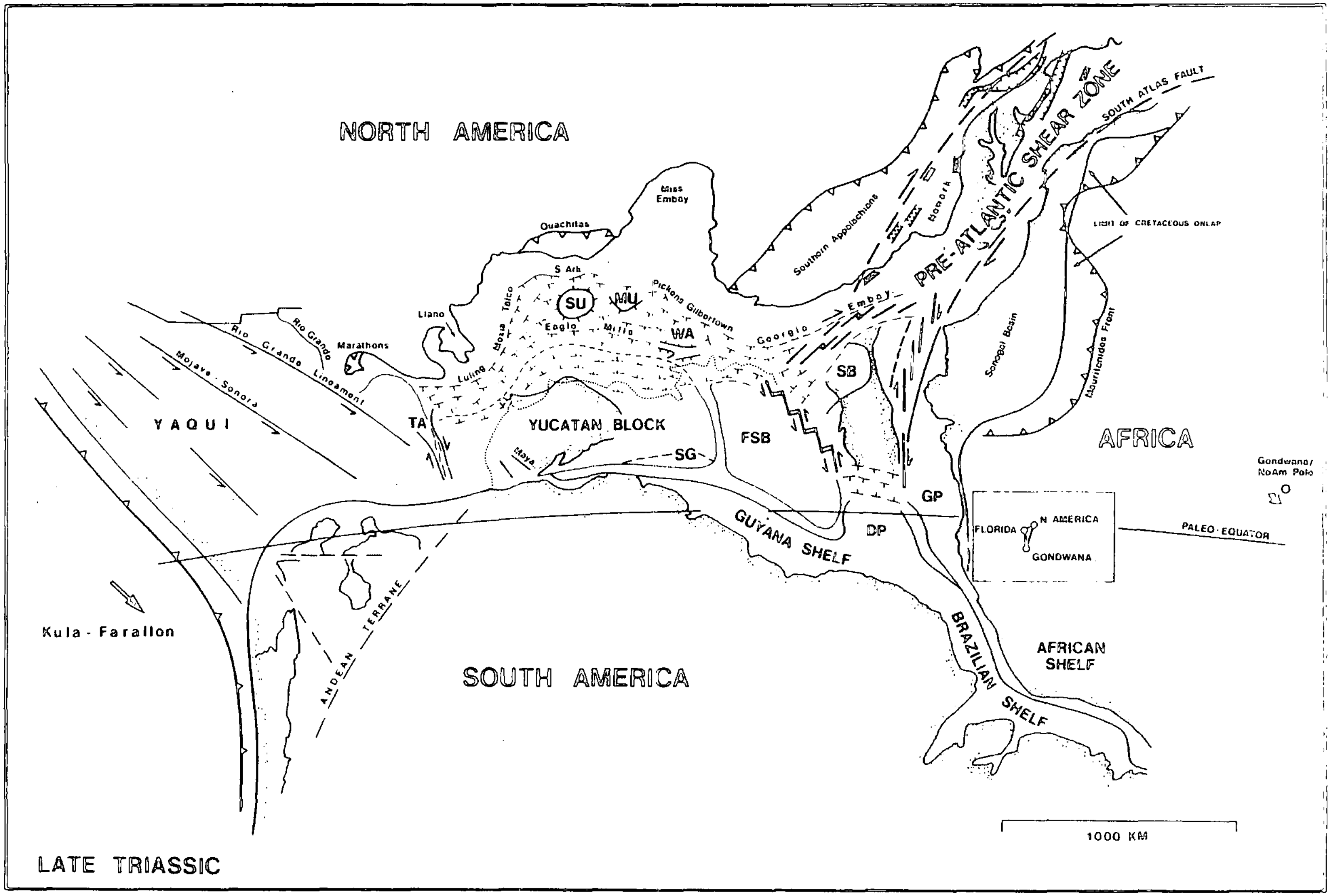


Figure 3.2. Late Triassic paleogeography of the Gulf region, at about 210 Ma. Relative motion of plates and blocks during time increment to next reconstruction defined by vector triangles in boxes. DP, Demerara Plateau; FSB, Florida Straits Block; GP, Guinea Plateau; MU, Monroe Uplift; SB, Suwannee Basin; SG, Sierra Guaniguanico of present-day western Cuba; SU, Sabine Uplift; TA, Tampa Arch; WA, Wiggins Arch. Strike and dip symbols schematic for rifting.

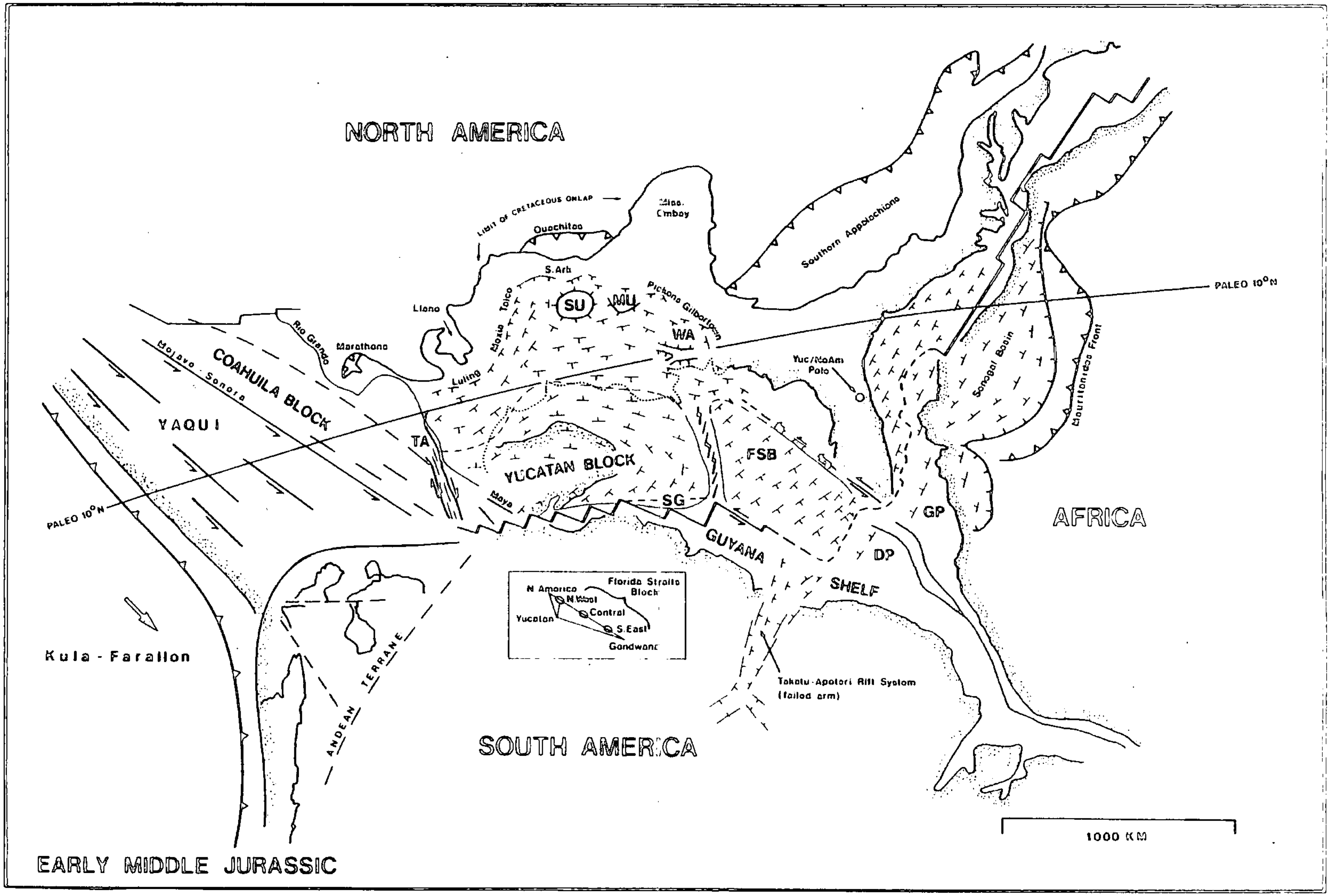


Figure 3.3. Early Middle Jurassic paleogeography of the Gulf region, at about 180 Ma. Abbreviations and conventions as in Figure 3.2.

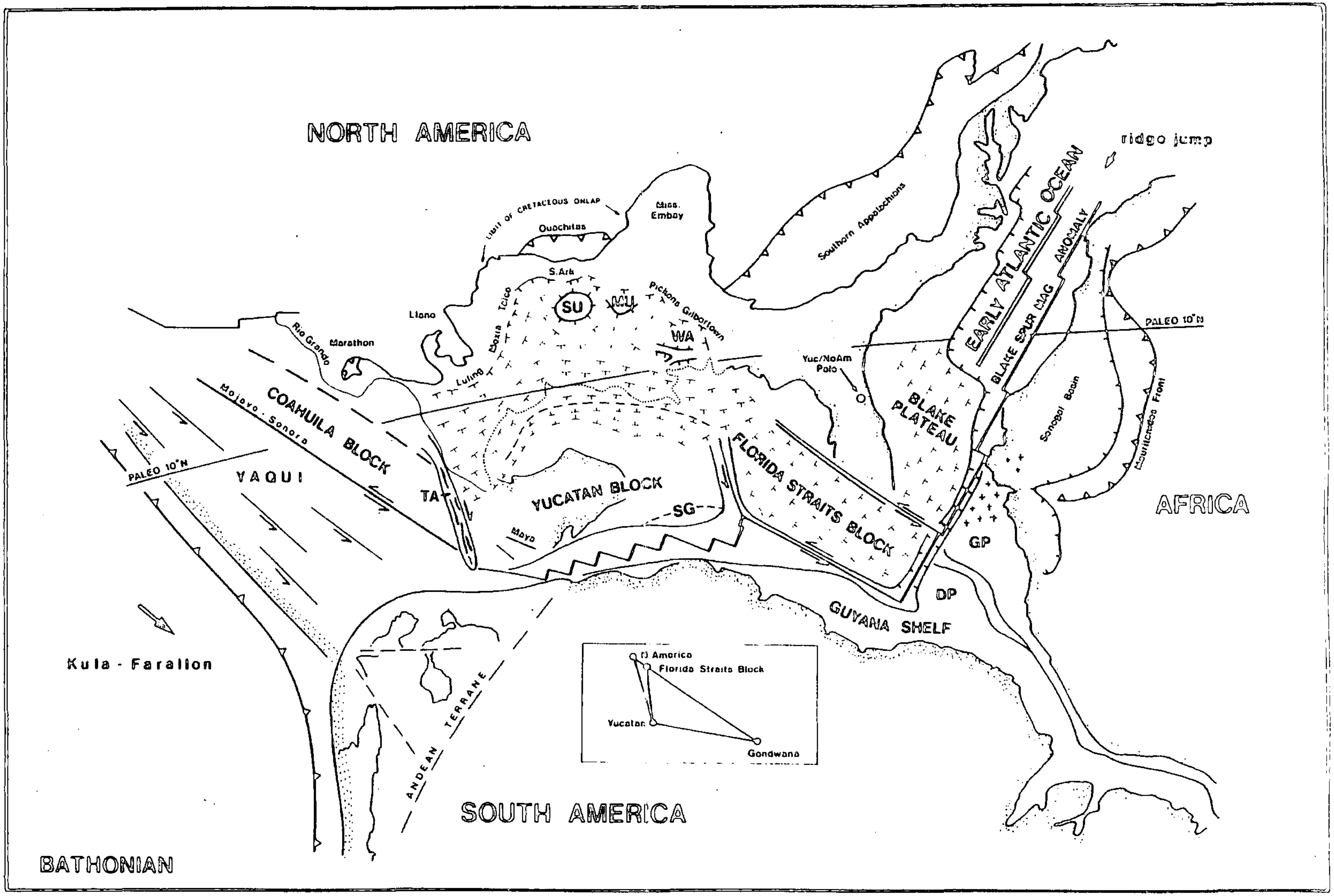


Figure 3.4. Middle Jurassic (Bathonian) paleogeography of the Gulf region, at about 170-172 Ma. Abbreviations as in Figure 3.2. Cross pattern represents rift-related salt of the African margin [Jansa and Weidman, 1982], which may correlate with Punta Alegre salt on the opposing margin.

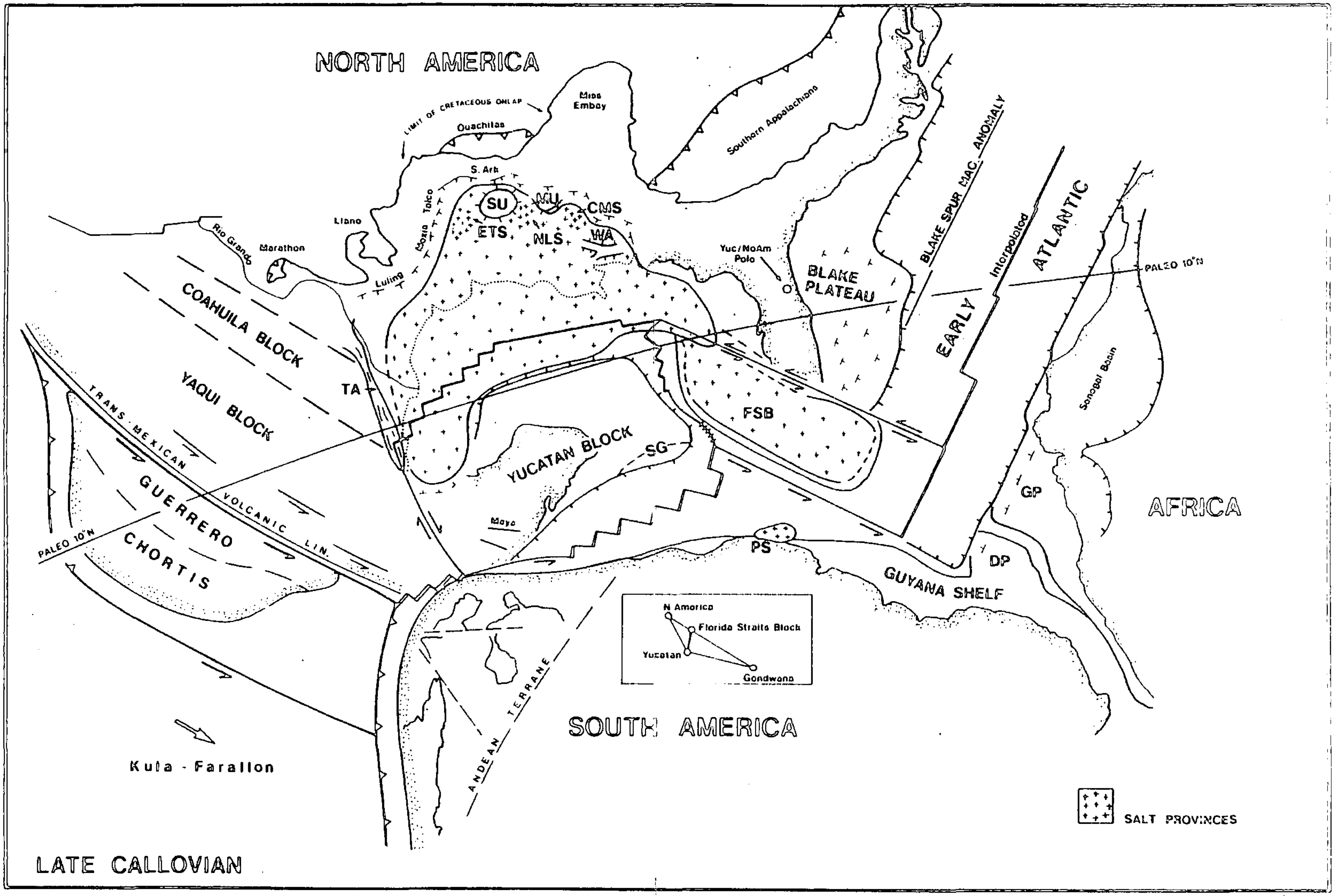


Figure 3.5. Late Middle Jurassic (Callovian) paleogeography of the Gulf region, at about 160 Ma. At this time, Gulf-wide salt had been deposited, seafloor spreading ridges were developed in the Gulf, and Chortis-Guerrero (Mexico south of Trans-Mexican Volcanic belt) may have been migrating along the Pacific margin from the north. CMS, Central Mississippi salt basin; ETS, East Texas salt basin; NLS, North Louisiana salt basin; PS, Gulf of Paria salt basin. Other abbreviations as in Figure 3.2. Whether the Florida Straits block migrated out of the Gulf as a single coherent block, or as a collection of smaller blocks separated by anastomozing transcurrent faults, is unknown.

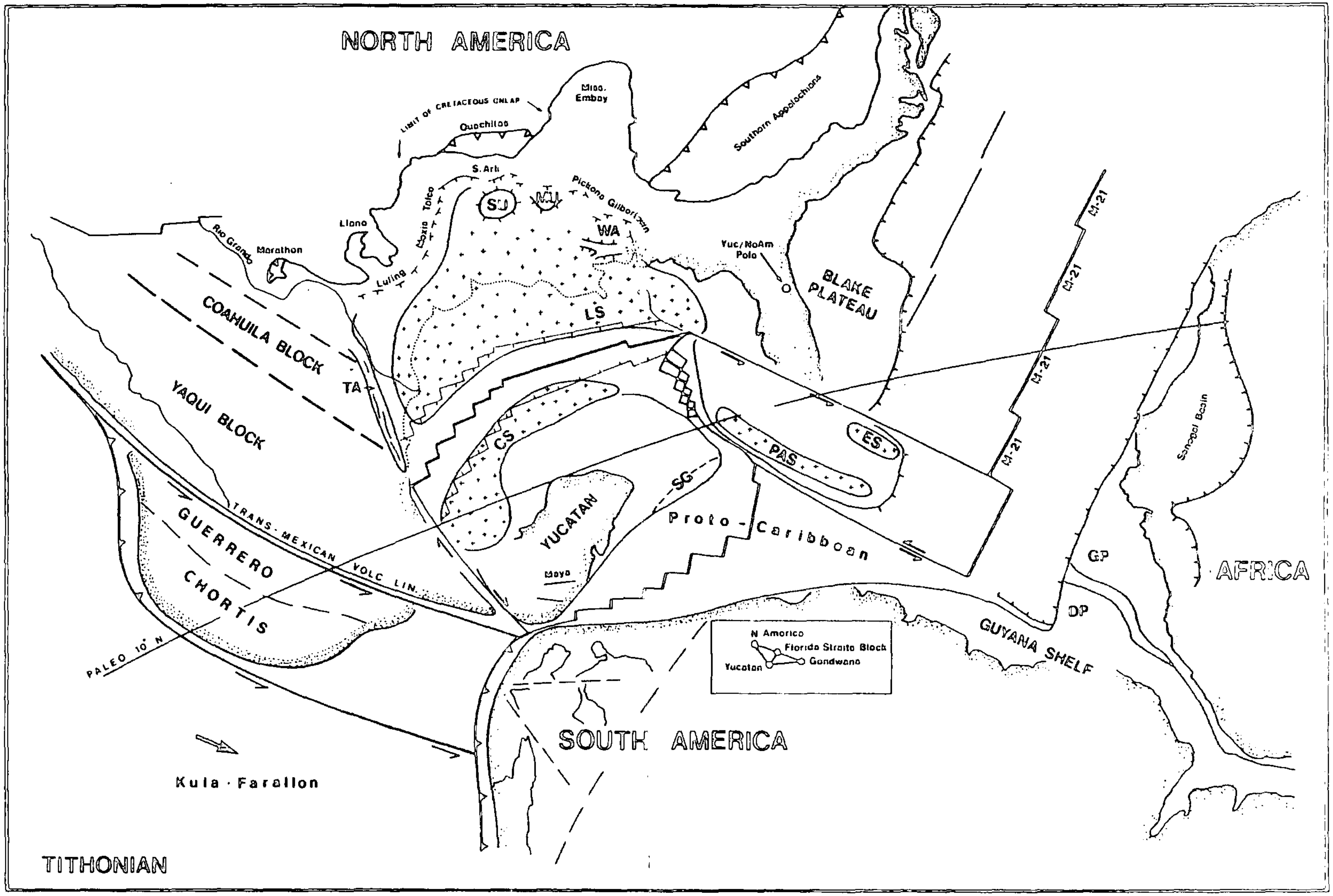


Figure 3.6. Late Jurassic (Tithonian) paleogeography of the Gulf region, at about 150 Ma. Seafloor spreading had separated the Gulf salt deposits into component members. CS, Campeche salt; ES, Exuma Sound? salt; LS, Louann salt; PAS, Punta Alegre Salt. Other abbreviations and conventions as in Figure 3.2.

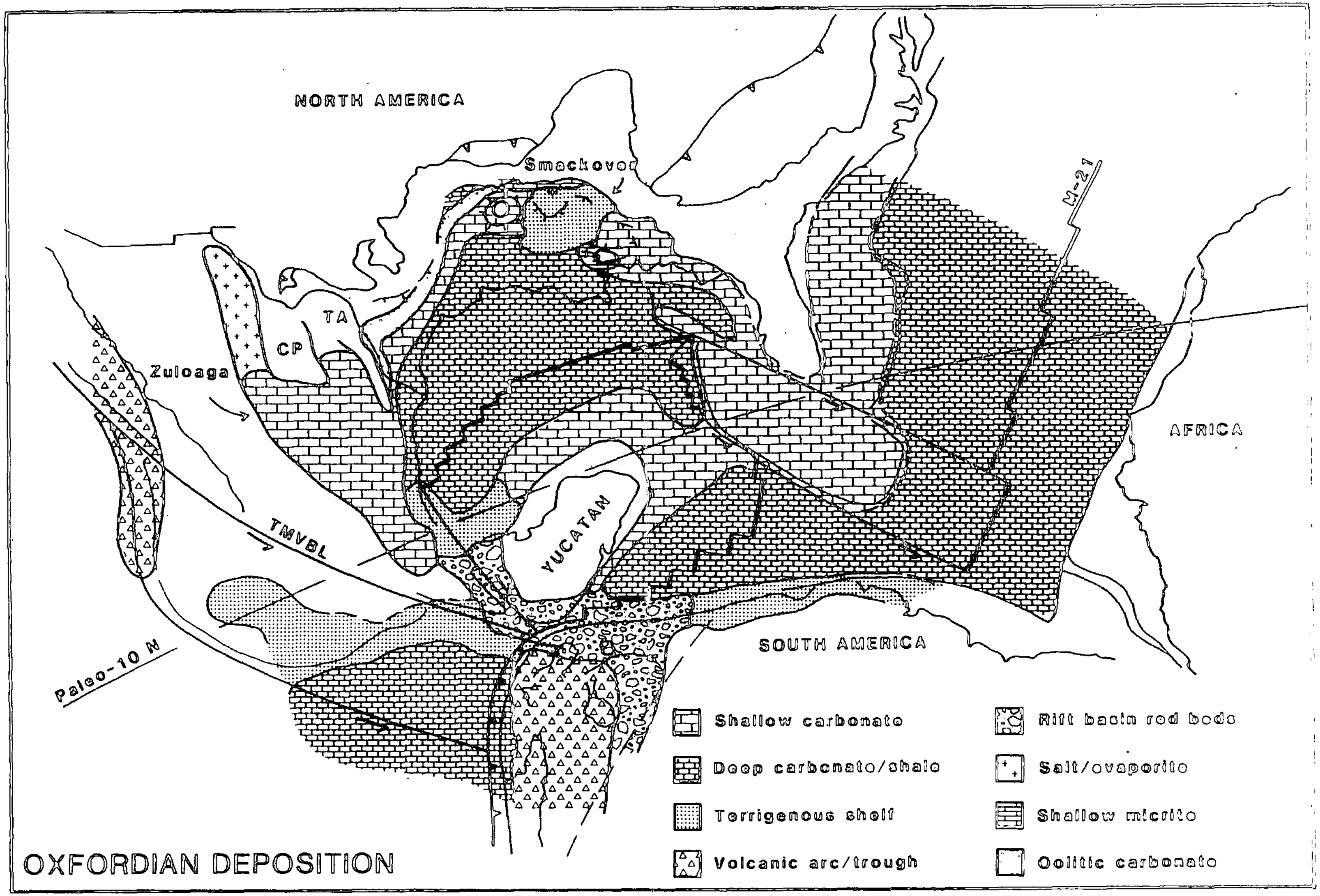


Figure 3.7. Depositional facies of the Gulf region at the close of Smackover deposition, plotted on Late Jurassic plate reconstruction. CP, Coahuila Platform. Other abbreviations and conventions as in Figure 3.2.

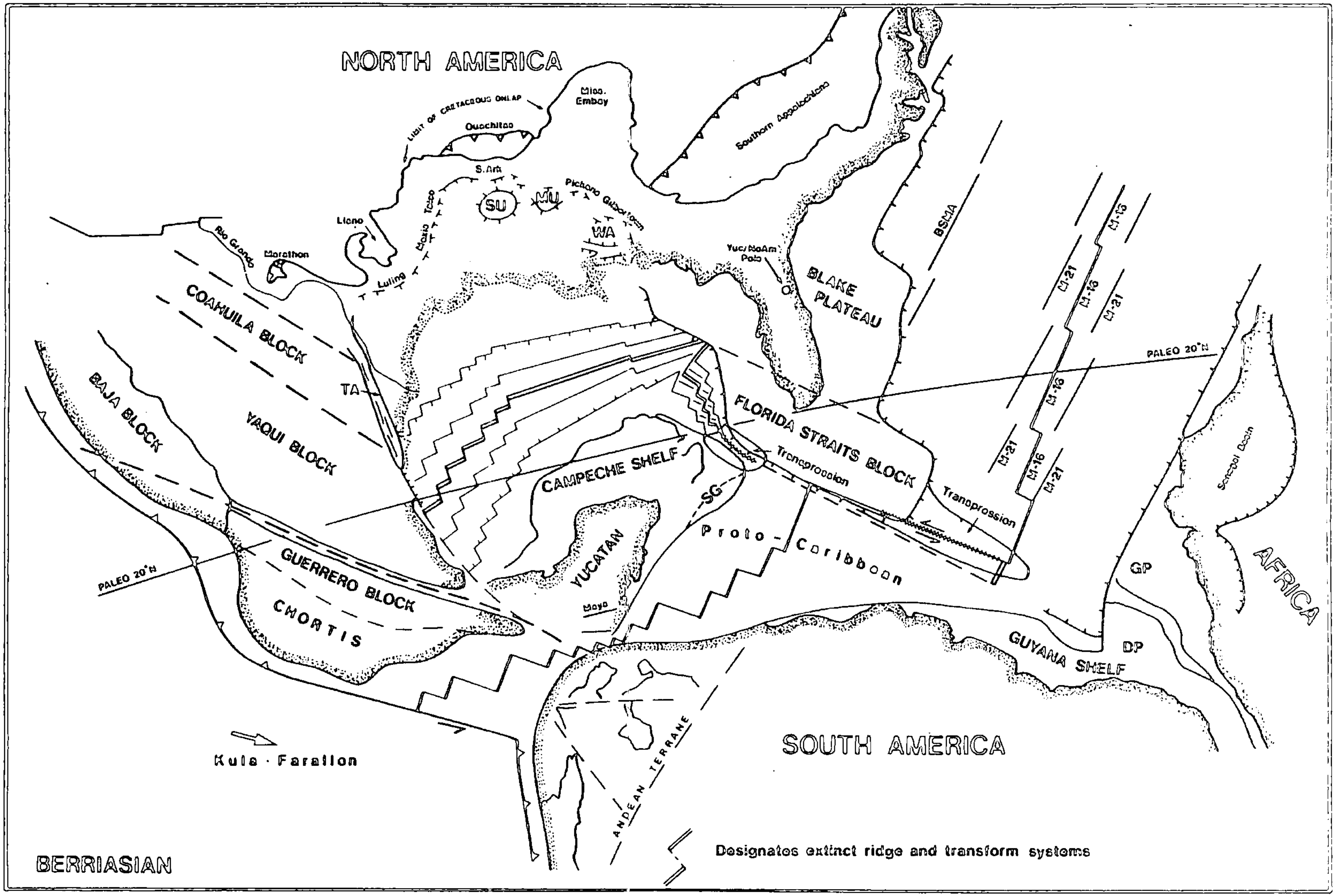


Figure 3.8. Early Cretaceous (Berriasian) paleogeography of the Gulf region, at about 140 Ma. Seafloor spreading terminated in the Gulf, and a change in position of the Central North Atlantic opening pole produced compression and uplift at the long left-lateral transform bounding the eastern Bahamas. Carbonate deposition has since kept pace with subsidence. Abbreviations and conventions as in Figure 3.2.

(Figure 3.2). Minor associated motions, recorded by the Late Triassic geology outlined below, led to the pre-Atlantic, early Middle Jurassic relationship between Africa and North America defined by Klitgord and Schouten [1982].

In the southern U.S., north-south extension produced an extensive graben system filled with Eagle Mills red beds. Along the eastern United States, red beds, dikes and sills of the Newark Group and its equivalents were deposited and emplaced in a belt of grabens paralleling the coast from the Piedmont out to the continental shelf. This system has generally been attributed to extension between Africa and North America, but the distribution and orientation of dikes with 200-220 Ma isotopic ages fit a right-lateral shear model within the arcuate zone (pre-Atlantic shear zone, Figure 3.2) along the eastern U.S. and western African margins [Swanson, 1982]. It is not implied that each graben of the Newark series is a pull-apart basin *sensu stricto*, but extension was probably accompanied by dextral slip. Dextral shear along the eastern U.S. is compatible with north-south extension along the U.S. Gulf Coast; thus, the Eagle Mills and Newark systems may be attributed to the same relative plate motion.

Lying amidst the eastern and southern U.S. shear and extensional zones, respectively, is the complex region of continental blocks comprising Florida, the Blake Plateau and the western Bahamas. Dextral shear between North America and Africa would have led to convergence between western Africa and the Blake Plateau/eastern Florida. A possible origin for the Georgia Embayment is that of dextral shear, which would have allowed Florida to escape laterally from the Africa-Blake Plateau convergence. A shearing origin for the Georgia Embayment is suggested by the absence of a Jurassic sedimentary section there, as would be expected had the Embayment formed purely by extension.

To the west, southeasterly tectonic transport of blocks within Mexico began at this time, along one or several shear zones, or by internal strain. This motion effectively maintained a land bridge from North to South America until the Callovian or later. A speculative cause of this early motion is left-lateral oblique subduction of the Kula/Farallon Plate.

Middle Jurassic

Divergence between Africa and North America began in early Middle Jurassic time (Figures 3.3 and 3.4) by seafloor spreading north of the Blake Plateau, and by continental attenuation in the area of Blake Plateau and the Florida Straits block. The Blake Spur Magnetic Anomaly (BSMA) can be symmetrically correlated about the mid-Atlantic Ridge with the western margin of Africa, and aligns with the eastern boundary of the present Blake Plateau (Figure 3.5). Hence, it is evident that internal stretching within the Blake Plateau accounts for divergence between North America and Africa until the time when the BSMA formed (approximately 170 Ma). The crust to the north of the Blake Plateau that was created by seafloor spreading prior to formation of the BSMA produces the Central North Atlantic's asymmetry with respect to the present ridge axis. An early ridge apparently was abandoned as the spreading center jumped to the site of the BSMA [Vogt et al., 1971], associated with which was ridge formation between the Blake Plateau/Florida Straits block and Africa.

In the developing Gulf of Mexico, Yucatan's progressive divergence from Texas-Louisiana was achieved by severe continental attenuation until probably Callovian or early Oxfordian time (postsalt), at which time seafloor spreading began. During this early rift stage in Gulf development, intracontinental extension was accompanied by subaerial deposition of the

Werner red beds and anhydrites.

In the west, southeastward transport of blocks accelerated along the Mexican Cordillera, preventing the incursion of significant amounts of sea water into the Gulf realm (Figures 3.4 and 3.5). One probable zone of concentrated sinistral shear strain was the Mojave-Sonora megashear [Anderson and Schmidt, 1983; Silver and Anderson, 1974], although strike-slip fault motions may have occurred as far inland as the "Rio Grande lineament" (Figures 3.2 and 3.3), a term used here to define that line which bounds the northeastward extent of significant basement reactivation during both Jurassic and Laramide tectonism. Based on geometry, it is suggested that left-lateral offsets within the "Chihuahua block" north of the Mojave-Sonora megashear total approximately 100 km. Motions upon the Mojave-Sonora and faults within the "Yaqui" block (of Anderson and Schmidt [1983]) to the south collectively approach 800 km. These motions were completed by Oxfordian time, as the Zuloaga Group (Oxfordian) masks the Mojave-Sonora Megashear. Collectively, the motions are responsible for the emplacement of central Cordilleran Mexico into its present position with respect to North America, which was formerly occupied by northwest South America.

The postulation that Yucatan rotated out of the Gulf realm in turn implies that a corresponding, synchronous rotational opening occurred between Yucatan and Venezuela (Figures 3.3, 3.4, 3.5, and 3.6). This basin may be termed the "Proto-Caribbean" as it occupied the future site of the Caribbean but was subducted beneath the true Caribbean Plate during Late Cretaceous to Eocene time [Dickinson and Coney, 1980]. Whether this Proto-Caribbean basin was marine during Middle Jurassic time is debatable. However, the fact that it was subducted suggests that it was floored by oceanic or extremely attenuated continental crust, and that it was probably marine. Furthermore, Middle Jurassic marine sedimentary rocks are found in the northern ophiolitic belt of Cuba [Pardo, 1975] that were accreted to the Greater Antilles arc at the leading edge of the Caribbean Plate during subduction of the Proto-Caribbean. Finally, Middle Jurassic, terrigenous, partially marine sediments (San Cayetano Group) [Meyerhoff, 1964] are preserved upon presumably continental crust in the Sierra Guaniguanico of western Cuba; the latter is interpreted here as a portion of the rifted margin of the northeastern corner of the Yucatan block which was obliquely overridden and caught up by Cuba in the early Paleogene Bahamas-Cuba collision. That this margin of Yucatan bordered the Proto-Caribbean also suggests a marine character for the latter during the Middle Jurassic.

Late Callovian (Latter Middle Jurassic)

By the Callovian (Figure 3.5), basement attenuation had produced a broad, subsiding platform of continental crust. As this platform subsided through sea level, or as a tectonic barrier to either the Pacific or juvenile Atlantic was repeatedly breached, flooding of the Gulf realm and subsequent evaporation produced the once continuous Louann and Campeche salt deposits. The Punta Alegre and (?) Exuma Sound salt may also correlate with the Louann and Campeche, but may be slightly older (Bathonian) and relate to the opening of the Central North Atlantic rather than to the Gulf of Mexico. Such is the case with the Senegal Basin and Guinea Plateau salt deposits [Jansa and Weidmann, 1982] which opposed the eastern Florida Straits block at that time (Figure 3.4). A Callovian age has not actually been proved for the Louann, but the majority of evidence [Scott, 1984] indicates a late Middle Jurassic age, probably late Callovian [Salvador and Green, 1980; Imlay, 1980]. Seismic stratigraphic studies [Buffler et al., 1980; 1981] suggest correlation of the Campeche and Louann deposits. The Louann (and probably Campeche) possesses shallow-water characteristics, and its clean, pure nature implies rapid deposition, perhaps on the order of only hundreds of thousands of years (C. Schreiber, personal communication, 1983). The environmental picture associated with the Gulf of Mexico's salt deposits consists of a subsiding basin/platform, during lithospheric attenuation, located possibly behind a foundering tectonic barrier to marine waters. Flooding by marine waters produced numerous or continued shallow-water influxes perhaps only tens of meters deep, the evaporation of which led to salt deposition at rates which easily matched rates of basin subsidence [Schreiber and Hsu, 1980].

Except for local occurrences such as the Buckner facies of the northern Gulf Coast, Gulf-wide salt deposition ceased by early Oxfordian time at the onset of relatively open marine circulation. Open circulation probably was structurally controlled by the establishment of the Gulf of Mexico's mid-ocean ridge system (Figures 3.5 and 3.6). The emplacement of oceanic crust at typical isostatic elevations (2.6 km below present sea level) produced a deep central Gulf trough sufficient to allow open marine circulation throughout most of the Gulf. Furthermore, it was not until about this time that the separation between South America and North America, incorporating all basement attenuation in the Gulf, was sufficient to accommodate the present size of the Yucatan block. The most likely entrance for marine waters into the Gulf during the Late Jurassic is between Florida and Yucatan, as DSDP leg 77 documented Jurassic extension and marine sedimentation [Schlager et al., 1984], and the Proto-Caribbean certainly had become marine by Oxfordian time. Spreading at the Gulf's ridge system isolated the main salt provinces, accounted for continued separation between the U.S. Gulf Coast and Yucatan without further basement attenuation within their margins, and translated the Florida Straits block to its present position beneath the western Bahamas and southern Florida (Figures 3.5, 3.6 and 3.8). Norphlet and Smackover deposition on the attenuated Gulf Coast margin was not controlled by primary rifting but by differential subsidence between adjacent basement blocks within a thermally subsiding platform and, perhaps locally, by salt migration where salt accumulations were sufficiently thick. The Norphlet effectively blanketed basement features (except Sabine and Wiggins) [Nicholas and Waddell, 1982; Cagle and Khan, 1983] so that by Smackover time (Figure 3.7), a true platform or carbonate ramp had developed [Budd and Loucks, 1981].

In the western Gulf, migration of Yucatan continued along the Tamaulipas-Golden Lane-Chiapas transform fault (TGLC). To the north of the migrating junction of TGLC and the central Gulf ridge system, TGLC evolved as a fracture zone separating zones of differential subsidence. The abrupt topographic low, or freeboard, east of TGLC received enormous volumes of sediment (for example, Burgos Basin east of Monterrey, Mexico) [King, 1969]. As strike-slip motion ceased along the Tamaulipas Arch, it subsided and eventually was onlapped by Upper Jurassic carbonates.

Along the Pacific margin, left-lateral migration of Mexican blocks may have occurred at this time primarily along the Trans-Mexican Volcanic Belt lineament. The portion of proto-Mexico referred to here as the Guerrero block could not yet have reached its present position because of overlap with South America. Models of Caribbean evolution [Wadge and Burke, 1983; Pindell and Dewey, 1982] assume a pre-Eocene connection between the Guerrero and pre-Mesozoic Chortis blocks; Chortis may have been connected to the Guerrero block at this time. Arc-related volcanism within Colombia is indicated for the Late Jurassic by radiometrically determined ages on presumably subduction-related plutons [Tschanz et al., 1974]. Convergence between Chortis-Guerrero and Colombia is, therefore, seen as likely. This can only have been possible if the Kula/Farallon Plate convergence rate with South America exceeded the spreading rate between North and South America, and if the Guerrero-Chortis terrane essentially belonged to the Kula/Farallon Plate, as the Salinian block of western California essentially belongs to the Pacific Plate today.

Early Tithonian (Late Jurassic)

Oxfordian-Kimmeridgian seafloor spreading at the Gulf ridge system separated the Yucatan and Florida Straits blocks from the U.S. Gulf Coast margin and isolated the three main salt provinces (Figure 3.6). Figure 3.7 summarizes lithofacies deposited in the Gulf area during Oxfordian time (primarily Smackover and equivalents). A postulated triple junction off of DeSoto Canyon in the northeast Gulf connected the central and eastern Gulf ridge systems to the postulated Florida Elbow transform. Southeastward migration of the Florida Straits block may have produced the steep, east-southeast trending portion of the northwestern Florida Escarpment. In the southwestern Gulf, transform motion between Yucatan and Mexico continued along the Golden Lane-Chiapas portion of TGLC. No salt was deposited along the Mexican margin where seafloor spreading occurred. In Chiapas, deposits of the Todos Santos "rift facies" are younger (Late Jurassic-Early Cretaceous) [Anderson et al., 1973] than those to the north (Werner and equivalents). This is because shear along the active transform between Yucatan and eastern Mexico continued longer in the south; that portion of TGLC north of the southward migrating central-Gulf spreading center progressively became a fracture zone, typified by simple thermal subsidence and carbonate deposition.

In Mexico, shearing, perhaps localized along the Trans-Mexican Volcanic Belt lineament, brought the Guerrero block (and Chortis-Nicaragua Rise) near its final position. To the east, the Proto-Caribbean continued to open, fanlike, between Yucatan and Venezuela. The mid-ocean ridge in this basin must have been joined in some way with the plate boundary separating the Florida Straits and Yucatan blocks, which in turn must have been connected to the Central North Atlantic ridge system by a long transform along the south side of the Bahamas. A single triple junction is portrayed connecting these plate boundaries in the northern Proto-Caribbean, for simplicity, but Paleogene subduction of this crust has eliminated direct evidence for this proposition.

Berriasian (Earliest Cretaceous)

Horizontal plate motions associated with the opening of the Gulf of Mexico were completed by the earliest Cretaceous (Figure 3.8). It was not until this time that South America had migrated sufficiently far from North America to accommodate Yucatan in its present position relative to North America. Likewise, the Guerrero block of Mexico may have reached its final position as well. Termination of its southeastward migration may relate to cessation of a shearing component of subduction, as is indicated by the Early Cretaceous initiation of head-on convergence (northeast relative motion) between North America and the Kula/Farallon plates [Engebretson, 1982]. The termination of seafloor spreading in the Gulf of Mexico affixed the Yucatan and Florida Straits blocks to the North American Plate. Thus, continued spreading between North and South America occurred at a ridge system(s) in the Proto-Caribbean basin. This ridge must have been connected to the mid-Atlantic Ridge via a long left-lateral transform, and must have extended out into the Pacific realm between Yucatan and Colombia.

Beginning at magnetic anomaly M-21 and ending at M-4, fracture zone traces in the western Central North Atlantic possess a kink that indicates a westward shift in the Central North Atlantic pole position during that time interval [Klitgord and Schouten, 1982; Klitgord et al., 1984]. Such a shift should produce compression at any long left-lateral transforms in the system. Assuming that the eastern Bahamas is underlain by oceanic crust (see above), the origin of the eastern Bahamas platform (oceanic basement uplift to, or formation at, the photic zone) is probably due either to compression at the long left-lateral Bahamian transform zone, similar to the Recent 3 km uplift of basement at the Mussau Ridge of the eastern Caroline Plate [Hegarty et al., 1983] and the Pliocene emergence of the Romanche transform in the Equatorial Atlantic [Bonatti et al., 1977], or to extension and volcanism along the transform zone (hot spot). However, due to the kinematic prediction of compression arising from the temporary change in the Central Atlantic pole position, a compressional origin is favored here (Figure 3.8).

Post-Berriasian Development

Throughout most of the Cretaceous, carbonate banks developed across nearly all of the shelf margins of the Gulf, Bahamas, and Proto-Caribbean region. Terrigenous clastics were largely restricted to the arc systems bounding the Pacific realm. During the Late Cretaceous, however, thrusting, uplift and erosion in the Sierra Madre Oriental and the Rocky Mountain Overthrust provided enormous quantities of terrigenous clastics to the western and central Gulf realm. The Florida Banks were protected from clastic deposition by the DeSoto Canyon bathymetric barrier. The probable cause of this orogenesis is the arrival of buoyant oceanic masses at western North American trench systems. Also associated with this orogenesis was the migration of the present Caribbean Plate, led by the Greater Antilles arc system, into the Proto-Caribbean realm, thereby subducting crust of the Proto-Caribbean basin. The Caribbean Plate as a whole probably is a buoyant mass of Pacific origin [Burke et al., 1978]. Orogeny and clastic deposition has occurred progressively eastward along the margins of the Proto-Caribbean, associated with the northeastward migration of arc systems at the leading edge of the Caribbean Plate [Dickinson and Coney, 1980; Pindell and Dewey, 1982]. The Caribbean Plate interacted with the southern Yucatan shelf and northwest South America in the Late Cretaceous, the western and central Bahamas and north-central Venezuela in the Paleogene, and with northeastern Venezuela and the easternmost Bahamas in the Neogene.

CHAPTER 4

FIELD STUDIES, PUERTO PLATA AREA, NORTHERN DOMINICAN REPUBLIC

FIELD STUDIES, PUERTO PLATA AREA, NORTHERN DOMINICAN REPUBLIC

Introduction:

This chapter summarizes the author's field studies conducted in the Puerto Plata area of northern Dominican Republic, on the Greater Antillean island of Hispaniola. The results of the field work are important to the definition of first-order Caribbean plate-tectonic elements made in Chapter 6 (Outline of Primary Caribbean Geological Features); therefore, this chapter precedes that one.

This chapter is organized as follows. First, the field problem is defined. This is done by briefly reviewing the geology and history of Cuba, Hispaniola, and the Cayman Trough. It is shown that an important discrepancy exists between the commonly accepted models for the history of the Greater Antilles arc and for the Cayman Trough. Then, on a more detailed scale, the morpho-tectonic provinces of Hispaniola and their histories are reviewed. This provides a framework, on the scale of Hispaniola, into which can be placed the significance of the field work. The results of the field studies in the Puerto Plata area are then summarized. From the field work, conclusions are drawn that are later integrated into the Caribbean plate-tectonic elements defined in Chapter 6.

Definition of the field problem:

The Greater Antilles islands comprise Cuba, Hispaniola, Puerto Rico and

Jamaica (Figure 4.1). Each of the islands is predominantly composed of Cretaceous to Eocene volcanic and sedimentary rocks of island-arc association. The volcanic island-arc constituents of western and central Cuba, however, may have been obducted onto, and hence overlie, stretched pre-Mesozoic continental crust of the Bahamian margin. The basements of Hispaniola and Puerto Rico appear to be composed entirely of oceanic crust [Bowin, 1975; Banks, 1975]. That of Jamaica is unknown, but also is probably oceanic crust [Arden, 1975]. The island arc associations of Cuba, Hispaniola and Puerto Rico are very similar; the three islands very likely constitute portions of one original island arc system. The rocks of Jamaica show a similar history too, but Jamaica's geographic position relative to the other islands makes it necessary to question Jamaica's genetic relationship with the other islands.

To the north of Cuba and Hispaniola are situated the Jurassic to Recent carbonate banks of the Bahamas. Stretched pre-Mesozoic crust underlies Great Bahama Bank, whereas oceanic crust probably underlies the eastern portion [Schlager et al., 1984; Uchupi et al., 1971]. From the literature, a strong case may be made that Cuba collided with the Bahamas Platform in the early Paleogene; the collision culminated in the Middle Eocene. A detailed account of this collision is given in Chapter 6, but the main lines of evidence for the timing of this collision are (1) Paleocene-Early Eocene orogenic sediments and the Jurassic-Eocene carbonates of the Bahamas Platform are overthrust by the north Cuban ophiolite sheets [Pardo, 1975], (2) island arc volcanism ceased in the Early Tertiary, (3) orogenic flysch units of Early Eocene age contain, for the first time, debris from both a volcanic source and the Bahamas Platform [Pardo, 1975], (4) seafloor spreading ceased in the Yucatan intra-arc basin in probably the Eocene [Rosencrantz et al., in press, 1985], (5) subsidence in the Andros Island well of the Bahamas was

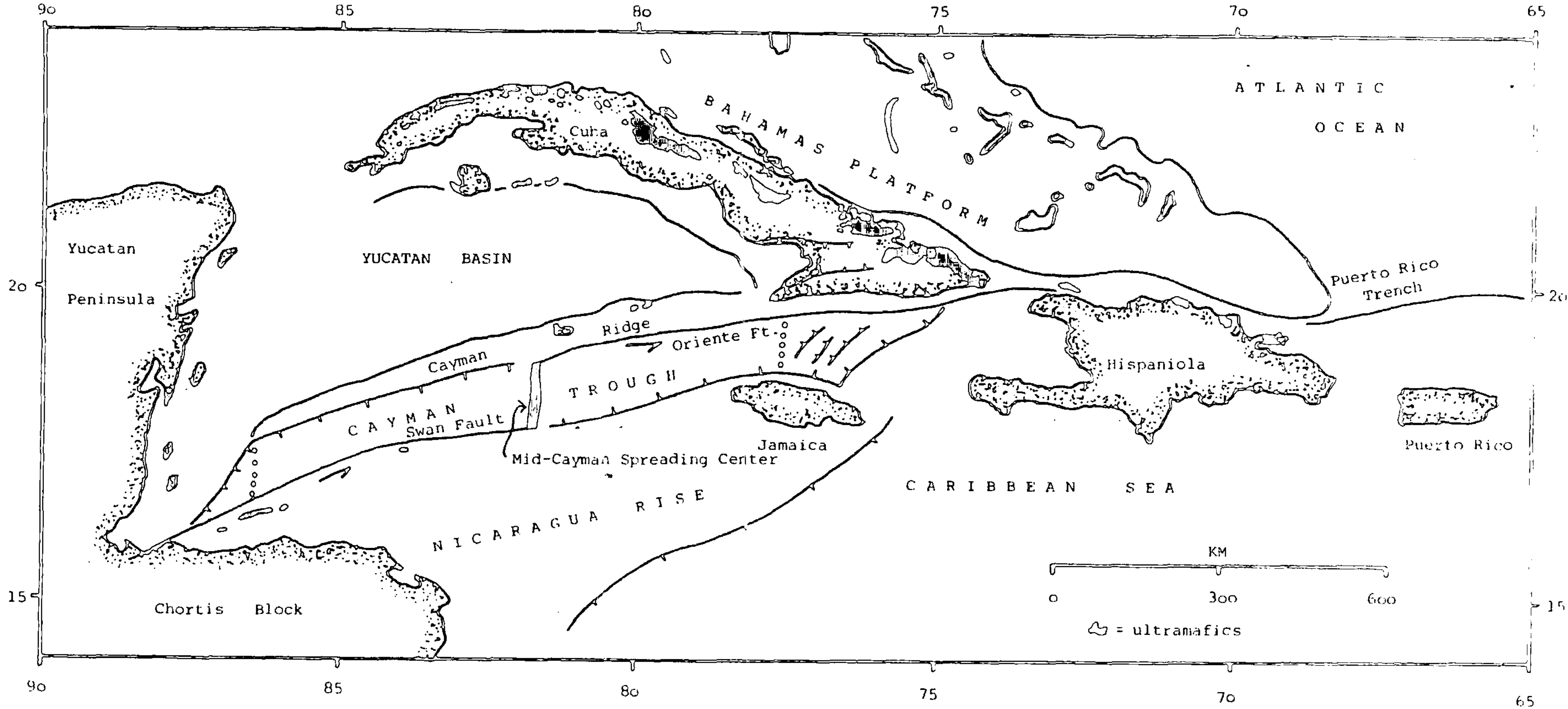


Figure 4.1. Geographic map of some major features of the northern Caribbean described in text. Total bathymetric length of the Cayman Trough is 1400 km, but oceanic portion where 100 % extension has occurred (between heavy solid dots) is only about 980 km. Additional extension within the ends of the Trough during initial rifting is probably 70 to 100 km. Realignment of the volcanic axes and of the ultramafic belts of Cuba and Hispaniola requires only 400 km of restoration along the Oriente Fault. Hence, much of the motion associated with the opening of the Cayman Trough must have bypassed to the south the arc and forearc complexes of Hispaniola.

five times faster during the Paleocene-Early Eocene than it was before and after this period [Paulus, 1972], and (6) Upper Eocene flat-lying carbonates overlie unconformably the deformed collision belt.

Along the north coast of Hispaniola, the remains of an uplifted subduction zone indicate arc-Bahamas collision there, as well as in Cuba [Nagle, 1966; 1979; Bowin and Nagle, 1980]. Like Cuba, these authors suggest that the Hispaniolan collision occurred during the Eocene, based on the argument that the nearly flat-lying Luperon sandstone of the north coast (Upper Eocene) overlies the subduction complex unconformably, and that the subduction complex contains rocks as young as Eocene. If so, it could be concluded that Cuba and Hispaniola collided together as single arc with the Bahamas in Eocene time.

However, between Cuba and Hispaniola lies the Cayman Trough, a 1400 km-long basin that is defined by the Swan and Oriente transform faults and the Mid-Cayman Spreading Center (Figure 4.1). The Cayman Trough began to form in the Early Oligocene [Wadge and Burke, 1983], is floored by oceanic crust [Perfit and Heezen, 1978], and is presently seismically active primarily along the above-mentioned fault zones [Molnar and Sykes, 1969]; the migration of the Caribbean Plate past North America is being recorded by the formation of the Cayman Trough. Assuming the entire Trough formed in this way, it would appear that the basin is an exceptionally extended pull-apart basin, whose offset is probably 1050 to 1100 km, taking into account the fact that the ends of the 1400 km-long basin are underlain by attenuated crust rather than true ocean.

Meyerhoff [1966] believed that Cuba and Hispaniola are sufficiently similar to propose that the two islands must have been once aligned, and that the alignment of their metamorphic belts (arc and subduction complex) provides a measure for the total offset along the Cayman Trough. He concluded that the islands have been offset by only 180 km along the Oriente Fault, which presently leaves the eastern end of the Cayman Trough and passes between Cuba and Hispaniola (Figure 4.1). If the offset across the Cayman Trough was any greater, such as 1050 or 1100 km, restoration of the offset would place Hispaniola far to the west of Cuba prior to the Oligocene opening of the Trough, and would bring into question the genetic relationship between the Cuban and Hispaniolan arc segments.

Although the offset across the fault cannot be directly proved from observation of the Trough itself, the period of active subduction at the Lesser Antilles (Eocene to Recent) is strong evidence to suggest that the Caribbean Plate has been migrating east with respect to North America since that time. Further, for lack of any other probable mode of origin for the Cayman Trough, it appears that the pull-apart model of opening explains the geology of the Trough most satisfactorily. We are left, therefore, in a quandary: how can Cuba and Hispaniola have belonged to the same arc prior to the Eocene, and the Cayman Trough have opened by 1050 to 1100 km of strike-slip offset since the Oligocene?

To further complicate matters, recent work by Bourgois et al. [1980] shows that the subduction complex of northern Hispaniola contains rocks as young as Middle Miocene, and that the subduction complex is not overlain anywhere by the Upper Eocene Luperon Formation. They conclude that emplacement of the subduction complex is Middle Miocene in age, following the deposition of the Luperon. The main implication of this would be that the Hispaniolan and Cuban portions of the Greater Antilles belong to different arcs, which collided with the Bahamas at different times. This would have profound influence upon the possible plate-boundary mosaics that could be derived to explain the evolution of the Caribbean region. Unfortunately, no wells exist in the eastern Bahamas to measure a possible

Miocene subsidence phase due to a Hispaniolan collision.

Thus, the premise of an originally continuous Cuba-Hispaniola arc is brought into question. It was decided that the critical area to test the various alternatives was the subduction complex along the north coast of Hispaniola. Was the age of emplacement Eocene or Miocene? If Eocene, did Cuba and Hispaniola belong to the same arc? If so, how can the great offset across the Cayman Trough be explained? Also, is 180 km the best estimate of the Cuba-Hispaniola offset? Finally, when did the offset occur? These questions define the object of the field work undertaken by the author. Most of these can be answered from the study of the north coast. The reconciliation of the Cayman Trough offset, however, must be studied in the southwestern part of Hispaniola; these studies are defined and summarized in Chapter 5.

The morpho-tectonic provinces of Hispaniola:

The physiographic map of Hispaniola (Figure 4.2) shows that the island is formed of four linear cordillera that are separated by three intervening troughs. Lewis [1980] divided Hispaniola into 9 physiographic provinces. In Figure 4.3, a modified version of Lewis' provinces is shown which highlights the various structural elements of the island. Each of the provinces of Figure 4.3 is geologically distinct from the next, and the provinces roughly follow the topographic variations of the island. Hence, the provinces may be considered as morpho-tectonic provinces. The general stratigraphies of the morpho-tectonic provinces of Figure 4.3 are shown in Figure 4.4. The general geology of each province is summarized below in reference to the stratigraphic compilations of Figure 4.4.

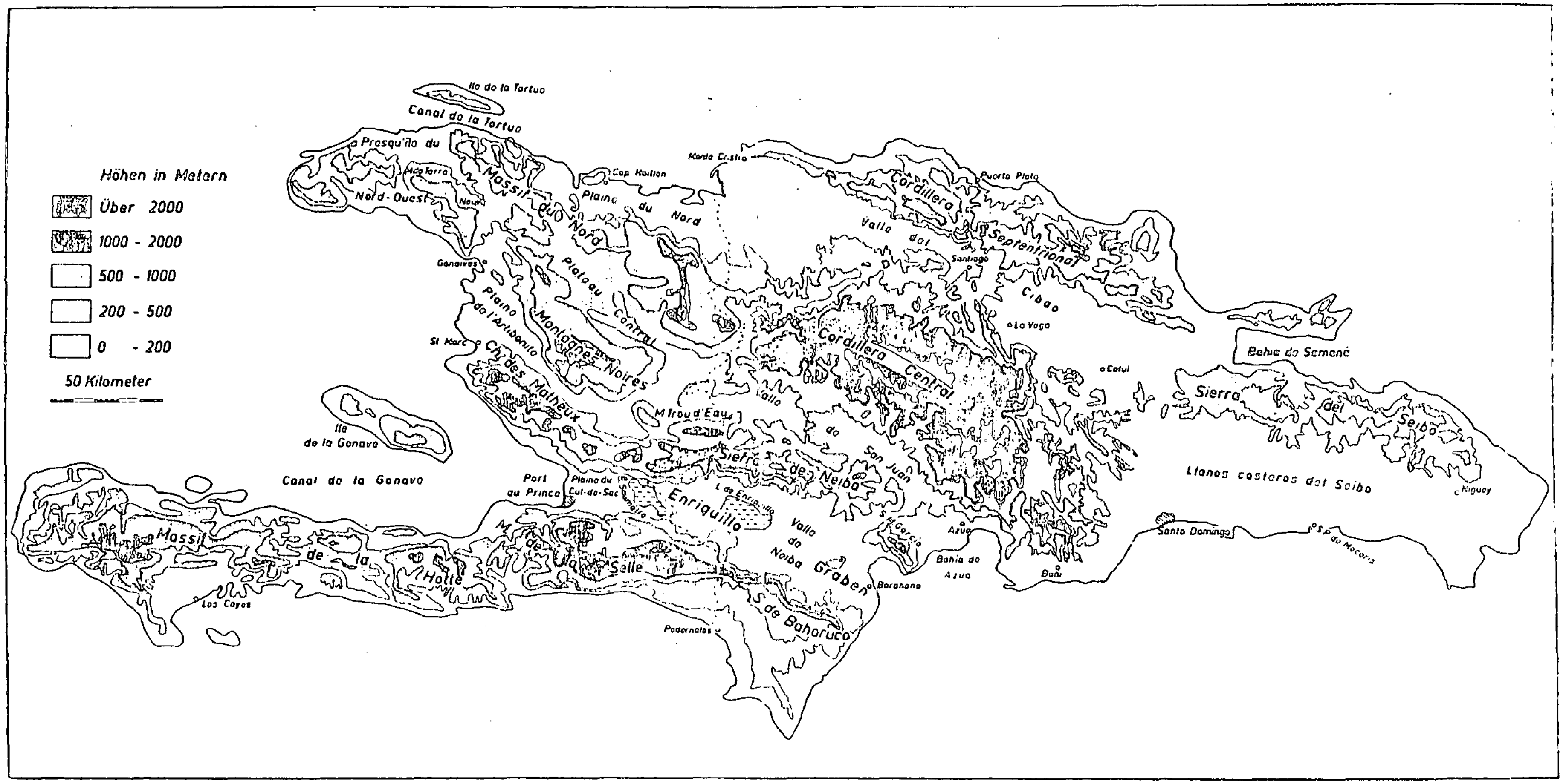


Figure 4.2. Physiographic map of Hispaniola showing general morphological features [after Bowin, 1975].

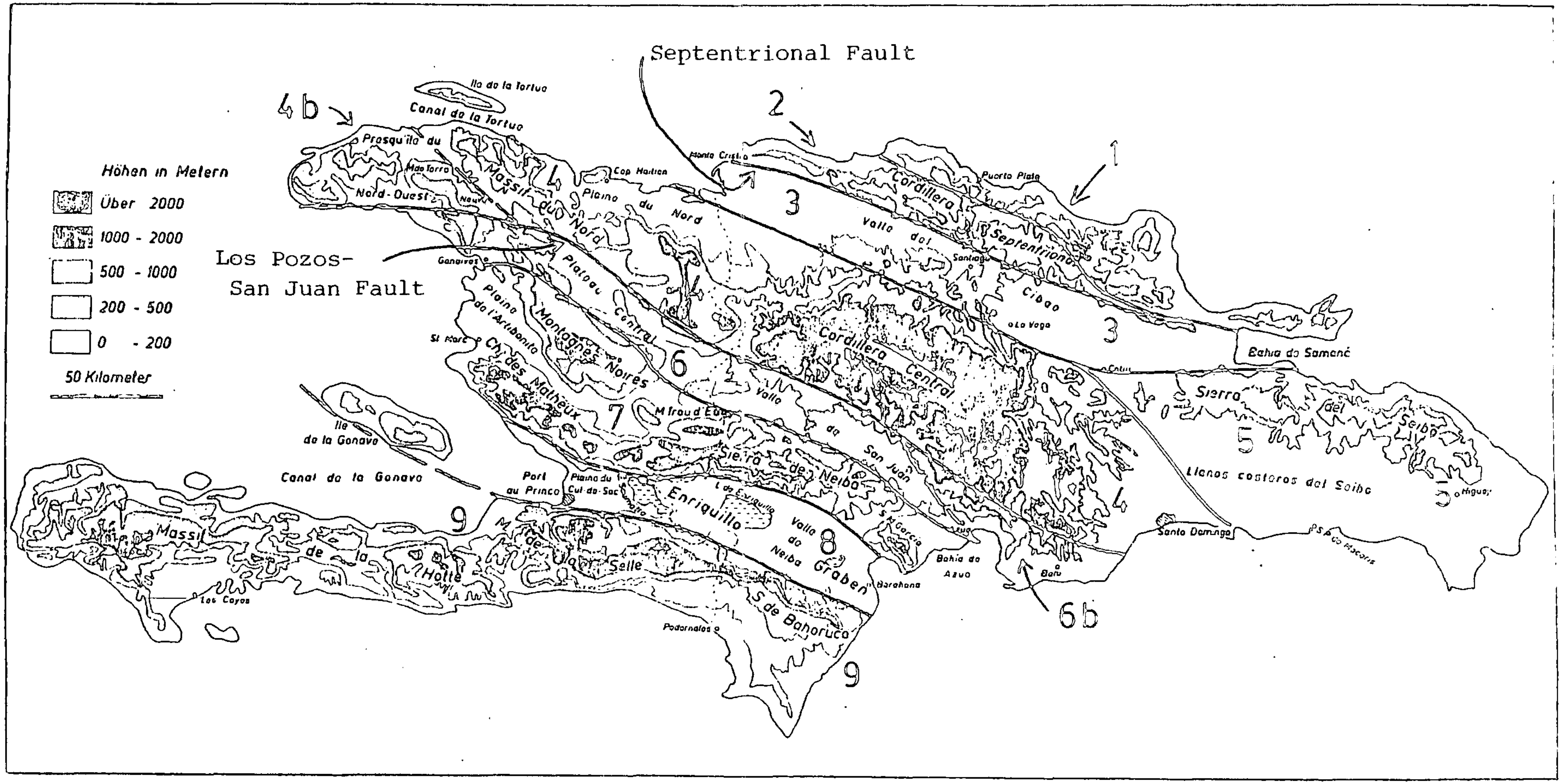
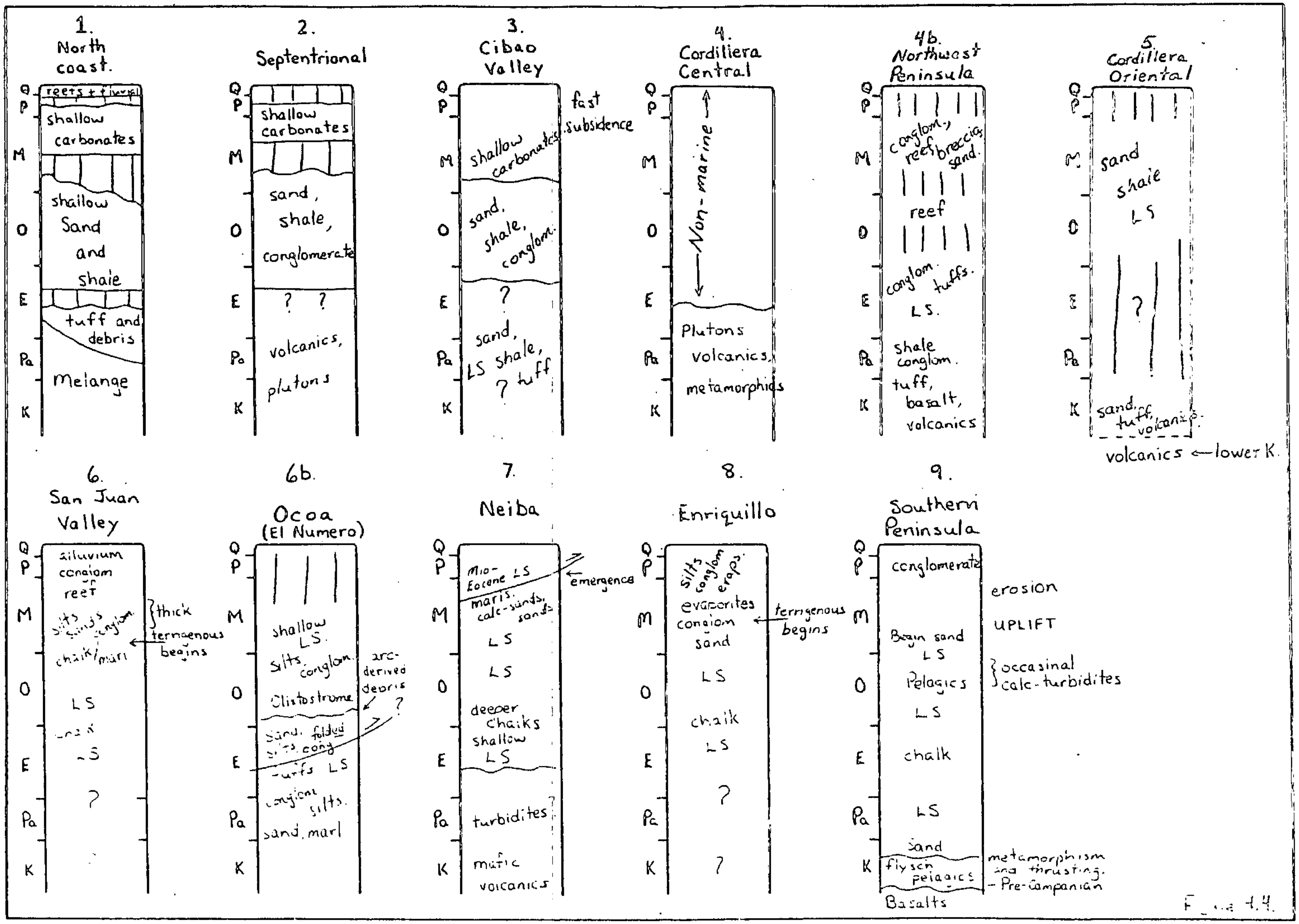


Figure 4.3. Morpho-tectonic provinces of Hispaniola as described in text. Provinces are keyed to generalized stratigraphic columns of Figure 4.4. 1, North Coast subduction complex; 2, Cordillera Septentrional; 3, Cibao Valley; 4, Massif du Nord/Cordillera Central arc complex; 4b, Northwest Peninsula of Haiti; 5, Cordillera riental; 6b, Sierra El Numero; 7, Plateau Central/San Juan Basin; Montagnes Noires/Chain des Matheux/Sierra Neiba; 8, Cul de Sac/Enriquillo Basin; 9, Sierra Bahoruco/Massif de la Selle/Massif de L'Hotte.

Figure 4.4. Generalized stratigraphic columns for the morpho-tectonic provinces defined in text. Columns are keyed to provinces of Figure 4.3.



1) Subduction complex, north coast (this outline essentially summarizes the results of the author's fieldwork). Pre-Middle Eocene ultramafics, gabbro, serpentinized peridotite, sheared volcanics, pillow basalt, limestone, sandstone, tuff, minor chert, and mud comprise a melange that was emplaced in the Early to mid-Eocene. This complex was transgressed by the Upper Eocene-Upper Oligocene Luperon shallow-water sandstone. Luperon-equivalent deposition may have persisted into the Early Miocene towards the east of the province. An Early to Middle Miocene period of folding, faulting, and erosion was followed by the deposition of the the Upper Miocene-Middle Pliocene Villa Trina carbonates. Since the Middle Pliocene, the province has been uplifted, and Quaternary reefs now flank the shores. Reactivation of the melange (mud- and serpentinite-diapirism) has occurred since the Middle Miocene, intruding the Luperon and Villa Trina and incorporating those rocks into the mud matrix.

2) Cordillera Septentrional. A pre-Upper Eocene basement of silicic volcanic rocks and plutons is unconformably overlain by Upper Eocene to Early Miocene sands, shales and conglomerates of the Altamira Formation, which is correlative to, but a deeper-water facies of, the Luperon Formation of province 1. A Middle Miocene period of folding, faulting, and erosion was followed by the deposition of the Upper Miocene to Middle Pliocene carbonates of the Villa Trina Formation. The Cordillera became emergent in the Middle Pliocene, and elevations are now in excess of 2000 meters.

3) Cibao Valley. The pre-Oligocene is poorly known, but the site of the Cibao was probably the proximal forearc basin of the Cordillera Central Arc during the final stages of the convergence between the Greater Antilles and the Bahamas. Volcaniclastics, dirty limestones, and minor

volcanics might be expected. Upper Eocene-Early Miocene depositon is inferred to be similar to that along the basin margins; i.e., conglomerates, sands and shales equivalent to the Tabera Group southwest of Santiago, and to the Altamira Formation of the Cordillera Septentrional. In the Neogene, various shallow-water carbonate units were deposited, such as the Cercado, the Gurabo, and the Mao Formations [Saunders et al., 1980]. These units are thought to be shallow-water equivalents of the Villa Trina Formation in the Cordillera Septentrional to the north. An important aspect of the Cibao Valley is the exceedingly fast subsidence rates for the Neogene. As much as 7 km of Miocene-Pliocene sediment fills the basin asymmetrically, deepening to the north against the Septentrional Fault. Loading by a reverse component of at least 9 km, in association with strike-slip, on the steeply north-dipping Septentrional Fault clearly has played a role in the subsidence mechanism for the basin.

4) Cordillera Central-Massif du Nord. This province is the backbone of the island arc. A metamorphosed Early Cretaceous basement of mafic volcanics (probably a form of oceanic crust) is intruded by Aptian through Eocene intermediate to silicic plutons. This time-span is assumed to indicate the period of active subduction beneath the Hispaniolan arc complex. The flanks of the Cordillera are onlapped by various Cretaceous and Tertiary sedimentary strata, but the range has remained emergent, probably since the Campanian, certainly since the Eocene. Minor Quaternary? alkalic volcanics occur near Constanza [Donnelly and Rogers, 1980], but these are probably not related to subduction.

4b) Northwest Peninsula of Haiti. Some physiographic outlines of Hispaniola (e.g., Lewis [1980]) consider this area a part of the Plateau

Central-San Juan Province, but the thin Tertiary sediments there are underlain by intermediate to silicic volcanics like those of the Cordillera Central-Massif du Nord. Hence, this area is considered part of the main arc system, and is not strictly a separate province. In the Tertiary, conglomerates and sands were deposited in the Eocene and in the Miocene, separated by an Oligocene unconformity with local reefs.

5) Cordillera Oriental. It is not known whether or not to include this area with the Cibao Valley or the Cordillera Central. Lower Cretaceous acidic volcanics, tuffs and basalts are overlain (at El Seybo at least) by an Aptian-Albian rudist-bearing limestone [Hernandez, 1980]. These are followed by sandstones, limestones, tuffs and basic lavas into the Upper Cretaceous. A stock dated at 75 Ma cuts most of the volcanics. Minor terrigenous and marine Oligocene? to Upper Miocene beds locally transgress the older complex. It appears that volcanism largely ceased in Cordillera Oriental by the beginning of the Tertiary; this is distinctly different to the Cordillera Central, where plutonism continued into the Eocene, although indications of Paleogene plutonism may be obscured by poor exposure and erosion. It is likely, however, that the axis of arc volcanism migrated from this area in the Late Cretaceous, perhaps related to changes in the geometry of subduction beneath the arc (see Chapter 8). Although unknown, this history may be more similar to the Cretaceous history of the Cibao Valley rather than to the Cordillera Central.

6) Plateau Central-San Juan Basin. This valley, and all areas to the south as well, possesses no record of Cretaceous to Eocene island arc volcanism. This is clearly distinct from the northern half of the island. Basement in this province is poorly known; Eocene pelagic

limestones and cherts (Neiba or Plaisance Formations) form the lowest known section. Basalts exposed in the Sierra Neiba and the Southern Peninsula to the south are probably indicative of basement throughout the southwest of the island. The basalts are correlative to the B" seismic horizon of the Caribbean Plate. In the valley proper, no sign of a terrigenous (arc-derived) sedimentary input is known until the Early Miocene [Michael, 1979]. This is discrepant with the Cordillera Central's emergence throughout the Tertiary. Hence, the northern boundary of the basin (Los Pozos-San Juan Boundary Fault) is proposed as a site of large strike-slip offset which occurred during the Middle Eocene to Early Miocene period (see chapter 6). This motion brought the San Juan Basin into juxtaposition with the Cordillera Central, so that arc-derived material is present only in the basin's post-Oligocene sections. Carbonates with a terrigenous component continue in the basin through the Miocene, after which time the basin became mostly emergent. Pliocene to Recent thrusting along the Los Pozos-San Juan Fault zone obscures direct assessment of the strike-slip history.

6b) Sierra El Numero. In the southeasternmost part of the San Juan Basin, in the Sierra El Numero, Eocene conglomerates and sandstones (formerly thought to be Upper Cretaceous) are folded and thrusted in a fashion that is similar to that in an accretionary prism. The south-verging complex is overlain by Oligocene conglomerates and olistostromes. The clasts of the Eocene and of the Oligocene sediments are arc-derived. It is suggested that this "accretion" is due to a convergent component associated with the Eocene-Early Miocene strike-slip motion along the Los Pozos-San Juan Fault Zone, rather than due to subduction <u>sensu stricto</u>. No deposits other than sand, shale and conglomerate are known, there is no known metamorphism in the belt, and there are no known Upper Eocene-Oligocene subduction-related volcanics to the north. The Eocene sediment was probably deposited along the length of the arc, but the subsequent strike-slip motion is responsible for its present position in the southeast only. Hence, the magnitude of offset may exceed 250 km.

7) Montagnes Noires, Chaine des Matheaux, Sierra Neiba. Basement in this area consists of Cretaceous mafic volcanics, with no indication of subduction-related volcanism. Volcaniclastic turbidites of possible Paleocene age occur only locally. The first regional sedimentary succession includes Middle Eocene bioclastics (Plaisance Formation), and Upper Eocene pelagic carbonates (chalks) and cherts (Neiba Formation). These sediments show no indication of the Eocene orogenesis that was occurring synchronously in the Cordillera Central. As stated above, large strike-slip offset probably has since occurred along the Plateau Central-San Juan Basin. Pelagic limestones (Sombrerito and Lemba Formation) were deposited through the Lower Aquitanian (Lower Miocene). In Sierra Neiba, these are followed by marls, bioclastics, and volcano-sedimentary sediments of Upper Aquitanian to possibly Serravallian age [Bourgois et al., 1979a]. These are the first sediments to show a terrigenous source (Cordillera Central). During the Tortonian or later, these Middle Miocene sediments were overthrust from the north (south-vergent thrusting) by Eocene-Miocene sedimentary rocks. Uplift and possibly emergence of Sierra Neiba occurred in the Late Miocene, in association with the subsidence of the Enriquillo Basin to the south. This uplift geographically separated the San Juan and Enriquillo Basins, probably for the first time.

8) Enriquillo-Cul de Sac Basin. Basement in this basin is unknown, but by analogy with surrounding areas it is probably oceanic crust affected by

the B" basaltic extrusions. Likewise, the Plaisance and Neiba Formations probably occur, but this is unconfirmed to the author. Late Oligocene pelagics and minor shale are known, and deep-water conditions generally persisted into the Middle or Late Miocene. The Fondo Negro subbasin in the east was filled at rapid rates during the Late Miocene [Cooper, 1983], probably in association with the southward advance of thrust sheets of Sierra Neiba (foredeep relationship). The Enriquillo-Cul de Sac has since remained near sea-level, with periodic transgressions and regressions.

9) Sierra Baharuco, Massif de la Selle, Massif de l'Hotte. Basement consists of Caribbean seafloor with B" basalts which was extremely deformed during the Campanian-Maestichtian [Maurrasse et al., 1979]. Various Upper Cretaceous-Early Eocene turbidites containing basaltic and carbonate clasts (Beloc, Marigot Formations) locally overlie the basement locally. In Sierra Bahoruco, an Upper Cretaceous to Lower Paleocene unconformity occurs, and these clastics are absent. In the Paleogene, carbonate banks separated by deeper areas developed, as recorded by the Plaisance, Neiba and Jeremie Formations [Maurrasse, 1982]. Carbonate deposition continued into the Miocene. A regional uplift or shallowing of the peninsula began in the Middle? Miocene, associated with the general north-south compression witnessed across southern Hispaniola.

<u>Geology and evolution of the Puerto Plata area:</u>

The location of the field area relative to other geographic features in northern Dominican Republic is shown in Figure 4.5. The area circumscribes a fenster through the Miocene-Pliocene marls and limestones of the Cordillera Septentrional, which exposes a portion of the subduction complex of the north coast (province 1 of Figure 4.3).

Previous work

Most of the work to date on the area was done by F. Nagle [Nagle, 1966; 1969; 1971; 1972; 1974; 1979; Bowin and Nagle, 1980]. Prior to Nagle's PhD mapping in the early 1960's, the Puerto Plata area had been included in a general paleontological study by Bermudez [1949], in which several Tertiary ages on the sandstones and carbonates in the area were reported. Gabb [1873] and Vaughan et al. [1921] give general descriptions, but only from an historical perspective. More recently, the area has been included in the general mapping by Eberle et al. [1980] of the entire Cordillera Septentrional. Also, some paleontological work by Bourgois et al. [1980] has provided some new information on the age of certain units in the area.

Until the time of this writing, no geological map of the area on a proper topographic basemap exists. Filling this void is one of this work's objectives. Nagle's work provides, however, quite detailed assessments of the lithologies present. His petrographic analyses are quite complete; few additional lithologies were discovered by the writer. The work of Eberle et al. [1980] for the Puerto Plata area relied heavily upon the previous work of Nagle. As indicated above in the definition of the field problem, the main point of contention regarding the Puerto Plata area is the timing and style of emplacement of the subduction complex. The Nagle "school" argues for Eocene, while Bourgois et al. [1980] calls for Maestrichtian-Paleocene elevation of peridotite to the photic zone, followed by a Miocene phase of

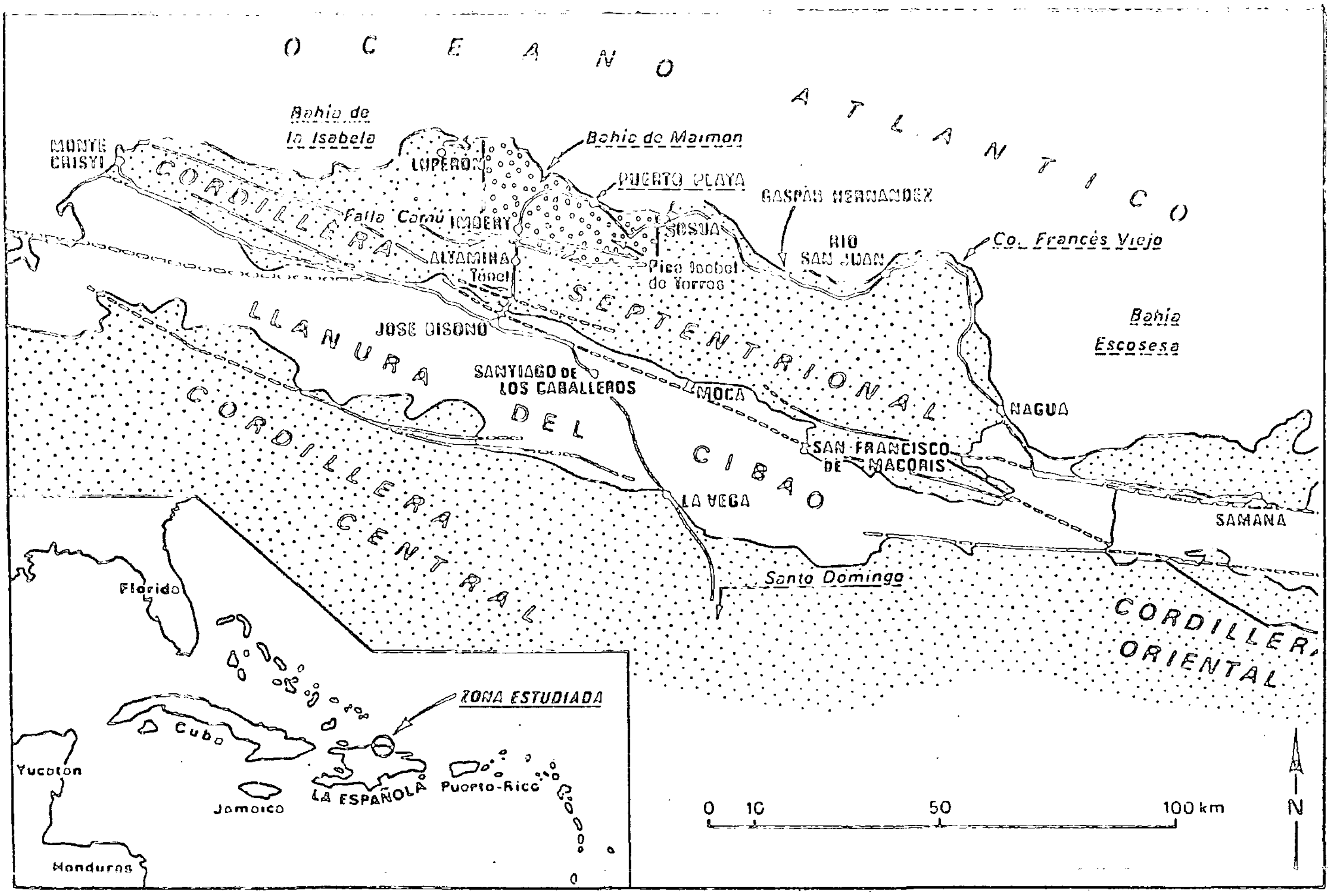


Figure 4.5. Location map of the field area (box with heavy stipple), placed in a regional context (modified after Bourgois et al. [1980]).

"tangential" tectonics (emplacement). Hence, the author's mapping concentrated on (1) the definition of the lithologic units present, and (2) the relationships in time and space between the various units, rather than on a comprehensive lithologic examination. The findings are interpreted under the more recent philosophy of subduction complexes in general [Cloos, 1982; Karig, 1980; Becker and Cloos, 1985; Williams et al., 1984; Murphy and Aalto, 1983].

Geological map of the Puerto Plata area

About 15 weeks of field work around Puerto Plata were done over the summer of 1982 and the winter of 1985. The resulting map is shown as Plate 1. Mapping was done at a scale of 1:25,000 on 100% blowups of the 1:50,000 sheets (20 meter contour interval) of the U.S. Army map service topographic maps of the Dominican Republic (updated in 1966-1967). Plate 1 is presented at a scale of 1:50,000.

Localities in the map area are identified in the text by the Universal Transverse Mercator Grid System (UTMG). This coordinate system is shown on Plate 1. Using this system, and mapping at a scale of 1:25,000, outcrop localities may be defined to a realistic accuracy of 50 meters. By convention, the north-south trending coordinate is cited before the east-west coordinate.

Fossil localities where age determinations were available to the author are identified on the map in large circled numbers. These are defined below, after the description of the stratigraphy.

Stratigraphy of the Puerto Plata area

The single most striking characteristic of most of the rocks around Puerto Plata is that they occur in a chaotic manner [Nagle, 1972]. Nagle [1966; 1972] referred to one of the units present as an olistostrome (San Marcos) to explain the occurrence of a variety of allochthonous blocks, or knockers, sitting in a mud matrix. More recently, the idea of the area representing a tectonic melange of trench fill has emerged [Bowin and Nagle, 1980; Bourgois et al., 1980]. Figure 4.6 compares the stratigraphy of the area from the views of Nagle [1966] and Bourgois et al. [1980].

The present work acknowledges strengths to both compilations. The author's stratigraphic column for the Puerto Plata area is shown in Figure 4.7. Briefly stated, it is agreed that a tectonic melange with mud encompassing all of the lithologies in the Puerto Plata area exists, following Bourgois et al. [1980], but it is contended that the melange was initially emplaced during the Eocene, following the Nagle "school," and has been reactivated and remobilized since the Middle Miocene. The rock units, as they have been mapped in Plate 1 and schematically shown in Figure 4.6, are described below.

<u>The Puerto Plata Melange</u>:

The Puerto Plata Melange consists of serpentinized peridotite, layered cumulate ultramafics and gabbros, sheared volcanics of intermediate composition, pillow lavas, minor pelagic sediments, blueschists, amphibolites, marble, greenshists, andesitic tuffs, shallow-water limestone, and a variety of sedimentary rocks including turbiditic limestone, sandstone and mudstone, all set in an unconsolidated, weakly foliated, sheared mud matrix. In some areas, the solid constituents volumetrically dominate, but never exclude, the mud matrix. In other places, the mud dominates, but never excludes, the solid constituents. These various lithologies are

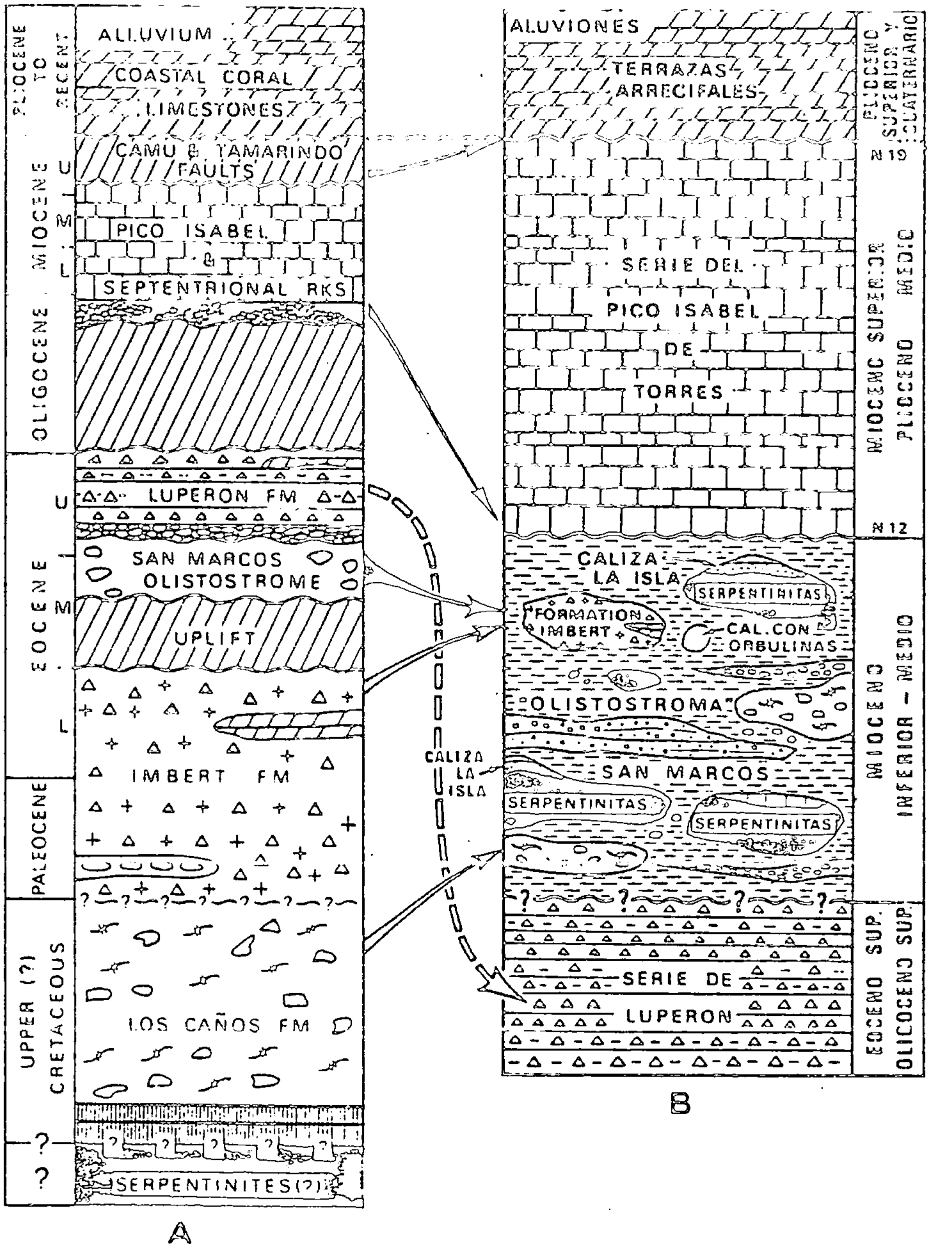


Figure 4.6. Previously defined stratigraphic columns for the Puerto Plata area. A, after Nagle [1972]; B, after Bourgois et al. [1980].

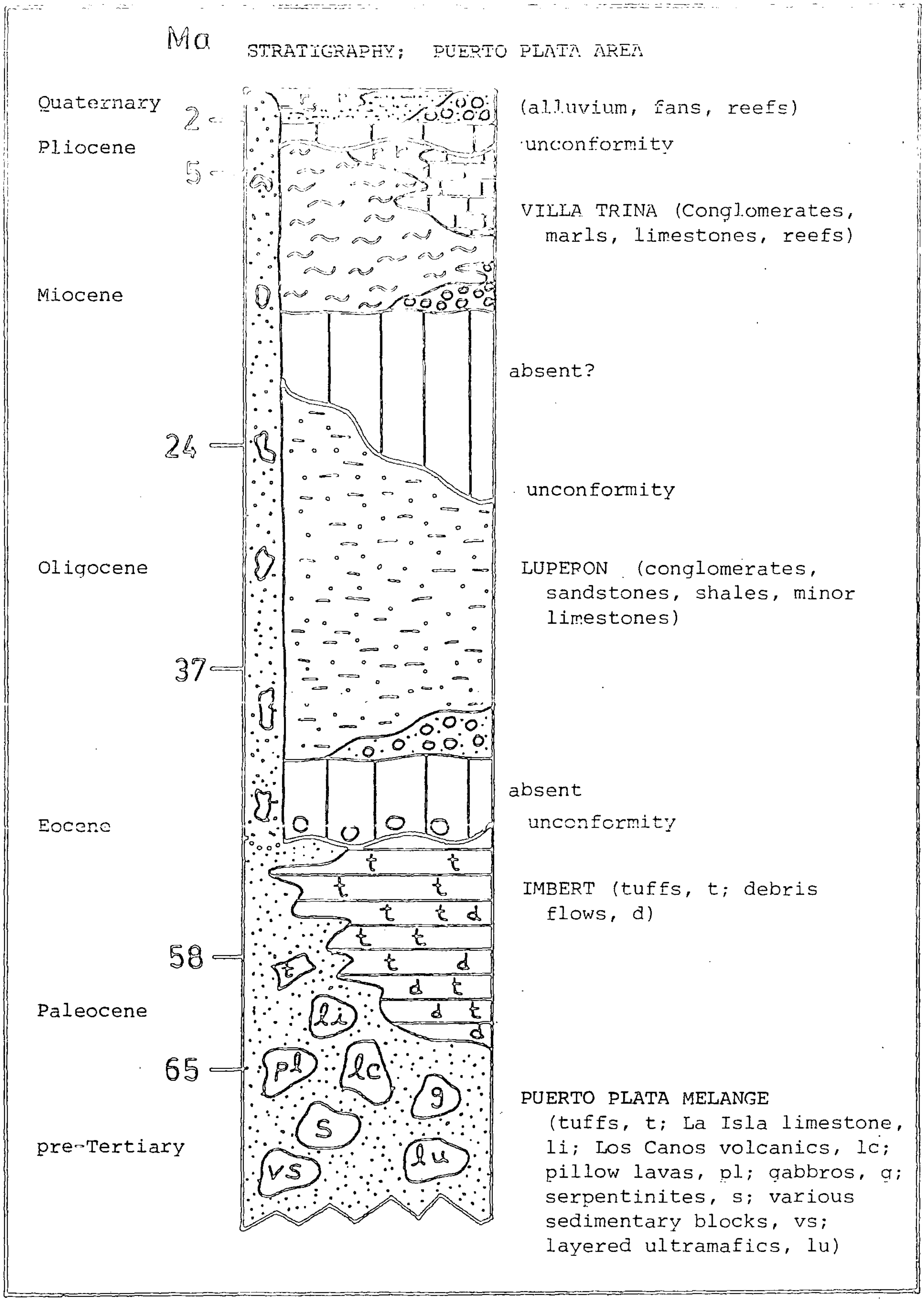


Figure 4.7. Stratigraphic column of the Puerto Plata area as defined in this study.

better described below.

Serpentinized peridotites: Nagle [1966] defined two types of serpentinite in the field area, massive and brecciated, neither of which was emplaced magmatically. This is confirmed by this study. Roadcuts along the new Santiago-Puerto Plata highway show that the massive serpentinites are internally sheared in a phacoidal fashion, on a centimeter to decameter (10 meters) scale. Blocks of other lithologies such as sheared volcanics and schists are occasionally interlaced with the phacoids, indicating intense tectonic shearing.

The brecciated serpentinites are more diverse. Well-rounded fragments of serpentinized peridotite are encased in a lighter green to grey matrix consisting mostly of serpentinite gouge. The fragments range in size from sand to boulders one meter in diameter. The matrix ranges from about 10% to as much as 70% of the total rock volume. The rounding of the fragments is probably tectonic in origin, but a sedimentary origin for the "breccias" cannot be ruled out. In certain areas, they look very much like a typical conglomerate consisting of peridotite clasts within a serpentinite-derived matrix of clay to sand sized particles. In at least three separate localities west of Maimon Bay (31270/219350, 31020/219540, and 31050/219710), limestone clasts of 3 cm to 25 cm are among the fragments in the peridotite breccia. This, however, could be due to either sedimentary or tectonic mixing.

Layered cumulate ultramafics: In a cliff section along the coast to the east of Maimon Bay (31650/219420), tightly folded layered ultramafics occur. The occurrence is fault-bounded and unrelated to the neighboring rocks; it seems to be a block exceeding 40 meters in diameter, jumbled in with the gabbros and volcanics also present along this portion of the coast. The hills immediately south of this occurrence may be layered cumulate ultramafics as well, but the exposure in these hills is insufficient to tell for certain. No other definite layered cumulate ultramafics were seen in the Puerto Plata area.

Gabbros: Along the coast to the east of Maimon Bay (31620/219420) and along the new highway (31720/219280), cliff sections of coarse-grained leucogabbro are exposed. On the coast, these are clearly layered cumulates. Some horizons are darker that the rest, and contain a greater percentage of pyroxine. The gabbros are very fractured and hydrothermally altered, with calcite veins filling the fractures. As with the ultramafics, the gabbro bodies appear to be blocks, two of which exceed 100 meters in length.

In one locality on only a meter scale (31580/219420), a dark diabase intrudes the gabbro. In another locality on a meter scale (31680/219430), the gabbro intrudes the ultramafics. Both instances probably occurred in the original magma chamber as perturbations in the convective cooling process.

Sheared volcanics of intermediate composition: This constituent correlates with the volcanic flows of Nagle's [1966] Los Canos Formation. In most occurrences, they are extremely sheared and so weathered that decent samples could not be taken. Nagle [1966] considers them intermediate volcanics, and in some localities true andesites. The sheared volcanics are the only known rocks north of the Camu fault that could be related to island-arc volcanism. The basement of the Cordillera Septentrional immediately south of the Camu Fault, however, is composed of intermediate volcanics. It is suggested that the sheared volcanics (Los Canos) owe their source to the hanging wall of the overriding plate, and were torn off during the subduction accretion process prior to collision with the Bahamas.

Pillow lavas: Unquestionable pillow lavas occur at only one locality, which is completely surrounded by the mud matrix (31750/218810). Micritic limestone cement between the pillows gives an Early Cretaceous (pre-Aptian) age [Bourgois et al., 1980]. The pillows are not sheared or highly fractured, suggesting that they were probably emplaced into the melange late in the subduction history. They probably originated from the downgoing Atlantic crust just before collision with the Bahamas. Plate reconstructions (see Chapter 3) predict an Early Cretaceous age for the proto-Caribbean seafloor to the south of the Bahamas.

Minor pelagic sediments: At the locality of the pillowed lavas mentioned above, one finds fragments of red and gray pelagic sediments in the surrounding mud matrix. Other pelagics occur randomly throughout the melange. However, no ages have been reported for these fragments, and their significance to the history of subduction is unknown.

Blueshist, greenschist, amphibolite, marble: Blocks of various metamorphic grade occur throughout the melange. Blueschists are common in the area immediately east of Imbert (around 31240/218580). Here, Nagle [1966] reports that they occur within the serpentinite, which is confirmed by the author, but occurrences were also found resting on the mud matrix. They probably occur in both the mud and the serpentinite. There is perhaps a relationship between this area of blueschists and the proximity of the Camu Fault. Discussed later is a case for the Camu Fault defining the contact between the subduction melange and the basement of the overriding forearc complex. If true, the Camu is the logical vent by which originally deep-seated blueschist could have escaped quickly to the surface, particularly if subduction was oblique prior to collision.

Greenschist and amphibolite grade rocks occur erratically. One good example of an amphibolite is at (30540/218940), where a ten meter block rests on mud matrix. This area was previously unrecognised as being underlain by the mud matrix. An area dominated by greenschists is at (30580/218890), where serpentinite is also present.

Marble was reported along the new highway by Bowin and Nagle [1980], but the author was unable to find this occurrence. Other recrystallized limestones approaching marble grade are common in the mud matrix.

Tuffs: The Puerto Plata Melange also contains blocks and fragments of usually beige to white tuff. Nagle [1966] included coarser-grained tuffs in the Los Canos Formation, whereas he placed finer-grained tuffs into the Imbert Formation. In this study, all tuffs are considered to be of the Imbert Formation.

The presence of tuff fragments and blocks interspersed throughout the melange suggests that (1) tuff deposition occurred in the trench during active subduction and accretion, or (2) tuff deposition post-dated subduction, and the melange became remobilized, and then intruded the tuffs sometime after the collision. The degree of interworking is considerable, however, and mud diapirism into an overlying tuff sequence probably couldn't have achieved the degree of interworking observed. Paleontological dating of the Imbert Formation indicates a Paleocene-Lower Eocene age [Nagle, 1966]; the Imbert thus predates the collision, whether it occurred in the Eocene or in the Miocene. Therefore, the former alternative is preferred here.

Shallow-water limestone: Large blocks of commonly brecciated, recrystallized white limestone occur on and within the melange across the west-central part of the field area. This is the La Isla limestone of Nagle [1966], which he inferred to be Miocene or younger. Bourgois et al. [1980] have determined a Late Maestrichtian or Paleocene age for a block of the La Isla that lies southwest of Maimon Bay (31270/219350). Eberle et al. [1980] disclaim this as the La Isla, but the author has mapped this occurrence as

La Isla as well. Further, probable cobbles and pebbles of La Isla occur in at least three exposures of the peridotite breccia (31270/219350, 31020/219540, and 31050/219710), and in another exposure of the sheared volcanics of the Los Canos (31030/219540). Hence, a pre-Middle Eocene age is preferred by the author. Amoco Production Company examined a sample of La Isla limestone from (31500/219370) and stated the following: the sample is Tertiary in age, was deposited in shallow water, and has been rotated with respect to its original depositional attitude [S. Barrett, pers. comm., 1983]. Hence, La Isla limestone is assumed to be of Paleocene age, deposited in shallow water, accreted to the subduction complex and brecciated therein, and brought to the surface during collision with the Bahamas. The original depositional environment may have been an atoll-like seamount in the proto-Caribbean Sea, or a shallow southward protrusion of the Bahamas carbonate bank.

Various sedimentary rocks: Included as well is a variety of finely laminated sandstones, limestones, and mudstones, most of which appear to be of turbiditic origin. Many show definite soft-sediment deformation, others appear as though they "ruptured" possibly due to instantaneous loss of hydrostatic pressure. All have depositional characteristics typical of trench fill.

In addition to these, both the author (at 31780/218810) and Bourgois et al. [1980] have obtained Middle Mioocene ages on some limestone blocks within the mud matrix. Bourgois et al. [1980] claim this as evidence for Miocene emplacement of the entire melange. However, a case will be made later that these inclusions were added to the melange during its Miocene to Recent phase of reactivation.

Argillaceous mud matrix: In variable percentages, a soft mud that is usually striated on breakage surfaces (equivalent to slickensides) occurs with all of the above lithologies. It is seen "intruding" volcanics and serpentinite (31580/219410), underlying serpentinite (31800/219230), and surrounding most of the blocks of the melange. It effectively forms a matrix to the unit as a whole. Rather than treating the serpentinite and volcanics as basement that is overlain by an argillaceous olistostrome, as did Nagle [1966], this author feels that all of the above described lithologies are incorporated as variably-sized blocks within the mud matrix.

The Puerto Plata Melange: the ophiolite suite? Most of the above constituents, which form sand-sized to greater than 100 meter (and possibly km) blocks in the mud matrix, form an association that could be interpreted as having originated from the ophiolite suite. The author favors this interpretation and suggests that the collection as a whole comprises a single unit of subduction-related melange. The various turbiditic lithologies, the La Isla limestone, and the Imbert tuffs were probably incorporated during the final stages of subduction, prior to final ocean closure between the arc and the Bahamas.

The Imbert Formation:

Although blocks and fragments of the Imbert tuffs are included in the Puerto Plata Melange, The Imbert Formation overlies the melange, with far more structural coherency, in a large area to the north of the town of Imbert (Plate 1). Hence, the Imbert is a mappable unit, distinct from the melange, and is given Formation status.

Nagle [1966] defined the Imbert Formation as: a series of tuffaceous rocks with minor thin-bedded, fine-grained, turbid limestones and rare green radiolarian cherts. The tuffs can be divided into light to dark gray calcareous crystal tuffs and white, fine-grained vitric tuffs. Pelagic forams from two tuffaceous rocks indicate a Paleocene-Lower Eocene age, but this may be the older part of the formation only. The fine grain size and the presence of radiolaria suggest deep-water deposition.

The author extends the definition of the Imbert Formation to include turbiditic beds, or debris flows, of sand and conglomerate that contain angular to sub-rounded lithic fragments up to 10 cm of foliated metamorphics, serpentinites, volcanics, and green siliceous fragments. These are interbedded with the white tuffs of the Imbert Formation to the east of Imbert (31020/218550), but the best exposure of this facies occurs along the new highway at (31250/219000). This latter exposure has been subjected to soft-sediment deformation. The general map pattern suggests that these debris beds are in the lower part of the Imbert Formation.

The Imbert can be interpreted as orogenic trench-fill sediment. The presence of serpentinite and metamorphic clasts in the debris flows indicates that some portion of the accretionary complex had been uplifted prior to the uplift of the Puerto Plata portion. This is suggestive of an oblique collision between the arc and the Bahamas. Some of the Imbert was incorporated into the Puerto Plata Melange, whereas the Imbert Formation *sensu stricto* (Plate 1) overlies the melange. Hence, it appears that the Imbert was deposited during the phase of active subduction, but was not entirely accreted. The author suggests that deposition of the Imbert spanned the period of collision between Hispaniola and the Bahamas. Much of the deformation within the Imbert Formation probably results from the uplift and emplacement of the trench fill against the Bahamas.

In the field area, the contact between the Imbert and the overlying units is usually littered with blocks of variable lithology and size. Serpentinite, contorted limestone, sandstones, metamorphics nodules of chalcedony (members of the melange) are typical. In some places, the mud matrix is present as well. It would seem that during the emplacement of the subduction complex, mud diapirs broke through the overlying tuffs and turbidites and locally spread out across them. This is not unlike the melange diapirs intruding the Cambria trench slope basin near Cambria, California [Becker and Cloos, 1985]. The resulting surface, after emplacement, was exposed to erosion and a profound unconformity atop the Imbert Formation and the diapiric areas of the Puerto Plata Melange developed. This unconformity is Middle Eocene in age, as indicated by the overlying, transgressive Upper Eocene to Upper Oligocene Luperon Formation.

The Luperon Formation:

Nagle [1966] defined the Luperon Formation as a "1 km thick repetitious sequence of poorly indurated buff and yellow-orange calcareous tuffaceous shales, bioclastic buff limestones, and calcareous tuffaceous sandstones. Generally thick-bedded calcareous sandstones alternate with thin-bedded shales and rare limestones. The beds are graded, both on the scale of the internal structure of individual beds and on a larger scale of several beds in sequence as seen in any one outcrop. Thick-bedded conglomerates form the base of the formation," which overlies the Imbert Formation with angular unconformity.

The present author follows this definition of the Luperon Formation, although the aereal extent of the formation has been extended to include the area north of the town of Imbert (Plate 1). In fact, the stream crossing beneath the new highway in this area (30780/218640, and continuations of stream) provides the best exposures of the Luperon, and its contacts with the Imbert, known to the author. Here, the contact can be identified to within 10 meters, and the horizon is composed of blocks up to 2 meters comprising a variety of lithologies. This collection of blocks can be regarded as a basal conglomerate of the Luperon, but the larger blocks are probably residual boulders of the unconformity that were not reworked by the waters of the transgression.

Paleontological age determinations reported by Nagle [1966] and by Bourgois et al. [1980] indicate a Late Eocene to Late Oligocene age for the Luperon. It is unknown how much of the upper part of the formation has been removed by erosion.

The Luperon is generally gently dipping, except near the traces of faults such as the Camu. Its deposition clearly post-dates the main period of emplacement of the Puerto Plata Melange, as it is often flat-lying and overlies the Imbert Formation. However, at (30830/218640) the mud matrix of the Melange overlies gently-dipping sands of the Luperon. As outlined later, this is interpreted as being due to Miocene-Recent remobilization and diapirism of the melange unit.

The Villa Trina Formation:

Throughout much of Cordillera Septentrional and the plains and hills along the north coast, latter Tertiary carbonates overlie the older rocks. Nagle [1966] did not name these rocks in the field area, but recovered Miocene ages from the carbonates of Pico Isabel des Torres. Bourgois et al. [1980] referred to these carbonates as the Pico Isabel Series. Eberle et al. [1980] followed Vaughan et al. [1921], and referred to the Septentrional carbonates as the Villa Trina Formation. The carbonates are similar in age and lithology throughout the Septentrional, the north coast, and the Puerto Plata area; thus, the term Villa Trina is adopted here for the Puerto Plata area, although the type locality is outside of the field area.

In the Puerto Plata area, north of the Camu Fault, the Villa Trina carbonates reach 500 meters in thickness and consist of three facies. At the bottom a basal conglomerate is often encountered (33410/218040, for example), and the rounded clasts of 3 to 50 cm in diameter include fine-grained, very hard, dark grey limestone, medium grained basic and intermediate igneous rocks, greenshist-grade rocks of igneous origin, and second-generation conglomerate cobbles. All of these could have originated from the Puerto Plata Melange or possibly the basement of the Cordillera Septentrional.

Overlying the conglomerates are soft, dirty limestones with varying contents of clay, or marls. These form the bulk of the formation in the field area. The marls are usually cream colored, and occasionally interbedded with 10 to 50 cm thick beds of beige to orange bioclastics (such as at 32450/218630). Some of these beds are conglomeratic and contain mafic pebbles. Burrows are common, and coral fragments occur in some of the bioclastic beds.

At the top of the unit, the limestone becomes cleaner and much harder. This hard cap is responsible for the way in which the occurrences of the Villa Trina stand out topographically. In the Cordillera Septentrional, these caps are usually composed of reefs or reef debris. Atop Pico Isabel des Torres, reef debris is present also, but no reefs <u>in situ</u> were seen.

Bourgois et al. [1980] have dated the Villa Trina of Pico Isabel; they report Serravallian at the base and Tortonian at the peak (upper Middle to lower Upper Miocene). Outside the field area at Cabo Frances Viejo to the east, the marly facies is of Messinian age. In the Septentrional, the reefy facies which records shallowing is Middle Pliocene, which is presumably the time at which the Cordillera Septentrional became emergent. It is unclear whether or not these younger sediments originally occurred on Pico Isabel des Torres and have been eroded. Their absence atop Pico Isabel suggests that the peak had a different history relative to the remainder of the Septentrional. Pico Isabel's odd position, standing alone within the Puerto Plata Melange, suggests gravity sliding from an unknown source [Nagle, 1966]. The sliding may have occurred any time after the Tortonian, and was probably related to reactivation of the Puerto Plata Melange.

Quaternary sediments; alluvium, reef terraces, and alluvial fans:

Much of the field area is covered by Quaternary alluvium. In particular, the mud of the Puerto Plata Melange and the tuffs and turbidites of the Imbert Formation are easily eroded, and rivers with wide floodplains cut into them. The deposits are typically muds, sands, and gravels, the gravels being composed mainly of constituents of the Puerto Plata Melange. Streams usually dissect the floodplains by 2 to 5 meters, indicating that the area is undergoing recent uplift.

Around the western, eastern, and northern fringes of Pico Isabel des Torres, broad carbonate alluvial fans have developed. Avalanches occur quite frequently from the cliff sections of the mountain.

In coastal areas in the northeast of the field area and at the mouth of Maimon Bay, Quaternary reef terraces and beachrock are exposed. This is further evidence of recent uplift in the area.

Fossil localities and age determinations

The following list is keyed to the large circled numbers on the map (Plate 1). For each, the age and source is identified.

1. Luperon Formation; Upper Middle Eocene to Upper Oligocene [P. Cepek, pers. comm., 1985].
2. La Isla Limestone; Late Maestrichtian or Paleocene [Bourgois et al., 1980].
3. Villa Trina (in Puerto Plata Melange); Middle Miocene or later [E. Robinson, pers. comm., 1984].
4. Villa Trina Formation; Upper Lower Miocene [P. Cepek, pers. comm., 1985]; Serravallian [Bourgois et al., 1980, but exact location uncertain].
5. Villa Trina Formation; Upper Lower to Lower Middle Miocene [P. Cepek, pers. comm., 1985].
6. Villa Trina Formation; Upper Lower Miocene; [P. Cepek, pers. comm., 1985]; Tortonian [Bourgois et al., 1980].
7. Villa Trina Formation; Upper Lower Miocene [E. Robinson, pers. comm., 1984].
8. Luperon ? or Villa Trina ? Formation; Upper Middle Oligocene to Lower Upper Oligocene [P. Cepek, pers. comm., 1985].
9. Luperon ? or Villa Trina ? Formation; Middle Tertiary and bathyal [E. Robinson, pers. comm., 1984]; Upper Oligocene [P. Cepek, pers. comm., 1985].
10. Fault gouge (from Luperon and/or Villa Trina; Upper Eocene and Lower Miocene, so Lower Miocene [P. Cepek, pers. comm., 1985].
11. Villa Trina Formation; Upper Lower Miocene to Lower Middle Miocene [P. Cepek, [pers. comm., 1985].
12. Luperon ? or Villa Trina ? Formation; Middle Tertiary and bathyal [E. Robinson, pers. comm., 1984].

NOTE: Recently, Saunders et al. [1980] have undertaken to refine the biostratigraphy of the Neogene of the Dominican Republic. Their work is not yet completed, but preliminary results show that all previous age determinations, including the above, are approximately 8 million years too old. This cannot be, and is not, applied as fact to these determinations, although the assigned ages are suspect and, perhaps, should be considered as age maxima.

Structural and tectonic considerations of the geology of the Puerto Plata Area:

Little structural information is available from the field area. However, the following considerations assist with the interpretation of the field area and of the region as a whole.

Foliations

The only rocks possessing foliations are within the Puerto Plata Melange. Many of the knockers have well developed cleavages (greenschists, blueschists, amphibolites) but, as knockers, these provide little insight as to the evolution of the actual field area. They do, however, indicate an origin with high P/T ratios of up to 20 km depth, and hence are integral to the interpretation of the Puerto Plata Melange representing a subduction complex [Nagle, 1966]. The mud matrix of the Melange is locally weakly foliated, with anastomozing cleavage surfaces bearing shearing striations (essentially slickensides). No one outcrop, however, possesses a consistent throughgoing attitide of cleavage. This may be due to the Middle Miocene to Recent reactivation of the matrix.

Folds

The few measurable attitudes of bedding in the Imbert and Luperon formation do not conform to folding about any preferred axis. Dome and basin structure is perhaps more the case, and may relate to the presence of mud and serpentinite beneath these horizons. It is felt that the Imbert and Luperon sediments more or less "ride" on the Puerto Plata Melange, and do not deform according to regional stress patterns. Deformation within the Villa Trina carbonates is brittle, and inclined attitudes are mostly due to faulting.

Blocks within the Puerto Plata Melange are often folded with an associated cleavage. Soft-sediment folds in a sandy greywacke lithology is also common.

Faults

Because of the mobile role of the Puerto Plata Melange, faults within it are difficult to recognize. They may be apparent locally, but in many areas they cannot be traced.

The Camu Fault is the most prominent fault of the field area, and is linear for over 40 km. The Camu defines the southern limit of the Puerto Plata Melange, and the northern limit of the intermediate volcanic basement of the Cordillera Septentrional. The majority of blueschist occurrences in the Puerto Plata Melange crops out within 1 km of the Camu; the fault may have served as a conduit along which the blueschists were emplaced by circulation within the melange. The Camu is interpreted as the boundary between the Greater Antilles forearc complex and subduction complex. Fault gouge and breccias are common along it, but no stretching lineations could be seen due to weathering. No markers are available by which to measure the offset along the Camu. Presently, the Villa Trina of the Septentrional is topographically higher than the coastal hills, but the fact that one can walk south from Melange to Pliocene carbonate suggests that the fault's motion is down on the south. The linearity of the fault and its residence in the sinistral North Caribbean PBZ would suggest that a sinistral component exists as well. The dip of the fault is steep, probably to the south.

Another prominent feature is the Maimon Graben, a topographic low extending southwest from Maimon Bay. Nage [1966] reports fault breccias along the west flank. The graben, if in fact it is a graben, is oriented in a trend that fits a sinistral shear model. However, the highest elevations of the field area flank the southern margin of the "graben," and it is possible that an obscure effect of mud diapirism plays a role in the observed topography.

Finally, Pico Isabel des Torres is an isolated occurrence of the Villa Trina carbonates. Its basal beds are contorted, commonly with shearing striations. Blocks of the limestone occur in the Matirx of the Melange surrounding the Peak. It appears as though Pico Isabel slid into its present position from an unknown place, but the direction of motion could not be deduced.

To summarize, east-southeast sinistral compressive faults and the chaotic reactivation of the Puerto Plata Melange are the primary causes of structural deformation in the field area. Most of the presently witnessed topography is probably Late Pliocene or later.

On larger scale, the pre-Upper Eocene paleogeographic relationship between Hispaniola and Cuba needs to be addressed. Outlined earlier were arguments for an Early to Middle Miocene initiation of strike-slip motion along the Oriente Fault System. Using North America/Caribbean slips rates of 3 to 4 cm/yr [Macdonald and Holcombe, 1978], Hispaniola can be returned approximately to colinearity with Cuba within this time period. Eocene markers on both Cuba and Hispaniola that can be realigned to measure subsequent offset include the arc plutons, the uplifted subduction complexes, and possibly the Guantanamo and Cibao forearc basins of the two portions of the arc.

Figure 4.8 shows a realignment off these features. Total offset is predicted to be 400 km, with 300 or 350 km having occurred offshore northern Hispaniola in the Bahamas Channel, and the remainder of 50 to 100 km having occurred along the various faults that are north of the Cordillera Central (Tabera, Septentrional, and Camu). Hence, it is concluded that Cuba and Hispaniola were part of the same arc complex until the Middle Miocene, at which time strike-slip offset amounting to 400 km began to separate the two. It is interesting to note that terrestrial fauna of Cuba and Hispaniola diverged in the Late Miocene (R. Buskirk, pers. comm., 1983).

Geological history of the Puerto Plata area:

The following account is the author's interpretation of the geologic history of the Puerto Plata area. Most of the story is derived from field relationships and the paleontologic ages of the various formations, but in some instances regional aspects have been incorporated to complete or augment the story. The area's evolution is schematically shown in Figure 4.9.

The critical phase in the evolution of the Puerto Plata area is the Early to Middle Eocene. This is when the subduction complex was uplifted

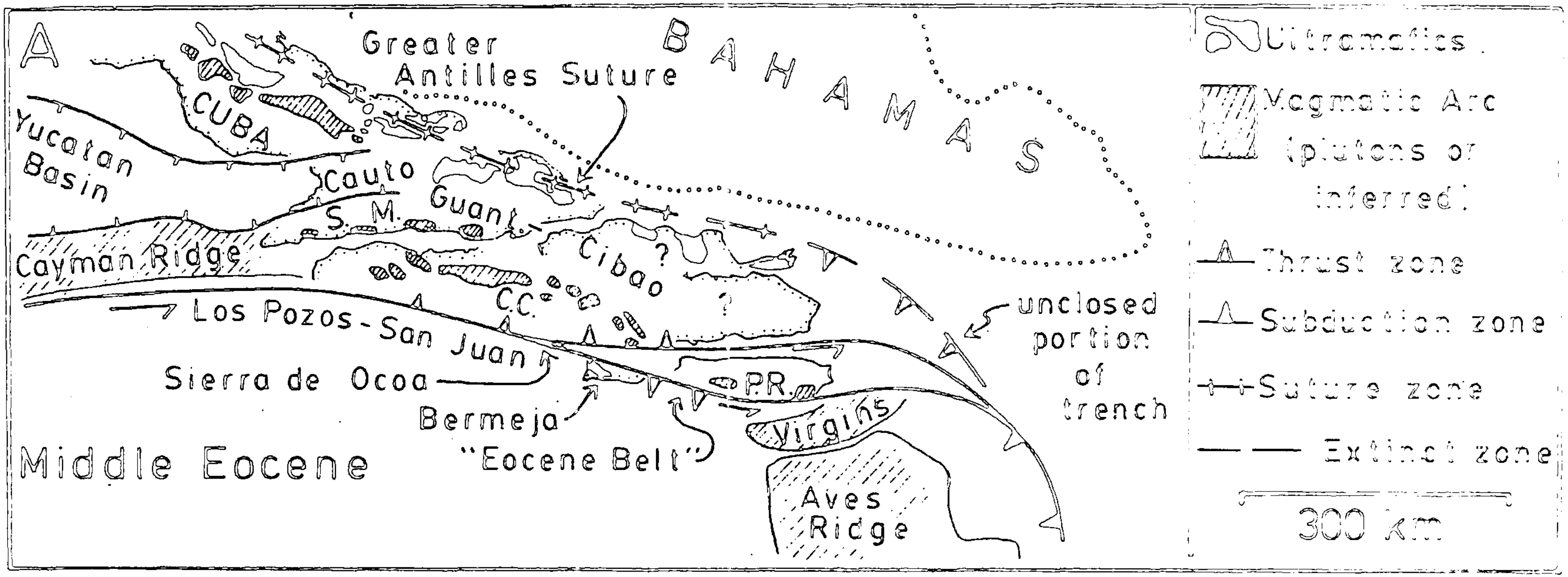


Figure 4.8. Best reconstruction between the Eocene plutons and ultramafic provinces of Hispaniola and Cuba. The Guantanamo-Cibao Basin also aligns, and may represent the original pre-Middle Eocene forearc basin. Total offset along the Oriente Fault System is about 400 km. 300 to 350 of this occurred offshore in the Bahamas Channel; the remainder was shared along the Septentrional, Camu and Tabera Faults. The Tabera Fault was active in the Oligocene, but motion was probably minor.

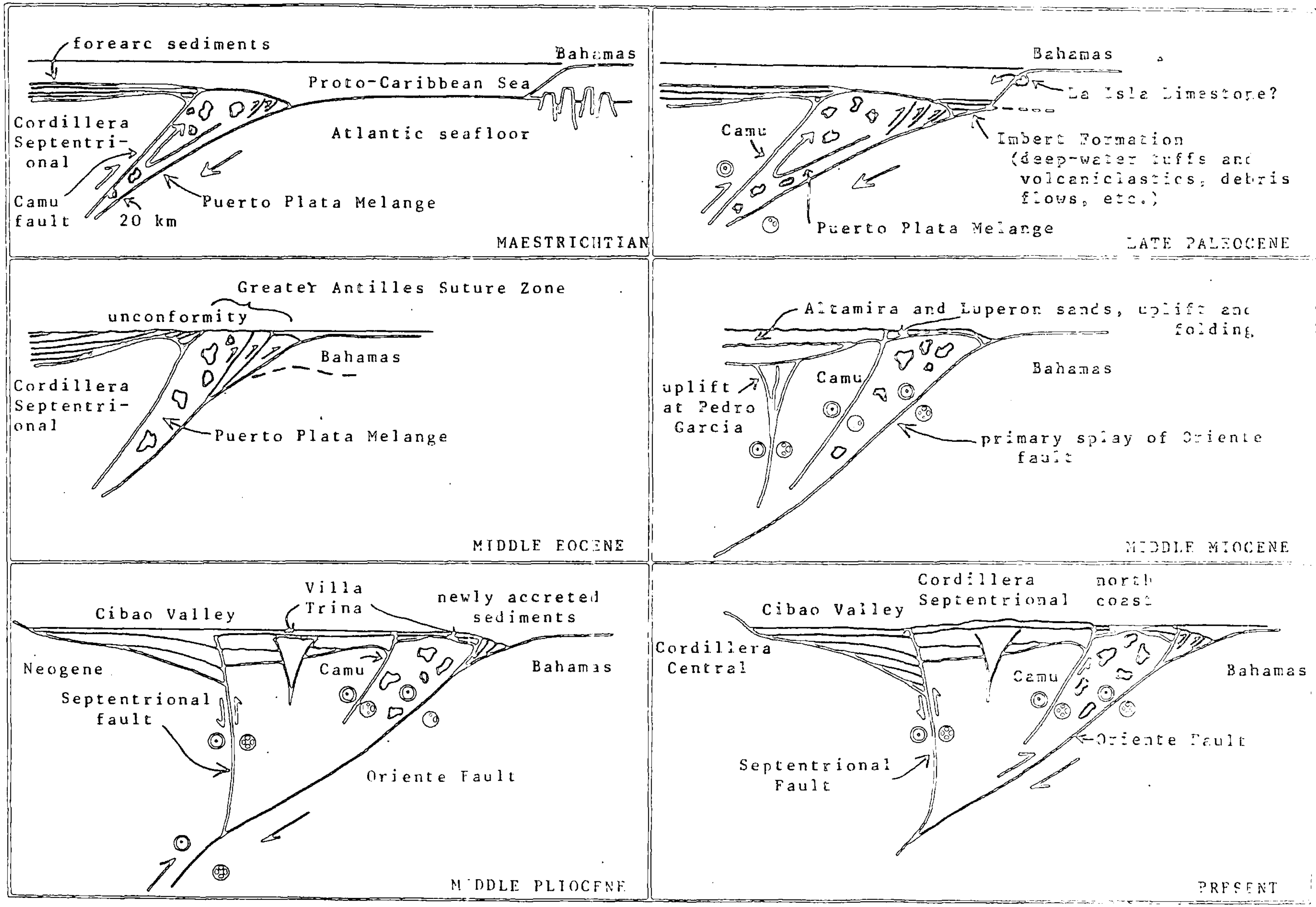


Figure 4.9. Caption on next page.

Figure 4.9. Six-stage cross-sectional evolution of the north coast subduction complex. Camu Fault defines the boundary between the forearc and the subduction complex. Late Cretaceous to Paleocene, subduction and development of the Puerto Plata Melange. Paleocene to Early Eocene, deposition of Imbert tuffs and debris flows, with partial accretion into Puerto Plata Melange. La Isla shallow-water limestone may have been derived from the Bahamas or an atoll-like seamount in the proto-Caribbean. Early to Middle Eocene, primary period of collision eith the Bahamas, uplift and erosion. Upper Eocene-Oligocene deposits (Luperon and Altamira) transgress the orogen. Early to Middle Miocene, sinistral strike-slip motion begins with uplift and minor folding as northern Hispaniola separates from Cuba. Puerto Plata Melange is reactivated, and renewed subsidence through the Middle Pliocene allows deposition of Villa Trina shallow-water carbonates. Middle Pliocene initiation of strike-slip and reverse motion on Septentrional Fault separates Cordillera Septentrional from Cibao Valley, and Neogene deposits of the Cibao are over 7 km thick. Present, uplift continues and the Puerto Plata Melange remains highly mobile.

from abyssal depths to above sea level. The uplift is interpreted as resulting from the collision of the Hispaniolan portion of the Greater Antilles Arc with the Bahamas carbonate platform. Therefore, the geologic history prior to the Eocene is that of subduction. The available information allows the following statements to be made about this subduction phase.

The oldest known sediment is the pre-Aptian micritic limestone cement between pillow lavas in the Puerto Plata Melange. This cement may or may not have survived submergence below the calcium carbonate compensation depth. Whether the basalts originated from the downgoing Atlantic Plate or from the overriding forearc of the Caribbean Plate cannot be determined; both sources should be of Early Cretaceous age.

No other Cretaceous fauna have been reported, except for the Upper Maestrichtian to Paleocene age of La Isla limestone, which here is believed to be Paleocene based on other age determinations suggesting a Tertiary age. Thus, little is known of the Cretaceous history, and the field area contains no information as to the time of the initiation of south-dipping subduction. The only dates on the blueschists come from the Samana Peninsula, where Joyce [1983a] reports ages ranging from 50 to 400 Ma, which helps but a little. The volume of material in the Puerto Plata Melange that has not been dated as Tertiary is very small. This may indicate that subduction began relatively late in the Cretaceous, but terrane migration or underplating may be responsible for this as well. In Chapter 8 on the geological evolution of the Caribbean, a variety of other reasons are invoked to suggest a Campanian age for the initiation of subduction along the north coast of the Greater Antilles Arc. This age is fully compatible with observations made in the field near Puerto Plata.

In the Paleocene, the shallow-water La Isla limestone and the tuffs and turbidites of the Imbert Formation were deposited. The metamorphics and serpentinites in the Imbert turbidites suggest that part of the Greater Antilles subduction complex was exposed by then. The Imbert and La Isla facies are discordant (shallow versus deep), and a mechanism must be responsible for their final juxtaposition. Three alternatives for the La Isla seem reasonable: (1) La Isla limestone was deposited on an atoll-like rise within the proto-Caribbean basin and was accreted to the Puerto Plata accretionary wedge just prior to collision, (2) La Isla was deposited on parts of the subduction complex that were already uplifted, and (3) La Isla was deposited on the Bahamas Banks or some portion of the the leading edge of the Caribbean Plate and tumbled down into the trench at the onset of the collision and was then accreted to the subduction complex with the coeval Imbert Formation. Option (2) is least likely because of the intimate intermixing of La Isla with the melange, but evidence is insufficient to choose between options (1) or (3).

Oblique closure of the trench is suggested by the turbiditic deposition of serpentinites and metamorphics in the Imbert Formation. An along-strike portion of the subduction complex, once uplifted to the surface, would naturally provide detritus of these sorts to the, as yet, unclosed portions of the trench system. In Cuba, collision is accompanied by the propagation of sinistral tear-faults across the orogen, indicating west-to-east oblique closure there (younging to the east). This scenario, if extended to Hispaniola, suggests that the Oriente Province of Cuba should have been uplifted before the Puerto Plata area, and that it may be the source of the Imbert lithic clasts. Indeed, both serpentinites and metamorhics occur in Oriente Province [Case and Holcombe, 1980]. The author suggests that the serpentinite and metamorphics of the deep-water Imbert Formation is evidence for a collision with some degree of obliquity, younging to the east, and that the uplift of the Puerto Plata area post-dates the uplift of Oriente Province, Cuba.

Uplift of the Puerto Plata melange occurred in the Early Eocene while the deposition of tuffs continued on the floor of the rising trench. Mud diapirism probably accompanied this uplift. By the time the complex became subareal, mud diapirs had penetrated the overlying tuffaceous section, and the mud matrix and its exotic blocks flowed, perhaps only locally, onto the tuffs of the Imbert Formation. The Imbert was deformed heterogenously during uplift, but the area to the north of the town of Imbert, although west-dipping and in angular unconformity with the Upper Eocene Luperon, retained coherency and thus is mappable as a Formation.

In the Middle Eocene, the entire area was exposed to erosion, and the mud matrix was largely removed from above the Imbert. The erosion surface consisted of the Imbert Formation, the Puerto Plata Melange, and residual boulders of the eroded diapirs. This unconformity marks the culmination of the arc-Bahamas collision.

In the Late Eocene, the sea transgressed the eroded surface and the Luperon Formation was deposited. The quartz of the Luperon sandstones probably originated from the basement of the Cordillera Septentrional or even the Cordillera Central. Much of the Luperon, however, originated from the ground over which the sea transgressed; the Luperon's sands are largely composed of the older lithologies, plus calcite. Deposition was in primarily shallow water, as evidenced by occasional gypsum beds and the fauna of the sandstones, but shalier layers contain fauna of typically deeper water. The coeval Altamira Formation of the Septentrional to the south is coarser, and abundant marine fans of conglomerate [Redmond, 1980] indicate deeper water conditions.

These conditions prevailed until at least the Upper Oligocene. Early

Miocene to early Middle Miocene rocks are not clearly present, although the very dirty marls (or calcareous mud-/siltstones) of the easternmost area have been assigned "middle Tertiary", which may indicate Oligocene-Early Miocene (E. Robinson, pers. comm., 1984). To the south, the Early Miocene is represented in the upper Altamira Formation of the Septentrional. These horizons contain the well-known Dominican amber deposits. In both the north coast and in the Septentrional, uplift and folding produced a second regional unconformity during the Early or early Middle Miocene. The Luperon and Altamira Formations are more deformed than are the Villa Trina carbonates, the latter of which commonly have a basal conglomerate. Nowhere in the Puerto Plata area, however, was a definite angular unconformity observed by the author.

In addition to the unconformity between the Luperon/Altamira and the Villa Trina, there are major sedimentological differences between the two. In particular, the source areas for the respective deposits were different. The Eocene to Early Miocene deposits are primarily arc-derived, probably from the Cordillera Central/Sierra Madre of Cuba. The Luperon/Altamira depocenter flanked the Cordillera, and bathymetric gradients must have been down to the northeast, in the direction of fan and turbidite deposition in the Altamira [Redmond, 1980]. The Villa Trina, in contrast, has little terrigenous input other than clay and a small volume of conglomerates, both of which were derived from the underlying Luperon and Altamira. Tectonic motions associated with the Early to Middle Miocene uplift (early-Middle Miocene unconformity) apparently changed the geometry of the depocenter so as to prevent continued terrigenous input to the Septentrional and north coast areas from the Cordillera Central. Deposition in the Cibao Valley at this time is not entirely clear: the paleobiostratigraphy is presently being redefined [Saunders et al., 1980]. However, it appears that an assemblage of conglomerates, sands and silts from the Cordillera Central was shed into shallow-water carbonate domains, where the terrigenous and carbonate components were mixed by marine currents (Janico and/or Cercado Formations). It seems likely that prolific carbonate production during shallowing of the basin played a role in preventing terrigenous material from reaching the Septentrional.

The Early to Middle Miocene uplift and subsequent change in deposition is interpreted to be the result of the initiation of stike-slip motion between Cuba and northern Hispaniola. The uplift is indicative of compression which was either related to, or relieved by, strike-slip motion along the Oriente Fault. Hispaniola began to migrate eastwards relative to Cuba, more or less as a part of the Caribbean Plate; the total offset is about 400 km. The Septentrional and north coast regions subsided to neritic to upper bathyal depths (10 to 200 meters) during this migration and the deposition of the Villa Trina.

Important during this second phase of plate motion is the remobilization of the Puerto Plata Melange. The northern coast has experienced far more deformation than has the Septentrional. Also, the present topography is related to this phase. The highest hills and ridges are composed of the softest, most-easily weathered, material in the stratigraphic column, the Puerto Plata Melange (Plate 1).

Perhaps the most dramatic effect of this remobilization is the sliding of Pico Isabel to its present position. It is clearly an isolated occurrence of the Villa Trina. It is suggested that Pico Isabel was detached from another area of Villa Trina deposition and translated to its present position by gliding across, or being carried by, the mud matrix of the Puerto Plata Melange. Such a translation may be responsible for incorporation of Middle Miocene fragments in the mud matrix to the west and south of Pico Isabel. The exact origin of Pico Isabel is unknown, and it is unclear exactly when this may have occurred. The sliding probably post-dates the deposition of the Pica Isabel carbonates, which is Tortonian (14 Ma). Carbonates at the base of Pico Isabel are folded and have retained primary bedding; no soft-sediment deformation was seen and it is felt that sliding occurred after diagenesis, probably in the Late Miocene or Pliocene. Melange reactivation occurs to the present, as witnessed by ever-present mud slides and rapid break up of the highways crossing the melange. The formation of Maimon "Graben" is related to this phase of deformation as well, and has probably occurred in Quaternary times.

Conclusions:

1) Cuba and the northern half of Hispaniola are portions of the same Greater Antilles island arc complex. Their histories are the same.

2) The Greater Antilles collided with the Bahamas in the Late Paleocene to Early Eocene. Collision had terminated by the Late Eocene.

3) The ophiolites of northern Hispaniola are part of a tectonic melange. Mud diapirism from the trench sediments during and just after the collision, and again in the Middle Miocene has occurred.

4) 400 km of strike-slip dissection of the Cuba-Hispaniolan portion of Greater Antilles arc has occurred.

5) The separation has occurred since the Middle Miocene, with a corresponding Middle Miocene unconformity and a subsequent change in deposition.

CHAPTER 5

FIELD RELATIONS IN THE SOUTHERN DOMINICAN REPUBLIC

FIELD RELATIONS IN THE SOUTHERN DOMINICAN REPUBLIC

Introduction:

In chapter 4, morpho-tectonic provinces of Hispaniola were defined and outlined. It was shown that southern Hispaniola is composed of several generally east-west trending belts possessing different stratigraphies. The tectonic style of the northern Caribbean during the Cenozoic suggests that the differing terranes were juxtaposed by strike-slip motions with varying amounts of a compressional component. Reconnaissance observations were conducted by the author at various localities to learn more about the apparent strike-slip juxtaposition story. The observations suggest that facies discontinuities exist between major shear zones, forming a "San Juan Block" and a "Southern Peninsula Block" of Hispaniola. The San Juan block appears to have arrived in juxtaposition with the Cordillera Central in the Early Miocene, whereas the Southern Peninsula block appears to have arrived in the Late Miocene (10 Ma). Transforms associated with the early Cayman Trough are responsible for these motions.

Faunal ages on collected samples:

Below is a listing of faunal ages and paleoenvironments determined by E. Robinson (pers. comm.) for various rock samples collected by the author in the southern Dominican Republic. Notes have been added where pertinent. The findings have augmented the general stratigraphic columns for the morpho-tectonic domains defined in Figures 4.3 and 4.4, and have helped to define the structural limits of the San Juan and Southern Peninsula Blocks. The sample numbers are keyed to localities drawn on Figure 5.1. The basemap for Figure 5.1 is after the Blesch [1966] geologic maps at a scale of 1:250,000. Many more samples were originally submitted, but these could not be dated because of a lack of faunal content or of diagenetic recrystallization.

Carbonates of Sierra Bahoruco

1. ?pelagic.

2. Middle Tertiary, pelagic.

3. Middle Tertiary, pelagic.

4. Middle Eocene shelf, or Oligo/Miocene with reworked Eocene material.

5. Middle to Upper Eocene shelf.

6. Middle Eocene shelf.

7. Middle to Upper Eocene, bioclastic, ?shelf.

8. Upper Eocene, bioclastic.

9. ?pelagic.

10. ?pelagic.

11. Upper Paleocene, shelf, or bioclastic layer.

12. Paleocene or Lower Eocene, sparite, ?shelf.

13. oolitic, ?shelf.

14. ?pelagic.

15. ?shelf.

16. Spicalite, ?bathyal.

17. Submarine breccia.

18. Lower Miocene, bioclastic in pelagic.

Carbonates of Sierra Neiba and Martin Garcia

19. ?Miocene/Pliocene, pelagic.

20. Middle Miocene to Pliocene.

21. Miocene/Pliocene, pelagic.

22. Upper Eocene, bathyal turbidite.

23. Paleocene or earliest Eocene. Note: this occurrence is interbedded with and baked by lavas. The age defines the age of volcanism in the central S. Neiba and may represent the age of the entire volcanic area on the Blesch map.

24. Lower Miocene, ?shelf, but turbiditic (may be reworked).

25. Upper Paleocene/Lower Eocene, pelagic.

26. Lower Eocene, pelagic.

27. Miocene/Pliocene?.

Comments:

Sierra Bahoruco appears to have been a shallow carbonate bank during the Paleocene to Eocene. Little is known of the Oligocene, and the rare occurrence of Oligocene sediments is bothersome and may suggest a fundamental problem in the understanding of Oligocene biostratigraphy in the northern Caribbean. No evidence of unconformity has been detected between the Eocene and Miocene sections, and potentially Oligocene sections are micritic, deep-water limestones. The Miocene began with predominantly deep-water deposition, but uplift, probably controlled by strike-slip faulting, was underway by Early Miocene time as indicated by bioclastic

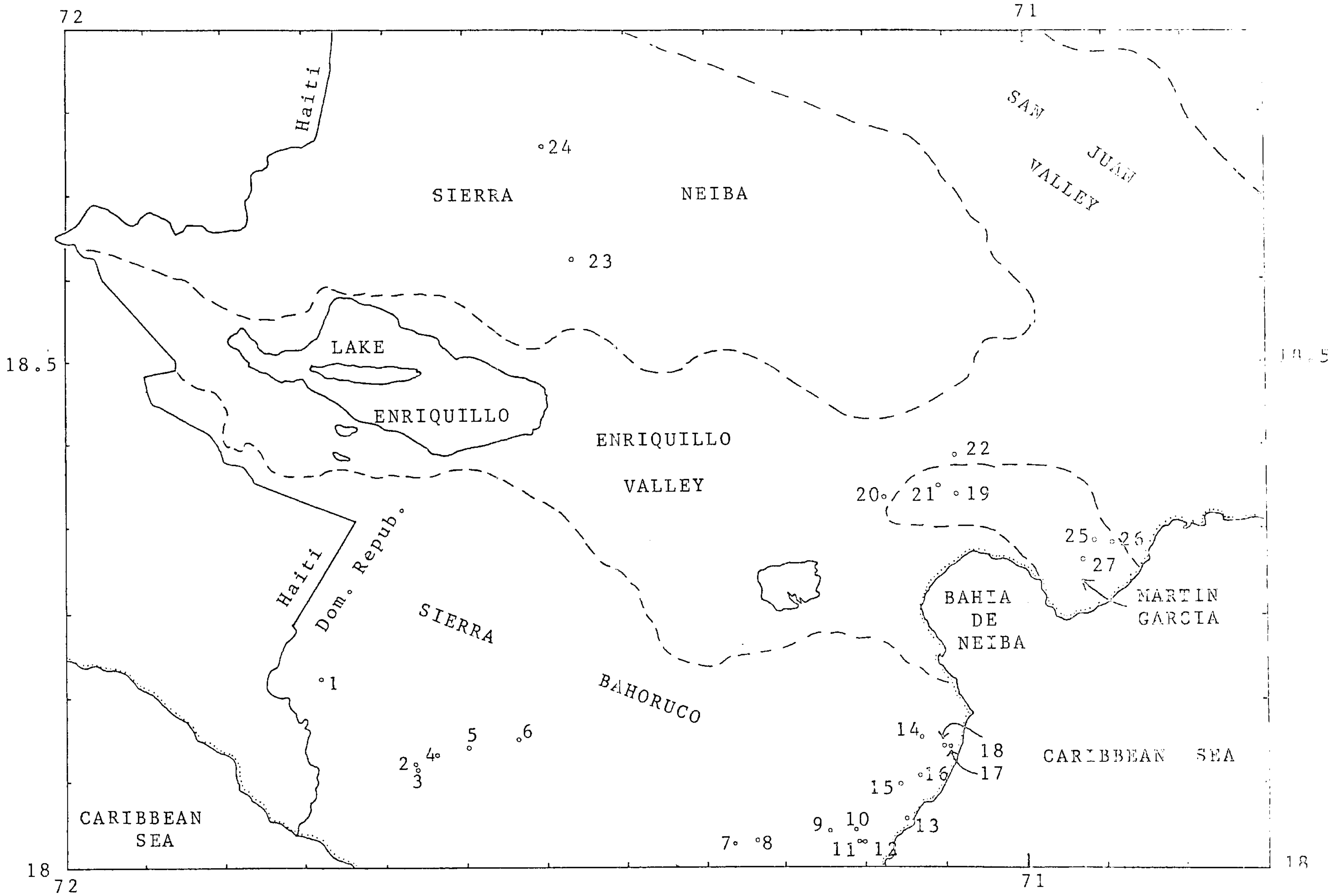


Figure 5.1. Sample localities in southern Hispaniola.

layers in sediments of that age. Middle Miocene or younger marine sediments are restricted to the coastal areas.

Volcanism apparently occurred locally in Sierra Neiba during the Paleocene/Eocene, but the cause of the volcanism is unknown. The petrology of these lavas is not necessarily indicative of island arc volcanism. The lavas may have come up along strike-slip faults. Deep-water sedimentation continued into the Early Miocene, but uplift began at this time and emergence occurred at least locally by the Late Miocene, forming the structurally distinct San Juan and Enriquillo Basins that we see today.

The Paleocene/Eocene section of Sierra Martin Garcia is pelagic like the Sierra Neiba, in contrast to the Paleocene/Eocene shelf facies of Sierra Bahoruco. Hence, Sierra Martin Garcia is viewed as being more closely related to Sierra Neiba than to Sierra Bahoruco. The structural boundary between the Sierra Neiba and Enriquillo Valley provinces, therefore, trends offshore into Bahia de Neiba at Barahona, rather than through the Fondo Negro Basin to the north of Martin Garcia.

Structural development of southern Hispaniola:

In the north, the Cayman Trough suggests a minimum of 1050 km of transcurrent motion; this is well-defined in the Trough itself, whose eastern end lies directly south of Sierra Maestra of eastern Cuba. However, continuations of composite splays to the east of the Trough must be inferred from geologic relationships in and around Hispaniola and Puerto Rico. Presently, the Oriente Fault between Cuba and Hispaniola is the active transform which connects the Mid-Cayman Spreading Center with the Lesser Antilles subduction Zone, although the Oriente splays into the Bahamas Channel, Camu, and Septentrional faults in northern Hispaniola. In chapter

4, it was shown that a total of 400 km of strike-slip motion has occurred sincel the Early Miocene along these composite faults. If 400 km of motion have occurred along the Oriente Fault system in the north, fault systems responsible for the remaining 650 km of the Cayman Trough's indicated offset must pass to the south of the arc complex of Hispaniola. Further, as the eastern end of Cayman Trough lies to the north of the Southern Peninsula of Hispaniola, this motion must pass through the island of Hispaniola itself.

Southwestern Hispaniola consists of three physiographic highs separated by two intervening lows which are Neogene sedimentary basins (Figure 4.3). The Late Miocene to Recent history of these terranes is predominantly that of north-south compression [Biju-Duval et al., 1983a; Bourgois et al., 1979a,b]. The northern block (Cordillera Central-Massif du Nord-Massif du Nord'ouest) has overthrust the San Juan-Plateau Central basin to the south. The middle physiographic high (Sierra Neiba-Chain des Matheaux-Ile de la Gonave) has overthrust the Cul de Sac-Enriquillo Basin. The relationship of the Cul de Sac-Enriquillo Basin to the southern high of Sierra Bahoruco and Massif de la Selle is poorly known. Little evidence exists in these areas for large-scale transcurrent offsets since the Middle Miocene, although minor strike-slip faults such as the Plantain Garden-Enriquillo fault [Mann et al., 1984] do occur. Hence, large-offset transcurrent motions (totalling 650 km) that splayed from the Cayman Trough through southwestern Hispaniola are probably resticted to the Oligocene through Middle Miocene interval (about 3 cm/yr, if motion was continuous).

If an Oligocene to Middle Miocene strike-slip phase preceded the Late Miocene to Recent southward vergent overthrusting phase, then the thrusting probably nucleated upon the pre-existing transcurrent fault zones. Thus, the southern hanging walls of the overthrusting complexes may preserve evidence of the former strike-slip phase, as well as evidence constraining the timing of thrust initiation. Conversely, the northern flanks of the intervening basins have been overthrust, thereby obscuring any record there of strike-slip deformation. Following this logic, the northern San Juan Boundary Fault and the northern boundary of Enriquillo-Cul de Sac Basin are the likely locations to find evidence for large-scale strike-slip offset.

Along the south flank of Cordillera Central and Massif du Nord, the Los Pozos-San Juan boundary fault zone separates two distinct terranes. To the north lie metamorphosed Cretaceous to Paleogene arc-related rocks and volcanogenic sands of the Cordillera Central and Massif du Nord, and to the south lie Paleogene deep-water micrites, silts and slates which possess little arc derived volcanic debris [Michael, 1979; Bowin, 1975]. Only in Sierra de Ocoa in the southeast of the San Juan Basin does one find the Eocene-Oligocene volcaniclastic and conglomeratic facies expected to have been deposited along the length of the arc complex during and following Late Paleocene-Eocene orogensis [Biju-Duval et al., 1983a; Bourgois et al., 1979b; Dolan, in prep., 1985]. Not until the Early Miocene did significant amounts of terrigenous sand enter the San Juan Basin [Michael, 1979]. This suggests that the basin had arrived in juxtaposition with the arc terrane by that time, although the basin probably continued to migrate along the length of the arc unitl the Middle Miocene. On sedimentologic grounds, the facies discrepancy across the Los Pozos-San Juan Boundary Fault Zone suggests that at least 350 km of transcurrent motion has brought the floor of San Juan Basin into its present position from the west, as measured from western Sierra Ocoa to the tip of the northwest peninsula of Haiti, which itself was a source of volcanogenic sediment in the Paleogene. Structural studies along the fault [Michael, 1979] acknowledge the existence of a strike-slip component of offset, but the magnitude is unknown. The Late Miocene(?) to Recent section along the northern San Juan Basin margin is relatively undeformed in comparison to the older sediments deposited before and during the strike-slip phase.

Along the southern margin of Sierra Neiba, evidence for large-scale strike-slip motion has not been reported. However, the maximum age for the initiation of thrusting is Upper Miocene (Lower Tortonian, about 11 Ma; Bourgois et al., [1979a]), as indicated by Eocene rocks thrust onto Tortonian rocks. To the south of Sierra Neiba, the Fondo Negro Basin, an exposed part of Enriquillo Basin, was filled predominantly during Late Miocene-Pliocene time, in accordance with southward-directed thrusting in Sierra Neiba [Cooper, 1983]. The difference between the minimum of 350 km offset along the Los Pozos-San Juan fault and the total of 650 km that passed through southern Hispaniola may have occurred along Sierra Neiba-Chaine des Matheaux-Ile de la Gonave, before it overthrust the Enriquillo-Cul de Sac Basin in Late Miocene time. Before this overthrusting, the Plateau Central-San Juan Basin and the Cul de Sac-Enriquillo Basin may have been one and the same.

Both the Los Pozos-San Juan Fault and the southward limit of thrusting in the Sierra Neiba-Matheaux-Ile de la Gonave thrust belt have logical bathymetric continuations into the northeast corner of Cayman Trough [Case and Holcombe, 1980]. These probable former transcurrent fault zones are now draped with Neogene sediment, attesting to their dormancy and/or evolution into thrust zones [E. Rosencrantz, pers.comm., 1983].

CHAPTER 6

OUTLINE OF PRIMARY CARIBBEAN GEOLOGICAL FEATURES

OUTLINE OF PRIMARY CARIBBEAN GEOLOGIC FEATURES:

Introduction:

The principle geographic features and the present plate boundary configurationn of the Caribbean region are shown in Figure 6.1. Put simply, the Caribbean Plate is migrating eastward with respect to North and South America at about 3-4 cm/yr [Sykes et al., 1982]. Subduction occurs at the leading (Lesser Antilles Arc) and trailing edge (Middle American Arc) of the plate, while generally strike-slip deformation and offsets occur along the northern and southern plate boundary zones.

This chapter defines a set of Caribbean plate-tectonic elements which must be integrated into models outlining the region's evolution. This is done by reviewing Caribbean geology and deducing primary magmatic arcs, ultramafic-mafic complexes representing uplifted oceanic crust, arc-continent collision zones, the paleogeographic significance of the Yucatan Basin, Grenada Basin and the Cayman Trough, and the concept and development of northern and southern Caribbean plate boundary zones. These features are shown in Figure 6.2.

Magmatic Arcs and Uplifted Occurrences of Oceanic Crust: Keys to the Timing and Polarity of Subduction in the Caribbean:

General significance of arcs:

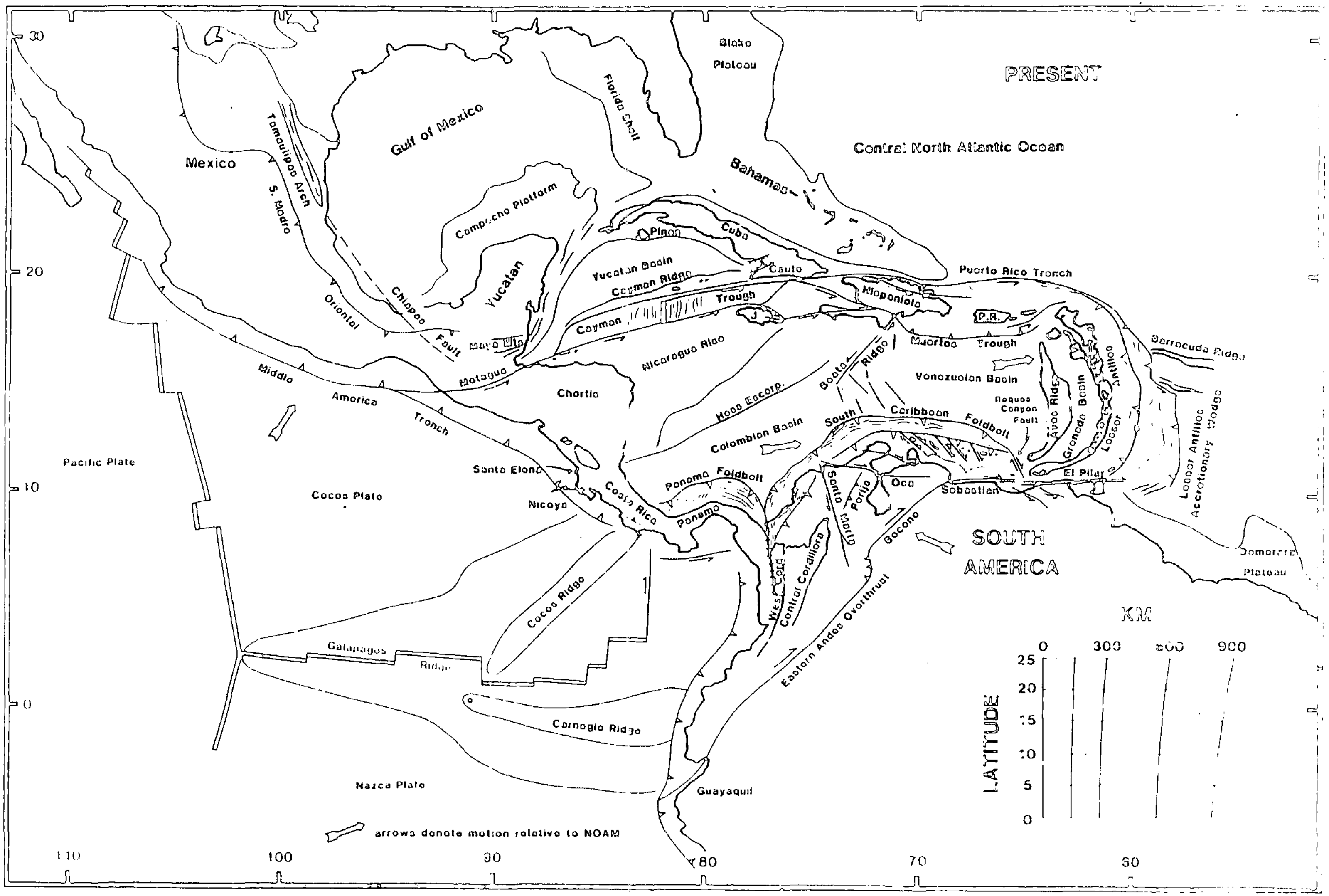


Figure 6.1. Geography and present plate boundaries of the Caribbean region. Heavy arrows indicate directions of plate motions relative to North America. J, Jamaica; PR, Puerto Rico. Adopted from Case and Holcombe [1980].

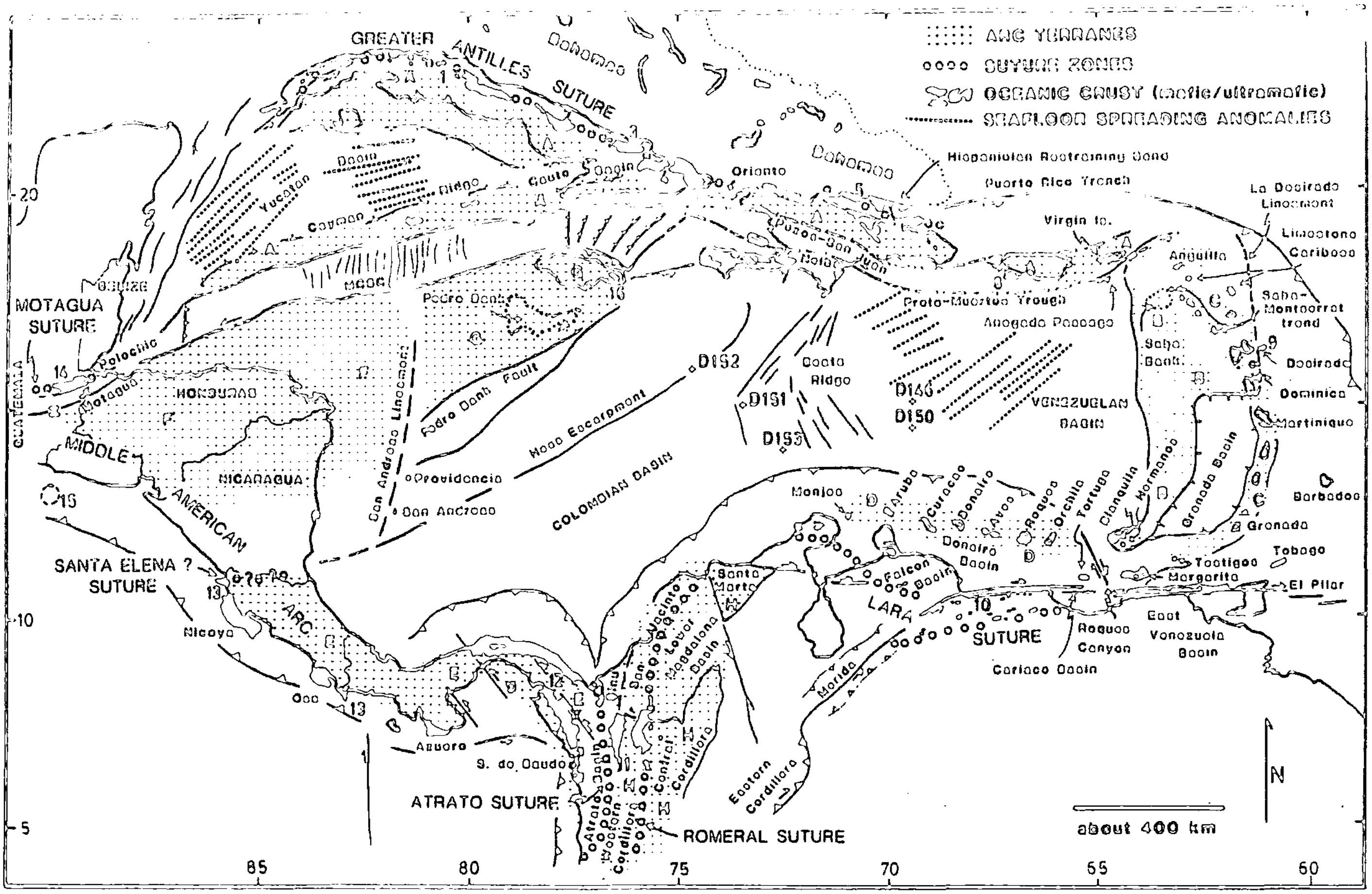


Figure 6.2. General tectonic map of the Caribbean showing features described in text. Heavy letters A through H refer to the following arcs: A, Greater Antilles; B, Aves Ridge; C, Lesser Antilles; D, Leeward Antilles; E, Costa Rica-Panama; F, Chortis; G, Nicaragua Rise-Jamaica; H, Central and Western Cordillera. Numbers refer to occurrences of oceanic crust and are keyed to and described in Table 6.1. D146 and D150-D153 are DSDP holes (Leg XV) that reached medial Cretaceous basalts of the B" seismic horizon. Traces of transform systems through Hispaniola are shown: Los Pozos-San Juan-North Puerto Rico was active Late Eocene or Early Oligocene through Early Miocene; Neiba-Proto Muertos Trough-Anegada was active primarily during Miocene; Oriente-North Hispaniolan-Puerto Rican Trench active since Middle Miocene. Lara Suture has been dextrally offset since the Late Miocene by transcurrent motions along the Merida Andes. NW trending sinistral faults across Panama have allowed escape of blocks from the Panama-Western Cordillera collision zone (Atrato Suture). Significance of trends of magnetic anomalies in Yucatan [Hall and Yeung, 1980] and Grenada Basins (Westbrook, pers. comm) shown in Figures 6.4 and 6.5, respectively. MCSC, Mid-Cayman Spreading Center, with flanking structural trends. Oceanic portion of Cayman Trough bounded by first normal fault symbol at each end.

Subduction-related igneous rocks in the Caribbean (generally Jurassic to Recent intermediate magmatic rocks) can be grouped into eight primary magmatic arcs (Figure 6.2). The eight arcs have different histories and can be defined by comparing the record of subduction-related magmatism, volcanogenic sedimentation, and the evidence for arc polarity at various times for various Caribbean geomorphologic provinces. Some of these arc terranes have been dissected by transcurrent plate motions during and following their formation and, therefore, no longer possess their original arc geometry. The terms "Andean" and "intra-oceanic" are used here to refer to subduction-related magmatic arcs which have developed upon pre-existing continental and oceanic crust, respectively.

Some of the intra-oceanic arcs, such as the Greater and the Leeward Antilles arcs, contain mafic metamorphic rocks that are older than all plutonic rocks of the respective arcs. It is unclear whether these rocks are ophiolitic basement or whether they represent primitive arc material. This is a critical question when attempted to determine the period of active subduction at each arc, becasue the crystallization age of ophiolitic basement has little to do with the period of active subduction. In this study, only those magmas and volcanics that are truly intermediate or calc-alkaline are assumed to be indicative of subduction, and the older metamorphic mafics are regarded as basement. Concerning the polarity of arcs, the direction from the arc plutons to the forearc and accretionary prism is referred to as the "facing" direction (e.g. an east-facing arc overlies a west-dipping Benioff Zone).

The subduction-related igneous activity (plutons, lavas, tuffs) of seven of the eight arcs through time is summarized in Figure 6.3; the eighth arc of northwest South America has been operative since the Jurassic, but the volcanic axis has shifted intermittently from area to area (see later

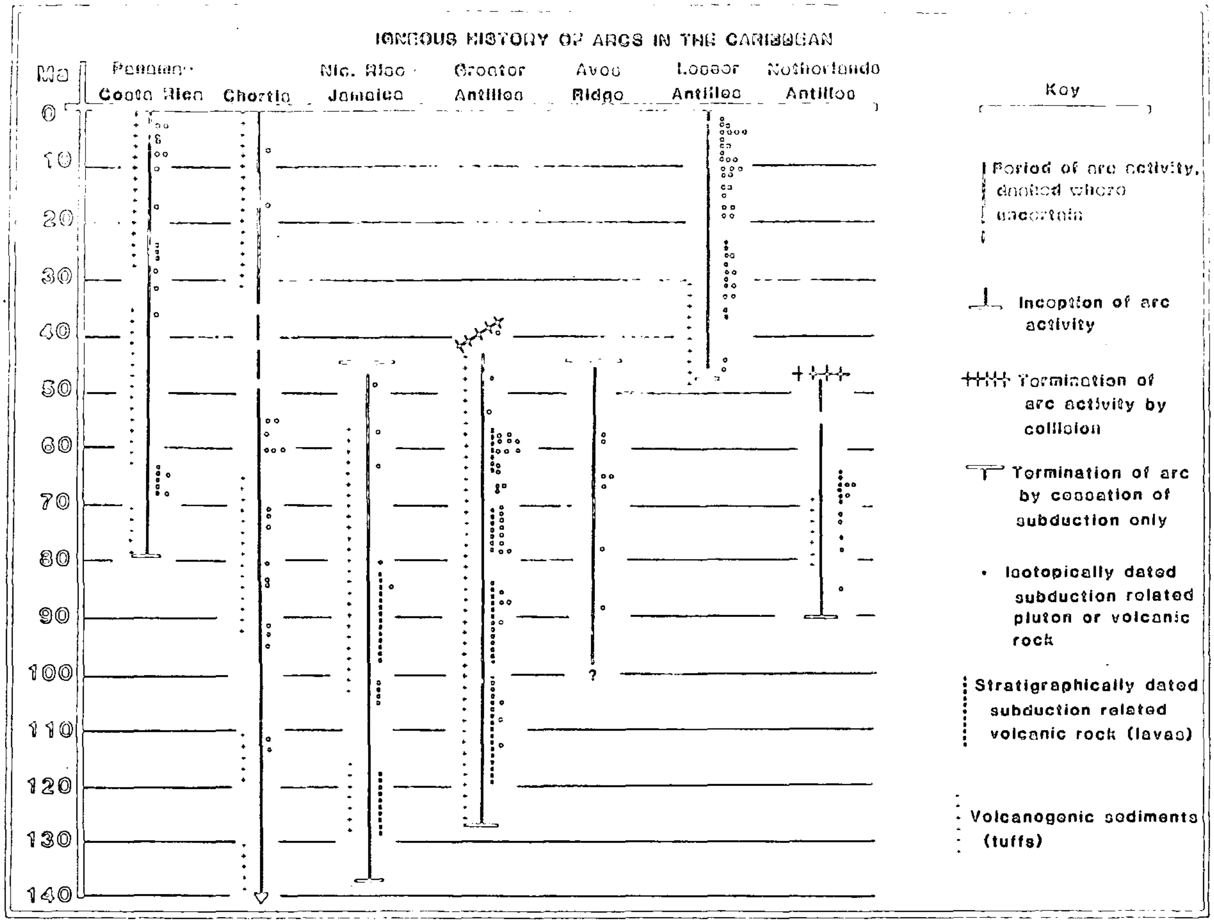


Figure 6.3. Time-space compilation chart of subduction-related igneous activity (tuffs, intermediate plutons, and lava flows) for each arc (designated in Figure 6.2) within the Caribbean region. The inclined collision symbol for the Greater Antilles indicates slightly oblique collision. Sources used, by arc, as follows. Greater Antilles: Bowin [1975]; Cheilletz et al. [1978]; Cox et al. [1977]; Feigenson [1978]; Kesler [1971]; Kesler and Sutter [1979)]; Kesler et al. [1977]; Khudoley and Meyerhoff [1971]; Laverov [1967]; Meyerhoff et al. [1969]; Pardo [1975]. Aves Ridge and Saba Bank: Andrieff et al. [1979]; Bouysse et al. [1981]; Fox and Heezen [1975]; Fox et al. [1971]; Nemec [1980]; Santamaria and Schubert [1974]. Lesser Antilles: Briden et al. [1979]; Nagle et al. [1976]; Tomblin [1975]. Panama-Costa Rica: Bandy and Casey [1973]; Bellon and Tournon [1978]; Escalante [1966]; Fisher and Pessagno [1965]; Guidice [1979]; Henningsen [1965]; Restrepo and Touissant [1977]; Terry [1956]; Weyl [1980]. Venezuelan Antilles: Maresch [1974]; Santamaria and Schubert [1974]. Chortis: Clemons and Long [1971]; Horne et al. [1974; 1976]; Levy [1970]; Mills et al. [1967]; Mills and Hugh [1974]; Paz Rivera [1964]; Piniero and Romero [1962]; Weyl [1980]; Wilson [1974]; Zoppis Bracci [1960]. Jamaica-Nicaragua Rise: Arden [1975]; Kashfi [1983]; Lewis et al. [1973]; Meyerhoff and Kreig [1977].

section) and cannot be shown as a single arc in Figure 6.3. Most radiometric dates on plutons are K-Ar dates, which generally indicate uplift rather than emplacement; nevertheless, K-Ar dates on plutons are indicative of arc activity. Furthermore, general correlation exists between stratigraphic and radiometric ages for arcs where data from both methods are available. The period over which each arc in Figure 6.3 was volcanically or magmatically active correlates approximately with the period of active subduction. When deriving paleogeographic reconstructions, however, it should be noted that subduction may precede initial magmatism, and volcanogenic sedimentation, by perhaps 2 to 8 million years, depending on the subduction rate.

General significance of occurrences of oceanic crust:

Ultramafic-mafic complexes are another primary source of information for plate-tectonic syntheses; they often represent ophiolites that are preserved pieces of oceanic crust which may indicate the former existence of now vanished oceans [Burke et al., 1977]. Ultramafic-mafic complexes that are representative of oceanic crust occur throughout the Caribbean [Burke et al., 1984; Case, 1980; Wadge et al., 1982].

Oceanic rocks of the ophiolite suite usually are formed below sea level and are brought to the surface primarily in compressive regimes. They occur on land generally in three tectonic settings; 1) in uplifted accretionary complexes as clipped off seamounts or fragments of the former downgoing plate, 2) as diapirs or sheets in flower structures within strike-slip zones where oceanic crust forms the basement, and 3) in belts separating pairs of arcs or continents, occurring as slabs, blocks and slivers obducted onto the subducting plate during collisions. The obducted complexes may be bits of accretionary prisms which had developed prior to collision, or may be preserved pieces of ophiolitic forearcs of overriding plates, or both. All three tectonic settings occur in the Caribbean region. For the purposes at hand, the important aspects of occurrences of ocean crust around the Caribbean (Figure 6.2) are the age of formation (ophiolite generation), and the timing of emplacement into their present tectonic settings.

The age of formation (magmatic crystallization) of ophiolites can be determined by radiometric dating, and by faunal dating of the earliest pelagic sediments of the ophiolite suite. In accretionary complexes, the formation age of ultramafic-mafic rocks indicates the age or age-range of the crust which has been subducted at that margin. Uplifted rocks within compressional strike-slip systems may include ultramafics whose age may be indicative of ophiolitic basement. Finally, formation ages of ultramafic-mafics in collision zones may relate to either of two original tectonic settings; either indicating the age or age-range of pieces clipped from the subducting plate prior to collision, or indicating the age of the forearc of the overriding plate.

The emplacement age of mafic and ultra-mafic complexes can define the time of development of one or more of the three tectonic settings where ultramafics are found. In all three settings, emplacement age, which is always younger than formation age, can be constrained between the youngest age of sediments overthrust by the ultramafic body (maximum age of emplacement), and the age of the first post-orogenic sediments to onlap the ultramafic body unconformably (minimum age of emplacement). Care must be taken that the development of forearc successor basins during subduction are not interpreted as post-orogenic sedimentation.

These simple relationships, however, often become more complex. For example, development of accretionary complexes is time-progressive and often long-lived relative to collision or formation of flower structures. Hence, a range of emplacement ages may arise during accretion. In collision zones where ophiolite-bearing accretionary prisms had formed, two ages of emplacement may be recognized: one during the subduction stage, and a second during the collision. In northern Cuba, for example, commonly quoted emplacement ages of Late Cretaceous and Eocene probably represent subduction accretion and collision, respectively. Finally, if an accretionary terrane which has been obducted during collision is then subjected to transform motion, such as along the north coast of Hispaniola, yet a third period of structural emplacement may be recognizable.

The various structural setting(s) and potential complexities of belts containing bits of oceanic crust must be established in order to determine the true significance of the ophiolite in question. Figure 6.2 identifies mafic-ultramafic occurrences around the Caribbean. For each, Table 6.1 outlines the age of formation and the age(s) of emplacement, and the probable tectonic setting that it represents (tectonic significance), and the mode(s) of emplacement.

Arc systems of the Caribbean:

Greater Antilles Arc: This arc consists of central and eastern Cuba [Pardo, 1975], the Cayman Ridge [Perfit and Heezen, 1978], Cordilleras Central [Bowin, 1966; 1975] and Oriental [Hernandez, 1980] of Hispaniola, Puerto Rico [Donnelly and Rogers, 1980] and the Virgin Islands [Donnelly, 1966; Donnelly and Rogers, 1980]. Intra-arc spreading in Yucatan Basin separated Cuba from Cayman Ridge [Gealey, 1980], and motion upon splays of the Cayman Transform System have dissected the once continuous arc into its present configuration [Pindell and Dewey, 1982]. Basement in Sierra Guaniguanico of western Cuba, west of the Pinar Fault, appears to be

TABLE 6.1: OCCURRENCES OF SEAFLOOR CRUST IN THE CARIBBEAN

Name/location (keyed to Figure 6.2), and tectonic significance	Formation age	Age and mode of emplacement	Selected references
1. Central Cuba. A or C	Pre-Albian	Paleocene to Mid-Eocene. E	Wadge et al. 1984; Gealey 1980; Mossakovskiy and Albear 1978; Soto 1978.
2. Las Villas, Cameguey, Cuba. A or C	Pre-Aptian	Early or Mid-Eocene. E	Wadge et al. 1984; Thayer and Guild 1947; Mossakovskiy and Albear 1978; Pardo 1975; Flint et al. 1948; Meyerhoff and Hatten 1968.
3. Holguin, Cuba. A or C	Pre-Aptian	Mid-Eocene. E	Kozary 1968; Knipper and Cabrera 1974; Wadge et al. 1984; Brezsnyansky and Korpas 1973.
4. Nipe-Purial, Cuba. A or C	unknown	Maestrichtian-Paleocene. E?	Boiteau and Michard 1976; Cobiella et al. 1977; Cobiella 1978; Lweis and Straczek 1955; Wadge et al. 1984; Boiteau et al. 1972.
5. Northern Hispaniola. A or C	Pre-Aptian	Paleocene-Mid Eocene, E. Post-Mid Miocene, F.	This study; Bourgois et al. 1980; Bowin and Nagle 1980; Eberle et al. 1980; Wadge et al. 1984; Nagle 1966.
6. Central Hispaniola. B or C	Early Cretaceous	Albian?, D. Oligo-Miocene, F	Bowin 1960; Palmer 1963; Haldemann et al. 1980; Draper and Lewis 1983; Theyer 1983.
7. Southern Hispaniola. B and C	Turonian or Campanian	Pre-Paleocene, D, and Neogene, F	Maurrasse et al. 1979; Mercier de Lepiney et al. 1979; Pindell et al. 1981.
8. Bermeja, Puerto Rico. C	Tithonian	Pre-Turonian, D.	Mattson, 1973; Mattson and Pessagno 1979.
9. La Desirade, C, see note 1.	Late Jurassic-Early Cretaceous	Pre-Eocene, D	Mattinson et al. 1980; 1973; Briden et al. 1979; Fink 1970.
10. Villa de Cura A	Hauterivian-Albian	Paleocene-Miocene, E	Gealey 1980; Maresch 1974; Stephan et al. 1980; Beck 1977; Santamaria and Schubert 1974.
11. Western Cord., Colombia. C see note 2.	Pre-Cenomanian	Campanian to Maestrichtian, D	Barrero 1979; Shagam 1975; Irving 1975; Henderson 1979; Bourgois et al. 1982; Moody 1980.
12. Eastern Panama, B	Campanian	Neogene E	Case et al. 1971; Bandy and Casey 1973; Bandy 1970; Bourgois et al. 1982.
13. Nicoya, Santa Elena. C	Late Jurassic-Late Cretaceous	Campanian to Paleogene, D	Bourgois et al. 1982; Azema and Tournon 1980; de Boer 1979; Galli-Olivier 1979; Lundberg 1983; Gursky et al. 1983; Schmidt-Effing, 1979.
14. Santa Cruz, Guatemala. A	Late Valanginian-Early Cenomanian	Campanian-Maestrichtian, E. Cenozoic, F.	Williams 1975; Rosenfeld 1981; Bertrand et al. 1978.
15. Pacific margin, DSDP Leg 84. C	Pre-Cenomanian	Paleocene. D	Auboin et al. 1982.
16. Blue Mountains, Jamaica. C	Campanian?	Maestrichtian to Paleocene?. D	Wadge et al. 1984; Wadge et al. 1982.

A Basement of forearc of associated intra-oceanic arc.
B Basement of arc or local terrane.
C In accretionary terrane, clipped from downgoing plate at associated trench.
D Accreted to subduction complex by obduction.
E Arc/continent collision and forearc/accretionary prism obduction.
F Flower structure or diapir in compressional strike-slip domain.
Note 1. Possibly basement of Lesser Antilles, initially accreted to Aves Ridge Arc.
Note 2. Cenozoic arc has developed upon the ophiolitic basement subsequent to its obduction/accretion.

continental and is probably the (par)autochthonous mass with which the intra-oceanic arc collided in the Paleocene-Eocene (see chapters 3 and 8).

In Cuba and Hispaniola, subduction-related magmatism began in the Early Cretaceous, increased during the Late Cretaceous, and ceased in the Middle Eocene. In Puerto Rico and the Virgin Islands, magmatism started in the latest Early Cretaceous or early Late Cretaceous and persisted into the Late Eocene (Puerto Rico) and Oligocene (Virgin Islands). However, the young magmatism in the Virgin Islands may relate to subduction beneath the younger Lesser Antilles arc (Figures 6.2, 6.3), rather than to subduction beneath the Greater Antilles arc. The foundation on which the Greater Antilles arc was built seems to have been oceanic crust; the metamorphosed mafic Duarte Formation of Hispaniola [Bowin, 1975] and the Domingo Belt of Cuba [Pardo, 1975] are probable examples of such basement.

The early polarity of the arc is unknown; the ultra-mafic bearing Bermeja Complex of Puerto Rico may be oceanic crust accreted to the south side prior to the Campanian [Mattson, 1973; Mattson and Pessagno, 1979], suggesting initially south-facing polarity. After the Santonian or Campanian, the arc was definitely north-facing, as indicated by the Late Cretaceous to Eocene ultramafic-bearing subduction complexes along the north side of the arc [Case et al., 1984], which overthrust the early Paleogene shelf carbonates of the Bahamas [Pardo, 1975; Wadge et al., 1982].

The Aves Ridge Arc: Included with the Aves Ridge are Saba Bank to the north and, probably, La Blanquilla and Los Hermanos to the south (Figure 6.2). Dredging of this largely submarine ridge has produced granodioritic rocks indicative of arc magmatism [Fox and Heezen, 1975]. Gravimetric and seismic studies [Kearey, 1974] suggest that the crustal structure of Aves ridge is similar to that of an island arc.

Radiometric age determinations on the dredged samples (Figure 6.3) yield Late Cretaceous through Early Eocene dates; Eocene to Recent shallow- and deep-water limestones form the bulk of the known sediment. Nothing is known of a pre-Late Cretaceous arc-related history; it appears as though subduction occurred only during the Late Cretaceous to Paleocene or Eocene. The slight eastward convexity of the ridge's geomorphology, and of the magnetic expression of plutons of Aves Ridge [Speed and Westbrook, et al., 1984], are the only indicators of arc polarity; it was probably east-facing, forming an arcuate continuation of the Greater Antilles arc during Late Cretaceous-Eocene time. No crustal break between the ridge and the Venezuelan Basin to the west is indicated. The Aves Ridge is probably a continuation of the crust of the Venezuelan Basin that has been intruded with plutons.

The Lesser Antilles Arc: This arc comprises Anguilla to Grenada (including Saba Island and the Limestone Caribbees — St. Eustace, etc.). The ridge upon which the arc is built extends to Los Testigos and Margarita, but those islands are far removed from the belt of modern Lesser Antillean volcanism.

The beginning of magmatism in the present Lesser Antilles arc is difficult to date. At least in the north, the arc has been built upon older volcanic rocks which may have been associated with the pre-Eocene Aves Ridge arc. Early Paleogene intra-arc spreading have have dissected the Aves Ridge arc complex and created Grenada Basin; the Lesser Antilles frontal arc thus may be underlain in places by older arc-related rocks. Subduction-related arc magmatism has been continuous in the northern Lesser Antilles Arc since the Eocene, whereas in the south it may have started only in the Oligocene [Briden et al., 1979]. However, older plutons in the south may be buried.

The northern third of the arc is broader than the rest; east-west? crustal extension during the Paleogene widened the Aves Ridge foundation in the north [Nemec, 1980], whereas backarc spreading in Grenada Basin (discussed later) has left a well-defined trough of presumably oceanic basement in the south. Also, the axis of magmatism moved some 50 km to the west (Saba Island-Montserrat island trend) in the latest Miocene [Briden et al., 1979]. The polarity of the Lesser Antilles arc is and always has been east-facing; Barbados is a subaerially exposed portion of the Lesser Antilles accretionary ridge, which has been forming at least since the Eocene. Basic rocks of La Desirade appear to be a piece of oceanic crust with isotopic ages of 149 to 85 Ma and Late Jurassic fauna [Mattinson et al., 1980; Briden et al., 1979; Bouysse et al., 1983], either accreted to the Lesser Antilles or uplifted from the forearc. Hence, these rocks are not included in Figure 6.3.

The Leeward Antilles Arc: The islands of Los Monjes, Aruba, Curacao, Bonaire, Aves, Los Roques and Orchila are high-standing parts of an east-west offshore ridge that is cross-cut by northwest trending extensional grabens [Silver et al., 1975; Case, 1974a; Biju-Duval et al., 1983b]. Pre-Middle Eocene rocks consist of metamorphosed mafic and intermediate igneous rocks suggestive of an island arc [Maresch, 1974; Beets et al., 1984].

Radiometric age determinations on intermediate-composition plutons range from 85 to 64 Ma (Figure 6.3), but the plutons apparently intrude a basement of Early Cretaceous mafic rocks with K-Ar dates of 135 to 115 Ma [Gonzalez et al., 1981; Santamaria and Schubert, 1974]. Both igneous series have been related to island arc volcanism [Maresch, 1974], but the age and petrological differences between the plutons and the older mafic rocks suggest that a Late Cretaceous arc developed upon pre-existing Early Cretaceous oceanic crust [Pindell and Dewey, 1982]. This suggestion is further supported by sedimentological studies; the oldest known deposits are deep-water pelagics of Albian age from Curacao and Bonaire [Beets, 1975], and volcanogenic sediments were not deposited until the Late Cretaceous.

The origin and polarity of this arc are not clear; an important consideration is whether or not the Villa de Cura ultramafic- and blueschist-bearing klippe of the Lara Nappes [Stephan et al., 1980] in northern Venezuela is part of the forearc of the Leeward Antilles arc, as several authors have proposed [Maresch, 1974; Gealey, 1980; White and Burke, 1980]. The Lara Nappes were emplaced into a Paleogene flysch trough [Maresch, 1974]. This emplacement age immediately follows the magmatic termination and uplift of the arc; therefore, this history fits temporally and geometrically with the scenario of a south-facing arc colliding with continental crust of South America.

However, this is not the only possibility. Paleogeographic reconstructions retracting post-Middle Miocene fault-zone offsets within northern South America (see Chapter 7) restore the offshore islands to the west of, and nearly along strike of, the Lara Nappes, prior to the Late Miocene. This brings into question the apparent arc-forearc relationship, and also whether the islands were adjacent to northern South America as early as the Eocene. One alternative possibility is that the blocks comprising the islands were progressively clipped off the southern end of the Aves Ridge magmatic arc and concomitantly accreted to the Venezuelan margin during the Eocene to Miocene, as the Caribbean Plate migrated along northern South America by more than 1000 km (see section on Cayman Trough). If so, the supposed Paleogene compression that caused obduction of the Lara Nappes need not have been of sufficient magnitude to have produced the subduction-related magmatism of the offshore islands. In this case, the polarity of the arc is that of the Aves Ridge, or east-facing (prior to obduction and probable block rotation).

A second alternative, which is preferred here to the others and discussed further in the section on collision zones, is that both the Leeward Antilles islands (arc) and the Lara Nappes (forearc) were progressively obducted onto the Venezuelan margin throughout the Eocene-Miocene interval, as a result of the progressive eastward migration of the Caribbean Plate. This emplacement led to the reconstructed pre-Late Miocene configuration of arc, forearc and continent presented in chapter 7). In this case also, arc polarity is east- or southeast-facing prior to obduction and rotation, and the arc/forearc complex is derived from crust of the Caribbean Plate, rather than of the Atlantic Ocean.

The Costa Rica-Panama Arc: This intra-oceanic arc includes the Central American Isthmus from northern Costa Rica (Santa Elena Peninsula) to the western border of the Western Cordillera of Colombia, and forms the trailing edge of the Caribbean Plate. The arc was built on Upper Jurassic(?) to Lower Cretaceous oceanic crust since the Late Cretaceous; tuffs, shallow-water limestones, and volcanogenic sandstones in Costa Rica suggest initiation of subduction and associated uplift sometime during the Campanian [Lundberg, 1983; Galli-Olivier, 1979; Schmidt-Effing, 1979]. The arc's oldest subduction-related plutons, in the Azuero Peninsula, have yielded radiometric ages of 69 Ma [Kesler et al., 1977]. Paleogene, shallow-water sediments are commmon from central Panama to southeast Costa Rica, indicating a well-developed arc foundation by that time, but in eastern Panama deep-water sediments were deposited upon oceanic crust into the Neogene [Bandy and Casey, 1973]. Accordingly, the eastern end of the arc is less well developed than the rest.

To the northwest of Azuero Peninsula, the arc has been west-facing since its inception; forearc development on the western margin (Nicoya Complex) occurred in the Late Cretaceous [Lundberg, 1983; Galli-Olivier, 1979]. Since Miocene time, however, the Panamanian portion of the arc has reversed its polarity from south-facing to north-facing; the arc's present northward convexity is the result of sinistral displacements on several throughgoing northwest trending faults crossing Panama [Case and Holcombe, 1980]. Seismic sections across the Panmanian foldbelt show that the arc is overriding the southwestern Colombian Basin [Lu and McMillen, 1983]. This deformation presumably is due to east-west compression from collision between eastern Panama and Colombia.

At present, and probably since the Eocene, the Costa Rica-Panama magmatic arc is aligned with the subduction-related magmatic arc of the Chortis block. These two arcs together comprise the Middle American arc system.

The Chortis Arc of Nuclear Central America: This "Andean" arc was built upon pre-Mesozoic continental crust and includes southern Guatemala (south of Motagua Fault Zone), Honduras, part of northern Nicaragua, and extends out the Nicaragua Rise an uncertain distance, probably to about the San Andreas Trough (Figure 6.2; Christofferson, [1983]; Case and Holcombe, [1980]). The Chortis arc pre-dates, and is geographically distinct from, the Eocene to Recent Middle American arc which extends through western Chortis and Costa Rica to Panama.

Radiometrically determined ages and volcanogenic sedimentation suggest that subduction-related magmatism had begun by the Late Jurassic (Figure 6.3). Most plutons yield Cretaceous and Paleocene ages, and arc magmatism apparently ceased in all but the western portion of the arc in the Late Paleocene. The Mesozoic and Paleocene plutons of the Chortis arc form a poorly defined but generally east-west trending ovoid with no apparent age progression.

The Cenozoic Middle American magmatic arc is superposed upon the western part of the Chortis arc, and it appears as though the Middle American arc developed after a southward shift in volcanism from the Chortis Arc during the Early Eocene. The Costa Rica-Panama Arc has formed the southern half of the Central American Arc with no apparent major post-Eocene structural offset. Therefore, the Chortis and Costa Rica-Panama arcs apparently merged tectonically during the Paleogene. The boundary between the two arcs may be through the ultramafics of the Santa Elena Peninsula (strike-slip suture?), which aligns to the east with the Hess Escarpment, a long, linear zone of crustal dislocation.

Polarity of the Chortis Arc is uncertain. One problem is the amount of post-Eocene offset assumed across the Motagua-Cayman transform fault system (estimates range from tens to 1400 km); polarity indicators such as ophiolitic forearcs and accretionary wedges may be allochthonous or missing. Suggested offsets of 200 to 850 km (e.g., White and Burke, [1980]) would juxtapose portions of the Chortis arc with possible forearc remnants that were thrust onto the southern Yucatan shelf of Guatemala (Santa Cruz ophiolite and other ultra-mafic bodies [Rosenfeld, 1980; Case and Holcombe, 1980]. Such a relationship suggests north-facing polarity for the Chortis Arc (south-dipping subduction beneath Chortis). Greater and, in our opinion, more realistic estimates of Cayman offset since the Eocene (1100 km, Wadge and Burke, [1983]) place the arc west of the Santa Cruz and related obducted ophiolites and into the realm of the Mexican Cordillera. In this case, it is likely that the arc was always southwest-facing (e.g., Pindell and Dewey [1982]; Wadge and Burke [1983]; Duncan and Hargraves [1984]).

Nicaraguan Rise-Jamaica Arc: This arc includes Jamaica and the shallow banks and intervening deeps of the eastern Nicaraguan Rise, and is bounded on the north by the Cayman Trough. A morphologic boundary to the south is poorly defined (Pedro Bank Fault?), as bathymetry grades erratically towards the Hess Escarpment along the Colombian Basin. The arc presently is aligned with the Chortis arc, but earlier relationships are unknown. The arc is probably intra-oceanic; in the Nicaraguan Rise, seismic velocities are typical of oceanic crust, and pre-Cretaceous continental crust is apparently absent there and in Jamaica [Arden, 1975; Perfit and Heezen, 1978].

Hence, a boundary may exist between this intra-oceanic arc and the Chortis arc (founded on continental crust). This boundary probably occurs at the faulted area bounding the edge of the Nicaraguan Shelf (about latitude 81.5 West, including the San Andreas Trough; Christofferson [1983]; Case and Holcombe [1980]). The boundary may be tectonic (a zone of plate juxtaposition), or may represent only a difference in basement of a single arc: continental in the west, oceanic in the east.

Basement in this arc is largely obscured, but volcanogenic sedimentation in Jamaica indicates that arc activity was underway by the Early Cretaceous. On the Nicaraguan Rise east of 81.5 W, the oldest arc-related igneous rocks (recovered from drilling) are Paleocene or latest Cretaceous [Arden, 1975]. Radiometric dating and stratigraphic studies indicate that magmatism continued locally (e.g., Wagwater Trough, Jamaica) until the Eocene [Kashfi, 1983; Chubb and Burke, 1963; Meyerhoff and Kreig, 1977], but this magmatism may be due to rifting [Mann and Burke, 1984a] rather than to subduction. The Nicaraguan Rise has no known pre-Paleocene volcanic or sedimentary history, but the broad development of the Rise and its apparent continuity between Jamaica and Chortis suggest that it too comprises arc-related rocks which probably date back to the Early Cretaceous.

The polarity of the Nicaraguan Rise-Jamaica arc is uncertain; no obvious forearc or accretionary prism exists, although this may lie offshore south of Pedro Bank. Perfit and Heezen [1978] viewed the Cayman Trough as a former trench, but the Cayman Trough is interpreted here as a very long post-Eocene pull-apart basin which has nothing to do with indications of arc polarity. Blueschists of the Blue Mountain Inlier of eastern Jamaica formed at a former trench environment, and these lie to the south of, but adjacent to, the Jamaican arc volcanics. This relationship may be taken to suggest south-facing polarity for Jamaica [Draper, in press, 1985; Burke et al., 1978]. However, the faulting which led to the juxtaposition of the blueschists and the arc volcanics is probably pre-Eocene, as the Tertiary Yellow Limestone is far less deformed than the older rocks. This suggests the possibility of pre-Eocene tectonic transport, possibly on a large scale, of fault-bounded blocks and terranes, which could have obscured the original relationships of tectonic environments to the point where they may now be misleading.

If a forearc or prism lies to the south of the island it must be located south of Pedro Bank as well, where arc plutons have been drilled [Meyerhoff and Kreig, 1977]. If the arc was north-facing, its forearc and prism, which are absent, must have been removed by strike slip faulting associated with opening of the Cayman Trough. Estimates of offset along the Cayman Trough of 1000 to 1300 km [Wadge and Burke, 1983; Pindell and Dewey, 1982] would align the Nicaraguan Rise-Jamaica Arc directly with the Santa Cruz ophiolite and related ultra-mafics of central Guatemala. Several of these ultra-mafic bodies have been dissected and structurally reactivated by Cenozoic shear along the Polochic and Motagua fault systems, but the Santa Cruz has remained relatively intact and was thrust northwards onto the southern Yucatan continental shelf in the Late Campanian-Maestichtian [Rosenfeld, 1980; 1981; Williams, 1975]. Hence, Pindell and Dewey (1982) considered these remnants of ophiolites as the original forearc of the Nicaraguan Rise-Jamaica Arc, rather than of the Chortis Arc.

In this work, however, it is acknowledged that sinistral shear probably was important for the Maestrichtian-Eocene interval as well as for the post-Eocene, and that the total offset may be even greater than 1300 km since the Maestrichtian collision. Thus, the Nicaragua Rise and Jamaica, as well as Chortis, probably lay to the west of the Motaguan collision zone, placing the Chortis and Nicaragua Rise-Jamaica Arcs along the Mexican Cordillera. In this case, both arcs were probably southwest facing through time. The arc that collided with the southern Yucatan shelf, then, is probably western Cuba or an extension of western Cuba.

Arcs along northwest South America: Arc activity indicative of subduction has occurred since at least Jurassic time within the Cordillera Central of Colombia and the main Cordillera of Ecuador (Figure 6.3), whose basements consist of a variety of pre-Mesozoic continental rock types [Shagam, 1975]. Mesozoic and Cenozoic arc plutons are also located in the Santa Marta block. The rocks of Santa Marta are very similar to those of Cordillera Central; basement may be continuous beneath the Lower Magdelena Basin, but strike-slip faulting during the ?Late Cretaceous or Paleogene may have offset the two. The existence of subduction over this long time-interval provides an important plate-tectonic constraint on evolutionary models for the Caribbean.

Arc plutonism and volcanism also has occurred within the Western Cordillera of Colombia since the Late Cretaceous [Henderson, 1979], but mainly during the Middle Cenozoic [Irving, 1975; Restrepo and Touissaint, 1977]. Basement of the Western Cordillera of Colombia and Ecuador is Cretaceous oceanic crust [Barrero, 1979; Case et al., 1971] which was accreted to the Cordillera Central in the Late Cretaceous [Mooney, 1980]. Basement accretion and initial arc-related igneous activity both took place in the Campanian, making it difficult to determine whether the Western

Cordillera was an arc prior to collision or whether subduction simply stepped out to the west after the oceanic basement had been accreted to the Cordillera Central. The latter is favored here, and the probable cause of accretion and subduction zone step-out was the choking of an east-dipping subduction zone beneath Cordillera Central by abnormally thick and buoyant crust now seen in the Western Cordillera.

The magmatic arc has developed primarily upon continental crust of the South American Precambrian craton, and has always been active to the east of known or inferred subduction zones. Hence, the polarity of subduction has been west-facing since the Jurassic.

Collision zones:

In this section, compressional deformation due to subduction, common and longlived throughout Caribbean evolution, is distinguished from orogenesis caused by collisions of buoyant masses such as arcs, which are relatively short-lived events at various times. At least five such collisional events can be identified (Figure 6.2): the Greater Antilles arc with the southern Bahamas Platform in the early Paleogene; the North Venezuela Nappes with northern South America throughout the Tertiary; an uncertain terrane possessing a volcanic source and an ophiolitic forearc with the southern Yucatan block (Motagua suture zone) in the Late Campanian and Maestrichtian; the Western Cordilleran oceanic complex with the Cordillera Central (Romeral suture zone) in the Late Cretaceous; and the eastern part of the Costa Rica-Panama Arc with the Western Cordillera since the Middle Miocene (Atrato suture zone). The brief review presented below of the timing and the vergence (direction of motion and emplacement in the orogenic zones) of each of these collisions further constrains the spatial and temporal motions of plates during Caribbean evolution.

Early Paleogene Collision of Greater Antilles Arc with the Bahamas Platform:

The effects of this collision are most pronounced in Cuba, where the edge of the Bahamas Platform has been involved in thrusting with the forearc. An uplifted subduction complex exists in northern Hispaniola [Bowin and Nagle, 1980], but evidence for large-scale horizontal motions of thrust sheets is lacking. Northern Puerto Rico has no evidence for collision at all. These differences are due to large-offset strike-slip motions associated with the opening of the Cayman Trough at an oblique angle to the Greater Antilles arc since the Eocene (discussed later).

In northern Cuba, three important tectonostratigraphic assemblages from south to north are: 1) a mafic basement complex (probably forearc) with primarily Cretaceous, but possibly Paleogene also, arc-related plutons, 2) an intermediate zone of Late Jurassic to Eocene deep-water strata (continental rise/slope, abyssal plain), and 3) a carbonate platform (Bahamas) of Late Jurassic through Middle Eocene age [Pardo, 1975; Mossakovskiy and Albear, 1978; Meyerhoff and Hatten, 1974]. The forearc was thrust northward over the carbonate platform by as much as 90 km [Shein and Kleshchev, 1977]. Olistostromes of Paleocene to Middle Eocene age formed during this thrusting, which also included thrusting of ophiolitic nappes.

Allochthon and autochthon both were folded and were overlapped unconformably by little-deformed Upper Eocene and younger sediments. Left-lateral strike-slip faulting accompanied the folding [Shein and Kleshchev, 1977]. The structure and stratigraphy suggest the following sequence of tectonic events: as the north-facing Greater Antilles Arc (Cuban portion) approached the edge of the Bahamas Platform, the forearc region began to override first the carbonate-bank edge and then the bank itself. Before compression ceased, the forearc was squeezed into a north-vergent nappe, and subsequently folded and sinistrally tear-faulted. Late Eocene and younger sediment was deposited unconformably upon thrusts of the collision zone.

The major structural effects of the collision began in the Paleocene, and development of the olistostromes and deposition of mafic/ultramafic bearing flysch in the slope deposits continued into the Early Eocene. The termination of the collision is constrained between the age of that flysch and the age of the transgressive sediments above the post-orogenic unconformity. Therefore, collision culminated in the Middle Eocene, an age corroborated by the Middle Eocene termination of arc magmatism in Cuba. Further evidence for this age of collision comes from the subsidence of the Bahamas Bank. In the Andros well to the north of central Cuba, Paleocene-Middle Eocene deposition occurred at rates 5 times faster than the rates before and after this interval [Paulus, 1972]. It is suggested that the Andros area effectively represents the foredeep basin of the Cuba-Bahamas collision.

The Cauto Basin (Figure 6.2) seems to be a fundamental structural break along the strike of Cuba. To the west of Cauto Basin, the Bahamas Platform is involved in the thrusting whereas to the east, obducted ophiolites and melanges occur but the platform sediments apparently are not significantly thrusted. This difference in response to the collision may be due to the original shape of the arc or the Bahamas, or simply to differing degrees of compression.

In Hispaniola, the effects of collision are much diminished relative to those in Cuba. The structure of the original arc-trench system is well preserved: volcanic arc (Cordillera Central), forearc basin (Cibao Basin, basement of Cordillera Septentrional), and subduction complex (north coast). The Eocene collision did not progress to the point where the forearc complex was thrust onto the Bahamas Platform, nor did thrusting propagate into the Platform itself at that time. The collision in Hispaniola is marked by an Eocene melange containing ophiolitic debris and Cretaceous to Paleocene rocks [Bowin and Nagle, 1980] which is unconformably overlain by relatively undeformed Late Eocene shallow-water sands (Luperon Formation) along the north coast [Nagle, 1966, 1979; Bowin and Nagle, 1980]. Magmatic-arc volcanism generally ceased in Hispaniola in Middle Eocene time, coeval with collision.

Seismic reflection and seismological studies, however, indicate that Hispaniola has overthrust the Bahamas somewhat during the Neogene to Recent [Austin, 1983; Bracey and Vogt, 1970; Sykes et al., 1982]. This younger development is due to post-Middle Miocene transform motion along the Oriente Fault between Cuba and Hispaniola, which has resulted in transpressional convergence between Hispaniola and the eastern Bahamas [Pindell and Dewey, 1982]. Minor, Neogene alkalic volcanism has occurred in west-central Hispaniola [Vespucci, 1980] which could relate to this convergence, but this volcanism probably results from local extension accompanying sinistral shear in the North Caribbean Plate Boundary Zone [Burke et al., 1980].

In Puerto Rico, a forearc region and subduction complex are not exposed north of the arc. Consequently, the end of arc-continent convergence is identifiable only by the marked Mid-to-Late Eocene unconformity [Meyerhoff et al., 1983; Glover, 1971] and by the termination of arc magmatism [Cox et al., 1977]. Oligocene dates on granodiorites and quartz diorites in the Virgin Islands [Kesler and Sutter, 1979; Cox et al., 1977] may represent the last gasp of Greater Antillean magmatism, or may relate to initial subduction along the Lesser Antilles Arc.

In summary, the collision between the Greater Antilles and the Bahamas Platform occurred in the Late Paleocene to Mid-Eocene, and the Cuba portion of the arc shows the most dramatic evidence for the collision. Most thrusts there are north vergent, which supports the concept of south-dipping subduction of the Proto-Caribbean beneath the Greater Antilles Arc prior to collision. There is no strong diachroneity of the time of collision along strike, although the sinistral tear-faults in Cuba [Skvor, 1969; Case and Holcombe, 1980] may indicate slightly oblique, eastward progressive suturing.

An Uncertain Arc Terrane with Southern Yucatan (Motagua zone):

This Late Cretaceous collision occurred between a volcanic arc possessing an ophiolitic forearc, and the continental crust and carbonate shelf of southern Yucatan (Figure 6.2). During the Late Campanian to Maestrichtian, the Neocomian to Santonian carbonate shelf (Campur and Chemal Formations, Vinson [1962]) was depressed and buried by up to 2500 meters of turbiditic flysch from a southern source (Sepur Formation, Rosenfeld [1981]). This flysch is an argillaceous foredeep facies which extends from the Mexican/Guatemalan border to the Caribbean.

The Early Cretaceous Santa Cruz ophiolite was thrust northward onto the Sepur foredeep [Williams, 1975; Rosenfeld, 1980; 1981]. Overlying the ophiolite are deep-water sediments of the Aptian-Albian (and possibly younger as well) Tzumuy Formation [Rosenfeld, 1980], which grade from tuffaceous pelagics to increasingly coarse and proximal turbidites with volcanic debris of andesitic composition and ash. These sediments suggest a volcanic arc as the source area. The timing of emplacement of the ophiolite, dating the collision, is Campanian and Maestrichtian; the ophiolite was initially thrust onto Campanian Sepur sediments, and the Sepur depocenter migrated northward during the Maestrichtian, presumably as thrust loading progressed. Subsequently, uplift and large-scale transform motion has occurred along the Polochic and Motagua faults [Burkart, 1983].

Because of the uncertainty of the amount of Cenozoic offset along the Cayman Trough and Polochic-Motagua transform, the identity of the piece that collided with southern Yucatan is not clear. The sediments derived from the unknown terrane indicate that it was a volcanic arc of intermediate composition. Because southern Yucatan was a stable shelf up to the time of collision, the vergence of the collision was northwards, that is, subduction prior to collision was south-dipping, in accordance with the emplacement of the Santa Cruz. Assuming the Santa Cruz is the remnant of an ophiolitic forearc, one can speculate that the colliding mass was an intra-oceanic arc rather than an Andean-type arc; few Andean arcs possess ophiolitic slabs as forearcs. If one accepts more than 1000 km of east-west opening along the Cayman Trough [Pindell and Dewey, 1982; Wadge and Burke, 1983], a possible candidate for the colliding mass is the Nicaragua Rise-Jamaica arc. However, as pointed out earlier, sinistral shear probably occurred prior to, as well as during, the opening of the Trough. Hence, western Cuba is a possible candidate as well.

North Venezuelan Nappes with Northern Venezuela:

The northern margin of South America from Guajira to Trinidad has been overthrust from the north or northwest during Cenozoic time (Lara Suture of Figure 6.2). The thrust sheets comprise mainly Cretaceous to Paleogene variously metamorphosed volcanics, ultra-mafics and volcanogenic sands and shales, with minor limestones. These rocks collectively form an assemblage typically associated with accretionary wedges at subduction zones. Rocks deposited in shallow-water of the Venezuelan shelf are sometimes incorporated into the thrusting as well. The assemblage crops out in northern Guajira and Paraguana [MacDonald, 1964; Lockwood, 1965; Martin-Belizzia and Arozena, 1972; Case et al., 1984], but in Falcon Basin they predominantly lie in the sub-surface (except along the basin's southern margin at Siquisique and Aroa-Mision) beneath a 6 km sedimentary section [Muessig, 1984; Bellizzia et al., 1972]. In the Caribbean Mountain system of north-central Venezuela these rocks comprise the Lara Nappe of metasediments and the Villa de Cura Klippe of predominantly blueschist-facies mafic metavolcanics, metasediments and ultra-mafics; metamorphism is medial to Late Cretaceous in age [Stephan et al., 1980; Maresch, 1974; Gealey, 1980; Beets et al., 1984]. Similar assemblages of rocks occur in the Arraya-Paria Peninsula, Margarita, and in the Coastal Ranges of northern Trinidad [Vierbuchen, 1984; Speed, 1985]. Generally, it appears that an arc terrane with a well-developed forearc and accretionary complex collided with northern South America.

Two major questions, however, plague the understanding of this margin's development, and both must be considered before the apparent arc-continent collision can be properly interpreted. First, "How much deformation has occurred to northern South America during the Late Miocene to Recent Andean Orogeny?" Second, "When exactly were the thrust sheets of oceanic rocks emplaced onto the Venezuelan continental margin, and was this a diachronous process?" The first question is fully considered in chapters 1 and 7; it appears that the entire northwest corner of South America has migrated to the northeast by about 290 km over the last 9 million years. The 9 Ma paleogeographic reconstruction of northern South America (Figure 7.2) is assumed here to quantify and to retract the effects of this deformation.

Answering the second question is more difficult; assessing the timing of collision for any collision zone is difficult because of the potential geometric and volumetric complexities resulting from thrusting. Of particular importance is discerning the sediments of the accretionary wedge, and their deformation during subduction, from the foredeep sediments and the deformation resulting from the actual emplacement of the accretionary wedge upon the margin. In the case of northern South America, this is very difficult because both the accretionary wedge and the foredeep basin are composed of similar terrigenous sediments.

The best evidence for the timing of collision should come from the subsidence history of the autochthon, or the Venezuelan shelf to the south of the thrust sheets. The initiation of subsidence in the Maracaibo area is Paleocene-Lower Eocene [Zambrano et al., 1972]; subsidence south of the Venezuelan Mountain system (Piedmont Province of Beck [1978]) is Oligocene?-Miocene [Beck, 1978]; initiation of subsidence in the East Venezuelan Basin is Miocene-Pliocene [I. Rodriguez, pers. comm., 1985; Vierbuchen, 1984]. Thus it appears that the emplacement of the north Venezuelan Nappes was diachronous to the east. Much of the Paleocene-Lower Eocene flysch commonly referred to as the foredeep facies in front of the entire thrust belt (such as the Piemontine Nappe of central Venezuela [Beck, 1978]) was probably accreted at the accretionary complex from the floor of the Proto-Caribbean during the Eocene in the the subduction stage, prior to the actual emplacement onto the margin.

Assuming the Caribbean Plate has migrated eastwards with respect to North and South America since the Eocene, interactions between the Caribbean and northern South America should be diachronous to the east. In a later section (Development of the South Caribbean Plate Boundary Zone), the positions of the Caribbean Plate relative to South America are estimated. It is shown that close correlation exists between the probable position of the leading edge of the Caribbean Plate and the emplacement age of the North Venezuelan Nappes, at various points along the margin. Hence, it is suggested that the North Venezuelan Nappes were emplaced progressively during the Cenozoic as a result of a compressional component of the predominantly transcurrent relative motion which has exceeded 1000 km since the Eocene. Iteratively, this emplacement history provides an indirect check on the predicted motion history of the Caribbean Plate relative to South America. This model is in contrast to to the interpretations of some previous geologic-kinematic Caribbean evolutionary models [White and Burke, 1980; Pindell and Dewey, 1982] which attributed nappe emplacement to Late Cretaceous-Paleogene convergence between North and South America, rather than to interaction with the Caribbean Plate.

The arc complex that collided with northern South America is difficult to identify with certainty because of the large amounts of strike-slip offset. However, the Leeward Antilles probably form at least a part of the arc complex, particularly to the west of Golfo de Triste. The 9 Ma reconstruction (Figure 7.2) places the Leeward islands west of Golfo de Triste, along strike with a few calc-alkaline plutons of the Coast Ranges of the Venezuelan Mountain system. The Paleocene-Early Eocene collision to the west of Golfo de Triste matches the Paleocene termination of plutonism in the Leeward Antilles (Figure 6.3). The plutons of the Venezuelan Coast Ranges may be an eastward extension of the Leeward Antilles arc; if so, their younger age (Eocene, [Maresch, 1974]) may also testify to the diachronous nature of the collision.

Assuming that the diachronous collision was related to interaction between the Caribbean and South American Plates, the Leeward Antilles arc probably represents a portion of the Caribbean Plate. As shown later in a more complete model for the development of the northern South America, the Leeward Antilles Arc (and its forearc and accretionary wedge) was probably a southern extension of the Aves Ridge Arc, which was progressively clipped off and accreted to northern South America during the Cenozoic. The metamorphic rocks of the Lara Nappes that were obducted onto the Venezuelan margin were metamorphosed during the accretionary prism stage during subduction of the Proto-Caribbean, prior to their Cenozoic emplacement. The apparent geochemical affinity between the basalts defining seismic horizon B" of the Caribbean Plate and the tholeiitic units (basement) of the Leeward Antilles [Donnelly, in press, 1985] is further evidence for this model of arc-continent collision.

Western Cordillera/San Jacinto Belt with Cordillera Central, Colombia and Ecuador (Romeral Suture):

In Colombia and Ecuador, the Romeral Suture (Figure 6.2), a major fault zone along which occur ultramafics, melanges and blueschists, separates two distinct geological provinces [Case et al., 1971]. To the east occur Precambrian through Cenozoic continental rocks of Cordillera Central and the Santa Marta Massif. Directly to the west, the Western Cordillera and San Jacinto Belt consist of Cretaceous oceanic crust, Late Cretaceous through Paleogene highly-deformed, deep-water sediments, and Cenozoic magmatic rocks [Case et al., 1971; Case and MacDonald, 1973; Irving, 1975; Duque-Caro, 1979; 1984]. Oceanic basement does not actually crop out in the San Jacinto Belt, but it is inferred from gravity studies to exist at depth along strike of the Western Cordillera. Structural and facies relations indicate that the Western Cordillera and San Jacinto terranes were accreted to Cordillera

Central during Late Cretaceous-Paleogene time; suturing was completed by the Eocene, and may have been diachronous towards the north. The most intense deformation (subduction zone choking) seems to have occurred during the Late Cretaceous in southern Colombia and Ecuador [Henderson, 1979]. In contrast, deformation climaxed in the Eocene to the north in the San Jacinto Belt [Duque-Caro, 1979; 1984]. The predominantly Cenozoic (mostly Oligocene) magmatism within the Western Cordillera post-dates the suturing to Cordillera Central.

The Western Cordillera's brief history as an island arc, therefore, bears no relationship to subduction geometries prior to its Late Cretaceous accretion. Composite faults along the Romeral dip to the east, and the only significant subduction-related magmatism prior to collision occurred in the Cordillera Central/Santa Marta Arc. These considerations suggest east-dipping subduction at the site of the Romeral Suture prior to accretion.

It appears as though accretion resulted from a non-subductable area of oceanic crust arriving at and choking the Cordillera Central subduction zone during Late Cretaceous-Paleogene time. Subduction then stepped outboard of the Western Cordillera for the remainder of the Cenozoic. It has been suggested [Duque-Caro, 1979] that oceanic crust in the San Jacinto/Western Cordillera belt is related to the abnormally thick crust of the Caribbean Plate (basaltic extrusives, seismic horizon B", atop oceanic crust). If the Caribbean Plate is of Pacific provenance [Malfait and Dinkelman, 1972], the Western Cordillera could be an accreted remnant that was emplaced during the Plate's entrance into the gap between the Americas.

After the development of a new east-dipping subduction zone to the west, the Western Cordillera became a logical site for subduction-related magmatism through the Cenozoic. It did not evolve as an axis of plutonism, however, until the Oligocene. Figure 2.1 notes that the rate of Farallon-North America convergence, and therefore very nearly the rate of Farallon-South America convergence, dropped from 140 to 75 mm/yr at the end of the Eocene. If it is accepted that convergence rate controls the dip of the Benioff Zone, then a cause-and-effect relationship may exist between the slowing of subduction and the westward shift of the axis of northwestern South American volcanism (steepening of Benioff Zone). Conversely, the Miocene return of volcanism to the Cordillera Central may have resulted from the subduction of the more buoyant crust of the eastern part of the Panama Arc (see below), and the young crust which has been created at the Galapagos Ridge spreading center since the Early Miocene only [Klitgord and Mammerickx, 1982].

Eastern Panama with Western Colombia:

Supposition of a collision between the eastern part of the Costa Rica-Panama Arc and the Western Cordillera of Colombia arises from consideration of Cenozoic plate motions and modern plate boundaries in the area. If the Caribbean Plate, with Costa Rica-Panama forming its western edge, has moved eastward with respect to South America, the southwestern portion of the Caribbean must have converged with South America. Hundreds of kilometers of relative motion implies that the Costa-Rica-Panama Arc was formerly distant from Colombia and, therefore, that the two have collided at some time during the period of migration.

However, a suture between the Western Cordillera and the Panama Arc has not yet been clearly recognised in the field. Despite this, the Atrato-San Juan Basin (Figure 6.2) is a strong candidate for the predicted suture [Case et al., 1971]. First, structural and lithologic trends continue from eastern

Panama into the Serrania de Baudo, which defines the western flank of the basin. The Atrato-Rio San Juan Valley separates this basement high from the Western Cordillera. Second, a strong relatively negative gravity anomaly with steep gradients characterizes the basin, which is filled with 5 to 10 km of sediments that are moderately to complexly folded [Case et al., 1971]. Finally, post-Eocene sediment accretion has occurred along strike to the north in the Sinu Belt, and is due to subduction of, and sediment offscraping from, the Colombian Basin beneath northern Colombia [Duque-Caro, 1979; 1984]. In so much as the Costa Rica-Panama Arc is viewed as the trailing edge of the Caribbean Plate, which comprises both the Colombian Basin and Panama, it appears that collision has yet to occur in the Sinu Belt. Hence, the collision appears to be northward progressive, and is presently occurring in the area of Golfo de Uraba.

Collision appears to have begun in the Miocene (Middle?), as suggested indirectly by 1) uplift of the Panamanian forearc in Middle to Late Miocene [Bandy and Casey, 1973; Bandy, 1970], 2) uplift of the Serrania de Baudo in the Miocene [Bourgois et al., 1982], 3) the Miocene plate-boundary reorganization within the Nazca Plate off western Colombia [Lonsdale and Klitgord, 1978], 4) the regional hiatus of Late Miocene age in the Atrato Basin [Duque Caro, 1972; 1984], 5) Late Miocene tectonic instability in northwest Colombia that changed the drainage course of the Magdalena River [Duque-Caro, 1979], 6) divergence of planktonic species in the Pacific and Caribbean in the early Pliocene [Keigwin, 1978], and 7) major faunal exchange between North and South America during the Plio-Pleistocene [Webb, 1976]. This collision is the primary cause of the Andean Orogeny in northwest South America (see Shagam [1975]; and Irving [1975], for reviews), and has triggered the Late Miocene to Recent northeastward mobilization of Andean Cordilleran Terranes (chapter 7).

Accompanying this collision is the northward escape of blocks of the eastern part of the Panama Arc along north to northwest trending sinistral faults [Case and Holcombe, 1980]. This has given the arc its northward convexity. The body of the arc itself has been obducted onto the Colombian Basin. Seismic studies trace the Colombian Basin underriding the Panamanian Foldbelt of recently accreted sediment [Lu and McMillen, 1983]. In essence, it appears as though the eastern portion of the arc is reversing polarity in its attempt to resist subduction at the Atrato-San Juan suture. The collision began with the impingement of the eastern end of the south-facing Costa Rica-Panama Arc (Serrania de Baudo) into the Sinu Trench (subduction beneath Western Cordillera) approximately at Buenaventura. Cessation of subduction at the now-buried, east-trending Panama Trench [Lowrie, 1978] caused northward migration of blocks of the eastern portion of the arc along the sinistral transcurrent faults. Motion of the eastern arc with respect to the Western Cordillera has remained essentially convergent, as it too has a northward component of motion with respect to the Colombian Basin (chapter 7).

Paleogeographic significance of the Yucatan Basin, Grenada Basin, Cayman Trough:

The Yucatan and Grenada Basins and the Cayman Trough (Figure 6.2) are the only oceanic portions of the Caribbean region that are not remnants of Cretaceous ocean crust of Pacific provenance. These basins have opened to accommodate geometric imperfections of the Caribbean Plate's Cenozoic migration between the Americas. The Grenada and Yucatan Basins are probably back-arc basins, whereas the Cayman Trough is probably a very long pull-apart basin. All three provide critical input to the deduction of

Caribbean evolution.

Yucatan Basin:

The Yucatan Basin usually has been considered (1) Mesozoic ocean crust formed due to radial separation of the Chortis Block from the Yucatan Peninsula [Dillon and Vedder, 1973], (2) a piece of Cretaceous ocean crust rafted in from the Pacific [White and Burke, 1980], or (3) an intra-arc basin fromed by separation between Cuba and the Cayman Ridge during the Paleogene [Gealey, 1980]. Seismic reflection and heat-flow data, and age-depth relationships, suggest an Early Paleogene (55-65 Ma) age for a seafloor spreading origin [Rosencrantz et al., in press], consistent with Gealey's [1980] intra-arc spreading origin. Seafloor spreading related magnetic anomalies trend northeast, suggesting northwest-southeast directed spreading [Hall and Yeung, 1980]; however, identification of anomalies is suspect.

For the Yucatan Basin to be an intra-arc basin between Cuba and the Cayman Ridge, Cuba must have migrated north-northeastward along the eastern Yucatan margin. However, this trend is discordant with the trend of opening suggested by magnetic anomalies. This discordance suggests that plate motions occurred between three plates; North America, the Caribbean, and Cuba. A model which satisfies the geometric and age requirements for the basin is shown in Figure 6.4. It is concluded that the Caribbean Plate (Cayman Ridge) migrated about 377 km east-northeast with respect to North America during the period of Yucatan Basin formation. This accords with the concept of the Caribbean Plate entering the North America-South America gap during the Late Cretaceous through Eocene.

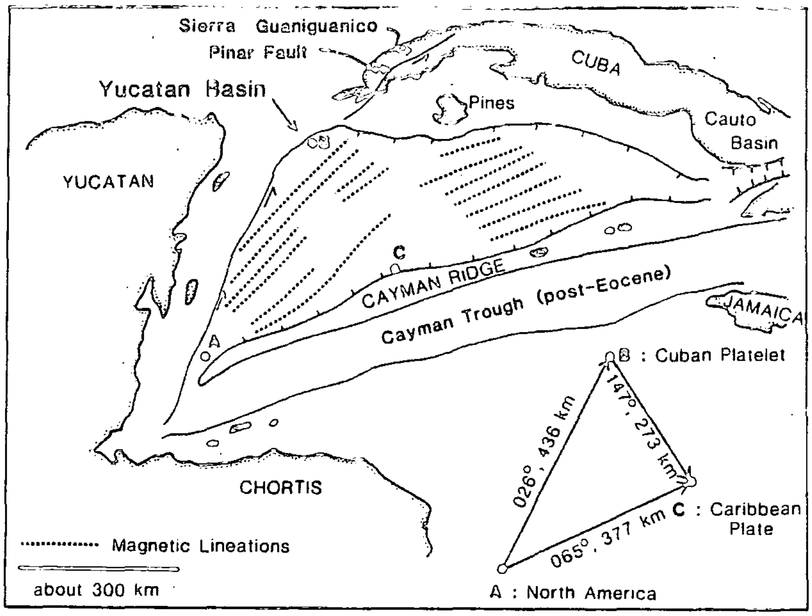


Figure 6.4. Magnetic anomaly lineations in Yucatan Basin, after Hall and Yeung [1980]. In the three-plate system of North America, Cuba and the Caribbean (Cayman Ridge), Cuba migrated, with respect to North America, from point A to point B, as defined by the trend and length of the eastern Yucatan sheared margin. The Cayman Ridge migrated, with respect to Cuba, from point B to point C, as defined by the perpendicular to the mean trend of the magnetic anomalies and the width of the basin. Completing the vector triangle from point A to point C gives the motion of the Caribbean with respect to North America during the period of back-arc spreading in the basin (Maestrichtian to Mid-Eocene, 15-20 My), assuming no subduction has occurred at any margins of the basin. Vector triangle in lower right defines trends and magnitudes of these apparent motions. Pinar fault in western Cuba separates the Cuban arc (or forearc) to the east from continental crust of the northeastern Yucatan block to the west.

The Grenada Basin:

This basin (Figure 6.2) separates the Lesser Antilles and Aves Ridge arcs. In the south, basement lies beneath 4000 m of quite undisturbed sediment [Biju-Duval et al., 1978] and is probably oceanic. In the north, however, a block-faulted basement is indicated on seismic lines which is composed of Cretaceous arc-related volcanics [Nemec, 1980]. This basement extends to the east and forms the basement of the northern Lesser Antilles. The block faulting is primarily Paleocene or Early Eocene [Nemec, 1980]. Likewise, heat-flow measurements in the deeper southern portion of Grenada Basin (G. Westbrook, pers. comm., 1984) suggest a Paleocene-Eocene age for the formation of oceanic crust. Thus, it is assumed that the Grenada Basin formed during the Paleocene to Early Eocene interval. The Paleocene-Early Eocene is also when plutonism ceased in the Aves Ridge Arc and began in the Lesser Antilles Arc (Figure 6.3). Hence, it appears as though the Grenada Basin is a Paleocene-Early Eocene back-arc basin which formed within the Late Cretaceous-Paleocene Aves Ridge Arc.

A simple east-west extensional origin for the basin is possible [Tomblin, 1975; Pindell and Dewey, 1982], but north-south trending magnetic anomalies or fractures have not been identified. On the contrary, magnetic lineations radiate generally east-west across the basin (G. Westbrook, pers. comm., 1984). If these are, in fact, seafloor spreading anomalies, an oblique opening for the basin is indicated.

Several other indirect lines of evidence also suggest an oblique opening for the Basin. First, a very steep and straight scarp defines the southeastern margin of Aves Ridge [Case and Holcombe, 1980]. The scarp appears more like a transform scarp than a rifted margin. Second, a positive gravity anomaly axis occurs along the eastern scarp of the Limestone Caribbees of northern Lesser Antilles [Tomblin, 1975]. As with southeastern Aves Ridge, this scarp is also bathymetrically defined. Further, the island of La Desirade aligns with this scarp, on which is found Jurassic pillow basalts and radiolarites which may represent oceanic crust [Mattinson et al., 1980]. The close proximity of La Desirade to the Lesser Antilles volcanic axis makes it unlikely for La Desirade to have been accreted to the arc; the accretionary complex lies about 100 km to the east. It is possible that the La Desirade and southeastern Aves Ridge scarps are remnants of transform faults which were connected by a short ridge segment in the southern Grenada Basin. Seafloor spreading in the basin, and dextral motion on the faults, could have opened the basin by north-south extension, as recorded by the magnetic anomalies in the basin.

Figure 6.5 outlines a model for such an oblique opening. In addition to the geometric arguments above, this model is made more attractive by two further lines of reasoning. First, If the Caribbean Plate entered the North America-South America gap in a northeast direction from the Pacific region during the Late Cretaceous-Eocene, then the eastern portion of the leading edge of the plate must have subjected to dextral shear, which is a requirement for the proposed model. Second, the model suggests that a significant portion of the original Aves Ridge and its forearc should have progressively protruded southwards as result of the opening of the Grenada Basin. There is geochemical affiliation between the basement of the Leeward Antilles Arc offshore northern Venezuela and the crust of the Caribbean Plate [Donnelly, in press, 1985]. It is suggested, as outlined in more detail later, that this southern protrusion of arc and forearc has become the Leeward Antilles arc and the Lara Nappes (including the Villa de Cura forearc) of northern Venezuela. Emplacement of the nappes resulted from a compressional component of many hundreds of km of dextral shear between the Caribbean and South American Plates since the Paleocene.

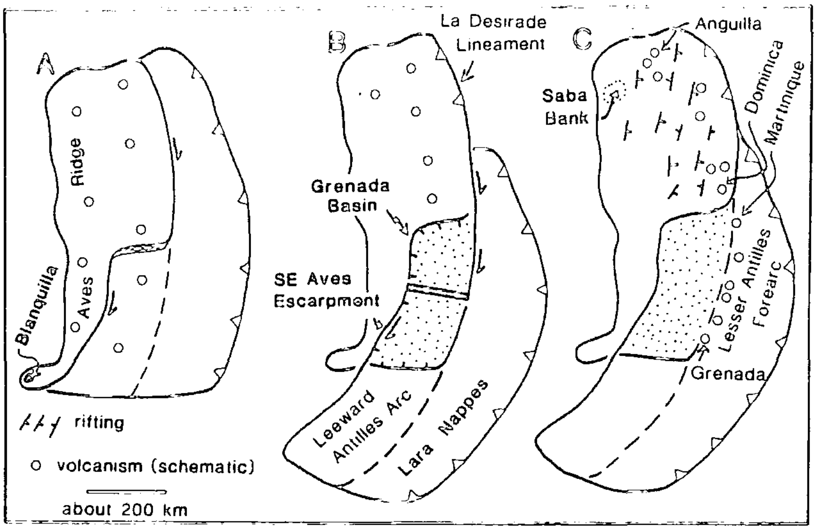


Figure 6.5. Proposed model of formation for Grenada Basin, based upon recently defined magnetic anomaly trends (G. Westbrook, pers. comm., 1984). A: pre-opening phase (Late Cretaceous), dextral oblique subduction of Proto-Caribbean (Atlantic) crust induces shear stress and mobilization of the eastern Aves Ridge arc and subduction complex. B: opening of Grenada Basin led to progressive protrusion to the south of southeastern Aves Ridge arc and forearc, which would become accreted to northern South America (Leeward Antilles and the Lara Nappe (see Figure 6.7) from Paleocene through Pliocene [Stephan et al., 1980]. Opening of Grenada Basin was probably completed by the end of the Eocene. La Desirade and the southeast Aves lineaments are remnants of transforms. C: Extensional adjustments in the north during the Eocene-?Oligocene [Nemec, 1980] allowed a restraightening of the Caribbean-Atlantic trench interface, as it is today beneath the accretionary complex [Speed and Westbrook et al., 1984]. Youngness of volcanism in the south, relative to that in the north (see arc section, above), is due to a lag in subduction as the Lesser Antilles forearc remained essentially fixed to the South American plate during the opening of the Grenada Basin.

The Cayman Trough:

This long, deep basin (100 by 1400 km) forms the North America-Caribbean plate boundary in the northwest Caribbean [Perfit and Heezen, 1978; Sykes et al., 1982]. Much of the basin is floored by oceanic crust produced at the north-trending Mid-Cayman Spreading Center, a short mid-ocean ridge segment that links the sinistral Swan and Oriente transform faults (Figure 6.2). The amount of oceanic crust which has been created at the Mid-Cayman Spreading Center defines the minimum relative motion between the Caribbean and North American Plates since seafloor spreading began; hence, the Cayman Trough is critical in unraveling the plate-tectonic history of the Caribbean.

Wadge and Burke [1983] reviewed the evidence from the Cayman Trough and concluded that it began to form (deep-water sedimentation accompanied by strike-slip motions in the northern Caribbean) in the Oligocene. However, the plate boundary configurations responsible for its origin are not definitive; seafloor spreading anomalies parallel to the Mid-Cayman Spreading Center [MacDonald and Holcombe, 1978] only indicate an offset of 284 km; anomalies further from the spreading center have not been identified. Debate focuses upon whether the remainder of the Trough has formed by spreading at the Mid-Cayman Spreading Center or by another mechanism such as north-south extension.

Estimates of east-west offset across the Trough range from 180 km, based on displacement of similar metamorphic belts (subduction complex and arc) of Cuba and Hispaniola [Meyerhoff, 1966], to 1400 km, based on the assumption that the entire bathymetric expression of the trough is due to spreading at the Mid-Cayman Spreading Center [Sykes et al., 1982; Macdonald, 1974]. However, it is unlikely that total offset could be as much as 1400 km; that portion with depths typical of true oceanic crust is only 980 km. The western and eastern ends of the Trough (420 km) are floored by block-faulted, probably arc-related basement (as shown by dredging; Perfit and Heezen [1978]), which has subsided to abyssal depths. Although extensional, these areas cannot be related directly to relative motion between the North America and Caribbean plates. Given that large amounts of subsidence arise from small amounts of extension in arc terranes due to a large crust to lithospere thickness ratio [Cooper, 1983], it is estimated that only 70 to 100 km of basement extension during an initial stage of Trough formation is sufficient to have produced the subsided, yet non-oceanic, ends of the Trough. Hence, if one accepts the pull-apart model of formation, total extension across Cayman Trough is between 1050 and 1100 km.

Macdonald and Holcombe [1978] identified magnetic anomalies flanking the Mid-Cayman ridge axis which indicate at least 284 km of spreading since the Late Miocene (anomaly 4'), although the spreading rate has decreased from 4 cm/yr to 2 cm/yr over the last 2.4 Ma. Based upon the fact that seismicity (sinistral strike-slip motion) occurs today along the southeast margin of the Trough, Sykes et al. [1982] suggested that Caribbean-North America relative motion continues to the present at 4 cm/yr by motion occurring at both the spreading center and along the southeast margin. Anomalies beyond anomaly 4' have not been identified in the Trough, but basement lineations attributable to east-west seafloor spreading suggest an offset of at least 560 km [Holcombe and Sharman, 1983]. Because these estimates of Late Miocene to Recent relative motion far exceed Meyerhoff's [1966] 180 km offset of the Cuban and Hispaniolan metamorphic belts (arc and subduction complexes), alignment of the metamorphic belts does not constrain the total east-west offset across the Cayman Trough [Pindell and Dewey, 1982]. Much of the motion must have occurred along faults which pass south of the Cordillera Central arc of Hispaniola.

Furthermore, motion between Cuba and Hispaniola along the Oriente Fault probably has occurred only since the Middle Miocene (see chapter 4). Oligocene-Middle Miocene transcurrent motion along the Cayman Trough, therefore, apparently bypassed the Hispaniolan arc to the south. Farther west, Oligocene-Middle Miocene relative motion probably occurred occurred along bounding faults of the early Cayman Trough, just as it has since the Late Miocene; no other similarly oriented fault systems are apparent. Thus, the transcurrent offset and crust generated by seafloor spreading over the last 8.3 Ma (about 284 km) or 15 Ma (560 km) probably represent only a portion of the total.

The Cayman Trough's initiation as a deep-water depocenter accompanied with strike-slip deformations by the Early Oligocene [Wadge and Burke, 1983] is an important piece of evidence indicating Oligocene eastward migration of the Caribbean Plate relative to North America. However, the beginning of subduction-related magmatism in the Lesser Antilles Arc is Late Eocene (see Figure 6.3), and evidence of strike-slip faulting within the Greater Antilles indicates initiation as early as Middle or Late Eocene. The "Eocene Tectonic Belt" of west-central Puerto Rico is composed of fault-bounded blocks of Late Cretaceous limestone and Eocene "flysch" and volcanics which was severely sheared during the Eocene, prior to the Late Eocene? and Oligocene [Barabas, 1983; J. Joyce, pers. comm., 1985]. Similarly, Eocene flysch sequences along the northern margin of the San Juan Valley of central Dominican Republic were deformed possibly at a restraining bend of a transform system during the Middle to Late Eocene [J. Dolan, pers. comm., 1985].

In light of 1) Oligocene deep-water sedimentation in portions of Cayman Trough, 2) east-west Caribbean-North America relative motion since the Eocene, 3) the alignment of Cuban and Hispaniolan metamorphic belts bearing no constraint upon total offset across the Trough, 4) the position of Mid-Cayman Spreading Center half-way between the ends of the Trough, and 5) no other apparent mode of formation for the Trough, it is concluded that seafloor spreading at the Mid-Cayman Spreading Center has produced the entire oceanic portion of the trough. Hence, the Cayman Trough may be considered as an exceptionally extended pull-apart basin which records a sinistral offset of 1050-1100 km between the Caribbean and North American Plates since the end of the Early Oligocene, but additional sinistral relative motion is indicated for the Middle or Late Eocene from other sources.

This interpretation is especially important in regard to the evolution of the Caribbean. The entire concept of the Caribbean Plate originating within the Pacific realm and entering the North America-South America gap prior to the Eocene is dependent upon this conclusion. If smaller estimates of offset are assumed (e.g., <800 km), an in situ formation of the Caribbean Plate is required.

Caribbean Plate Boundary Zones:

The Middle Eocene separates two phases of development in the northern and southern Caribbean: 1) an Early Cretaceous-Early Eocene phase of arc volcanism and compressional tectonism which culminated in orogeny in both the north and south, and 2) a subsequent phase of predominantly transcurrent faulting associated with more than 1000 km of eastward migration of the Caribbean Plate relative to the Americas. The latter phase has produced complicated northern and southern plate boundary zones (PBZs) with anastomozing faults and diffuse seismicity [Burke et al., 1980]. In this section, the locations of faults, and the timing of movement upon them, pertaining to this migration are defined.

Northern Caribbean PBZ:

In the north, the Cayman Trough suggests a minimum of 1050 km of transcurrent motion; this is well-defined in the Trough itself, whose eastern end lies directly south of Sierra Maestra of eastern Cuba. However, continuations of composite splays to the east of the Trough must be inferred from geologic relationships in and around Hispaniola and Puerto Rico. In addition to offset due to Cayman Trough spreading, the seismically active southeast margin of the Trough, continuing through Jamaica and into, and along, the southern peninsula of Hispaniola, has been the site of up to 48 additional km of transcurrent motion during the last 2.4 Ma [Sykes et al., 1982; Mann et al., 1984].

Presently, the Oriente Fault between Cuba and Hispaniola is the active transform which connects the Mid-Cayman Spreading Center with the Lesser Antilles subduction Zone, although the Oriente splays into the Bahamas Channel, Camu, and Septentrional faults in northern Hispaniola. Meyerhoff [1966] derived a Cuba-Hispaniola offset of 180 km by realigning the metamorphic belts. Figure 6.6 shows a realignment of the Cuban and Hispaniolan segments of the arc, forearc, and subduction complex, which suggests 400 km of cumulative offset along fault splays to the north of Cordillera Central-Massif du Nord of Hispaniola. The difference arises from recognition of sinistral offsets along the Tabera, Camu, and Septentrional faults. Collectively, restoring these offsets produces a geometry of

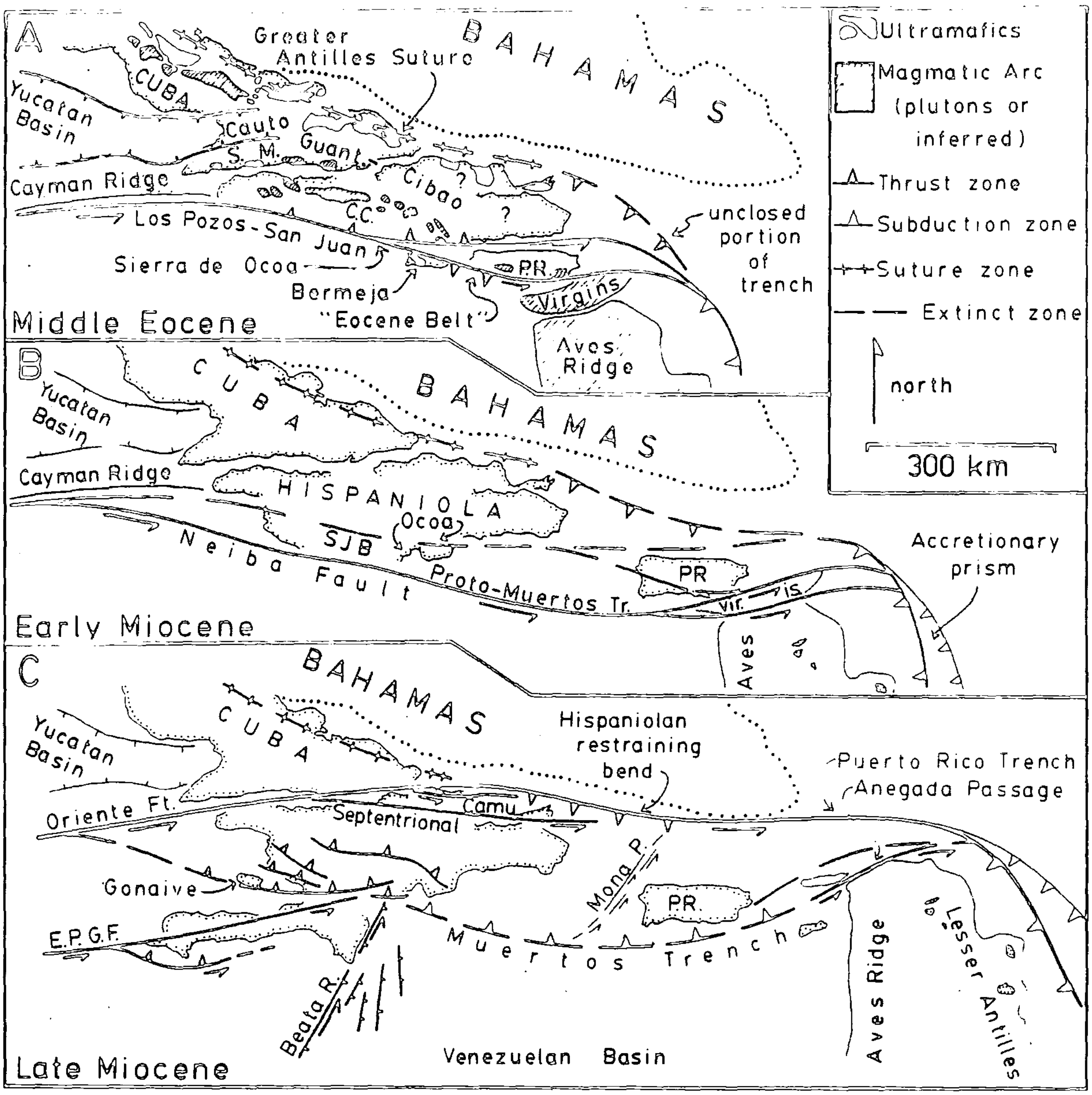


Figure 6.6. Caption on next page.

Figure 6.6. Three-stage development of the northern Caribbean PBZ, showing probable primary locations of transcurrent offset. A: Mid-Eocene time, immediately following Greater Antilles-Bahamas collision (suture zone). Magmatic arc (plutons) and subduction complex (ultramafics) are realigned, except for Paleocene-Middle Eocene dislocation of the arc across Cauto Basin (hinge of opening for Yucatan intra-arc basin). Guantanamo-Cibao forearc basin is inferred. Paleogene tectonic setting of the Eastern Cordillera and Septentrional basements are unclear. Transcurrent motion along the Los Pozos-San Juan fault system separated Puerto Rico (P.R.) from Hispaniola by about 300 km during Eocene-Early Miocene, and juxtaposed the San Juan block (SJB) with the Massif du Nord-Cordillera Central (C.C.) of Hispaniola. Offscraped sediments at this transcurrent zone have accumulated in the Sierra de Ocoa in southern Dominican Republic. The Bermeja area of Puerto Rico may have been displaced an uncertain distance along the "Eocene Tectonic Belt" of Puerto Rico. B: Early Miocene time. Transcurrent offset began on Neiba-Proto Muertos Trough-Anegada system, which dislocated the Aves Ridge, Virgin Islands and Puerto Rico, and juxtaposed the southern block of Hispaniola with the previously accreted San Juan block (SJB). Separation of Puerto Rico from Dominican Republic is responsible for development of the Arecibo Basin along northern Puerto Rico. C: Late-Middle to Late Miocene. Juxtaposition of southern block nearly completed, and primary zone of transcurrent motion jumped to the Oriente-North Hispaniolan system. General north-south compression across the Caribbean since the Late Miocene (see chapter 7) put the Neiba, San Juan and Proto Muertos Trough fault systems into compression, and they have become overthrust to the south. In the Pliocene, northeastern Hispaniola converged upon and overthrust the southern flank of Silver Bank, producing the Hispaniolan restraining bend. Motions have since occurred along the Septentrional, Camu and Enriquillo-Plantain Garden (EPGF) faults, as well as offshore in the north. The Septentrional and Camu are young faults that bypass the restraining bend, straightening out the migration path of the Caribbean Plate past the Bahamas. Probable sinistral faulting along the Mona Passage has allowed Puerto Rico to migrate north with respect to Hispaniola after having passed the Bahamas constriction. The Puerto Rico Trench is essentially a compressional transform [Schell and Tarr, 1978] that developed primarily in Late Miocene to Recent time [Monroe, 1968]. Metamorphics dredged in the trench relate to the pre-Late Eocene phase of subduction prior to the collison with the Bahamas.

tectonic provinces for northern Hispaniola that fits better with Cuba. If 400 km of motion have occurred in the north, fault systems responsible for the remaining 650 km of the Cayman Trough's indicated offset must pass to the south of the arc complex of Hispaniola.

Southwestern Hispaniola consists of three physiographic highs separated by two intervening lows which are Neogene sedimentary basins (Figure 4.3). The Late Miocene to Recent history of these terranes is predominantly that of north-south compression [Biju-Duval et al., 1983a; Bourgois et al., 1979a,b]. The northern block (Cordillera Central-Massif du Nord-Massif du Nord'ouest) has overthrust the San Juan-Plateau Central basin to the south. The middle physiographic high (Sierra Neiba-Chain des Matheaux-Ile de la Gonave) has overthrust the Cul de Sac-Enriquillo Basin. The relationship of the Cul de Sac-Enriquillo Basin to the southern high of Sierra Bahoruco and Massif de la Selle is poorly known. Little evidence exists in these areas for large-scale transcurrent offsets since the Middle Miocene, although minor strike-slip faults such as the Plantain Garden-Enriquillo fault [Mann et al., 1984] do occur. Hence, large-offset transcurrent motions (totalling 650 km) that splayed from the Cayman Trough through southwestern Hispaniola are probably resticted to the Oligocene through Middle Miocene interval (about 3 cm/yr, if motion was continuous).

If an Oligocene to Middle Miocene strike-slip phase preceded the Late Miocene to Recent southward vergent overthrusting phase, then the thrusting probably nucleated upon the pre-existing transcurrent fault zones. Thus, the southern hanging walls of the overthrusting complexes may preserve evidence of the former strike-slip phase, as well as evidence constraining the timing of thrust initiation. Conversely, the northern flanks of the intervening basins have been overthrust, thereby obscuring any record there of strike-slip deformation. Following this logic, the northern San Juan

Boundary Fault and the northern boundary of Enriquillo-Cul de Sac Basin are the likely locations to find evidence for large-scale strike-slip offset.

Along the south flank of Cordillera Central and Massif du Nord, the Los Pozos-San Juan boundary fault zone separates two distinct terranes. To the north lie metamorphosed Cretaceous to Paleogene arc-related rocks and volcanogenic sands of the Cordillera Central and Massif du Nord, and to the south lie Paleogene deep-water micrites, silts and slates which possess little arc derived volcanic debris [Michael, 1979; Bowin, 1975]. Only in Sierra de Ocoa in the southeast of the San Juan Basin does one find the Eocene-Oligocene volcaniclastic and conglomeratic facies expected to have been deposited along the length of the arc complex during and following Late Paleocene-Eocene orogensis [Biju-Duval et al., 1983a; Bourgois et al., 1979b; Dolan, in prep., 1985]. Not until the Early Miocene did significant amounts of terrigenous sand enter the San Juan Basin [Michael, 1979]. This suggests that the basin had arrived in juxtaposition with the arc terrane by that time, although the basin probably continued to migrate along the length of the arc unitl the Middle Miocene. On sedimentologic grounds, the facies discrepancy across the Los Pozos-San Juan Boundary Fault Zone suggests that at least 350 km of transcurrent motion has brought the floor of San Juan Basin into its present position from the west, as measured from western Sierra Ocoa to the tip of the northwest peninsula of Haiti, which itself was a source of volcanogenic sediment in the Paleogene. Structural studies along the fault [Michael, 1979] acknowledge the existence of a strike-slip component of offset, but the magnitude is unknown. The Late Miocene(?) to Recent section along the northern San Juan Basin margin is relatively undeformed in comparison to the older sediments deposited before and during the strike-slip phase.

Along the southern margin of Sierra Neiba, evidence for large-scale strike-slip motion has not been reported. However, the maximum age for the initiation of thrusting is Upper Miocene (Lower Tortonian, about 11 Ma; Bourgois et al., [1979a]), as indicated by Eocene rocks thrust onto Tortonian rocks. To the south of Sierra Neiba, the Fondo Negro Basin, an exposed part of Enriquillo Basin, was filled predominantly during Late Miocene-Pliocene time, in accordance with southward-directed thrusting in Sierra Neiba [Cooper, 1983]. The difference between the minimum of 350 km offset along the Los Pozos-San Juan fault and the total of 650 km that passed through southern Hispaniola may have occurred along Sierra Neiba-Chaine des Matheaux-Ile de la Gonave, before it overthrust the Enriquillo-Cul de Sac Basin in Late Miocene time. Before this overthrusting, the Plateau Central-San Juan Basin and the Cul de Sac-Enriquillo Basin may have been one and the same.

Both the Los Pozos-San Juan Fault and the southward limit of thrusting in the Sierra Neiba-Matheaux-Ile de la Gonave thrust belt have logical bathymetric continuations into the northeast corner of Cayman Trough [Case and Holcombe, 1980]. These probable former transcurrent fault zones are now draped with Neogene sediment, attesting to their dormancy and/or evolution into thrust zones [E. Rosencrantz, pers.comm., 1983].

In Figure 6.6, Puerto Rico is shown along the southeast flank of Dominican Republic in the Eocene (similar to the reconstruction of Sykes et al., [1982]). This reconstruction, which restores about 300 km of offset between Hispaniola and Puerto Rico, aligns the Late Cretaceous-Paleogene volcanic axes of the two islands, and places the Bermeja ultramafic complex on the south side of the arc, where it probably was accreted from the south prior to the Late Cretaceous.

The eastward continuation of the San Juan Fault system probably had two splays. A minor splay may have passed through the "Eocene Tectonic Belt" of west-central Puerto Rico, whereas the primary splay passed to the north of Puerto Rico, thereby translating it 300 km eastward from its initial position. The sum of the offsets agrees well with the suggested minimum offset for the San Juan fault system. The eastward extension of the Sierra Neiba fault system, therefore, must have passed south of Puerto Rico, probably along a proto-Muertos Trench transform system that has become compressional with southward vergence like Sierra Neiba [Biju-Duval et al., 1983a]. Further east, this fault probably passed through the sinistral Anegada Passage [Murphy and McCann, 1979] to the Lesser Antilles Trench.

In the Middle Miocene, transcurrent motion shifted from faults in southern Hispaniola to the Oriente Fault and related splays in northern Hispaniola (see chapter 4). The 400 km offset from Cuba could have occurred in the last 11.2 million years, given the spreading history of Cayman Trough [MacDonald and Holcombe, 1978] and extrapolating the 4 cm/yr spreading rate beyond the 8.3 Ma, for which magnetic anomalies have been identified. However, thrusting in Sierra Neiba is Tortonian or younger; the shift to the north therefore may have been gradual, with strike-slip motion occurring both at Sierra Neiba and the Oriente Fault during the late Middle Miocene. An age of 15 Ma seems reasonable for initial strike-slip separation of Hispaniola and Cuba, which matches the age of folding and unconformity in northern Hispaniola (chapter 4).

Southern Caribbean PBZ:

The southern Caribbean plate boundary zone is complex and poorly understood. Thrusting, transcurrent motions, and rifting have been important to its evolution throughout the Cenozoic. In a previous section (collision zones), a case was made for the progressive, diachronous emplacement of the North Venezuelan Nappes onto the Venezuelan margin since the Paleocene, in close association with the eastward advance of the Caribbean Plate. Beginning with a paleogeographic reconstruction of northern South America which accounted for the Late Miocene to Recent fault offsets (chapter 7), it was suggested that the North Venezuelan nappes were emplaced in the Maracaibo area during the Paleocene-Early Eocene, in the Central Venezuelan Mountain System area (Lara Nappes) in the Oligocene?-Miocene, and in the East Venezuelan Basin area in the Late Miocene-Pliocene.

In addition to the emplacement history of the North Venezuelan Nappes, the following points must be recognized by a model for development of the northern Venezuelan margin. First, eastward-dipping subduction of the Caribbean Plate has occurred at the Sinu Trench of western Colombia since the Paleocene-Eocene [Duque-Caro, 1979; 1984]. Second, Oligocene subsidence within the Falcon Basin, and possibly the Bonaire Basin too, was caused by dextral strike-slip motions and attendant pull-apart basin development [Biju-Duval et al., 1983b; Muessig, 1984]. The total east-west offset across the Falcon due to pull-apart opening is unknown. However, the occurrence of Paleogene metamorphosed sediments beneath an Oligo-Miocene sedimentary section of only 6 km that is intruded by alkaline, rather than tholeiitic, basalts [Muessig, 1984] suggests an origin by continental stretching with beta values of about 2 [Dewey, 1982]. Greater degrees of extension are typically associated with the intrusion of tholeiitic basalts. Therefore, dextral offset across the 200 km-long Falcon Basin is probably about half the basin's length, or about 100 km. Third, the Late Miocene to Recent Andean Orogeny has mobilized several blocks in the Venezuelan Borderlands, primarily by stike-slip motions along intervening faults. This deformation is treated in detail in chapter 7 rather than here, but the primary indication of that treatment is that the Andean Terranes and the Borderlands have migrated northeast by 290 km. The northward component of this motion is about 170 km, which is the amount by which the Borderlands have been obducted onto the Caribbean Plate to the west of La Blanquilla.

A crude model for the development of the south Caribbean PBZ which incorporates these considerations is outlined in Figure 6.7. The approximate position of the Caribbean Plate with respect to South America is shown in each reconstruction, as estimated from completing the vector circuit at each time for NOAM/CARIB and NOAM/SOAM. The NOAM/CARIB offsets are inferred from the Cayman Trough's opening history and interpolated within the following schedule of NOAM/CARIB relative motion rates: 0 to 10 Ma, 40 mm/yr; 10 Ma to 20 Ma, 30 mm/yr; 20 Ma to 36 Ma, 25 mm/yr. Integration of these rates over the last 36 Ma yields a total offset for the Cayman Trough of 1100 km. The SOAM/NOAM offsets are derived from Figure 2.5.

This model is highly speculative, but it is the first attempt to incorporate all of the points mentioned above and to acknowledge the flexural history of the Venezuelan foreland. Refinement of the model could come from palinspastic restoration of the thrust belts, and assessment of internal shear and block rotations. Existing paleomagnetic studies indicate that certain blocks within the southern PBZ have been rotated clockwise in association with dextral shear [Hargraves and Skerlec, 1980; Skerlec and Hargraves, 1980; Maze, 1984; Maze and Hargraves, 1984; Hargraves et al., 1984]. In general, the model recognizes these rotations. However, more work with specific tests in mind may further the understanding of shear tectonics as well as the southern Caribbean's evolution.

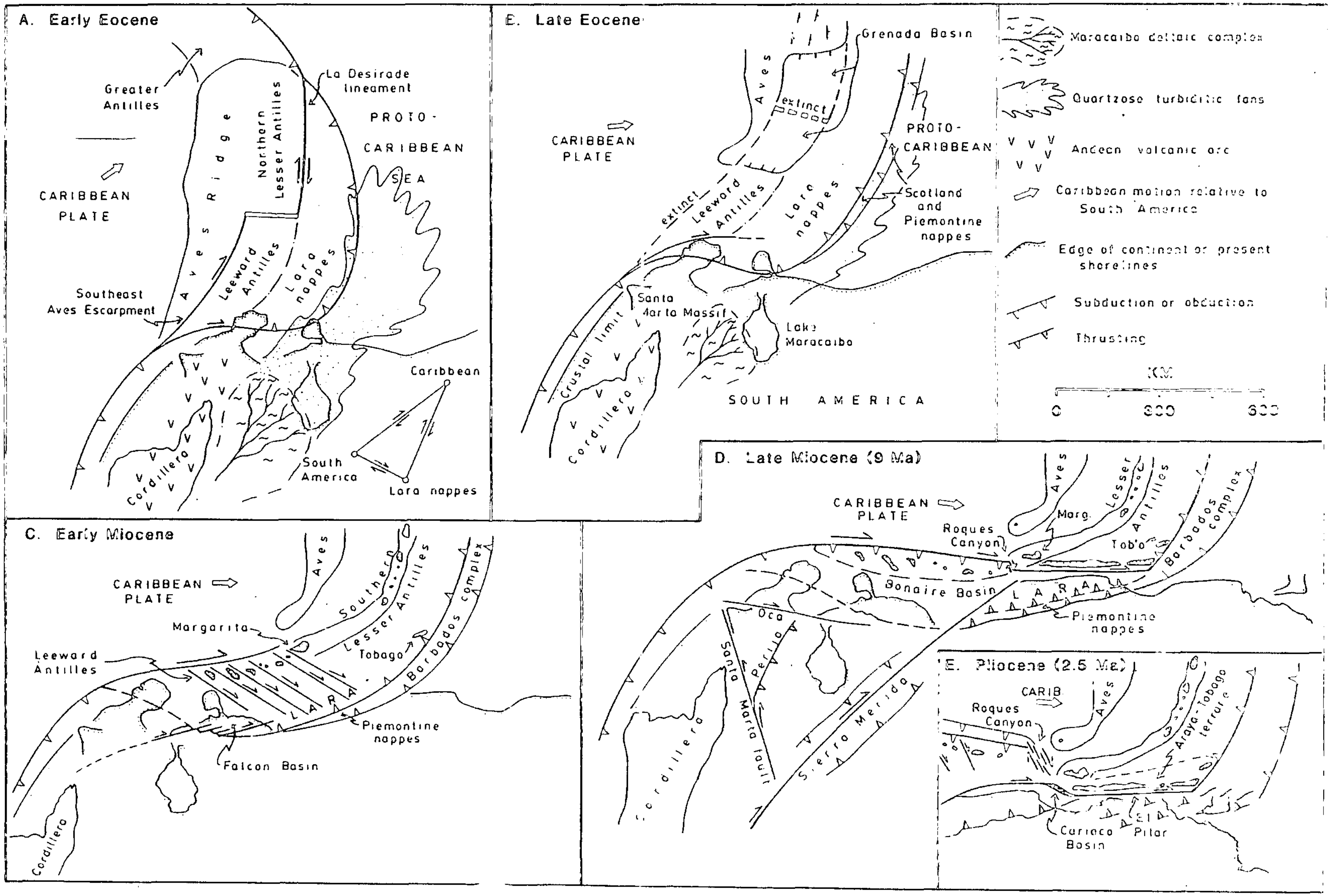


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Figure 6.7. Sequential Cenozoic development of northern South America, described further in text. A: opening of Grenada Basin by north-south back-arc spreading. Arc system along cordilleran Colombia fed clastics to a major drainage basin which emptied into the Maracaibo basin area, which was depressed by the advancing thrust sheets of the leading edge of the Caribbean Plate. Turbiditic flows continued far to the north. Vector triangle shows approximate relative plate motions. B: Eastward migration of the Caribbean Plate accreted the turbiditic flows to the already metamorphosed accretionary complex (Lara Nappes and Villa de Cura Complex). Continued motion progressively emplaced the accretionary complex onto the Venezuelan margin. C: Progressive southeastward emplacement of nappes may have been accommodated by general simple shear within the nappes complex. D: beginning of Andean Orogeny in the west mobilized blocks of the Cordilleran terrane. The Leeward Antilles islands were obducted onto the Caribbean Plate by perhaps 170 km (see chapter 7). E: in latest Pliocene, the Cariaco Basin started to form at a releasing bend; the basin records a dextral strike-slip offset of about 100 km since that time.

CHAPTER 7

NEOGENE BLOCK TECTONICS OF NORTHERN SOUTH AMERICA

NEOGENE BLOCK TECTONICS OF NORTHERN SOUTH AMERICA:

Introduction:

Since the early Late Miocene, much deformation has occurred within northern South America. This has led to the development of a South Caribbean plate boundary zone from the Gulf of Guayaquil to Trinidad, a wide zone of poorly defined seismicity, shear, extension and compression related to interactions between South America and the, relatively, eastward migrating Caribbean Plate. Generally, rocks within the plate boundary zone have migrated northeast and east with respect to South America as indicated by seismic studies [Pennington, 1981] and dextral offsets on many faults [Rod, 1956]. Because of the geological and topographical complexity of the zone, however, the offsets upon many of the fault zones have escaped direct assessment. Some of these offsets are indirectly assessed here, by completing the vector sums of relative motions between plates, platelets of the Caribbean and continental blocks within the plate boundary zone, all with respect to a stable South America (Guyana Shield). Possible tectonic rotations of blocks about vertical poles as suggested by paleomagnetism [Skerlec and Hargraves, 1980] are ignored because the rotations cannot be constrained to the Neogene, and because a clear understanding of these rotations has not yet emerged.

Timing:

Quantifying the northeastward migration of blocks within the Andean and

Venezuelan Cordilleran terranes from Guayaquil to Trinidad is achieved by construction of several velocity triangles between the blocks in question. Alternatively, if an age can be established for the initial motion along the faults separating the blocks and platelets, we need only construct vector triangles of fault offsets since that time.

For the present purposes, an initiation age of 9 Ma is assumed for several reasons. First, magnetic anomaly 5 has an age of 9 Ma and the North America-South America relative motion can be deduced by measuring the difference between the relative positions of these continents at anomaly 5 and the Present. Second, Caribbean-North American offset for the last 8.3 million years has occurred across the Cayman Trough at a rate of 40 mm per year [MacDonald and Holcombe, 1978; Sykes et al., 1982], and very little error is involved in extrapolating the 40 mm/yr total spreading rate to 9 Ma. Third, and most important, is that the primary features of the southern Caribbean plate boundary zone, which collectively comprise the deformation with which we are concerned, have developed within this period. In the east, the opening of the Cariaco Basin (pull-apart) by motion on associated dextral strike-slip faults (El Pilar and Moron) has occurred primarily during the Pleistocene [Schubert, 1982]. To the west, the uplift of the Cordillera Central, Oriental, Occidental, Perija and Merida of Colombia and western Venezuela is well constrained to the Late Miocene and Pliocene (the last 9 to 10 million years). Structural and depositional studies indicate uplift, large-scale thrusting and strike-slip faulting, coarsening of terrigenous sediment, and general emergence during that time [Irving, 1975; Bourgois et al., 1982; Duque Caro, 1972; Campbell, 1968]. Furthermore, paleobathymetric studies of facies and fauna [Bandy and Casey, 1973; Keigwin, 1978] indicate coeval shallowing of the Panama Basin and eventual emergence of the Panamanian Isthmus. The cause of this uplift and deformation is, most likely, the progressive collision of the Panama arc with western Colombia [Pindell and Dewey, 1982; Wadge and Burke, 1983], and the attempted subduction of young, buoyant crust produced at the Galopagos Spreading Center, which includes the Carnegie and Cocos aseismic ridges.

Construction of the vector triangle diagram:

With these considerations in mind, a vector triangle diagram for blocks of the southern Caribbean plate boundary zone may be constructed (Figure 7.1). Each block of the system is portrayed in Figure 7.1 as a point, with pairs of points being connected by tie lines representing the fault zones between the represented blocks. The trends of the tie lines (fault zones) are measured directly from geologic maps, and their lengths are defined by the amount of offset, where known, along each over the last 9 million years. Only those fault zones with proved offsets greater than 50 km have been used.

Both the trend and magnitude of relative motion of NoAm/SoAm and of NoAm/Colombian Basin for the last 9 million years are known fairly well. However, small circles defining the Cayman Trough and the North America/Colombian Basin relative motion [Jordan, 1975] cross cut the Puerto Rico Trench in such a way that one would expect extension. However, the Puerto Rico Trench is a transform fault with a compressional component [Schell and Tarr, 1978]. Furthermore, an unknown amount of convergence at the Muertos Trough [Ladd and Watkins, 1978] suggests even greater disparity between the trends of Cayman Trough small circles and North America/Venezuelan Basin relative motion. It appears that the Venezuelan Basin is behaving independently of the Colombian Basin. Seismic studies [Sykes et al. 1982] indicate present overthrusting by the northern Lesser

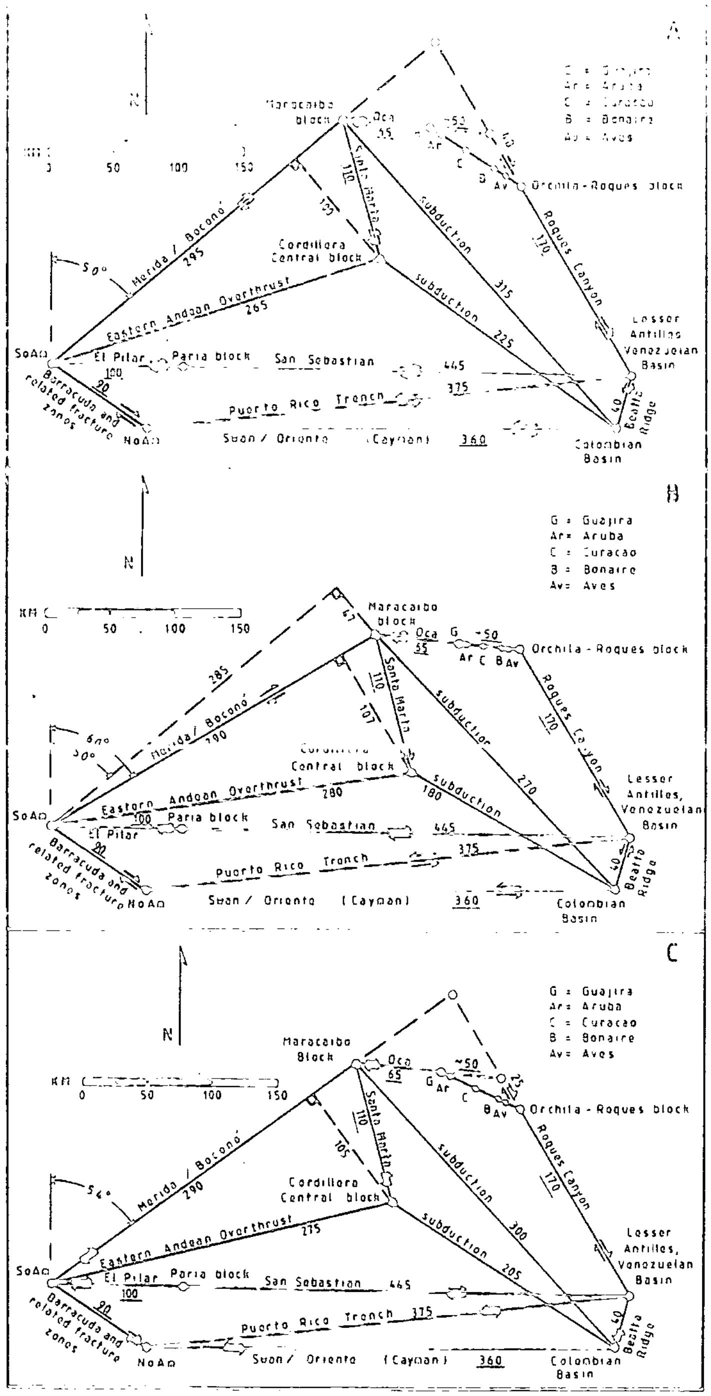


Figure 7.1. Caption on next page.

Figure 7.1. Vector triangle diagrams showing trend and magnitude of relative offsets between blocks of northern South America and Caribbean region, following methods of construction outlined in text. Where necessary, tie lines are chords to small circles of plate motions in the northern South American area. a) Assuming no compression in the Merida Andes. b) Assuming no dextral shear in basins separating Paraguana and Aruba-Orchila islands. c) Preferred vector triangle diagram using values and trends interpolated between the end members of a and c. Offset values in km; underlined values from published sources (below), others inferred from triangle construction. Only published values greater than 50 km are used, with an assumed east-west extension within Neogene basins separating Paraguana and Aruba-Roques islands of 50 km. Dashed line in (a) which connects Central Cordillera block to South America-Maracaibo tie-line is net compressional component in the Andean Cordillera, largely witnessed in Eastern Andean Overthrust. Sources: El Pilar, Schubert [1982]; Oca, Tschanz et al. [1974]; Santa Marta, Tschanz et al. [1974], Campbell [1968]; Cayman system, MacDonald and Holcombe [1978], Sykes et al. [1982]; Barracuda and related fractures, chapter 2; Roques Canyon, Case and Holcombe [1980] and inferred offset.

Antilles across the floor of the Atlantic at N70E. This trend is more northerly than that of the Puerto Rico Trench, and may indicate considerable underthrusting at the Muertos Trough, or may represent only a recently developed direction of Atlantic/Venezuelan Basin relative motion. Assuming an average trend of relative motion between North America and the Venezuela Basin over the last 9 million years that is approximately parallel to the Puerto Rico Trench (N83E), a North America/Venezuelan Basin tie line may be constructed whose trend differs from the North America/Colombian Basin tie line. The likely location for relative motion between the Colombian and Venezuelan Basins is the Beata Ridge. A tie line constructed from the point representing the Colombian Basin that is parallel to the Beata Ridge intersects the North America/Venezuelan Basin tie line at a point which theoretically defines the magnitudes of both the Puerto Rico Trench and Beata Ridge offsets over the last 9 Ma. By triangle completion with the North America/South America tie line, a South America/Venezuelan Basin tie line may be drawn that defines the trend (092) and magnitude (445 km) of the South America/Venezuelan Basin offset. This motion occurred on the San Sebastian off northern Venezuela, although, to the east and west, the motion has been spread across a number of faults, such as the El Pilar and the fault along the northern margin of the Arraya-Paria Peninsula [Case and Holcombe, 1980], whose component motions are of lesser magnitude.

An Orchila/Venezuelan Basin tie line may be constructed with the trend of Los Roques Canyon, which is defined on Case and Holcombe [1980] as a probable zone of dextral shear with questionable magnitude between the two. It is suggested that the magnitude of the offset is 170 km as measured from the northern margin of the Cariaco Basin (eastward continuation of the San Sebastian fault and southern edge of the Caribbean Plate proper) to the northern edge of the Orchila shelf, which is the distance by which the

Venezuelan Borderlands have overthrusted the Caribbean Plate. This is supported by seismic lines [Ladd and Watkins, 1978; Ladd et al., in prep.; Silver et al., 1975] which trace the Caribbean crust dipping beneath the accretionary complex just north of the islands, and also by the large negative gravity anomaly to the north of the islands [Bowin, 1976]. Furthermore, Neogene sediments are highly deformed to the west of Los Roques Canyon whereas to the east they are not [Silver et al., 1975].

At this point, further vector triangle construction may logically follow two different paths, depending upon the amount of shortening that has accompanied transcurrent motions within the Merida Andes. Method one is to assume no compression so that the present trend of the range is the direction of transcurrent motion (Figure 7.1a), and method two leaves the trend of relative motion as an unknown, so that the amount of compression may be deduced from vector triangle construction (Figure 7.1b). In both methods, an internal extension of 50 km is assumed within the basins between Paraguana and the Aruba-Orchila islands, based upon their number, size and post-Late Miocene sedimentary thicknesses [Feo-Codecido, 1971; Silver et al., 1975; Edgar et al., 1971]. In method one, a South America/Maracaibo block tie line of unknown length is drawn with the present trend of the Bocono-Merida system, and a line with the trend of the Roques Canyon tie line is extended beyond the Orchila-Roques block. The eastward displacement of the Orchila-Roques block with respect to the Maracaibo block must equal the sum of the offset upon the Oca Fault and the inter-island basin extension; hence, a tie line with a length of this sum and the trend of the Oca, is drawn in its singular position from the Merida tie line to the Roques Canyon tie line. Figure 7.1a shows that this tie line does not align with the point already defined for the Orchila-Roques block. Therefore, a dextral component of 48 km is inferred to have accompanied the extension within the inter-island basins. Method two constructs the Maracaibo/Orchila-Roques tie line directly west from the previously defined Orchila-Roques point, thereby eliminating the dextral shear implied by method one. Its western end, therefore, defines the position of the Maracaibo block, and construction of a South America/Maracaibo tie line defines the offset of Maracaibo block relative to South America. The transcurrent and compressional components within the Merida-Bocono system may then be deduced by measuring the difference between this constructed trend and the actual trend of the present day system, as shown in Figure 7.1b.

Once the main frameworks have been created, hypothetical tie lines may be drawn between any pair of points in either diagram to obtain the approximate trend and magnitude of the relative motion between the represented pair of blocks. For example, a tie line drawn from the Colombian Basin to the Maracaibo block defines the amount and trend of subduction during the last 9 Ma beneath the latter. The predicted average convergence trend over the last 9 Ma of N44W accords reasonably with the present convergence trend of N50W plus or minus 10 [Kellogg and Bonini, 1982]. Alternatively, the components of motion between blocks, i.e. transcurrent, extension and compression, may be approximated in other cases by comparing the constructed tie lines defining relative motion and the actual trends of the faults upon which the motion has occurred. This was done in method two (Figure 7.1b) to obtain the transcurrent and compressional components of the Merida system. Similarly, the compressional component of the Eastern Andes Overthrust can be determined by construction of a line normal to the Merida tie line and intersecting the Central Cordillera block (Figures 7.1a,b,c). Sinistral motion along the Santa Marta Fault and dextral motion along the Merida Andes has allowed the Maracaibo block to escape much of the compression seen in the Colombian Andes.

Dextral slip within the five basins separating the Paraguana Peninsula and the Aruba-Orchila islands is postulated due to their residence in the southern Caribbean dextral shear zone, which argues in favor of method one. However, a compressive component within the Merida Andes is well known [Shagam, 1975]. Thus, we suggest that the best vector triangle diagram for the southern Caribbean plate boundary zone is one that falls about half way between the end member constructions of methods one and two, which is depicted in Figure 7.1c. Figure 7.2, which depicts the pre-Andean paleogeography of the blocks in question, has been constructed from values and trends of Figure 7.1c.

Discussion:

The primary implication of this semi-quantitative analysis is that the Cordilleran Terrane of Ecuador, Colombia and western Venezuela has migrated about 290 km to the northeast with respect to a stable cratonic South America in the last 9 million years. Dextral offsets of 250 meters have been measured upon Quaternary sediments along the Bocono Fault [Giegengack et al., 1976)], and 33 km can be measured at a dextral offset of Rio Bocono [Rod, 1956], but no total transcurrent offset within the Merida Andes, of which the Bocono is only a presently active, relatively minor fault, has been postulated before. Studies of relative motions between South America and the oceanic plates offshore to the north and west should recognize the independent motion of this platelet.

Figure 7.3 outlines the complex interactions between the blocks of northern South America and the Caribbean Plate during the northeastward migration of the Cordilleran terrane. Simply put, the leading edge of the

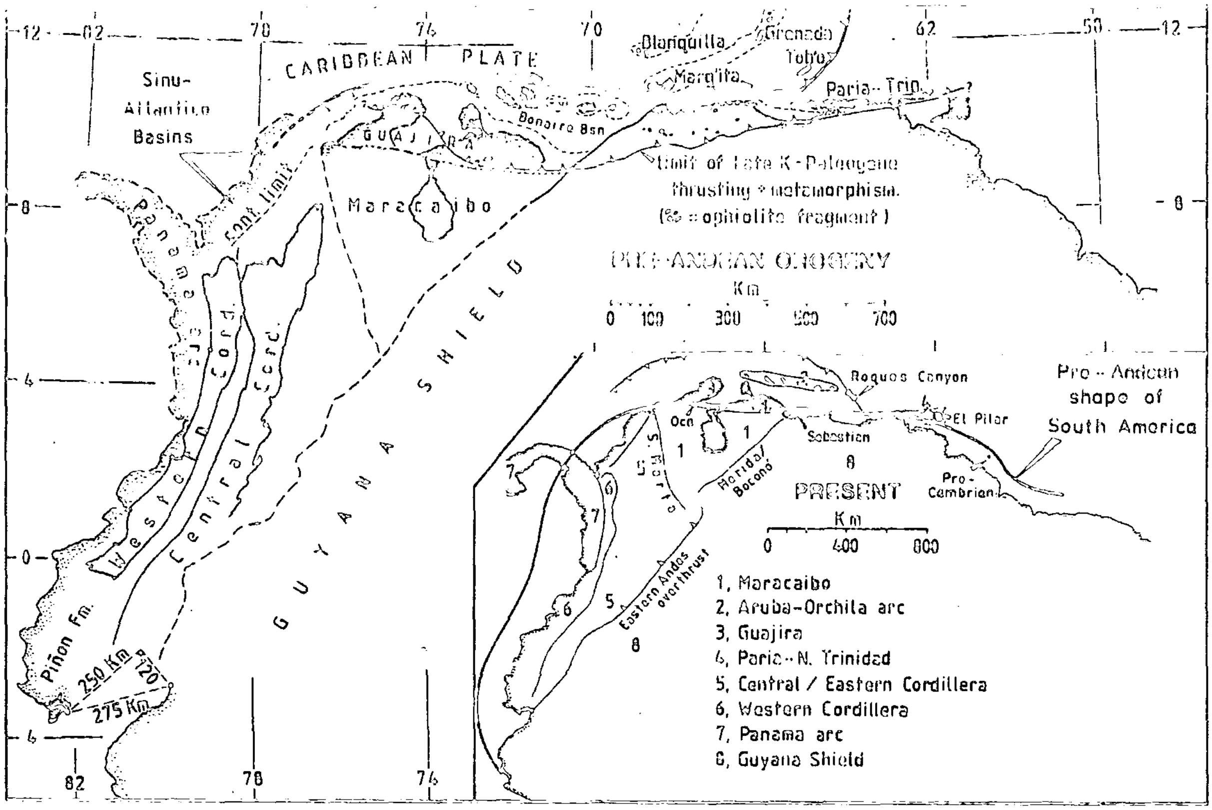


Figure 7.2. Pre-Andean (~9 m.y.b.p.) plate reconstruction of blocks comprising northern South America, removing offsets, relative to South America, defined in figure 7.1c. Shorelines of Panama and Sino-Atlantico Basins dashed because they are Pliocene-Recent additions. Proposed reconstruction produces a straight, continuous limit of pre-Late Miocene thrusting and metamorphism, and aligns Late Cretaceous arc related plutons of Aruba-Orchila islands with those of onshore eastern Venezuela, although the latter may be anatectic in origin. Thrust sheets include sporadic remains of ophiolitic fragments, as shown. Because Blanquilla presently lies east of Roques Canyon fault zone, it is considered part of Aves Ridge that hasn't collided with northern South America. No break between Margarita and the Lesser Antilles is indicated by geophysical data [Weeks et al., 1971]. Rocks similar to those on Margarita may form basement of the southern end of Lesser Antilles arc. Ophiolite fragments in Margarita may have been tectonically extruded during the more than 1200 km of shearing between Caribbean and South American plates prior to the plate boundary's jump to El Pilar Fault Zone in Pleistocene time. The Late Cretaceous plutons of Tobago at the leading edge of Caribbean Plate indicate that a "Tobago block" has been obducted at some point during migration of the Caribbean Plate (G. Westbrook, personal communication, 1984). The vector triangle at the Gulf of Guayaquil defines strike-slip and compressional components of the Cordilleran block relative to Guyana Shield. Inset: Present outline of South America, defining blocks and fault zones used in this analysis. Heavy line defines pre-Andean (pre 9 Ma) continental shape.

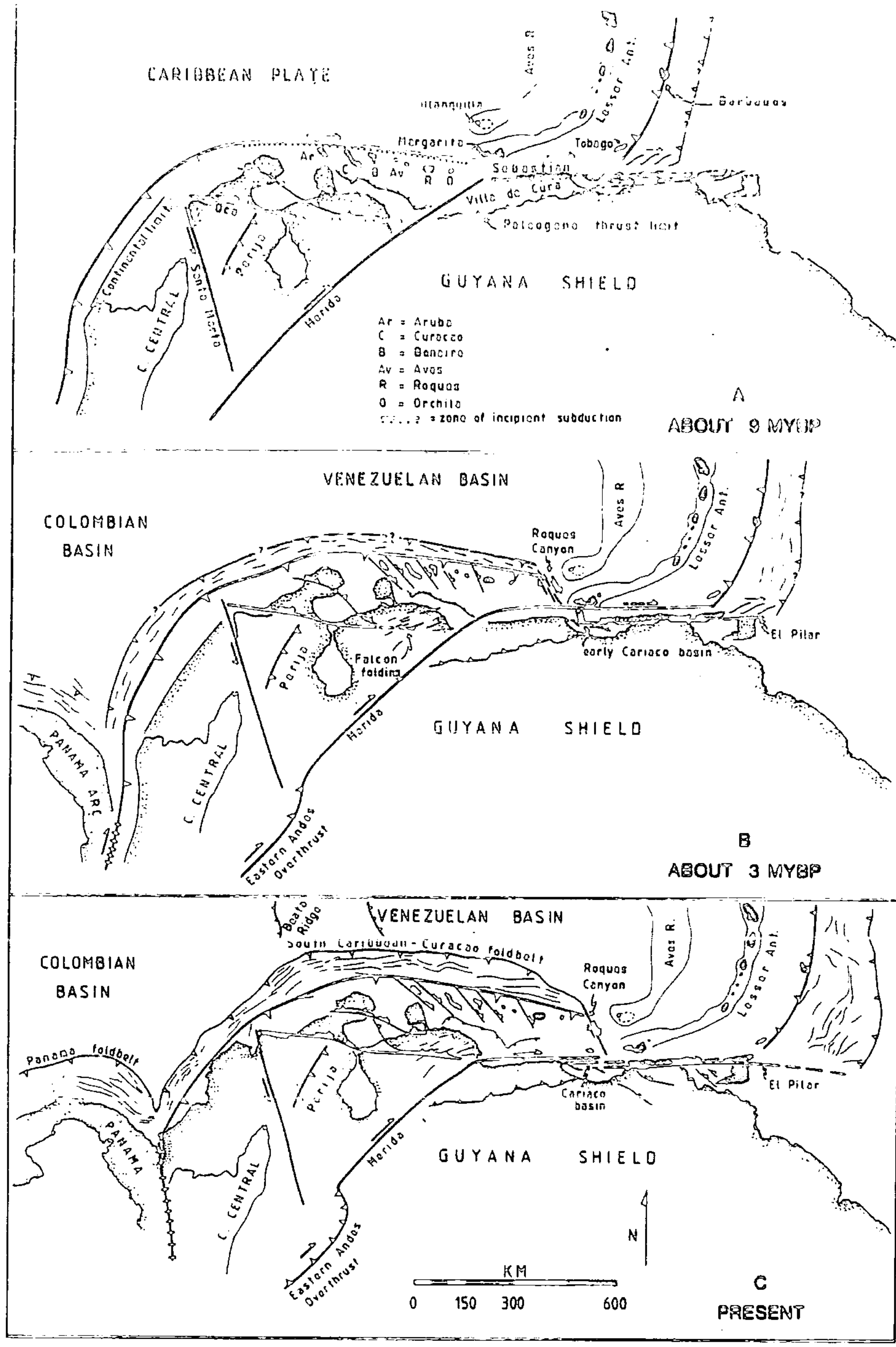


Figure 7.3. Schematic evolution of northern South America over the last 9 million years. A, B and C are, respectively, paleogeographic reconstructions at 9 Ma (pre-Andean), 3 Ma (pre-Cariaco Basin/El Pilar fault development), and Present. Reconstructed offsets in A are from figure 7.1c, those in B are interpolated between A and C. Caribbean/South America transform in A is the zone of incipient subduction, the site from which the Venezuelan Borderlands have been thrust onto the Caribbean Plate in Neogene time. Strike-slip faults within the Cordillera Central are ignored; their net offset over the last 9 Ma is inferred in Figure 7.1. Some of the motion upon the Oca and Santa Marta faults has been accommodated by thrusting in the Perija Andes, and the eastern continuation of the Oca may die out within decollement zones within the folds of the Falcon Basin. Progressive development of the Barbados, South Caribbean/Curacao and Panama accretionary complexes is inferred.

terrane has overridden the eastward migrating Caribbean Plate. However, because the Merida/Bocono fault zone changes trend so drastically as it merges with the San Sebastian fault zone, the Aruba-Orchila island arc has been internally deformed, resulting in the formation of the small basins lying between the islands. Dextral and extensional motion within these basins has prevented a gap from opening between the islands and northern Venezuela. Collectively, these motions have led to the obduction of the Aruba-Orchila islands onto the Caribbean Plate by about 170 km. The term obduction is preferred to subduction, owing to the buoyant nature of the Caribbean Plate [Burke et al., 1978] and to the fact that no subduction-related volcanism has yet occurred. It has, however, caused the development of the extensive South Caribbean-Curacao Ridge accretionary complex. The complex is composed of continental slope and rise sediments derived from South America and deposited upon the Caribbean Plate during its Cenozoic eastward migration. The site upon which the Cordilleran obduction was initiated was the original transform fault separating the Caribbean and South American Plates (Figures 7.1 and 7.2). Obduction was facilitated by the fact that the Caribbean crust sits isostatically at 4 to 5 km below the adjacent margin. Hence, this transform, and evidence for its large offset since the Eocene, cannot be directly seen. Only the eastern, recently developed, relatively minor portion of the South American-Caribbean offset (Cariaco Basin, El Pilar and related faults) can be directly studied [Speed, 1985].

The prospects for petroleum discovery in the basins separating the Aruba-Orchila basins are poor. At the surface, these basins have an appearence similar to the small productive basins offshore California. However, because the crust in which they formed overlies the underthrusting Caribbean Plate rather than mantle, the basins may be expected to be quite cold, with little chance of their sediments having reached thermal maturation during their short lives of only 5 to 8 million years.

CHAPTER 8

GEOLOGIC-KINEMATIC MODEL OF CARIBBEAN EVOLUTION

GEOLOGIC-KINEMATIC MODEL OF CARIBBEAN EVOLUTION:

Introduction:

A Caribbean evolutionary model is outlined in 11 paleo-reconstructions with accompanying text, incorporating plate-tectonic elements and interpretations described in chapters 4-7. The model continues the Gulf of Mexico evolutionary story presented in chapter 3, beginning in the Berriasian. The initial shape of northern South America is that obtained by restoring Neogene fault offsets defined in chapter 7, and by removing the Western Cordillera, eastern Costa Rica-Panama Arc, and the Leeward Antilles islands, until their times of respective accretion.

Vector triangle diagrams (chords to small circles of rotation) accompany each reconstruction to show the magnitude and direction of motion, relative to North America, of the plates until the next reconstruction. The relative motion data of the larger plates are from chapter 2, and motions of the Caribbean "Plate" are deduced from the reconstructions. For the Neogene, Caribbean motion is also constrained by the spreading history of the Cayman Trough [MacDonald and Holcombe, 1978].

The proposed evolutionary model: (figures at end of chapter)

Berriasian, 140 Ma, Figure 8.1.

Following the opening of the Gulf of Mexico, the Yucatan block became part of the North American Plate (NOAM) as spreading ceased in the Gulf. Starting in the Berriasian, additional NOAM-Gondwana plate separation occurred solely within a "proto-Caribbean Sea" of the Atlantic. Most margins of the proto-Caribbean Sea had been formed by rifting, and predominantly carbonate deposits accumulated on the subsiding shelves throughout the Cretaceous. The proto-Caribbean ridge system connected the Central North Atlantic ridge system to plate boundaries in the Pacific realm. The latter are poorly known, but sinistrally oblique subduction of the Farallon Plate beneath North America is indicated [Engebretson, 1982]. Early Cretaceous calc-alkaline volcanism in the Chortis, Niçaragua Rise-Jamaica, and Cordillera Central (of Colombia) arcs indicate subduction along the Pacific margin. Hence, it is unlikely that the proto-Caribbean ridge system continued out into the Pacific; it most likely merged in an unclear way with the subduction zones at the Pacific-proto-Caribbean interface.

From M-21 to M-11 time (Tithonian to Valanginian), a kink in Central Atlantic flow lines (Figure 2.5) indicates that the Central Atlantic's pole position shifted slightly to the west at that time. Such a shift theoetically should have caused compression at existing long left-lateral transforms in the Atlantic spreading system. The geometry of the early Atlantic suggests that the Bahamas probably was the site of such a transform. It is suggested that this compression uplifted the oceanic crust in the area of the eastern Bahamas to the photic zone, and is responsible for the Early Cretaceous initiation of shallow-water carbonate deposition.

Early Aptian, 119 Ma, Figure 8.2.

Spreading and plate divergence continued in the proto-Caribbean throughout the Early Cretaceous. Development of the Bahamas was structurally controlled by the Central Atlantic spreading system. The motion of the Farallon Plate relative to North America became convergent by Aptian time (Figure 2.1), and the previously sinistral trench/transform system along the Pacific realm lengthened and became the site of Greater Antillean subduction. Metamorphic ages of 127 Ma on basic basement rocks in the Greater Antilles [Pardo, 1975; Bowin, 1975] may relate to the Early Cretaceous transform phase or to the inception of subduction. Initial polarity of the arc was probably southwest-facing, like the arc complexes along western Mexico and South America, so that the early Greater Antilles arc remained at the Pacific--proto-Caribbean interface, subducting Farallon crust. The plate-boundary junction with the Central Atlantic Ridge is unknown, but may have been connected to the Pacific trench systems by a transform, much like the situation in the present Andaman Sea.

To the east, plate separation began in the Equatorial Atlantic between northern Africa and SOAM, producing for the first time a three-plate system of the Americas and Africa. This development, however, did not drastically affect NOAM-SOAM relative motion (Figure 2.5).

Late Albian, 98 Ma, Figure 8.3.

Carbonate platforms continued to develop along the margins of the proto-Caribbean, while arc activity continued along the Greater Antilles and Pacific margins of Mexico, Chortis and SOAM. The Bermeja ophiolitic complex probably was obducted from the Farallon Plate and accreted to the southern margin of the Greater Antilles at about this time [Mattson, 1979], but it may simply be an exposed portion of the oceanic basement of the Greater

Antillean arc system. The oceanic crust of the present Caribbean Plate, already created at Pacific ridge systems, was migrating toward the Americas as part of the Farallon Plate. Basalts beneath seismic horizon B" in the Venezuelan and possibly Colombian Basins apparently were extruded onto this crust during Cenomanian-Santonian time, giving the Caribbean crust its thick, buoyant nature [Burke et al., 1978]. Seismic velocities beneath B" are intermediate between those of sediment and igneous rocks [Ludwig et al., 1975; Stoffa et al., 1981] and may indicate interbedded sediment and lava. Off-axis volcanism, perhaps related to formation of seamounts as in the western Pacific, seems to have added this material to the pre-existing Caribbean seafloor. A speculative cause for the build up of heat which led to the widespread extrusion event is the Aptian to Albian period of little relative motion between the Farallon Plate and the mantle (see chapter 2).

Early Campanian, 84 Ma, Figure 8.4.

In the Early Campanian, relative motion between NOAM and SOAM effectively ceased due to congruency of pole positions and rotation rates defining the opening of the Central and South Atlantic Oceans. Hence, the ridge system in the proto-Caribbean probably became extinct, although it is possible that a subduction zone formed within the proto-Caribbean whose subduction rate and direction were equal but opposite to the rate and direction of seafloor spreading. The chances for this coincidental scenario developing, however, are seen as slim.

The oceanic crust that had been injected and flooded by basalt in the medial Cretaceous (B") began to arrive at the American and early Greater Antillean subduction systems in the Campanian. Along the Andean arcs of

Mexico, Chortis and SOAM, hindered subduction of the buoyant masses of crust led to Laramide orogenesis (hinterland thrusting) and to step backs of the subduction zone which obducted and accreted portions of Farallon crust to American forearc regions. Examples of such obductions include the basement of Western Cordillera of Colombia, the Pinon Formation of Ecuador, and possibly the ophiolite drilled on DSDP Leg 67 off Guatemala.

Along the Greater Antilles intra-oceanic arc, however, the buoyant crust choked the subduction zone, produced orogeny, and caused a polarity reversal or flipping of the arc. Remnants of the buoyant masses which choked the north-dipping subduction zone may be the basaltic basements of the southern Peninsula and San Juan block of Hispaniola. These areas possess no evidence of Cretaceous magmatism that is clearly subduction-related, and where it is exposed in the southern peninsula, basement was extremely deformed prior to the Maestrichtian or Paleocene. South-dipping subduction was initiated north of the Greater Antilles, probably during the Campanian, and crust of the present-day Caribbean Plate was allowed to enter the proto-Caribbean area, with the Greater Antilles at its leading edge. The new south-dipping subduction zone may have been initiated upon one of the pre-existing fracture zones of the proto-Caribbean, which were approximately parallel to the trench.

Subduction may have also began at about this time along the Pacific side of B" affected crust, whose volcanism upon the western edge of the Caribbean Plate has produced the southwest-facing Panama-Costa Rica arc. The continued deposition of carbonates on the rifted margins of the proto-Caribbean and no sign of volcanism (except for an odd intrusion in north-central Yucatan Peninsula) suggests that the active volcanism of the leading edge of the Caribbean Plate (Greater Antilles) was still located outboard of the proto-Caribbean. This, in turn, indicates that the trailing edge of the Caribbean Plate (Panama-Costa Rica) was located further out in the Pacific, as are the Tonga-Kermadec and Mariana arcs today. A cause-and-effect relationship perhaps exists between the compression resulting from the choking of the Greater Antilles arc and the inception of subduction at the Panama-Costa Rica.

Polarity reversal in the Greater Antilles arc must have been structurally and spatially accomodated within the arc; most island arcs are convex towards their facing direction, and a change in polarity implies a corresponding change in arc-convexity. The change is probably responsible for several geological relationships in the Greater Antilles. First, Campanian through Paleocene unconformities are widespread throughout the Greater Antilles, and Late Cretaceous exchange of terrestrial fauna between North and South America accompanied this phase of uplift [Bonaparte, 1984]. The Greater Antilles apparently provided an effective land bridge between Mexico and South America. Second, some areas of the Greater Antilles became volcanically inactive following reversal (Eastern Cordillera, Dominican Republic), whereas others became active (Aves Ridge). This observation relates to the termination of one plutonic axis and the initiation of another, within the same general arc terrane. Third, several radiometric ages on blueschists in Cuba and in northern Dominican Republic point to Late Cretaceous (Campanian) [T. Barros, pers. comm., 1985; Joyce, 1983]. The blueschist occurrences may relate to rapid uplift from subduction-zone choking in the south and also to subduction initiation in the north.

At this point, it is worth mentioning that the number of individual arc terranes, which formed at distinct subduction zones, that are present in the Greater Antilles arc complex is still unresolved. At least three major discontinuities exist which could be interpreted as potential sutures which closed at different times. These are the Cauto Basin of Cuba (Paleogene), the Loma Caribe Peridotite of Hispaniola (Late Cretaceous), and the "Eocene Tectonic Belt" of Puerto Rico (Eocene). However, they may also be interpreted simply as zones of extensional, compressional, or strike-slip displacement within a single original arc system. The Loma Caribe is of particular relevance to the Campanian, as this is when it and related structures primarily developed [Draper and Lewis, 1980]. The Loma Caribe is a linear belt of ultra-mafics that defines more or less the northern margin of the Cordillera Central magmatic arc of Hispaniola [Bowin, 1975]. No associated melange or blueschists typically associated with subduction are known in the area, although the development of the adjacent Amina Schist probably is closely associated [Draper and Lewis, 1980]. It appears most likely that the Loma Caribe formed by compression with possibly a strike-slip component, thereby raising a linear zone of arc-basement to the surface, as an accomodation structure during the period of arc-reversal. The uplift of the Loma Caribe and the development of the Amina Schist probably relate to backthrusting of rear-arc sediments during the reversal, rather than to subduction accretion during actual subduction. As for the other major arc dislocations, it was outlined in chapter 6 that the Cuban Cauto Basin is the pivot point for the opening of the Yucatan intra-arc basin, and that the Puerto Rican Tectonic Belt is a zone of strike-slip shear related to development of the north Caribbean plate boundary zone. Hence, the Greater Antilles is viewed here as a single arc complex which possesses no sutures defining the existence of former oceans.

Late Campanian, 72 Ma, Figure 8.5.

Northward migration of the Greater Antilles with respect to North

America by Late Campanian time is indicated by the onset of the northward-verging collision between an arc terrane and the previously south-facing stable carbonate shelf of southern Yucatan (see chapter 6). Continued convergence led to, by the Maestrichtian, the oblique obduction of the Santa Cruz and other ophiolitic bodies into the Sepur foredeep of the south-facing shelf of Yucatan (Motagua Suture Zone). The Santa Cruz may have been part of the forearc of the a western extension of the Cuban arc complex. The majority of the Greater Antilles arc, which had avoided collision with Yucatan, continued to migrate northeastward, toward the Bahamas.

In northwest South America, most of the oceanic crust of the Western Cordillera of Colombia and Ecuador had been emplaced by this time, and South America-Caribbean relative motion occurred outboard of the Western Cordillera at a new trench system.

Also important at this time was the formation of the Aves Ridge arc. As the eastern Caribbean Plate advanced northeastward beyond SOAM, the Greater Antilles trench progressively lengthened, and subduction at the new portion of the trench is responsible for the intrusion of plutons into the eastern margin of the Caribbean Plate (Aves Ridge). This subduction also is responsible for the emplacement of intermediate plutons of the Leeward Antilles arc. The Aves and Leeward Antilles "arcs" became separated probably in the Eocene by north-south intra-arc spreading in the Grenada Trough (chapter 6).

In the Costa Rica-Panama arc system, there is sufficient evidence to conclude that subduction was underway by the Late Campanian (Figure 6.3). Hence, by this time the Caribbean Plate was defined and was capable of independent motions relative to the other plates in the system.

Late Paleocene, 59 Ma, Figure 8.6.

The Caribbean Plate continued migrating approximately to the east-northeast, subducting oceanic crust of the proto-Caribbean. Yucatan Basin opened by intra-arc spreading, or extreme attenuation of Greater Antilles arc crust, between Cuba and Cayman Ridge; extension was northwest-southeast [Hall and Yeung, 1980]. The bulk of the Cuban portion of the Greater Antilles magmatic arc was left in the Cayman Ridge, or is presently submerged beneath Cuba's southern shelf. The Cretaceous volcanic assemblages of onshore Cuba may represent the forearc complex only, so that rifting occurred nearly along the arc-forearc interface. Trends and magnitudes of relative motions in the three-plate system NOAM-Caribbean-Cuba that describe the opening of Yucatan Basin may be defined by vector completion (Figure 6.4, inset).

The Yucatan block prevented simple east-northeast motion of Chortis and Nicaragua Rise/Jamaica with the rest of the Caribbean Plate; the sinistral shear-couple deformed the latter internally during the Paleogene (Wagwater, Montpellier Troughs, rifts of Nicaragua Rise) [Mann and Burke, 1984a]. Chortis-Caribbean relative motion may have occurred primarily along the Hess Escarpment, accompanied by compression in the Nicaraguan Rise, so that by Eocene time the Chortis and Costa Rica-Panama subduction zones aligned to form the Middle American Trench. Hence, volcanism ceased in the eastern Chortis and the Nicaragua Rise-Jamaica arcs by the Eocene. The original northeastward extension of the Hess may have traced between Jamaica and the southern peninsula of Hispaniola; the Cretaceous histories of the two are very different, and their present proximity may be due to strike-slip juxtaposition by motion on the Hess. Subsequent Cenozoic strike-slip within and along the south side the southern peninsula of Hispaniola possibly has obscured this initial relationship. Restoration of about 150 km of offset there realigns the Hess with the gap between Jamaica and southern Hispaniola, and provides an possible explanation for the significance of the problematic Hess Escarpment. As much as 48 km has occurred only since 2.2 Ma [Sykes et al., 1982; Mann et al., 1984]. About 100 km of earlier Tertiary (Early Eocene?) motion is postulated to have occurred along the southern peninsula.

In northern SOAM, obduction of the western equivalents of the North Venezuelan nappes (forearc complex of the Leeward Antilles arc) onto the South American margin was beginning, leading to load-induced subsidence and flysch sedimentation with a continental source in the Maracaibo area (Figure 6.7). This was accompanied by the initial development of the Grenada Basin (Figure 6.5). Late Cretaceous deformation and metamorphic ages from the North Venezuelan nappes relate to the accretionary prism stage of their development and to partial collision with the Western Cordillera, when the nappes were located to the west of northern South America, and not to the actual emplacement onto the Venezuelan margin.

The opening of both the Yucatan and the Grenada intra-arc basins was the mechanism by which the Caribbean Plate accomodated the shape of the proto-Caribbean basin, after having entered from the Pacific through a smaller gap. Intra-arc basin formation was probably initiated by subduction zone roll-back or trench pull; it seems that it was kinetically "easier" to rift the arc complexes than to tear new transforms, or "railroad tracks," into the proto-Caribbean oceanic crust.

Middle Eocene, 49 Ma, Figure 8.7.

The Middle Eocene is marked by the termination of collision between the Greater Antilles and Bahamas Platform. This reconstruction is constrained by 1) restoring 1100 km of offset in Cayman Trough, 2) alignment of Late Cretaceous-Eocene subduction-related plutons throughout the Greater Antilles and Aves Ridge, and 3) sedimentary facies changes across the islands (Figure 6.6). Extension in Yucatan Basin ceased as Cuba came to rest against the Bahamas. Volcanism shifted eastwards from the Aves Ridge at this time, and ensued in the Lesser Antilles arc. It began in the Eocene in the northern Lesser Antilles, but lagged until ?Oligocene or Early Miocene in the south. This may relate to the opening of Grenada Basin, which accounts for a lesser subduction rate in the south during the Eocene.

In the Middle to Late Eocene, just after the Antilles-Bahamas collision, east-northeast migration of the Caribbean Plate relative to NOAM continued along a new plate boundary system. This system became the northern Caribbean plate boundary zone (northern PBZ, Figure 6.6), in which the Cayman Trough nucleated by the Oligocene as a pull-apart basin between Yucatan and Jamaica. Cuba, the Cayman Ridge, and the Yucatan Basin were left as a part of the North American Plate with the development of this PBZ. Progressive development of the Barbados accretionary prism is due to this plate migration. Quartzose sands of the Piemontine nappes of central Venezuela and the Scotland District of Barbados were accreted to the arc complex in the Eocene, and probably originated from western Venezuela where Precambrian acidic massifs were exposed at the time.

By the Middle or Late Eocene, the trench systems the Chortis and Costa Rica-Panama arcs were aligned, and the Middle American Arc developed above a single subduction zone. Also, about 100 km of motion along the southern side of the southern peninsula of Hispaniola is postulated to have offset the Hess Escarpment from the gap between Jamaica and the Southern peninsula.

Early Oligocene, 36 Ma, Figure 8.8.

NOAM-SOAM relative motion since the Eocene has been minor when compared to motion of the Caribbean with respect to the Americas. Where the North and South American Plates are in contact along one or more fracture zones to the east of the Lesser Antilles Arc, relative motion of 200 to 300 km has been dextrally convergent (Figure 2.5), but a well-defined North America/South America plate boundary does not exist at present.

Eastward migration of the Caribbean Plate relative to NOAM and SOAM following the Eocene is indicated by subduction-related magmatism in the Lesser Antilles. In the northern PBZ, the Cayman Trough progressively developed by seafloor spreading at the Mid-Cayman Spreading Center, which linked transform faults that connected the Middle America and Lesser Antilles subduction zones. These transforms have wandered in the past, forming anastomosing fault systems across central Guatemala, in the west, and Hispaniola/Puerto Rico in the east (Figure 6.6). Large offset transcurrent motions between blocks of Hispaniola are indicated by drastically differing Tertiary sedimentary facies presently juxtaposed across fault zones. The primary offset during the Late Eocene and Oligocene occurred along the northern San Juan Boundary Fault, which juxtaposed the San Juan block with the Central Cordillera arc by early Miocene time, as indicated by flooding of arc-derived clastics into San Juan Basin [Michael, 1979; Cooper, 1983]. Motion upon eastward extensions of this fault system separated Puerto Rico from central Hispaniola, and brought an arc-derived fragment of southwestern Puerto Rico into juxtaposition with central Puerto

Rico (Eocene Tectonic Belt).

In the southern PBZ, eastward migration of the Caribbean progressively lengthened the zone of Caribbean-SOAM interaction. Thrusts of the North Venezuelan nappes have been emplaced southeastwards onto the Venezuelan margin since the Eocene, diachronously to the east. Subsidence of the Venezuelan shelf near Maracaibo is Paleocene-Eocene [Zambrano et al., 1972], and subsidence in the East Venezuela Basin is Miocene-Pliocene [Vierbuchen, 1984]. This thrusting was followed by shear tectonics at any given point along the zone of interaction between the Caribbean and SOAM. A transform between the Caribbean Plate and South America progressively developed after thrust emplacement. The Falcon (Oligocene) and Cariaco (Pleistocene) Basins are two examples of pull-apart basins developing in previously overthrust areas.

Early Miocene, 21 Ma, Figure 8.9.

Migration of the Caribbean Plate and associated plate boundary zone development continued. The predominant zone of motion through Hispaniola during the Miocene was along Sierra Neiba; by the Late Miocene, uplift in Sierra Neiba had structurally separated the San Juan and Enriquillo Basins. The logical eastward continuation of this fault system is the Muertos Trough, which recently has become an overthrust zone.

In the south, the Lara, Villa de Cura, and Piemontine nappes were emplaced onto the Venezuelan shelf, overthrusting the Oligocene-Miocene foredeep basin.

Out in the Pacific, spreading was initiated at the Galapagos spreading center, although its eastern portions have already been subducted and, therefore, its relationship to the Caribbean plate boundary circuit is unknown, and not shown.

Late Miocene, 9 Ma, Figure 8.10.

By this time, the blocks of Hispaniola were nearly assembled. For the first time, siliceous gravels derived from a plutonic source in the Cordillera Central reached the Fondo Negro Basin (Trinchera Formation) of the southern peninsular block [Cooper, 1983]. The Cayman Trough was well-developed and still extending, and the Panama arc had begun to collide with Colombia. The arc terranes of Cuba and Hispaniola were probably separating by this time by transform motion along the present-day Oriente Fault. Compression has dominated further development in the Sierra Neiba/Enriquillo systems, contributing to the present-day complexity of the northern PBZ. Seafloor spreading at Galopagos Ridge of the eastern Pacific since the Early Miocene had created a young, buoyant swath of oceanic crust with two aseismic ridges of excess volcanism, which entered the Panamanian and Ecuadorian subduction zones in the Middle and Late Miocene. The arrival of Panama and this buoyant crust has partially choked the subduction zones, thereby driving blocks of the Andean Cordillera northeastwards, from the Gulf of Guayaquil to the Venezuelan Andes. This mobilization has played a major role in the neotectonic development of the Caribbean.

Present, Figure 8.11.

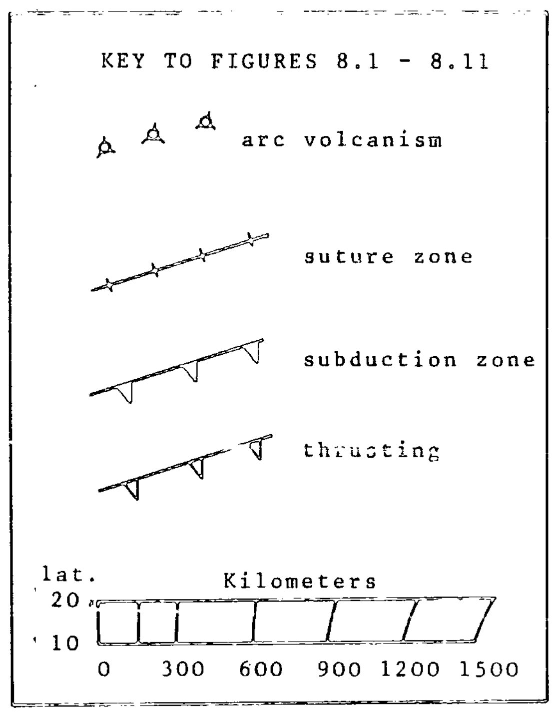
Over the last 9 to 10 million years, the Caribbean region, primarily the northern and southern Caribbean PBZs, has undergone severe deformation. The term "Neo-Caribbean Phase" is proposed to describe this Late Miocene to Recent period of Caribbean tectonism, as it continues today.

Neo-Caribbean tectonism is widespread. Throughout the northern PBZ, it is characterized by sinistral transpression and development of pull-apart basins along many sinistral strike-slip faults. Deformation is most severe in Hispaniola, where elevations exceed 3,000 meters and Pleistocene reef debris has been found at 650 meters [Cooper, 1983]. Also, drastic relief and active seismicity through the Motagua Valley of Guatemala indicate considerable plate motions there too. The eastern Cayman Trough and the Venezuela Basin have been recently overthrusted by Jamaica and an accretionary complex at Muertos Trough, respectively. Both compressional regimes are accompanied by sinistral components of strike-slip motion.

In the southern PBZ, Neo-Caribbean tectonism is more complex and variable. Like a mirror of the north, dextral strike-slip faults with transpressional and transtensional segments are common, and the Venezuelan and Colombian Basins have been overthrust by the South Caribbean and Panama accretionary complexes, respectively. However, these aspects merely accompany developments of greater significance. Onshore northern SOAM, many linear fault systems and mountain chains extend from the Gulf of Guayaquil to Trinidad. Some of the faults possess known offsets, whereas offsets on others and within some linear mountain belts are unknown, although strike-slip motions are usually indicated by seismic studies [Aggarwal, 1983]. In chapter 7, a 9 Ma (pre-Neo-Caribbean Phase) paleogeographic reconstruction of various blocks was derived by restoring known offsets and other displacements as determined by vector triangle completion. The Cordilleran Terrane of northwest SOAM has migrated 290 km northeastward relative to Guyana Shield during the last 9 Ma. This and associated motions have dominated development of the southern Caribbean PBZ. The cause of this motion presumably is the Late Miocene choking of the Atrato subduction zone by the Panama arc and the buoyant crust and aseismic ridges created at Galapagos Ridge (Figure 6.1). Associated with these translations is the morphologic change of eastern Panama as outlined by Wadge and Burke [1983].

Within the Caribbean Plate itself, uplift and possibly strike-slip motion are occurring at the Beata Ridge and Hess Escarpment, but few details are known. These features are probably sites of differential motion between the Venezuelan and Colombian Basins and the Nicaragua Rise.

The Neo-Caribbean Phase of deformation results from interaction between the Caribbean and the Americas and from general compression which can be related to at least three causes. First, NOAM-SOAM relative motion vectors (Figure 2.5) show convergence during the Neogene, which constricts the Caribbea Plate. Second, the restraining bend in the Oriente-Puerto Rico Trench transform fault northeast of Dominican Republic [Bracey and Vogt, 1970] constricts the eastward migration of the north-central Caribbean and is responsible for much transpression in Hispaniola. The eastern Dominican Republic, which has already passed this bend, has relatively subdued topography. Third, and most important, is the northeastward migration, relative to the Guyana Shield, of the SOAM Cordilleran Terrane. The compression induced upon the Caribbean Plate by this motion is north-northwest, because the Caribbean Plate possesses an eastward component of motion relative to the Guyana Shield that is slightly greater than that of the Cordilleran Terrane (chapter 7). Along the El Pilar fault zone of eastern Venezuela and Trinidad, where the Caribbean and South American plates come into contact, the relative motion of the two is slightly south of east [Aggarwal, 1983]. The strike-slip component is largely taken up along the El Pilar Fault, whereas the compression occurs on folds and thrusts in the basin to the south of the El Pilar [Case and Holcombe, 1980].


KEY TO FIGURES 8.1 - 8.11
arc volcanism
suture zone
subduction zone
thrusting
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20
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Kilometers
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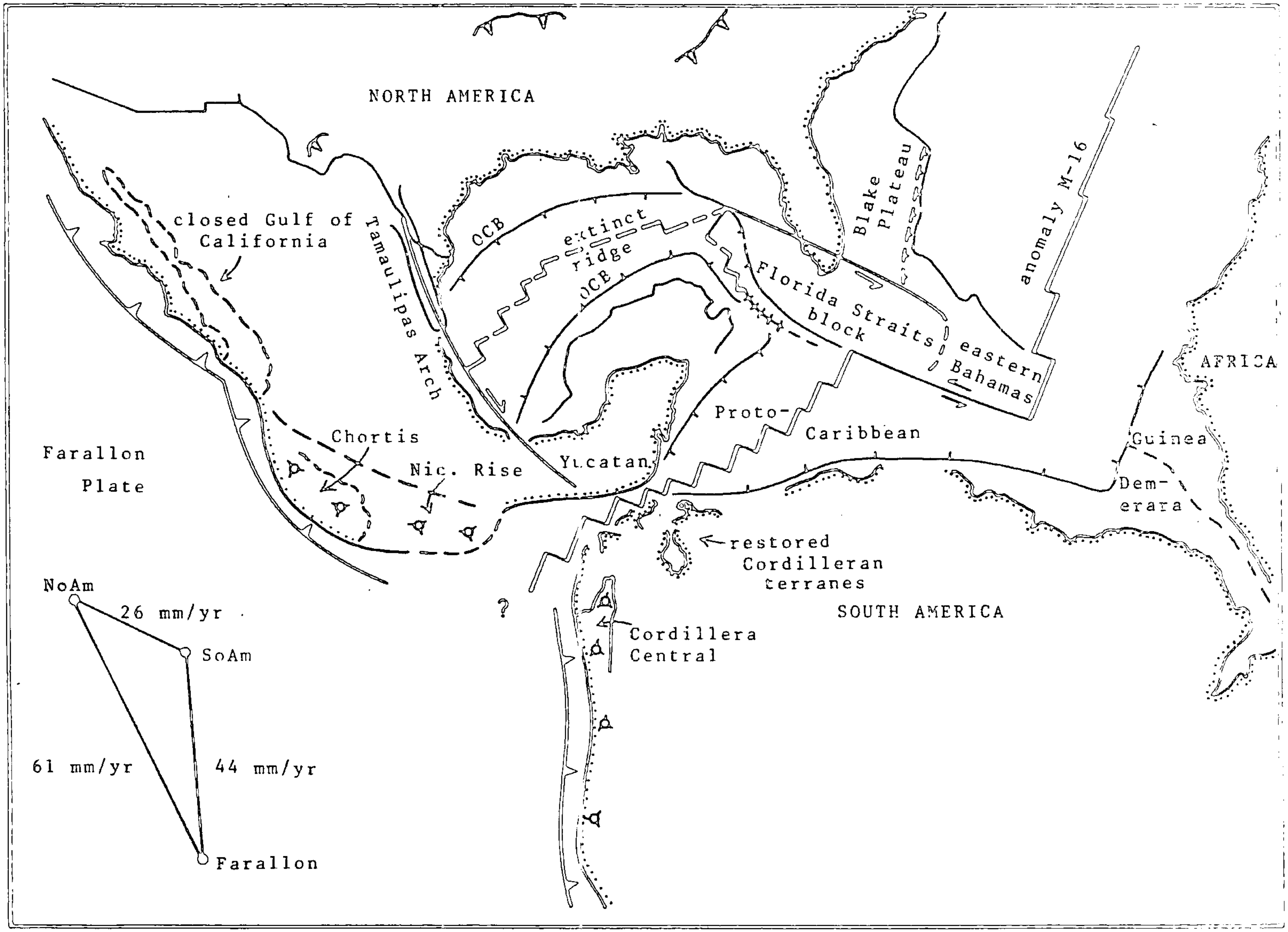

Figure 8.1. Plate-boundary map for the Berriasian (140 Ma).

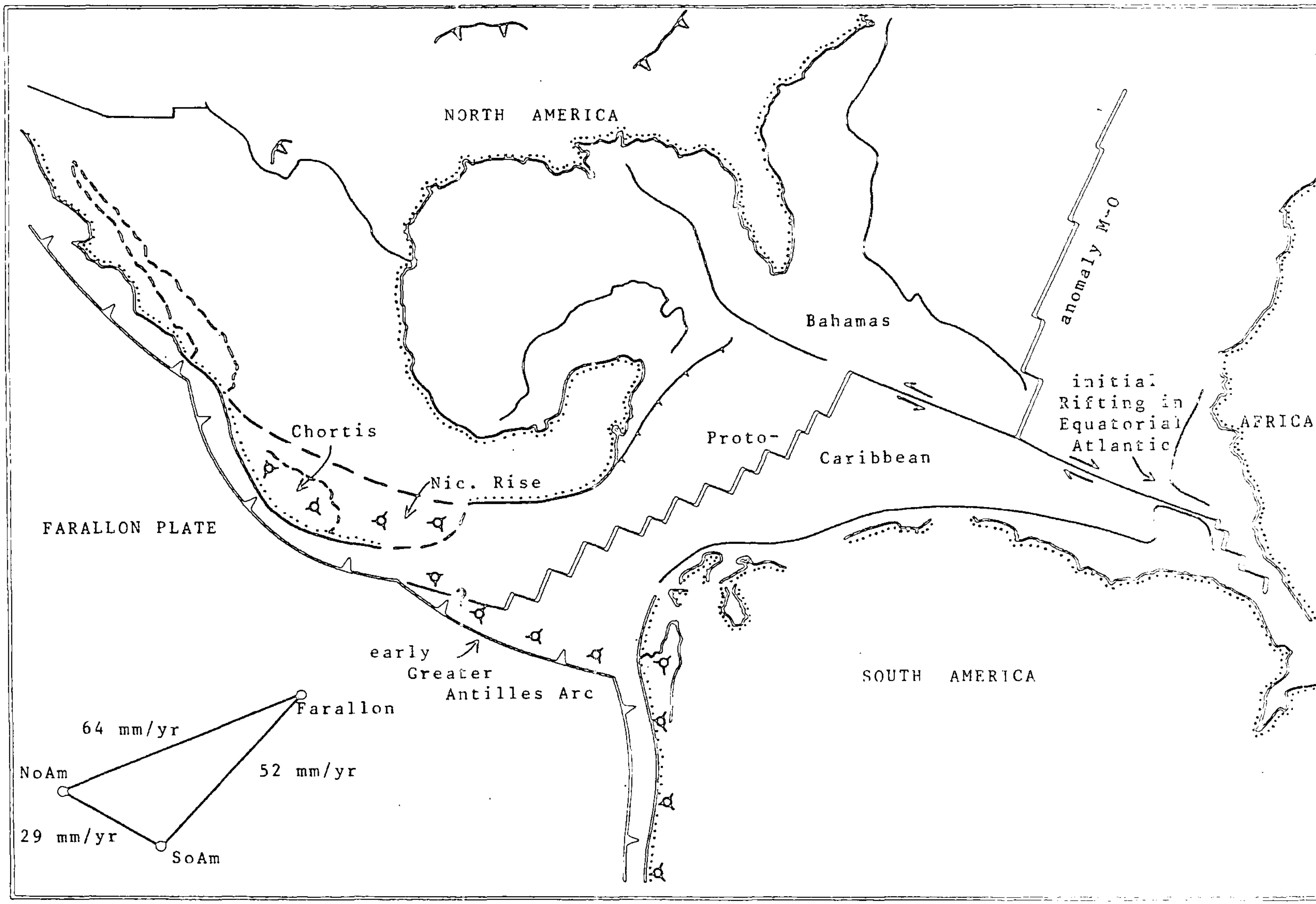

Figure 8.2. Plate-boundary map for the Early Aptian (119 Ma).

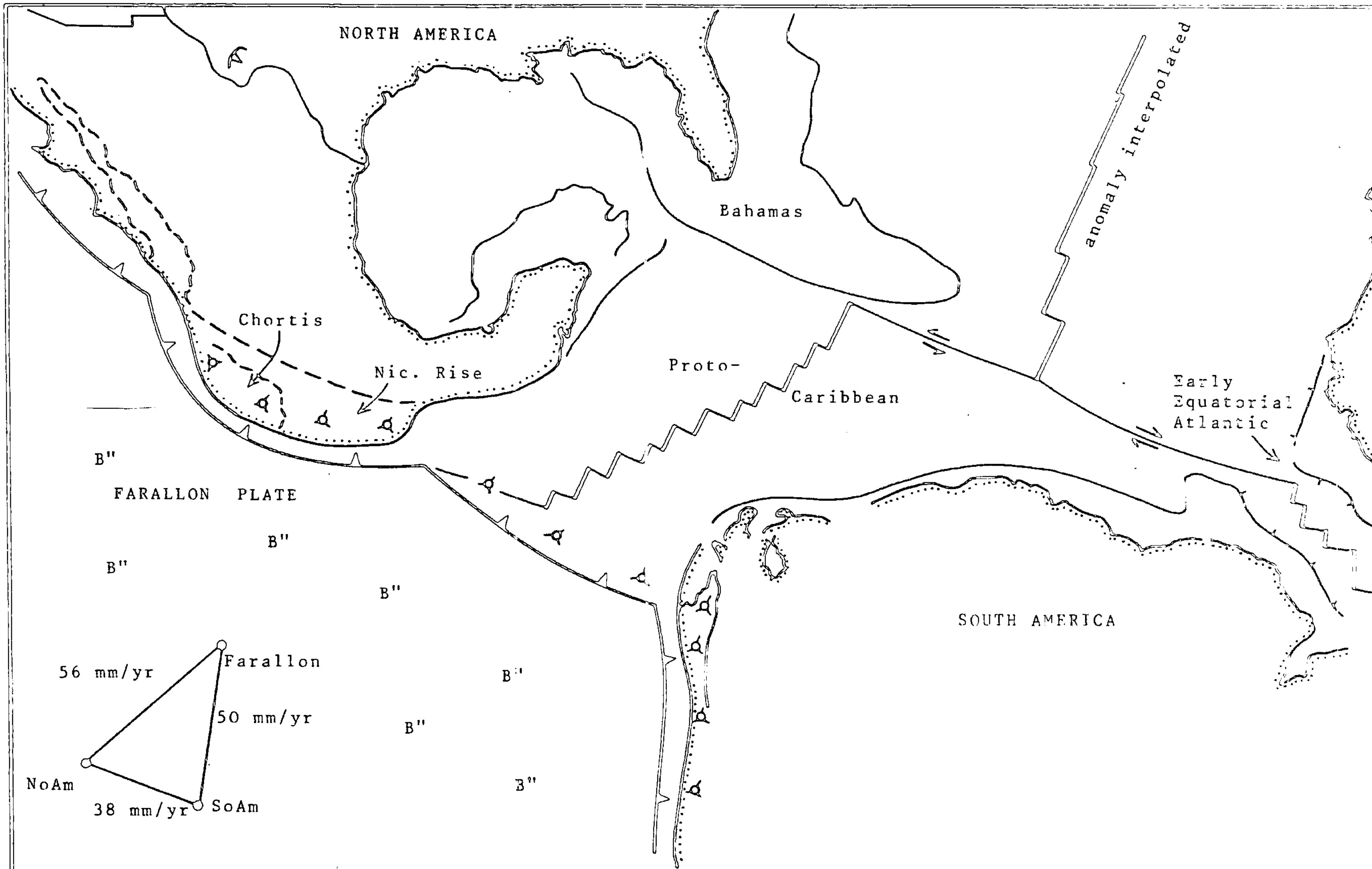

Figure 8.3. Plate boundary map for the Late Albian (100 Ma).

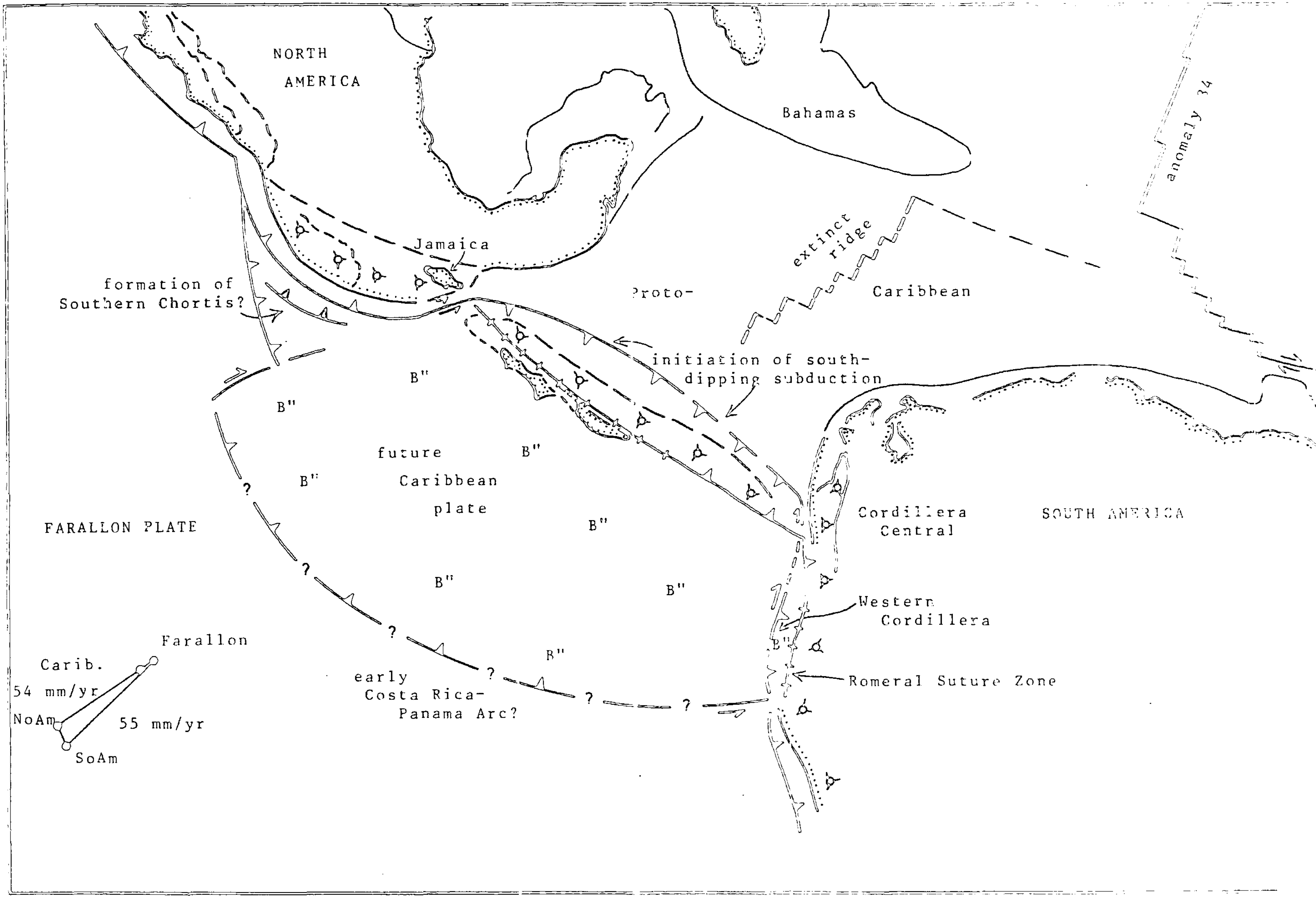

Figure 8.4. Plate-boundary map for the Early Campanian (84 Ma).

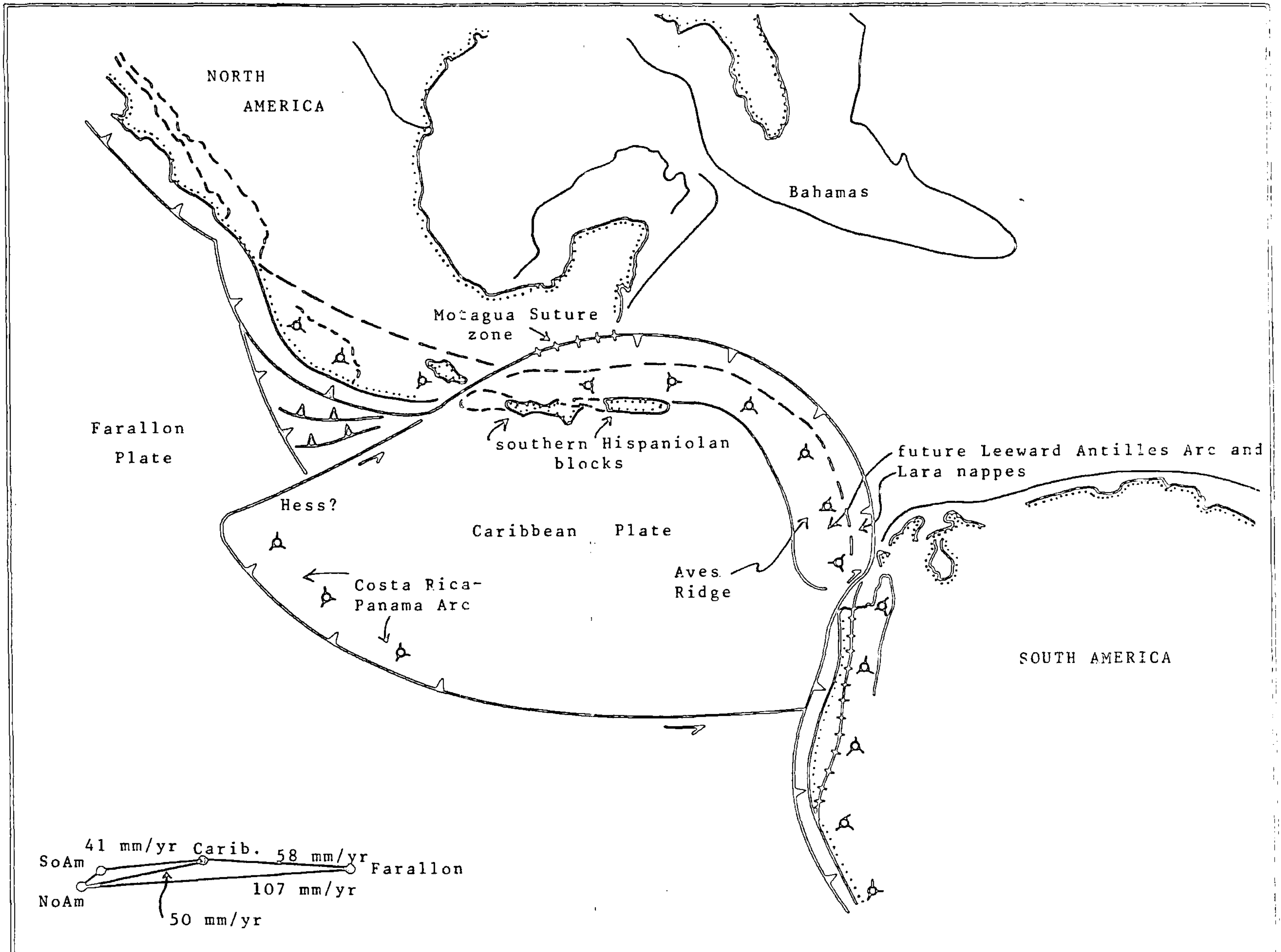

Figure 8.5. Plate-boundary map for the Late Campanian (72 Ma).

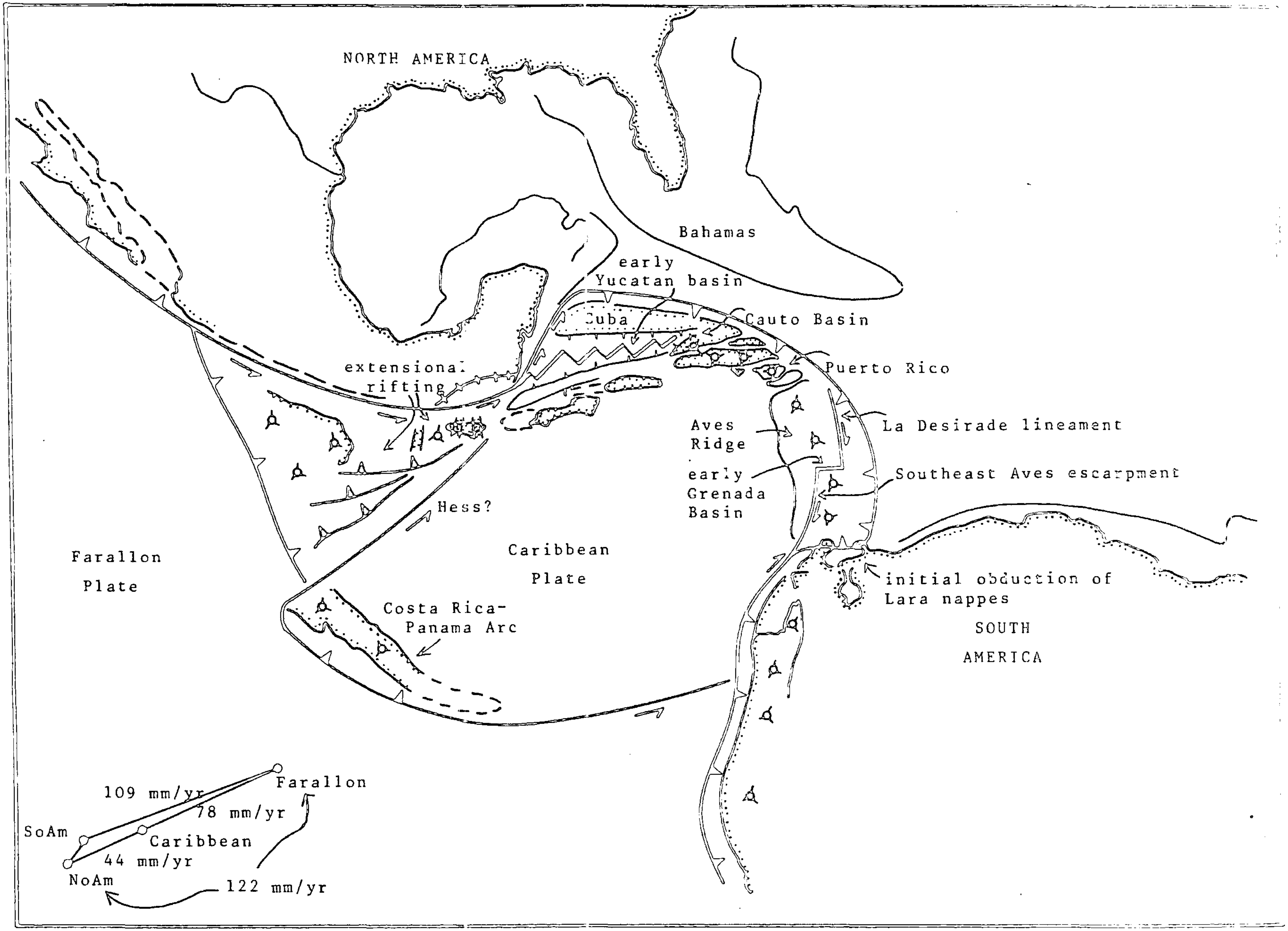

Figure 8.6. Plate-boundary map for the Late Paleocene (59 Ma).

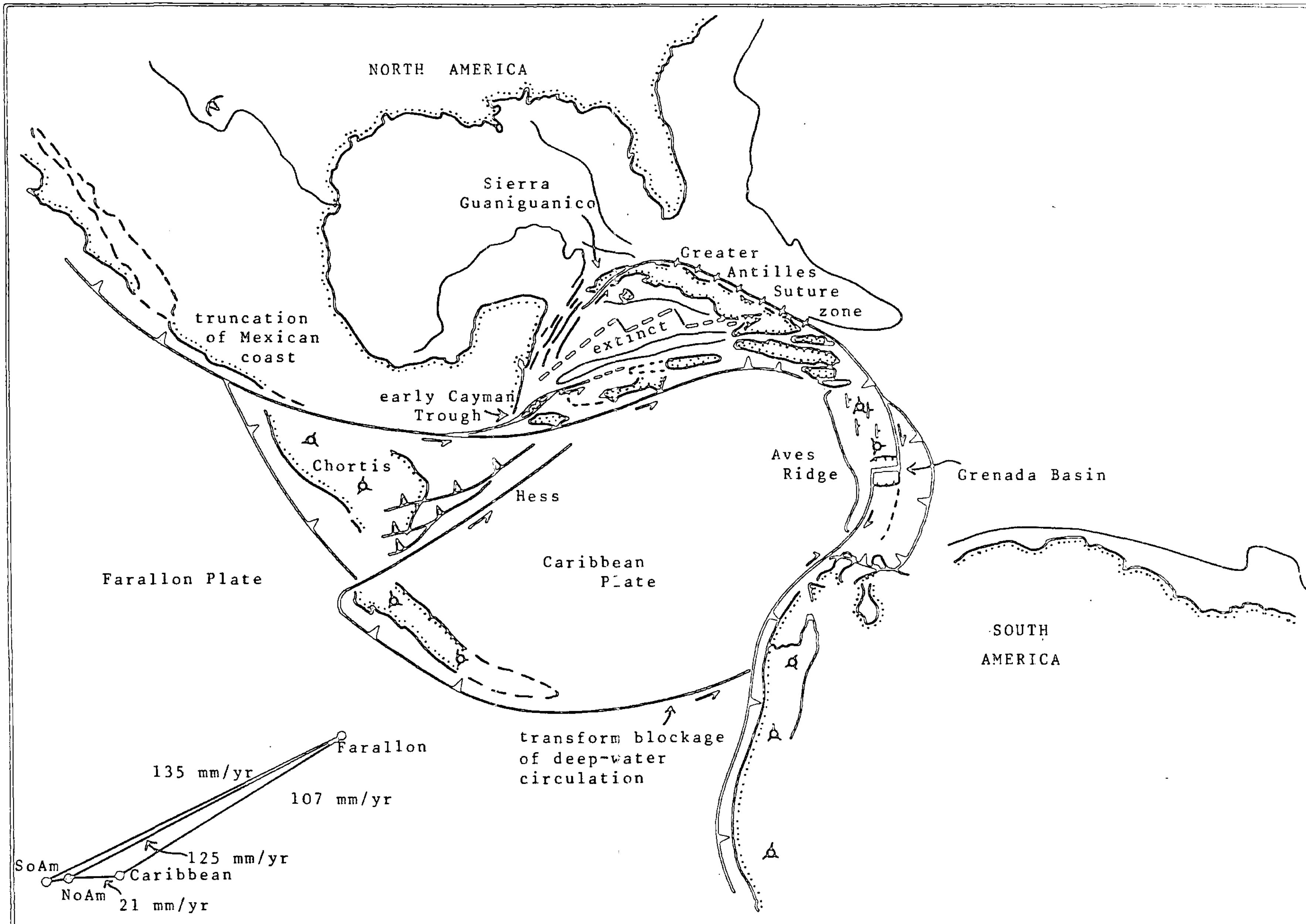

Figure 8.7. Plate-boundary map for the Middle Eocene (49 Ma).

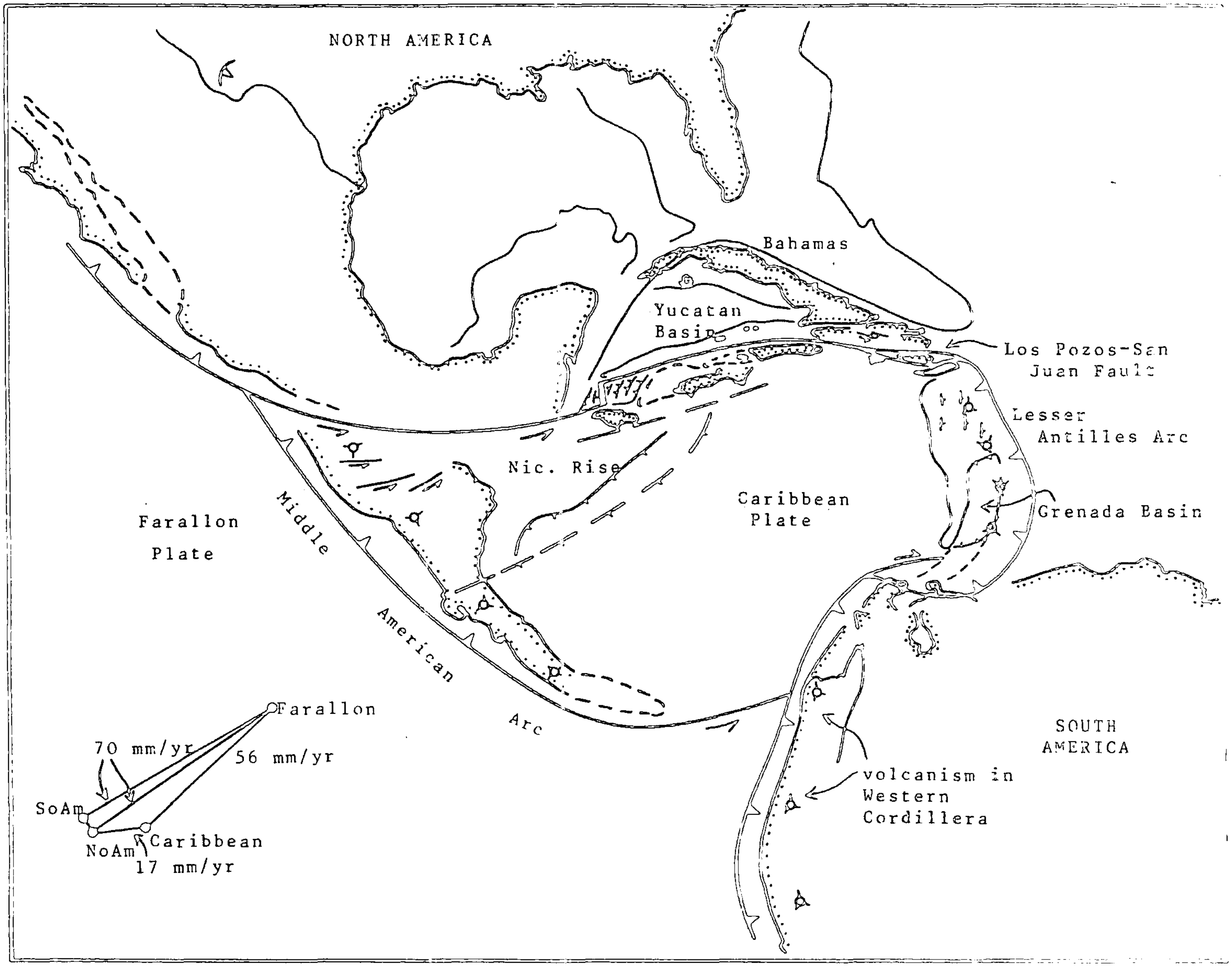

Figure 8.8. Plate-boundary map for the Early Oligocene (36 Ma).

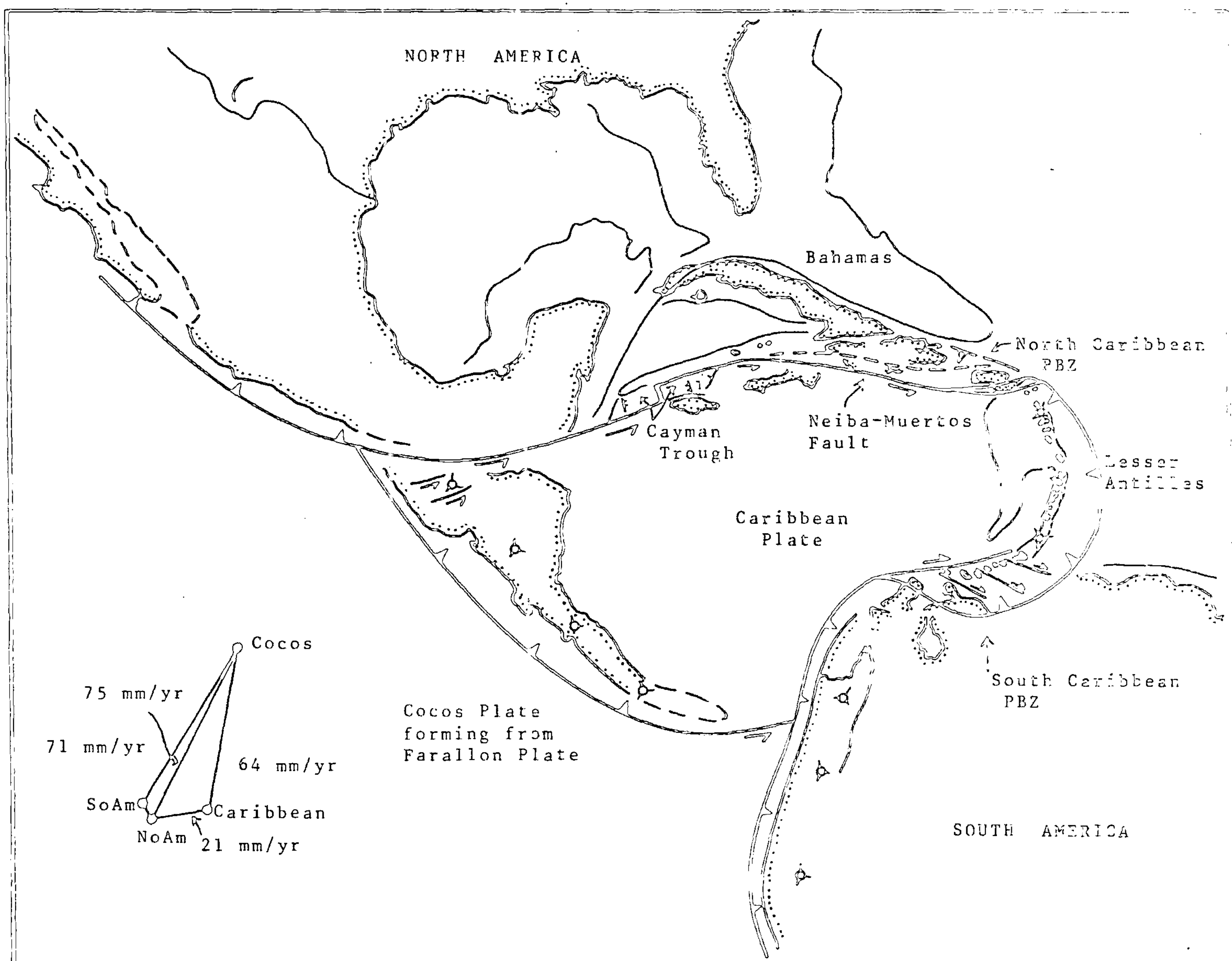

Figure 8.9. Plate-boundary map for the Early Miocene (21 Ma).

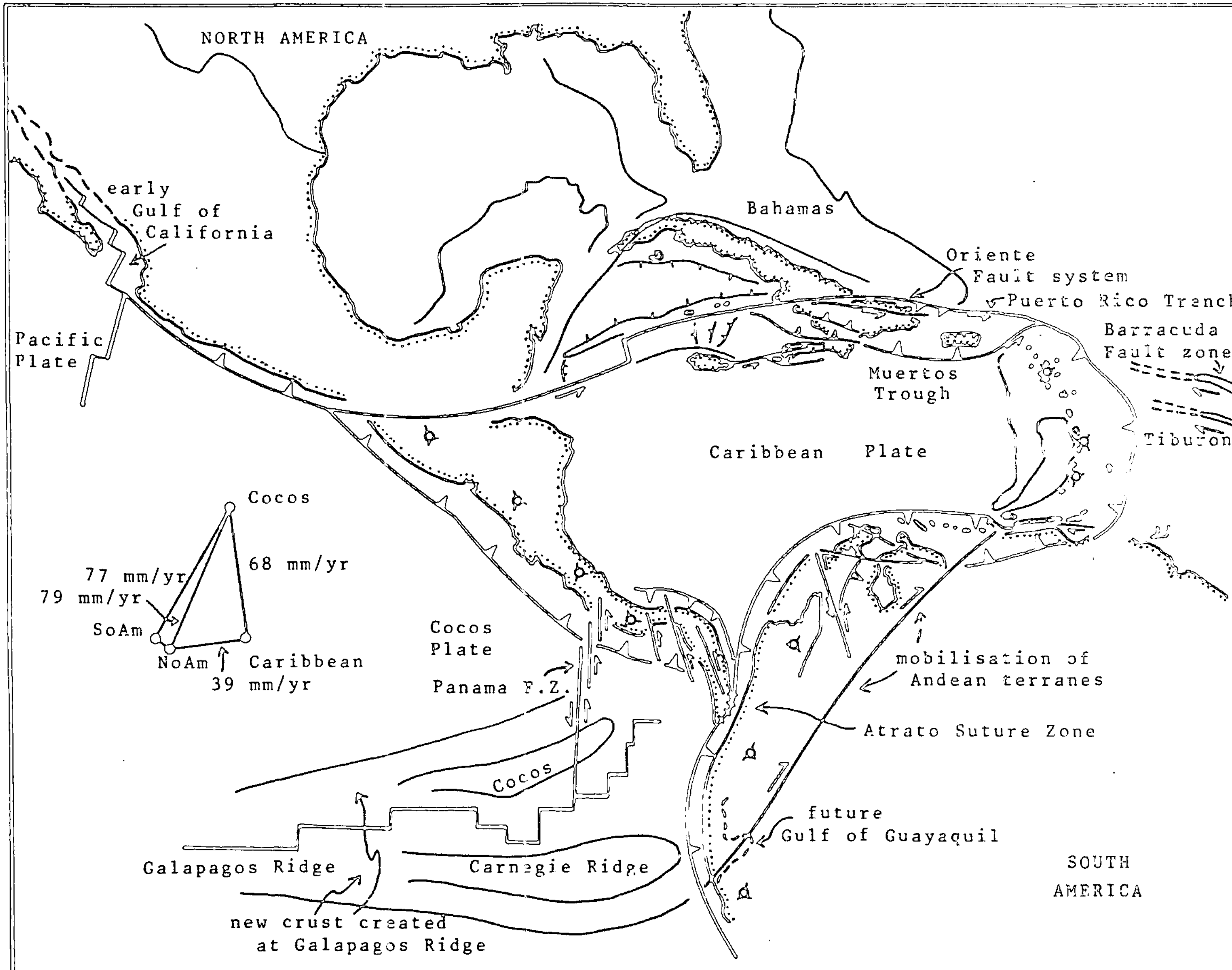

Figure 8.10. Plate-boundary map for the Late Miocene (9 Ma).

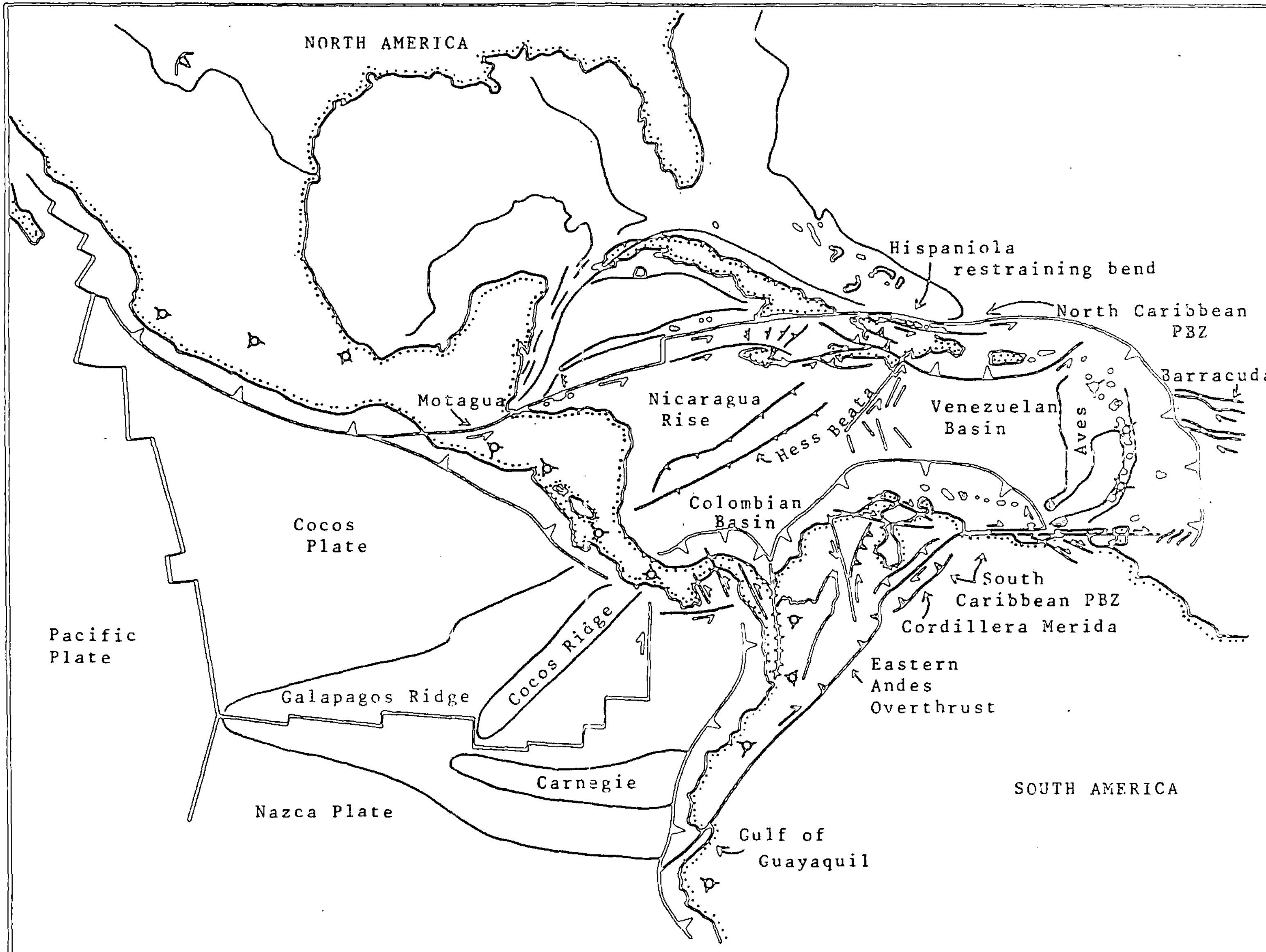

Figure 8.11. Plate boundary map for the Present.

CHAPTER 9

SUMMARY, DISCUSSION, CONCLUSIONS, UNRESOLVED PROBLEMS

SUMMARY, DISCUSSION, CONCLUSIONS, UNRESOLVED PROBLEMS:

Summary:

Despite several unresolved details regarding local evolution, it appears as though the Gulf of Mexico evolved by the rotation of the Yucatan block away from the U.S. Gulf Coast about a pole in northern Florida, and that the Caribbean Plate originated in the Pacific during the Cretaceous and migrated during Late Cretaceous-Cenozoic time into its present position between the Americas.

The relative motion history between North and South America (Figure 2.5) can be broken into three distinct phases. From the Middle Jurassic to the Campanian, South America diverged from North America in an east-southeast direction. From the Campanian to the Eocene, little or no relative motion occurred, but minor (200 km) sinistral convergence is perhaps suggested. Since the Eocene, South America has converged by about 200 to 300 km with North America in a west-northwest direction, achieved by convergent dextral slip along the Barracuda, Tiburon and other related fracture zones. The crust being subducted at present beneath the Lesser Antilles is Late Cretaceous in age.

Three general phases of Gulf/Caribbean evolution can be recognized that coincide with the three phases of North America-South America relative motion history. During the Jurassic to Campanian, rifting and development of passive carbonate shelves formed along circum-Gulf and proto-Caribbean continental margins, associated with the separation of North and South America, Yucatan and the Florida Straits block. Early Cretaceous crust of the Caribbean Plate formed in the Pacific Basin and was intruded by medial Cretaceous basalts (seismic horizon B"), to the west of South America. From the Campanian to the Middle Eocene, compressional tectonism, continued arc volcanism and development, and arc-continent collision was associated with the migration of the Caribbean "Plate" into the proto-Caribbean realm. Since the Middle Eocene, the Caribbean Plate has migrated eastwards relative to the Americas at 2.5 to 4 cm/yr, with much strike-slip deformation in the northern and southern Caribbean plate boundary zones. Arcs have developed along the leading (Lesser Antilles) and trailing (Middle America) edges of the Caribbean Plate by subduction of Atlantic and Farallon/Cocos oceanic crust, respectively. That the Caribbean Plate is exotic with respect to the proto-Caribbean, having formed prior to its entrapment amidst the Americas, precludes the possibility of magnetic anomalies in the Venezuelan and Colombian Basins having any direct bearing upon plate boundary schemes related to the evolution of the Caribbean region *sensu stricto*.

A Middle Miocene to Recent period of orogenesis, referred to here as the "Neo Caribbean phase of deformation," has developed across the Caribbean region. The Greater Antilles have undergone rapid rates of uplift, crustal blocks within the Andean terranes of South America have undergone large-scale (>100 km) transpressional and transtensional strike-slip movements, pull-apart basins have developed throughout the northern and southern plate boundary zones, and the Caribbean Plate has been deformed internally at Beata Ridge and Hess Escarpment. The deformation results from (1) interaction between the Caribbean and the American Plates, (2) compression caused by post-Eocene convergence between North and South America, (3) northeastward migration of the Andean Terranes of northwest South America, and (4) convergence at the Hispaniolan restraining bend along the Oriente-Puerto Rico Trench transform northeast of Samana, Dominican

Republic.

Discussion:

This study illustrates the methodology of developing models for regional evolution of complex interplate realms by integrating local and regional geology into a plate-kinematic framework provided by paleopositions and relative motions of major plates. The plate-kinematic framework provides first-order geometric constraints on proposed models. Paleopositions of continents encompassing the area in question provide the size and shape of the interplate realm at different times, whereas the relative motion history of encompassing plates provides the rate and direction of change of the size and shape of the interplate realm.

Plate-kinematic frameworks in general are constructed by determining poles of rotation for various times from marine magnetic anomalies and fracture zone traces that define former configurations of Atlantic-type oceans. However, these frameworks can be made significantly more accurate, particularly in the early stages of relative motion, by refining the original pre-rift reconstruction of all of the continents involved. This is achieved by considering and incorporating the palinspastic effects of (1) syn-rift extension within continental margins, and (2) post-rift changes in the initial shapes of the blocks and continents that are caused by compression, extension, strike-slip, accretion, or tectonic erosion. The Gulf of Mexico provides an example of the importance of syn-rift extension to the relative motion framework; only about half of the total separation between Yucatan and the U.S. Gulf Coast was accomplished by seafloor spreading, whereas the other half was achieved by intracontinental extension. Purportedly "closed" reconstructions using the present-day ocean-continent boundaries do not recognize this important factor. Likewise, northern South America portrays the importance of post-rift changes in shape; the Andean terrancs have moved northeastward by as much as 290 km, and thousands of square kilometers of land area have been accreted from the Farallon and Caribbean Plates. Plate-kinematic analyses for other interplate realms such as the Mediterranean also should integrate these considerations.

Using the plate-kinematic framework for North and South America and the Farallon/Cocos Plates as a basis, the Gulf of Mexico/Caribbean evolutionary model has been derived by the integration of first-order plate tectonic elements. In the Gulf of Mexico, the distribution of salt, oceanic crust, and attenuated continental crust, and fault zones and their styles, were used to show how the Yucatan and Florida Straits blocks migrated away from the U.S. Gulf Coast. No subduction assisted with the Gulf's development, and assessment of its formation is relatively simple when compared to the Caribbean.

In the Caribbean area, an early Atlantic-type rift and drift history of the proto-Caribbean Sea is overprinted by the entrance from the Pacific and eastward migration of the Caribbean Plate sensu stricto. The proto-Caribbean's tectonic history can be deduced from the relative-motion history of North and South America, but the tectonic history of the Caribbean Plate cannot. To achieve the latter, first-order plate-tectonic elements pertaining to the Caribbean Plate were integrated into the relative motion framework of the America's and the Farallon/Cocos Plate.

Such first-order plate-tectonic elements for the Caribbean include plate-motion indicators such as (1) the intervals of active subduction and polarity of the various Caribbean arcs, (2) the timing of collision between the arcs and the margins of the proto-Caribbean, as indicated by the geometry, timing and structural development of belts containing pieces of oceanic crust, (3) the mode and timing of formation of the Yucatan, Grenada and Cayman basins, all of which are, or have been, sub-basins of the larger, primary Caribbean Plate, and (4) the recognition of sites of large-scale transcurrent offsets since the Eocene.

The proposed evolutionary model for the Gulf and Caribbean is highly dependent upon two critical factors. The first is the proposed Equatorial Atlantic reconstruction that is tighter than that of Bullard et al. [1965]. This fit more satisfactorily rejuxtaposes the Guinea and Brazilian margins, but it also requires that Africa behaved as two plates during the Early Cretaceous. Evidence for this certainly exists [Burke and Dewey, 1974; Pindell and Dewey, 1982], but the total amount of internal deformation within Africa has yet to be quantified. The second critical factor is the assumption that the Cayman Trough records large amounts of relative motion between the Caribbean and North American Plates. Depending on the particular evolutionary model used, a minimum of 800 km of motion along the Cayman Trough is required in order for the Caribbean Plate to have originated in the Pacific realm. In the modelling presented herein, 1050 or 1100 km is assumed, for evidence to the contrary is non-existent, with additional transcurrent motions occurring elsewhere in the northern Caribbean plate boundary zone as well. If offsets greater than 800 km are eventually disproved, a Caribbean evolutionary model that is in situ with respect to the proto-Caribbean area must be accepted.

Conclusions:

Several conclusions may be drawn from the Gulf of Mexico/Caribbean evolutionary model. These are listed below.

1) The finite difference method [Dewey, 1975] has been employed throughout this study. It has been used to determine the relative paleopositions of North and South America through the three-plate circuit including Africa. It has also been used to predict the net motion upon fault systems or plate boundaries in poorly understood three-or-more-plate systems, such as the Neogene development of the southern Caribbean plate boundary zone, and the opening of the Yucatan and Grenada Basins. The finite difference method is useful at any scale, and provides an important tool for the deduction of paleogeographic scenarios.

2) The geometry of plate separation between North and South America was such that a sinistral shear couple existed across the early interplate realm. The Yucatan block rotated in a counter-clockwise sense during the period over which it was in contact with the margins of both North and South America, which is compatible with sinistral shear. Other rotations of this kind on a regional scale include Iberia between Europe and Africa in the Early Cretaceous, and the Jan Mayan Ridge between Europe and Greenland during the Tertiary. Such cause-and-effect relationships concerning the relative motions of tortionally rigid plates suggests that the underlying mantle has little control on the evolution of plate mosaics.

3) Rift-related salt deposits in the Gulf of Mexico occur primarily in realms of attenuated continental crust as defined by seismic reflection and refraction methods [Buffler et al., 1980, 1981; Ibrahim et al., 1981]. Thus, reconstructing boundaries defining the original depositional limits of salt, prior to halokinesis, is useful for estimating paleogeography at the onset of seafloor spreading. The Gulf of Mexico provides an example which, despite salt migration beneath

Sigsbee Escarpment, suggests that in areas of poorly known basement structure, the occurrence of salt deposits may be used to roughly approximate the extent of attenuated continental crust. However, this may not be the case for the eastern Brazilian margin.

4) Throughgoing, deep-water circulation between the Atlantic and the Pacific Ocean, via the Caribbean, might never have occurred. In the Early Cretaçeous, this would have depended upon the bathymetric expression of the early Greater Antilles island arc. During the Campanian to Maestrichtian interval the Greater Antilles most certainly should have prevented circulation. As the Aves Ridge, in its early stages as an arc, passed northwest South America in the Paleocene, deep circulation may have been possible, depending upon the bathymetric expression of the young Aves Ridge. Dredge hauls of Middle Eocene shallow-water carbonates from Aves Ridge, however, demonstrate that by that time the ridge had become shallow [Fox and Heezen, 1975]. Since the Eocene, the Aves and Lesser Antilles ridges have clearly prevented deep-water circulation. Circulation between the Pacific and the water above the Caribbean Plate was unhindered until the Middle Eocene, with the deposition of siliceous pelagics in both until that time. In the Middle Eocene, however, pelagic sedimentation in the Caribbean changed and the top of the Early Eocene siliceous cherts and micrites is observed as seismic reflector A" in the Caribbean Basins [Edgar et al., 1971]. No such change occurred in the Pacific. It is suggested that the transform/trench system connecting the Costa Rica/Panama arc to the South American trench system was of sufficient bathymetric expression to have interrupted deep water circulation and the entrance of silica rich waters into the Caribbean. Such bathymetric expression is common at long oceanic transforms [Hegarty et al., 1983].

5) Migrations of terrestrial vertebrate fauna between North and South America occurred in the Late Cretaceous, the Eocene, and the Pliocene. The Late Cretaceous migration of dinosaurs probably occurred across the Greater Antilles from Mexico to northwest South America, during the period of arc polarity reversal. The Eocene migration of rodents occurred along the Greater Antilles and ?Bahamas, and across the shallow-water (and partially exposed) Aves Ridge, again into northwest South America. The Pliocene migration of mammals such as the early horses occurred across the Isthmus of Panama, which was uplifted and exposed during the Panama-Western Cordillera collision.

6) A single mid-ocean ridge in the proto-Caribbean from the Jurassic to the Campanian is predicted. Proto-Caribbean ridge volume by stages can be calculated from the ridge length and the spreading rate, defined in the vector triangles, on each reconstruction.

7) The anomalously thick (12 to 25 km) Caribbean Plate owes at least part of its excess thickness to the extrusion and intrusion of basalts and diabases (seismic horizon B") onto and into pre-existing oceanic crust. The basaltic episode occurred mainly during the Cenomanian to Santonian period, possibly into the Campanian. This period follows an Aptian-Albian period of very little relative motion between the Farallon Plate and the mantle [Engebretson, 1982]. It is suggested that the cause of the basaltic event was the accumulation of excess heat and partial melting beneath the Caribbean portion of the Farallon Plate. The Caribbean Plate, as a trapped, buoyant piece of the original Farallon crust, may provide a clue to the cause of Laramide orogenesis in continental portions of the Americas [Livaccari et al., 1981]. Other areas of the Farallon Plate, had they been affected by such off-axis volcanism, should have begun to arrive at the Cordilleran margins in

the Late Cretaceous at the beginning of Laramide orogenesis. It is felt here that the medial Cretaceous basaltic event was more widespread than just the Caribbean portion of the Farallon Plate, and that the excess thickness and buoyancy of affected areas, coupled with high rates of Farallon convergence with North and South America [Engebretson, 1982], caused a shallowing in the dips of Cordilleran subduction zones which is responsible for much of the Laramide foreland orogenesis.

8) Engebretson [1982] predicted that Neocomian Farallon motion with respect to North America was convergent with a large component of sinistral shear, and that by Aptian time relative motion was nearly head-on in the Caribbean area. This accords well with the age of inception of the Greater Antilles arc. Before the Cretaceous, however, in the Middle Jurassic, blocks within Mexico migrated to the southeast into the zone that was occupied by South America prior to the Cretaceous. The likely mechanism for such southeastward transport of Mexican blocks is sinistral shear during subduction. Hence, it is suggested that the prediction of sinistral convergence between North America and the Farallon, or other Pacific plates, may be extended back to the Middle Jurassic.

9) The proposed model and the history of Farallon/North American relative motion make certain predictions about the origin of the allochthonous terranes of North America. For terranes presently in North America to have originated from South America, the 'jump' from South to North Ameirca must have occurred prior to the Middle Jurassic. In the Middle Jurassic, blocks of Mexico were migrating southeastwards. Southeastward migration apparently continued into the Neocomian, as predicted by relative plate motions. For the Aptian through

Maestrichtian interval, western Mexico and the western margin of Chortis could have been source areas for northward migrating terranes. For the Cenozoic, it appears that the present mouth of the Gulf of California has defined a point that to the north of which, terranes have migrated northwards (e.g., Salinia), and that to the south of which, terranes have migrated eastwards (e.g., Chortis). The wide distribution, variable paleogeographic environments, and large total volume of the North American allochthonous terranes suggest that their source area was also large and diverse. From the above, it appears that western Mexico is the only portion of the American Cordilleras from which the terranes could have been displaced northwards, and this could only have occurred since the Aptian. It is concluded that 1) the Cretaceous areal extent of Mexico was perhaps double its present size, and/or 2) many of the allochthonous terranes originated out in the Pacific (much like the western Pacific of today) and were carried for much of their northward migrations as parts of the Farallon Plate.

10) The relative motion vector derived herein for South America and North America (Figure 2.5) is extremely simple. South America diverged until the Campanian and then effectively stopped, with only minor motions occurring since that time that are related to reactivation of the original fracture zones. No well-defined North America-South America plate boundary can be defined at present from seismicity or for the Cenozoic from magnetics, and one probably has not existed since the Early Campanian. This coincides with the passive development of the proto-Caribbean Basin until the progressive entrance and eastward migration of the Caribbean Plate throughout the Cenozoic. Arc-margin interaction can be traced eastwards by structural deformation and facies development along the margins of the proto-Caribbean.

11) As the Caribbean Plate entered the proto-Caribbean through the neck between Yucatan and northwest South America, two intra-arc basins (Yucatan and Grenada) formed on the 'corners' of the leading edge of the Caribbean Plate. This was the mechanism by which the Caribbean Plate assumed the larger shape of the proto-Caribbean. The forces responsible for the initiation of the intra-arc basins were apparently smaller than those required to tear new transforms into the proto-Caribbean oceanic crust. the outward components of the forces that drove the Cuban and Lara forearcs away from the Caribbean Plate could be due to roll-back of the subduction zones, or to shearing caused by oblique subduction (subduction at Cuba was sinistral, at Lara was dextral). As roll back velocities are far slower than the opening of both basins [Taylor and Karner, 1983], the latter is assumed to be dominant. It is suggested that the trench-parallel shear component arising from oblique subduction is responsible for the outward force in arc systems that creates many intra-arc basins.

Unresolved problems:

In so far as the Gulf and Caribbean region is composed largely of a collection of allochthonous terranes [Case et al., 1984] whose exact relative motions are poorly known, the proposed evolutionary model is mainly an attempt to reconcile the existing geological data base with known or inferred plate motions. In that light, the model provides a framework which may be further substantiated or refined by continuing field and marine studies. At the same time, the model helps to identify problem areas that require further research. Such problems include the following.

(1) A new reconstruction and opening history for the Equatorial Atlantic has been suggested herein that seems to explain many problems plaguing the understanding of the opening history of the Atlantic and the Permo-Triassic paleomagnetism of the circum-Atlantic continents. However, this model is, as yet, speculative, and needs to be subjected to the rigors of specific geophysical tests tracing the fracture zones to the opposing margins.

(2) It is argued herein that the Yucatan Block rotated counterclockwise away from the U. S. Gulf Coast to its present position to form the Gulf of Mexico. However, it may have been translated along the Mojave Sonora Megashear, intact with central Mexico. The distribution of salt in the Gulf, the steep drop off of basement along eastern Mexico (transform margin?), the distribution of oceanic crust in the Gulf, and the trend of poorly-defined magnetic anomalies in the Gulf favor the former hypothesis. However, until the magnetics are better understood, a clear understanding of the origin of the Yucatan block cannot be determined.

(3) A "card-deck" simple shear model for the Jurassic emplacement of Mexico into the overlap position of South America is suggested herein, but evidence for this, or any other emplacement mechanism is highly speculative. More field-related studies within Mexico are required to better the understanding of the emplacement of Mexico.

(4) It is becoming clear that the western Bahamas (Grand Bank and south Florida Shelf) are underlain by thinned, pre-Mesozoic continental crust. However, the basement of the eastern Bahamas and its origin as a carbonate bank are unknown.

(5) The aereal extent of the continental portion of the Chortis block and its position relative to North America prior to the Aptian is unknown.

This hinders attempts to define plate boundary schemes for the Caribbean in the Early Cretaceous. Also, the nature of basement of the Nicaragua Rise and Jamaica is still unclear. In the Aptian and later, sedimentation in Chortis and southern Mexico was very similar, suggesting that Chortis was adjacent to Mexico in the medial and Late Cretaceous prior to entering the proto-Caribbean realm, as is suggested by the offset along Cayman Trough. In order for Chortis to have entered from adjacency with Mexico, compression must have occurred between Chortis and the Colombian Basin. This compression is suggested by the structures of the Nicaragua Rise and offshore southern Jamaica, but any specific understanding of the predicted compression is lacking. In addition, the significance of the Hess Escarpment, which is one of these structures, is unknown.

(6) The vast majority of evidence points to a pull-apart origin for the entire Cayman Trough, but this has yet to be proved by direct observation. At least 800 km of offset is needed for a Pacific origin of the Caribbean Plate. Until this is proved, some question will remain as to the provenance of the Caribbean Plate.

(7) Most accounts of the tectonic history of central Venezuela call for Eocene emplacement of the North Venezuelan nappes [Maresch, 1974]. An emplacement age of Oligo-Miocene is suggested here. Assuming large offsets along the Cayman Trough, an Eocene emplacement age requires that emplacement occurred upon plate boundaries of the proto-Caribbean. Such motions are not predicted by North and South American relative motion history. The differing interpretations hinge on whether the flysch of the Piemontine nappes in front of the Lara nappes was deposited and accreted during the subduction stage offshore, or whether it represents the sediment of the foredeep which developed

during obduction onto the margin. The former is favored here, as the wavelength of the Piemontine nappes is much too short to be a foredeep. In addition, an Oligo-Miocene foredeep did form, although only poorly, which was then overridden by the Lara and Piemontine nappes together [Beck, 1978]. Additional field studies in the Venezuelan Andes and a proper assessment of gravity and basin analyses of the Venezuelan foreland basins are needed in order to determine whether the Eocene or the Oligo-Miocene emplacement model is correct.

(8) The east-west magnetic anomalies across Grenada Basin may or may not be seafloor-spreading anomalies. That they are seafloor-spreading anomalies is important to ascribing the Leeward Antilles arc and Villa de Cura Klippe to the southeastern portion of a proto-Aves Ridge arc and forearc. The opening of the Grenada Basin is important to the emplacement of the Lara and Villa de Cura nappes (Figure 6.5); therefore, assessing this problem will aid in assessing the emplacement history of the north Venezuelan nappes.

(9) Because of insufficient data to warrant other alternatives, it has been assumed that the Greater Antilles Arc is composed of only one arc that reversed its polarity in the Campanian. The central Hispaniolan peridotite and associated metamorphosed pelitic sediments [Draper and Lewis, 1980], and the blueschist metamorphic areas of southern Cuba [Mattson, 1979] may be interpreted as pertaining to the polarity reversal. However, they may also indicate that the Greater Antilles are composed of more than one arc. Prior to the Campanian, paleogeographic relationships within and among the Greater Antilles are poorly known; a scenario similar to the western Pacific prior to a Campanian amalgamation 'event' is not unreasonable. Outlined herein is the simplest model for the Late Cretaceous that provides an

explanation, if nothing more, of most of the known major structural and depositional features. Continued efforts toward understanding the Campanian and older history of the Greater Antilles are desperately needed.

(10) The nature of basement in Cuba is still debated. It has been labeled as entirely arc-related, having collided with the Bahamas by arc-continent collision. However, basement in some apparently areas may be continental. It is possible that these bits were carried with the arc complex from a southern provenance. However, an alternative to the "allochthonous arc" concept is that the volcanic aspects of Cuba pertain to the Greater Antillean forearc, possessing the odd calc-alkaline intrusion, which was obducted onto the attenuated continental margin of the southern Bahamas in the Paleogene. In this case, basement is autochthonous, whereas the mafic volcanics and intrusions of Cuba are obducted forearc constituents. More analyses on the Cuban rocks are required to solve this problem.

(11) The nature and age of the floor of the Yucatan Basin is uncertain. A seafloor spreading origin in the Paleogene is suggested from seismics and heat-flow studies. If the floor is thinned continent, however, it becomes difficult to ascribe Cuba to having originated in the Pacific with the rest of the Caribbean Plate.

(12) Seafloor magnetic anomalies in the Venezuelan Basin trend northeast. Anomalies trend east in the Colombian Basin, but these may not be seafloor-spreading related. Differing trends in magnetism (and its sources) suggests that the intervening Beata Ridge may have had an early history of offset along it. The present uplift of the Beata Ridge is poorly understood, but probably relates to minor motion between the Colombian and Venezuelan Basins. It is suggested herein

that this motion possesses a sinistral strike-slip component.

(13) The apparent Campanian initiation of subduction at the Costa Rica-Panama arc is constrained only by 2 radiometric ages on plutons and some tuff horizons in the stratigraphic record. The timing of the inception of subduction at this arc is important to understanding when the Caribbean Plate became separated from the Farallon Plate, and when it was able to possess its own motion relative to the other plates.

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