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Plate-kinematics and crustal dynamics of circum-Caribbean arc-continent interactions: Tectonic controls on basin development in Proto-Caribbean margins

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ABSTRACT

The American margins of the Caribbean comprise basins and accreted terranes recording a polyphase tectonic history. Plate kinematic models and reconstructions back to the Jurassic show that Mesozoic separation of the Americas produced passive margins that were overridden diachronously from west to east by allochthonous Caribbean plate–related arc and oceanic complexes. *P-T-t* and structural data, sedimentary provenance, and basin-subsidence studies constrain this history. Caribbean lithosphere is Pacific-derived and was engulfed between the Americas during their westward drift as the Atlantic Ocean opened. This began ca. 120 Ma with development of a west-dipping Benioff zone between Central America and the northern Andes, now marked by the Guatemalan and Cuban sutures in North America and by the northern Colombian and Venezuelan "sutures" of South America, persisting today as the Lesser Antilles subduction zone. Most Caribbean high-pressure metamorphic complexes originated at this subduction zone, which probably formed by arc-polarity reversal at an earlier west-facing Inter-American Arc and was probably caused by westward acceleration of the Americas. The mainly 90 Ma Caribbean basalts were extruded onto preexisting Caribbean crust ~30 m.y. later and are not causally linked to the reversal. The Great Caribbean Arc originated at this trench and evolved up to the present, acquiring the shape of the preexisting Proto-Caribbean Seaway. The uplift and cooling history of arc and forearc terranes, and history of basin opening and subsidence, can be tied to stages of Caribbean plate motion in a coherent, internally-consistent regional model that provides the basis for further studies.

Keywords: Caribbean, plate tectonics, Pacific origin, paleogeographic evolution, arc-continent interaction, arc-polarity reversal, backarc spreading, Proto-Caribbean, Cuba, Venezuela, Trinidad, Mexico, Guatemala.

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INTRODUCTION

Basin and structural development at the Caribbean plate's boundaries with North and South America were strongly controlled by regional plate motions associated with the breakup of Pangea, opening of the Proto-Caribbean arm of the Atlantic Ocean, and the progressive engulfment of a swath of Pacific oceanic crust (now part of Caribbean plate) between the Americas as they drifted west from Africa. Thus, a passive margin phase of stratigraphic development preceded arc-continent interactions and syntectonic basin development along the Caribbean-American margins (Fig. 1).

As the American plates began their separation in the Triassic to Neocomian, passive margins formed along northern South America and southern North America, and an east-dipping subduction zone defined the western limits of continental crust along the Cordilleran margin, producing a continuous continental arc crossing from North to South America. However, as continental separation increased during the Neocomian, this trench became lengthened across the widening inter-American gap, giving rise to the hypothetical concept of an "Inter-American Arc" connecting the Andes and Mexico. This arc would have been situated, at first, above the east-dipping subduction zone, allowing Pacific crust to subduct beneath North and South America and the Proto-Caribbean, but in order for Pacific crust to eventually enter the widening gap between the Americas, subduction polarity must have reversed, becoming west-dipping.

We develop and review arguments that this subduction polarity reversal occurred during the Aptian and that subsequent plate motions allowed Pacific-derived crust to enter the Proto-Caribbean Seaway between the Americas. Thus, the hypothetical Inter-American Arc should form the roots of arc complexes that defined the leading edge of the Pacific-derived crust. Burke (1988) termed this arc the Great Caribbean Arc.

As circum-Caribbean plate kinematics have become better constrained (contrast Ladd [1976] and Pindell et al. [1988]), the Pacific origin of the Caribbean plate has become more firmly established (Burke et al., 1978, 1984; Pindell and Dewey, 1982; Burke, 1988; Pindell, 1990). Increasing confidence in a Pacific origin for the Caribbean plate, to the west of the Intra-American Arc, has led to many of the allochthonous complexes of southern

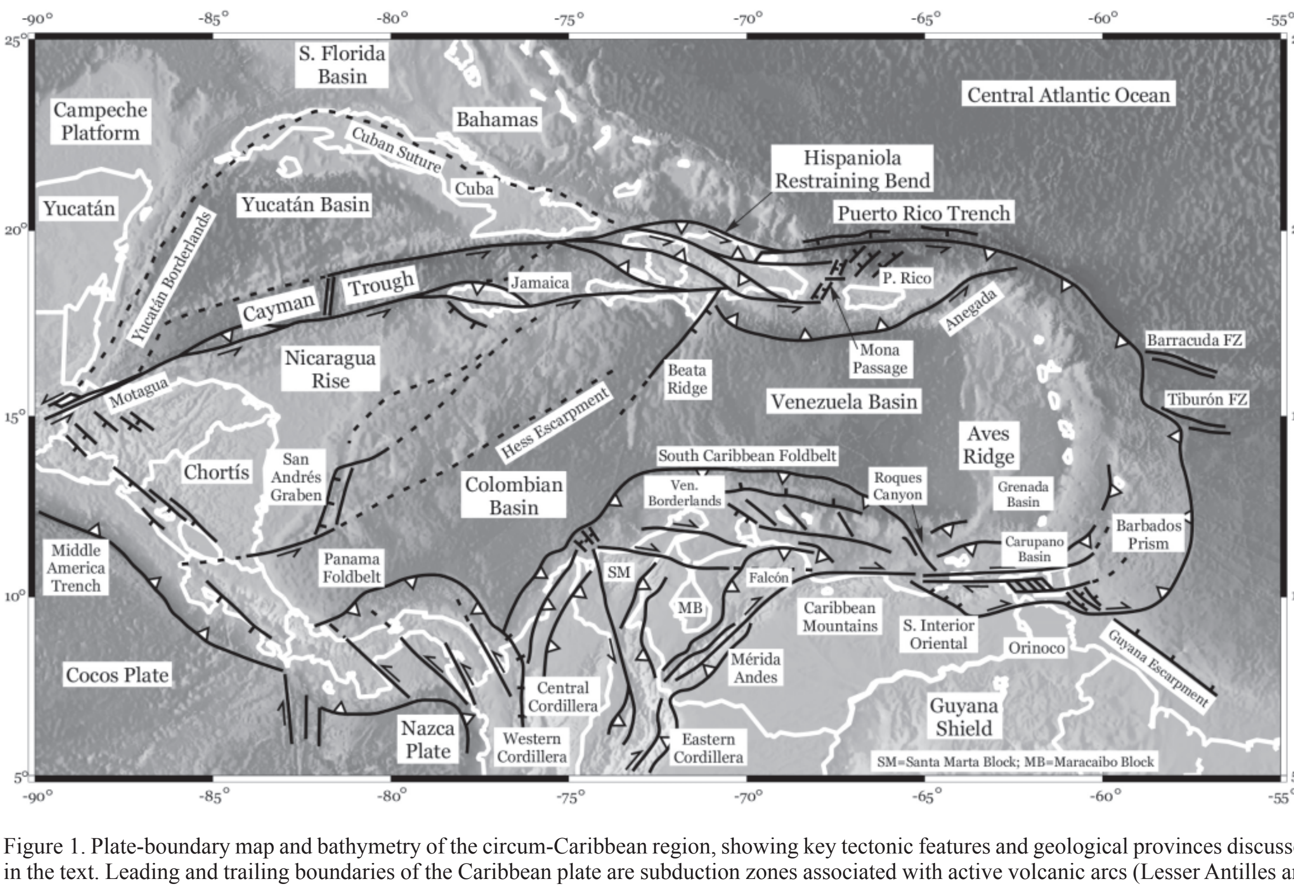

Figure 1. Plate-boundary map and bathymetry of the circum-Caribbean region, showing key tectonic features and geological provinces discussed in the text. Leading and trailing boundaries of the Caribbean plate are subduction zones associated with active volcanic arcs (Lesser Antilles and Panama–Costa Rica arcs, respectively). The southern plate boundary with Colombia, Venezuela, and Trinidad is wide, diffuse, and complex; strain is partitioned between thrust faulting and strike-slip faulting associated with development of pull-apart basins. Similarly, prior to the Eocene collision of Cuba with the Bahamas Bank, the northern plate boundary was also wide and complexly partitioned. Post-collision, the plate boundary was reorganized, with motion now concentrated on the relatively simple Cayman Trough in the west, whereas complexly partitioned thrust and strike-slip faulting continues in the Puerto Rico segment.

North America and northern South America, including much of the western flank of the northern Andes, being tied through lithology, chronology, and/or geochemistry to the Caribbean plate, to its arcs, and to its subduction-accretion complex (Donnelly et al., 1990; Snoke, 1991; Reynaud et al., 1999; Snoke et al., 2001). However, identifying a terrane as “Caribbean-related” is only the first step to understanding how that terrane fits into the long-lived Caribbean–South American interaction, and significant effort is required to reconstruct the various complexes into a coherent and detailed understanding of this progressive “Caribbean Orogeny.”

This paper examines the history of tectonic interactions of the Great Caribbean Arc and the Caribbean plate with the North and South American continental margins. We employ and integrate plate kinematics, pressure-temperature-time (*P-T-t*) data of metamorphic complexes in arc-related terranes, other geochronological data, periodicity of arc magmatism, seismic tomography, basin and arc stratigraphy, and regional structure, set within the above general evolutionary scheme.

Several of the allochthonous complexes on both of the American margins possess high pressure–low temperature (HP-LT) and other metamorphic rocks whose *P-T-t* paths can help to define the history of subduction and subsequent tectonic events leading to their obduction onto the American margins. Integration of this information with structural and stratigraphic data that show a fairly consistent west-to-east younging of terrane emplacement and basin subsidence along both northern and southern margins of the Caribbean allows us to constrain the origin of various Caribbean terranes and the timing and processes of continent-ocean interactions.

Here, we briefly review the history of progressive obduction of allochthonous metamorphic complexes recorded along the southern North American passive margin (Fig. 2). Before we can address such interaction, the Chortís Block (continental crust beneath southern Guatemala, Honduras, Nicaragua, El Salvador, and western parts of Nicaragua Rise) of today’s Caribbean plate must be restored to the southwest margin of Mexico for mid-Cretaceous and older times (prior to its incorporation into the Caribbean plate; Figs. 3 and 4; Pindell and Dewey, 1982; Johnson, 1990; Schaaf et al., 1995) to avoid overlap with the known Jurassic position of the South American plate. Once this is done, a northeastward younging of deformation, which we interpret as a result of Caribbean arc–continent interactions, can be documented starting in Chortís (mid-Albian; Horne et al., 1974), progressing to southern Yucatán (Senonian; Rosenfeld, 1993) and along the Belize margin (Rosencrantz, 1990). Finally, emplacement of the arc in the Bahamas (early to middle Eocene; Pszczólkowski, 1999) was made possible by the opening of the Paleogene Yucatán Basin, allowing the leading edge of the Caribbean plate to advance to the northeast while most of the Caribbean plate moved in a more easterly direction with respect to North America. The origin of some of the allochthonous complexes along this margin remains problematic and is discussed below.

The broad, diffuse Caribbean–South American plate-boundary zone (Fig. 1) is also strongly controlled by regional tectonics. In the west, continental crust of the northern Andean terranes is escaping the east-west convergence in Colombia and is extruding northward across Caribbean lithosphere at the north-vergent South Caribbean Deformed Belt (Dewey and Pindell, 1985; van der Hilst and Mann, 1994). Basement-involved deformation has occurred since at least 25^Ma, penetrating several hundred kminto the South American continent, allowing South America to acquire an arcuate, hanging-wall geometry necessary for long-term subduction of the southern part of the Caribbean plate (Pindell et al., 1998). In the eastern portion of the plate-boundary zone, primary structures more closely reflect Caribbean–South America relative motion (east-west shear) because the two plates have no large, independent Andean terranes between them.

Despite these complexities, this late Cenozoic tectonic setting for the South American margin tells only a part of the full history of Caribbean–South American interaction. Earlier development involves the emplacement of a fairly continuous belt of oceanic-related and arc-related rocks along northern South America (Fig. 2). It was thought previously that the collision of an arc directly with the northern South American margin had driven Cretaceous metamorphism of these arc complexes more or less in situ (e.g., Maresch, 1974; Beets et al., 1984). However, analysis of sedimentary provenance (e.g., van Houten, 1976) and subsidence history in northern South American sedimentary basins (Dewey and Pindell, 1986; Pindell et al., 1991) strongly supports the idea of a Pacific or Inter-American Arc origin for the allochthons. They occur in structural and basinal settings that record a protracted, west-to-east diachronous (Late Cretaceous along the Andes, Paleogene in northern Colombia–western Venezuela, and Neogene in Eastern Venezuela–Trinidad) history of arc-continent collision and obduction of allochthonous materials onto the South American margin, consistent with the Pacific-origin model (Pindell and Barrett, 1990; Pindell and Erikson, 1994; Pindell and Tabbutt, 1995).

It is generally considered that the South American margin onto which the Caribbean allochthons were obducted was an Atlantic-type passive margin (Speed, 1985; Pindell et al., 1988) that faced the Mesozoic Proto-Caribbean Seaway (Figs. 3 and 4; Pindell, 1985b). However, this generality now needs partial revision, owing to the fact that 60–200 km of north-south shortening has occurred across the Proto-Caribbean Seaway since the Paleocene (Pindell et al., 1988; Müller et al., 1999), starting well before the arrival of the Caribbean plate from the west. We follow Pindell et al. (1991, 1998), Higgs and Pindell (2001), and Pindell and Kennan (2001a, 2001b) and consider that this shortening was taken up by southward-dipping subduction of Proto-Caribbean crust beneath northern South America (see position of this Cenozoic trench on Figs. 5D–G), although subduction has not progressed enough to generate arc magmatism. Hence, at least the Venezuela portion of the South American margin was not passive at the actual time of emplacement of the Caribbean allochthons.

As developed in this paper, the refined tectonic, metamorphic, and geochronologic constraints on Caribbean evolution allow us to better define the controls on basin development along the American (Proto-Caribbean) margins and to separate the geo-

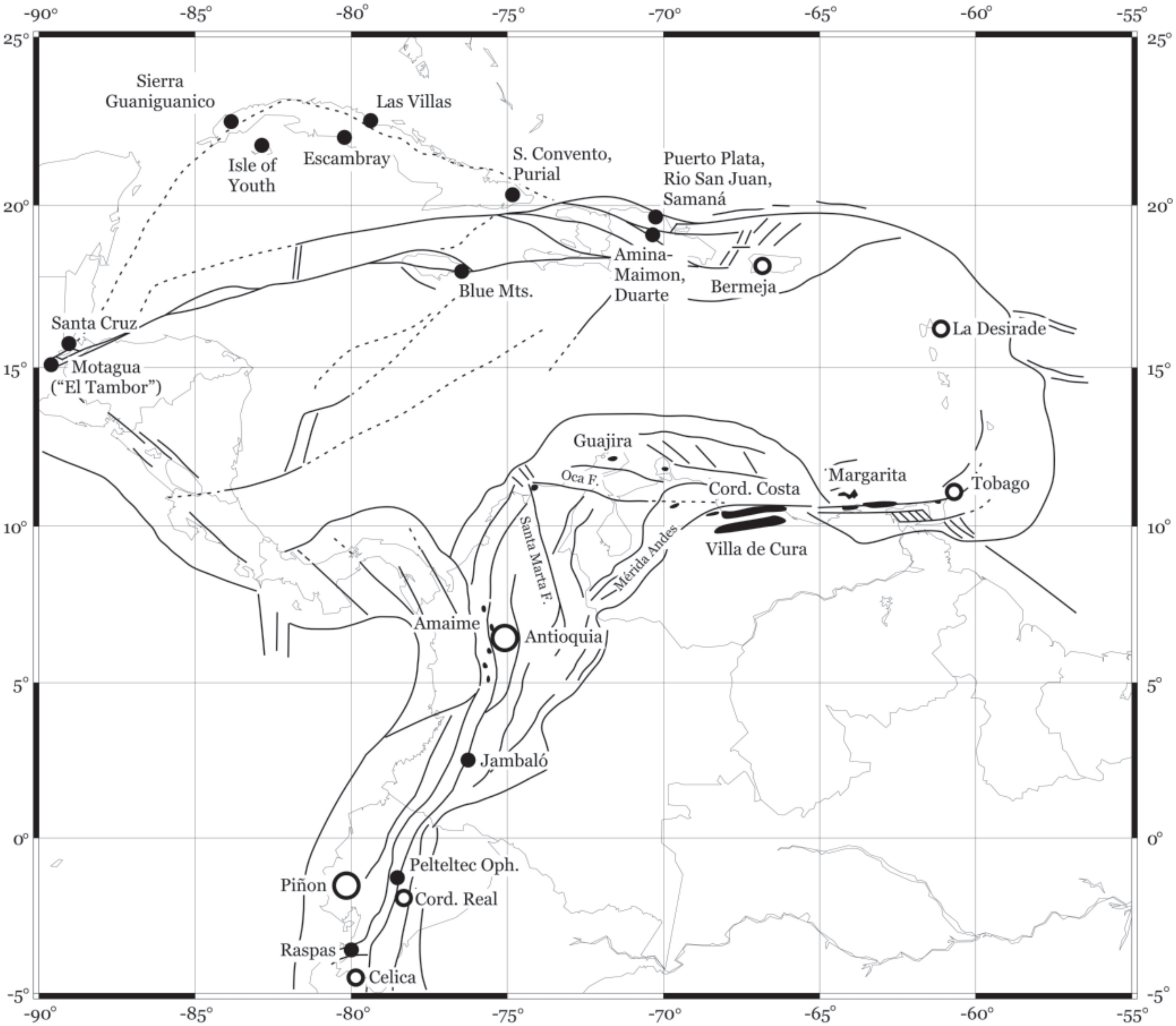

Figure 2. Map of circum-Caribbean region and northern Andes of Peru, Ecuador, and Colombia, showing the locations of high-pressure–low-temperature rocks (filled circles) and other key localities discussed in this paper (unfilled circles). Rocks as far south as −5° are included because the trailing edge of the Caribbean plate was initiated at approximately this latitude prior to northeastward migration during the Late Cretaceous and Cenozoic. F—fault.

logical developments that result from interaction with the Caribbean plate from those that pre-date any Caribbean influence. This paper will provide the framework for more detailed work and for other types of studies.

PROTO-CARIBBEAN RIFTED MARGINS AND CARIBBEAN EVOLUTION

The breakup of western Pangea involved Middle Jurassic to Late Jurassic rifting and seafloor spreading in the Gulf of Mexico, the Proto-Caribbean, and the Colombian Marginal seaways (Figs. 3 and 4). Although spreading in the Gulf of Mexico stopped in the earliest Cretaceous, it continued in the Proto-Caribbean and Colombian Marginal seaways into the Late Cretaceous producing, by that time, a wide oceanic gap between North and South America into which the Caribbean plate was progressively inserted from the Pacific realm. Figures 3 and 4 show modeled positions of the rifts and transfer zones of the various ocean-continent boundaries. The northern Venezuela-Trinidad margin rifted from eastern Yucatán, and the northwestern Colombia margin rifted from southern Mexico–eastern Chortís. The southern Colombia-Peru margin formed the eastern flank of an Early Cretaceous backarc basin

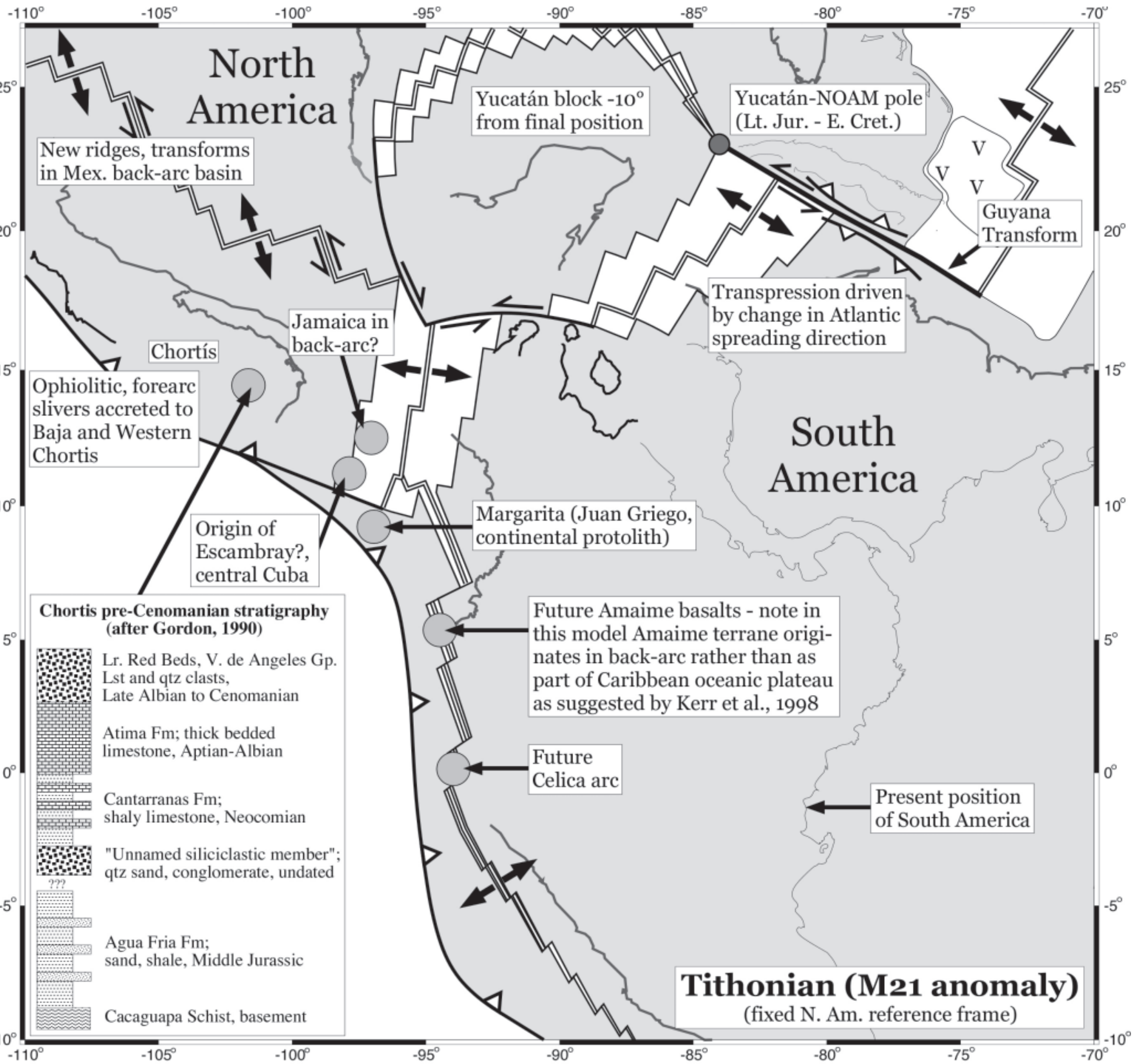

Figure 3. Paleotectonic reconstruction of the Americas at anomaly M-21 (ca. 150 Ma), drawn in a fixed North America reference frame; rotation parameters for the Americas after Pindell et al. (1988), Gulf of Mexico and Caribbean paleogeography modeled according to Pindell and Kennan (2001b), and Pacific plate boundaries schematic but in keeping with the motions of Engebretson et al. (1985). Separation of the Americas was not yet sufficient for Yucatán to have rotated into its final position (it was still constrained by Colombia). Zone of sinistral transform motion where Yucatán and northwest Venezuela interact connects the Proto-Caribbean and Colombian Marginal seaways to east and west. Reconstructed relative positions and contexts of some key localities are shown. Lr.—lower; Lst—limestone; qtz—quartz; V.—Valle.

between the South American autochthon and the continental roots of an arc system that later evolved into part of the Great Caribbean Arc. Incorporated into Figures 3 and 4 is a palinspastic restoration for the northern Andes that retracts Cenozoic transcurrent offsets of 150 km (dextral), 110 km (sinistral), and 120 km (dextral) on the Mérida, Santa Marta, and Oca fault zones, respectively, which collectively requires 180 km of associated orthogonal shortening across the northern part of the Eastern Cordillera of Colombia (Pindell et al., 1998, 2000). To the east, ~80 km of dextral transpressive shear has been removed from the parautochthonous Serranía del

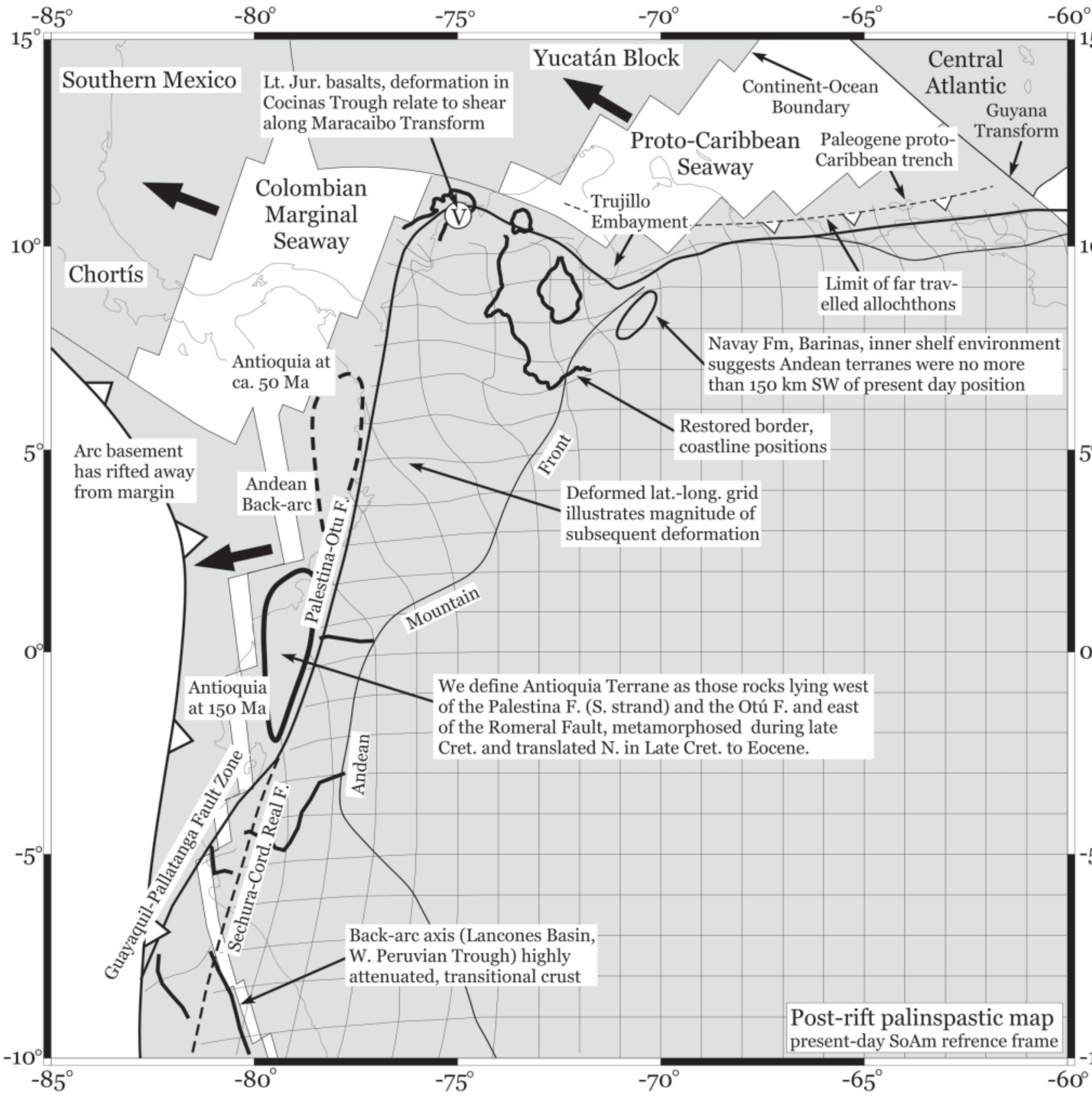

Figure 4. Palinspastic reconstruction of northern South America at 150^Ma (note fixed South America reference frame); rotations as in Figure 3. A deformed latitude-longitude grid (areas of known parautochthonous and autochthonous Jurassic-Cretaceous rocks only) shows the effects of undoing subsequent contraction and strike-slip deformation and provides a more reliable framework for modeling geological evolution. For example, the Guajira Peninsula of northern Colombia probably lay as much as 500 km west-southwest of its present position. The grid ends at the eastern limit of highly allochthonous terranes bounded by the Guayaquil-Pallatanga-Palestina-Otú faults. The Antioquia terrane is shown to the west of the Palestina-Otú fault, and some 500 km south of its present position, in order to lie adjacent to middle Cretaceous synorogenic deposits known in Ecuador but not at its present position in Colombia. To the west of Antioquia are shown the proposed Andean backarc and the arc, rifted entirely from the continental margin. The arc shape shown is drawn to connect with the trench along western North America. F—fault.

Interior–Trinidad foldbelt (Erikson and Pindell, 1998; Pindell and Kennan, 2001a). Along the Andes, for reasons discussed later, it is likely that the "Antioquia terrane" of the northwestern central Cordillera of Colombia has migrated northward an uncertain distance along the Palestina-Otú fault zone. This terrane was overridden by ophiolites and metamorphosed in the Late Cretaceous, but rocks immediately east of the Palestina-Otú fault were not (McCourt et al., 1984), and terrane allochthoneity is potentially large (several hundred km).

Inherent to Figures 3 and 4, and to any model outlining the breakup of Pangea, is the need for a west-facing arc system defining the western margins of the American plates in order to explain the well-known Triassic-Jurassic magmatic belt from at least California to Peru (Jaillard et al., 1990; Jones et al., 1995; Romeuf et al., 1997; Noble et al., 1997). In the Jurassic (Fig. 3), this west-facing arc system must also have spanned the widening gap between Chortís and the Andes, but relative motion along the trench must have become highly oblique by the Early Cretaceous, and arc-parallel transcurrent motions and internal extension were probably severe. It is not clear if a northeast-dipping Benioff zone was maintained, and eventually this portion of the Cordilleran arc system must have reversed its polarity to east-facing (west-dipping subduction) in order for the Pacific crust of today's Caribbean plate to become inserted between the Americas (Fig. 5). Estimates for the time of the reversal range from Aptian (e.g., this paper; Mattson, 1979; Pindell and Dewey, 1982; Snoke, 1991; Draper et al., 1996; Snoke and Noble, 2001), to Campanian (e.g., Burke, 1988; Kerr et al., 1998).

Our "Pacific-origin" model for Caribbean evolution is outlined in Figure 5 and honors the seven plate-kinematic, magmatic, stratigraphic, and paleobiogeographic arguments for the Pacific origin outlined by Pindell (1990). However, "intra-American" models for the origin of the Caribbean crust (i.e., derived from Proto-Caribbean lithosphere that formed between the Americas) persist (Meschede and Frisch, 1998; Kerr et al., 1999; James, 2002). Here, we further affirm the Pacific origin of the Caribbean plate by noting that intra-American origin models predict the following contradictions of Caribbean geology:

1. Inception of arc magmatism in the Greater Antilles Arc would be younger in intra-American origin models than that in the Costa Rica–Panama Arc, whereas the opposite is true (Maurrasse, 1990; Donnelly et al., 1990; Lebrón and Perfit, 1993; Calvo and Bolz, 1994; Stöckhert et al., 1995; Draper et al., 1996; Hauff et al., 2000; Stanek et al., 2000; Maresch et al., 2000).

2. Passive margin conditions in Colombia, western Venezuela, and Yucatán would have ceased by the Albian had the Caribbean arcs formed between the Americas, whereas they persisted well into the Late Cretaceous (Pindell and Erikson, 1994; Villamil and Pindell, 1998; Villamil, 1999; Erlich et al., 2000).

3. The Proto-Caribbean Seaway was not as large as the area of the Caribbean plate until Campanian (Pindell et al., 1988; Müller et al., 1999), but recent seismic data (e.g., Driscoll and Diebold, 1999) argue against seafloor spreading in the Caribbean plate as young as Campanian, which is required in intra-American origin models to explain the size of the Caribbean plate. Rather, the data indicate that ocean-plateau basalts were erupted upon a preexisting ocean floor at ca. 90^Ma, with no detectable plate growth thereafter.

4. If we consider the enormous area of Caribbean lithosphere that has been subducted beneath Colombia and western Venezuela as indicated by seismic tomography (van der Hilst and Mann, 1994), then the Proto-Caribbean was *never* wide enough to have housed the entire Caribbean plate, which must therefore have been situated west of Guajira (Fig. 2) until after Maastrichtian time.

Figure 5 portrays the arc-polarity reversal at the Inter-American Arc as Aptian. In the following section, we review existing Caribbean *P-T-t* data that are critical among our reasons for interpreting the Aptian as the time of initial southwest-dipping subduction in the Great Caribbean Arc, and, hence, of the arc-polarity reversal as well.

P-T-t DATA FROM CIRCUM-CARIBBEAN METAMORPHIC TERRANES

Metamorphic petrology and geochronology of Early to Late Cretaceous rocks of the circum-Caribbean region (Table 1 and references therein) point to an Aptian onset of southwestward dipping subduction beneath the Great Caribbean Arc, and hence to early Aptian subduction-polarity reversal. For several reasons, caution must be exercised when interpreting the data of Table 1. Geochronological systems such as Rb-Sr and K-Ar yield cooling ages only and require careful analysis. Incomplete resetting during multiple metamorphic events may lead to meaningless results. During HP-LT metamorphism, some minerals may incorporate excess argon, leading to higher apparent ages.

Blueschist and other HP-LT assemblages ranging in age from ca. 125 Ma to ca. 70 Ma are found in Guatemala, Cuba, Dominican Republic, Jamaica, northern Venezuela and Colombia, but not in Puerto Rico (the Bermeja complex is greenschist-amphibolite facies meta-ophiolite). Protoliths include fragments

Figure 5 (*following pages*). (A–H) Model for the evolution of the Caribbean region from 120 Ma to 19 Ma, drawn in North American reference frame and simplified from similar maps in Pindell and Kennan (2001b). The model is geometrically viable and satisfies primary geological constraints. In particular, the spreading between North and South America, and the need to keep the Andean margin compressive since 100–120 Ma while the northern Caribbean interacts with southern Mexico, Yucatán, and the Bahamas, places strong constraints on the position, rate of migration and rate of rotation of the Caribbean plate. NOAM—North America; SOAM—South America; CAR—Caribbean plate; FAR—Farallon plate; KUL—Kula plate; NAm—North American plate; SAm—South American plate; HS—hotspot; RRF—ridge-ridge-fault; V—volcanic; Oc—oceanic; CAY—Cayman; CC—Central Cordillera; DR—Dominican Republic; EC—Eastern Cordillera; ESC—Escambray; HA—Haiti; HUAL—Huallaga Basin; MAR—Margarita ; MOC—Mocoa; MS—Muertos Shelf; NC—Northern Cordillera; OCA—Oca Fault; PI—Isle of Pines or Isle of Youth; PR—Puerto Rico; SANT—Santiago Basin; SWH—Southwest Haiti; SMB—Santa Marta–Bucaramanga fault; TOB—Tobago; VIR—Virgin Islands.

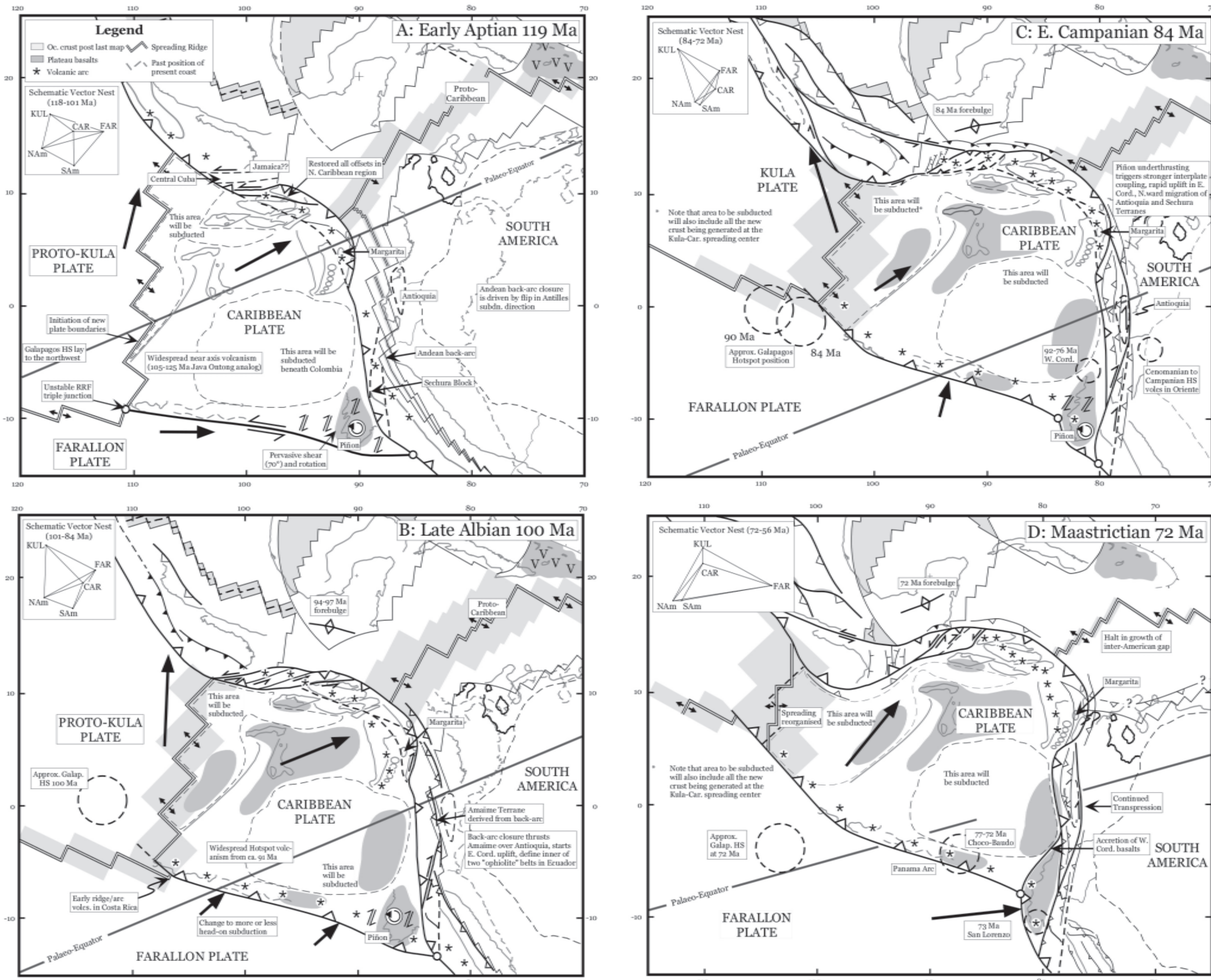

Figure 5 (*this and following page*).

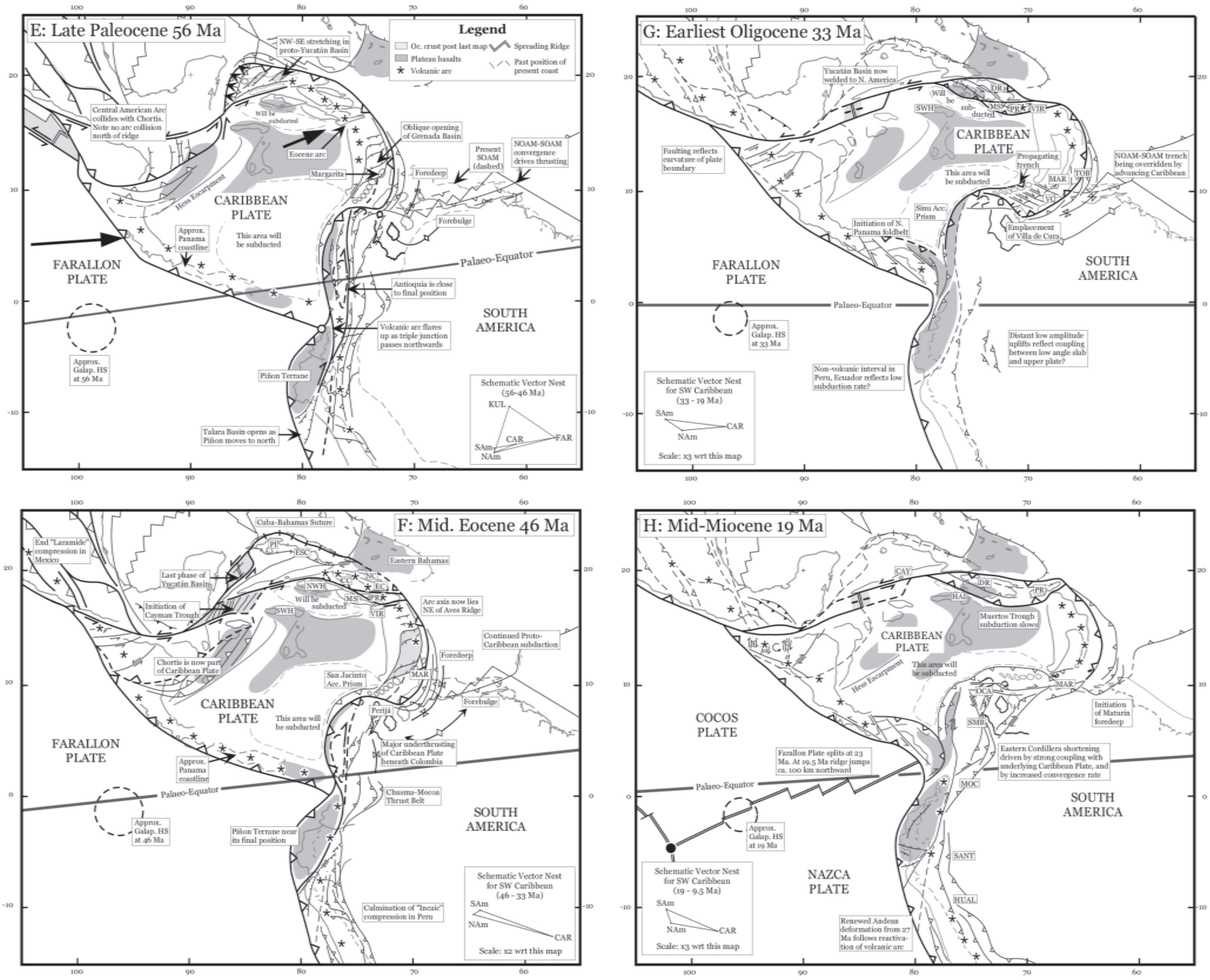

Figure 5 (*continued*).

TABLE 1. TIMING OF METAMORPHISM, IGNEOUS ACTIVITY, AND COOLING IN SUBDUCTION COMPLEXES AROUND THE CARIBBEAN

Complex	Location	Lithology	Protolith	Dating	Interpretation	References
Northern margin						
Motagua Fault Zone	Guatemala	HP blueschists, eclogites and jadeitites as inclusions in fault bounded serpentinites; eclogites more common on south side of MFZ	probably Proto-Caribbean oceanic crust formed at spreading center between southern Mexico and Colombia	116–125 Ma Ar-Ar Phen (south side Motagua); 65–77 Ma Ar-Ar Phen (south side Motagua); 78–63.7 Ma Ar-Ar; 58.5 ± 3.7 Ma K-Ar	The oldest ages may come from preexisting HP-LT rocks from the Pacific-facing trench south of Chortís. However, revised ages for Cretaceous stages indicate an Aptian age for metamorphism, matching both geochronologic and stratigraphic data elsewhere in the Caribbean (Fig. 6), suggesting these too formed during arc-polarity reversal to the southwest of their present location. Ca. 65 Ma ages are interpreted as time of collision between leading edge of Caribbean Arc and southern margin of Yucatán, unroofing HP-LT rocks that formed as Proto-Caribbean oceanic crust subducted beneath the Great Caribbean Arc. Older rocks now lie on the south side of the MFZ associated with slivers of Chortís basement, but regional reconstructions suggest this juxtaposition is probably as young as Oligocene-Miocene, related to eastward motion of Chortís relative to Yucatán during opening of the Cayman Trough farther east.	Bertrand et al., 1978; Harlow, 1994; Harlow et al., 2002; McBirney et al., 1967; Sisson et al., 2003; Sutter, 1979; Gradstein et al., 2004
Blue Mountains	Jamaica	greenschists, blueschists, amphibolites, serpentinites	MORB, probably of Proto-Caribbean origin	48.8 ± 1.3 K-Ar Bt; 52.9 ± 1.4 Ma K-Ar Bt; 76.5 ± 2.1 Ma K-Ar Hbl; all ages reset	Metamorphic rocks underlie Campanian-Maastrichtian arc rocks, cherts. Not clear if relationship is structural or stratigraphic. Overlain unconformably by Eocene strata (post-date latest cooling ages?). If unconformable, then metamorphism and uplift may relate to subduction-polarity reversal event.	Draper et al., 1976; Draper, 1979, 1986; Lewis et al., 1973
Cangre Belt	Western Cuba	blueschist-facies metabasite lenses in metacarbonate and quartz-mica schist	Jurassic (to Lower Cretaceous) passive margin sediments	119 ± 10 K-Ar WM	Blueschist-facies metamorphic belt overthrust north onto weakly metamorphosed Sierra Guaniguanico (allochthonous Late Jurassic passive margin strata) prior to Eocene.	Somin and Millán, 1981, 1972; Millán, 1988
Escambray	Central Cuba	eclogite, blueschist and marble lenses in carbonate-mica and quartz-mica schists; ferruginous metaquartzites	Jurassic to Lower Cretaceous carbonate and clastic sediments and volcanics deposited on southwest margin off the Bahamas platform	160–140 Ma U-Pb Zr; 106–102 Ma U-Pb Zr; 71 ± 5 Ma U-Pb Ti; 65–70 Ma Ar-Ar, Rb-Sr WM; 58–66 Ma FT Zr; 40–60 Ma FT Ap	Protolith includes Late Jurassic to Early Cretaceous marbles and quartzites derived from southeast margin of Chortis Block. Zircons in some eclogite lenses yield Proterozoic upper intercept age, Late Jurassic lower intercept: metamorphism in a pre-Caribbean Late Jurassic subduction environment? Otherwise HP-LT (up to 630 °C, 25 kbar) metamorphism in west-dipping trench at 118–102 Ma followed by uplift. Juxtaposed with Mabujina arc tholeiites before 80–90 Ma intrusion by non-metamorphic igneous rocks. Continued cooling to <350 °C by ca. 65 Ma. Exhumation may have been driven by tangential extension associated with opening of Yucatán basin. Unconformably overlain by middle Eocene chalks. Barrovian metamorphism on Isle of Youth is older than 72 Ma.	Somin and Millán, 1981; Renne et al., 1989; Somin et al., 1992; Stanek, 2000; Grafe et al., 2001; Maresch et al., 2003
Mabujina	Central Cuba	amphibolite-facies metavolcanics, gneisses, metasediments	PIA of the Greater Antillean Arc	>110 Ma Pb-Pb Zr; 87–80 Ma cross-cutting granitic intrusives; 74–72 Ma Rb-Sr, Ar-Ar Bt, WM		Somin and Millán, 1981; Bibikova et al. 1988; Stanek, 2000; Grafe et al., 2001

Continued

TABLE 1. TIMING OF METAMORPHISM, IGNEOUS ACTIVITY, AND COOLING IN SUBDUCTION COMPLEXES AROUND THE CARIBBEAN *(continued)*

Complex	Location	Lithology	Protolith	Dating	Interpretation	References
Northern Serpentinitic Mélange	Central Cuba	eclogite slivers in serpentinitic mélange	probably Proto-Caribbean oceanic crust	115–103 Ma Ar-Ar WM, amphibole; 118 Ma Rb-Sr isochron		Garcia-Casco et al., 2002
Isle of Youth	Offshore southwest Cuba	marble, carbonate-mica, and quartz-mica schists	Jurassic (to Lower Cretaceous) passive margin sediments	72–68 Ma Ar-Ar Bt, WM; 68–60 Ma cross-cutting subvolcanic dikes		Millán 1975; Somin and Millán 1977; Garcia-Casco et al., 2001
Purial	Eastern Cuba	HP-LT metabasite and gneiss	protolith unknown	105–103 U-Pb Zr; 82 Ma K-Ar isochron	Basement of the Oriente nappe stack	Hatten et al., 1989; Somin et al., 1992
Puerto Plata, Rio San Juan, Samaná	Northern Dominican Republic	garnet-bearing blueschists and eclogite, metagranitoids, serpentinite mélange	MORB, probably of Proto-Caribbean origin	ca. 100 Ma Ar-Ar on glaucophane (?); 120 ± 100, 125 ± 50, and 250 ± 100 Ma, all K-Ar on glaucophane; 78 ± 30 Ma Sm-Nd eclogite, 90 ± 10 Ma Rb-Sr eclogite, 98 ± 3 Lu-Hf eclogite, 70 Ma and 85 Ma Rb-Sr isochrons on metagranites; 85 ± 2 Ma Ar-Ar hornblende; 62 ± 2 Ma Ar-Ar phengite; 37–39 Ma K-Ar phengite; 25–50 Ar-Ar phengite	98 Ma Lu-Hf age pinpoints time of eclogite formation thought to be associated with post-flip westward subduction. Suggests polarity reversal is older than 100 Ma. Large spread in ages and peak metamorphic *P-T* conditions points to complex arrays of *P-T-t* paths in subduction zone. Mid-crustal levels generally reached by Eocene.	Joyce and Aronson, 1989; Draper and Nagle, 1991; Krebs et al., 1999, 2003; Perfit and McCulloch, 1982; Gonçalves et al., 2000; Catlos and Sorensen, 2003
Duarte	Central Dominican Republic	metagabbros, metabasalts, cherts	mixed ages, Oceanic Plateau Basalt affinities	86.1 ± 1.3 Ma Ar-Ar, 86.7 ± 1.6 Ma Ar-Ar, both on plateau basalts (Caribbean origin?); 123–148 Ma K-Ar on associated meta-tonalites and gabbros	Duarte complex is highly heterogenous. Oldest components may be Jurassic oceanic plateau with cherts (possible correlation to early Caribbean Piñon of Ecuador) or fragments of early Cretaceous forearc (basement to Amina-Maimón arc). Older ages may be misleading due to later structural juxtaposition, excess Ar, etc.	Lapierre et al., 1999; Draper and Lewis, 1989, 1991
Amina-Maimón Los Ranchos	Central Dominican Republic	greenschist metavolcanic mylonite, metavolcanics, volcaniclastics	Neocomian Primitive Island Arc	metamorphism and penetrative deformation must predate 110 Ma	Early Cretaceous arc is metamorphosed (>10 km depth), uplifted, and eroded prior to ca. 110 Ma Hatillo Limestone, In turn overlain by Tireo calc-alkaline arc is 93–89 Ma ± younger dacites.	Draper et al., 1996
La Desirade	Northern Lesser Antilles	pillow basalts, cherts, plagiogranite, basic dykes		147–140 Ma U-Pb Zr; end-Jurassic (faunal age)	Associated with Pacific affinity cherts. Probably formed at a Pacific spreading center and represents basement of Caribbean plate.	Mattinson et al., 1980; Montgomery et al., 1994

Continued

TABLE 1. TIMING OF METAMORPHISM, IGNEOUS ACTIVITY, AND COOLING IN SUBDUCTION COMPLEXES AROUND THE CARIBBEAN *(continued)*

Complex	Location	Lithology	Protolith	Dating	Interpretation	References
Southern margin						
North Coast schist	Tobago	greenschist metavolcanic rocks, volcaniclastic rocks	pyroclastic, epiclastic fringe to (Primitive?) Island Arc	120 Ma or older Ar-Ar from PIA; 105–103 Ma Ar-Ar from TVG and TPS; 113–102 Ma K-Ar from TVG, TPS	Aptian and older PIA rocks are metamorphosed prior to late Aptian–Albian plutonism and volcanism, also of “primitive island arc” character.	Snoke et al., 2001; Snoke et al., 1990; Speed and Smith-Horowitz, 1998
La Rinconada	Margarita, Venezuela	metabasalts, metagabbros and ultrabasics; trondhjemites	MORB, probably of Proto-Caribbean origin	116–109 Ma U-Pb Zr concordant	Continental fragment juxtaposed after 109 Ma with MORB formed in backarc or Colombian marginal seaway in a subduction zone. No similar aged MORB known from Caribbean Plate. Peak metamorphic conditions of up to 650 °C/20 kbar reached before 86 Ma when El Salado granite was intruded, still at ~25 km depth. We infer HP-LT metamorphism in west-dipping subduction zone immediately after subduction polarity reversal, followed by exhumation continuing into Oligocene.	Stöckhert et al., 1995; Maresch et al., 2000
Juan Griego	Margarita, Venezuela	quartz-feldspar gneiss and garnet-mica schists, granitoid intrusives, marbles	continental sedimentary and/or igneous rocks	315 Ma U-Pb Zr upper intercept augen gneiss; 86 Ma U-Pb Zr and 90–80 Ma Ar-Ar WM on unmetamorphosed El Salado Pluton; 66 Ma Ar-Ar amph in dyke; 55–50 Ma K-Ar WM in recrystallized schist; 53–36 Ma Zr FT; 23 ± 2 Ma apatite FT		Stöckhert et al., 1995; Maresch et al., 2000
Villa de Cura	Venezuela	metavolcanics,	volcanic island arc and associated clastic fringe	79.8 ± 0.4 Ma Ar-Ar WM for the three N zones; 96.8–96.3 Ma Ar-Ar barrositic amphibole; 91.1–89.5 Ma Ar-Ar WM for barroisite zone; 101–77 Ma K-Ar amph; 87–82 Ma K-Ar WM; 107–98 Ma K-Ar Hbl from ultrabasic intrusive complex (excess Ar?); 43.4–41.9 Ma Zr FT on Tinaco diorite 10 km to N; 6.1 Ma Ap FT on Tinaco diorite 30 km to W	HP-LT metamorphism of a >100 Ma protolith in forearc to west-dipping subduction zone. Adjacent bodies of rock reached peak metamorphism between 96 Ma and 78 Ma. The significance of 107–98 ages from associated non-metamorphic ultrabasic intrusives remains unclear. Ongoing cooling and unroofing prior to emplacement during Oligocene-Miocene. We consider it unlikely that a subduction polarity reversal could have occurred in the interval 96–78 Ma, ruling out trench “choking” by Caribbean Igneous Plateau at ca. 90 Ma as proposed by others.	Smith et al., 1999; Hebeda et al., 1984; Skerlec and Hargraves, 1980; Kohn et al., 1984

Continued

TABLE 1. TIMING OF METAMORPHISM, IGNEOUS ACTIVITY, AND COOLING IN SUBDUCTION COMPLEXES AROUND THE CARIBBEAN *(continued)*

Complex	Location	Lithology	Protolith	Dating	Interpretation	References
Cordillera de la Costa	Venezuela	serpentinites, leucotonalitic gneisses, amphibolites, actinolite-garnet-mica-chlorite schists, blueschists (“oceanic assemblage”); graphite-garnet-mica schists, marbles, quartzites, quartz-feldspar gneisses (“continental assemblage”); metagranites and augengneisses	Very heterogeneous “Oceanic assemblage” comprises dismembered meta-ophiolite, juxtaposed with “continental assemblage” of probable passive margin origin; deformed and metamorphosed intrusive rocks; some gneisses may be basement to passive margin strata	494 ± 52, 467 ± 52 Ma U-Pb on “basement gneisses”; 753, 735, 155, 32 Ma K-Ar on amphibolites; 77–76 Ma K-Ar from metadiorites, possibly originally cross-cutting intrusives?; AFT and ZFT data show protracted cooling history into the late Cenozoic	Blueschist metamorphism probably occurred at west-dipping subduction zone, as also inferred for Margarita and Villa de Cura HP-LT rocks. Relationships of HP-LT rocks to greenschist-grade “passive margin” protoliths are not clear, nor is the original depositional location of these “passive margin” rocks: are they para-autochthonous parts of the central Venezuelan margin or do they derive from farther west?	Morgan, 1967, 1970; Avé Lallemant and Sisson, 1993; Sisson et al., 1997
Northwest Guajira	Northern Colombia	garnet-bearing eclogite, greenschists, amphibolite, serpentinite mélange		48 ± 4 Ma K-Ar ?whole rock? on diorite stock that post-dates metamorphism, emplacement; unroofing recorded by eclogite clasts in ?Oligocene conglomerate	Metamorphism and initial unroofing during middle or late Cretaceous (unconformably overlain by Maastrichtian strata offshore) emplacement of leading edge of Caribbean in latest Cretaceous, prior to Eocene emplacement or unroofing of non-metamorphic stock.	Green et al., 1968
<u>Western South America</u>						
Antioquia	Central Colombia	HT-LP (Abukuma) pre–220 Ma	Paleozoic clastic sediments	locations for data not known; 120–108 Ma K-Ar; 75–57 Ma K-Ar	Early cluster may relate to proposed backarc closure. Late cluster may relate to exhumation driven by Caribbean underthrusting. Overlain by early Albian (ca. 110–100 Ma) sediments prior to overthrusting by Amaime Terrane.	McCourt et al., 1984; Bourgois et al., 1987
Amaime	Central Colombia	basalts, pillow lavas, ?komatiites?		163 ± 10–131 ± 9 Ma ?K-Ar WR?; 126 ± 12 Ma K-Ar Hbl (all above from Cauca Ophiolite); intruded by 113–99 Ma Buga Batholith (?post-metamorphic?)	Pre-Albian age and intrusion by arc (?) Buga Batholith suggest formation within narrow Andean backarc during Early Cretaceous. Nappes emplaced east onto the Central Cordillera after 110–100 Ma.	McCourt et al., 1984; Restrepo and Toussaint, 1974; Restrepo and Toussaint, 1976
Jambaló	Central Colombia	HP-LT metamorphics	not known	125 ± 15 Ma K-Ar WR; 104 ± 14 Ma K-Ar WR	Possibly protolith or metamorphic–uplift ages. Here we interpret driving mechanism for metamorphism to be eastward overthrusting of arc over backarc basin. Associated with eastward thrusting of Amaime (backarc basalts).	Orrego et al., 1980; Feininger, 1982; De Souza et al., 1984

Continued

TABLE 1. TIMING OF METAMORPHISM, IGNEOUS ACTIVITY, AND COOLING IN SUBDUCTION COMPLEXES AROUND THE CARIBBEAN *(continued)*

Complex	Location	Lithology	Protolith	Dating	Interpretation	References
Raspas	SW Ecuador	garnet-kyanite–bearing schists, omphacite-garnet eclogites, sheared serpentinites	N-MORB, E-MORB, quartzose and shaley clastic sediments of continental origin	221–228 Ma, method not known; continental protolith age; 132 ± 5 Ma K-Ar phengite; 80–70 Ma reset ages from adjacent non–HP-LT metamorphics	Highly sheared. Generally south-dipping structures may reflect Late Cretaceous accretion of Piñon Terrane. Context of ?Barremian-Aptian blueschist uncertain—does it indicate polarity reversal and terrane accretion earlier in south than north?	Malfrère et al., 1999; Arculus et al., 1999; Aspden et al., 1995; Feininger and Silberman, 1982
Other relevant data						
Cordillera Real	Ecuador	granitoids, metavolcanics, and clastic metasediments	backarc and arc granites and transitional crust	144 Ma U-Pb Zr is youngest intrusive age; 125–145 Ma K-Ar ages may be partial reset; 60–100 Ma K-Ar ages are widespread	We suggest that ca. 145 Ma marks time of final opening of backarc, rifting the arc away from the margin. From 100 Ma, backarc was closed, associated with dextral transpressive deformation, accretion of Piñon Terrane (overlain by late Cretaceous quartzose sediments), and large magnitude clockwise rotation.	Litherland et al., 1994 Noble et al., 1997
Celica-Ayabaca	Ecuador, Peru	andesite		113 Ma K-Ar ?WR? on cross-cutting pluton	Aptian or older andesitic Celica volcanics with continental signature are intruded by Late Aptian plutons. Overlain unconformably by Albian-Cenomanian volcaniclastics. No clearly Albian or younger arc (e.g., lavas, pyroclastics). Ayabaca volcanics of N. Peru are similar. Not dated but note no significant volcanic influence in Albian and younger sediments in nearby West Peru Trough.	Kennerley, 1980; Berrones et al., 1993; Jaillard et al., 1996
Casma	Peru	greenschist	andesites, volcaniclastics	102 Ma K-Ar on granitoids	Greenschist metamorphism of Aptian?-Albian volcaniclastics predates late Albian, 102 Ma, plutonism.	Cobbing et al., 1981

Note: HP-LT—high-pressure–low temperature; MORB—mid-oceanic-ridge basalt; E-MORB—enriched MORB; N-MORB—normal MORB; amph—amphibole; Ap—apatite; Bt—biotite; Hbl—hornblende; Lu-Hf—letetium-hafnium; Phen—phengite; Ti—titanite; WM—white mica; WR—whole rock; Zr—zircon; AFT—apatite fission-track age; FT—fission track; MFZ—Motagua Fault Zone; PIA—Primitive Island Arc; TPS—Tobago Plutonic Suite; TVG—Tobago Volcanic Group; ZFT—zircon fission-track age.

of Paleozoic continental crust (Margarita) and continent-derived sedimentary rocks that formed the basement or forearc of the Inter-American Arc (which formed as the Americas separated in the Late Jurassic and Early Cretaceous), as well as volcanic ocean floor and arc rocks.

In Guatemala, HP-LT rocks are found on north and south sides of the Motagua Fault Zone (Harlow et al., 2003), associated with the dismembered ophiolite, serpentinite bodies mapped the "El Tambor Group." Recent mapping shows that jadeitites and albitites are common in both north and south, but that eclogite and blueschist are much more common in the southern exposures. Ar-Ar ages on phengites are consistently 116–125 Ma in the south and, north of the Motagua Fault, are consistently 65–77 Ma (Sisson et al., 2003). In both cases, these ages are interpreted to be at or near peak-metamorphism and, together with the lithological contrast, suggest that two discrete belts of HP-LT rocks formed during two discrete events and may have been separated by up to several hundred kilometers prior to Cenozoic sinistral motion on the Motagua Fault. In both belts, the nature of the protolith is unclear (possibly volcanic rocks of primitive arc or oceanic origin). The original position of the southern belt with respect to the Great Caribbean Arc at the time of subduction polarity reversal is also unclear, such that this older belt may have formed very early in the polarity reversal event on the north flank of the arc or may contain rocks metamorphosed at the older west-facing trench prior to the reversal event. In neither belt is there any age data indicating a significant event occurring at ca. 90 Ma.

In Cuba, Late Jurassic equilibration of the U-Pb isotopic system is found in the zircons of some rare eclogites (Grafe et al., 2001), but the nature (i.e., magmatic, metamorphic, etc.) of the thermal event is at present enigmatic. We interpret this older material to have been incorporated into the Inter-American Arc (Fig. 3) as it swept past the southern end of the Chortís block (where Jurassic igneous rocks in a clastic and carbonate sedimentary sequence similar to the Cuban protoliths are present; Gordon, 1990). Primitive island-arc (PIA) volcanic protoliths are found in Cuba (Los Pasos Formation; Stanek et al., 2000) and Tobago (North Coast Schist; Snoke et al., 2001) with minimum ages of 110^Ma, and some older metatonalites and metagabbros are known. Pacific-derived Jurassic mid-oceanic-ridge basalt (MORB) crust and associated chert is present in La Desirade. MORBs (La Rinconada unit) erupted at 116–109 Ma are found in Margarita (Stöckhert et al., 1995), which is consistent with an origin at the spreading center in the Colombian Marginal Seaway, to the east of the Intra-American oceanic island arc (Fig. 5).

In central Colombia and Ecuador, blueschists are also found as highly sheared fault-bounded slivers along the Romeral-Peltetec Suture and in the El Oro Belt of southern Ecuador. K-Ar ages from Raspas (southern Ecuador) and Jambaló (southern Colombia) range from 125 to 132 Ma (Aspden and McCourt, 1986; Aspden et al., 1992, 1995). Both the field relationships and the ages are less easy to interpret than those in Cuba and Margarita, and it is not clear if the HP-LT rocks here directly relate to the polarity-reversal event. In the El Oro Belt of southernmost Ecuador, the Raspas blueschists are found north of both an Early Cretaceous arc and associated continental fragments and, if we account for at least 60°–90° of clockwise rotation (Mourier et al., 1988a), appear to lie on the west side (Pacific-facing) of the Early Cretaceous arc. Farther north, in central Ecuador, fragments of continental to transitional crust found to the west of the Peltetec ophiolite suture are inferred to be the trace of the "Andean backarc" or southern part of the "Colombian Marginal Seaway" (Fig. 4). However, the Jambaló blueschists are found farther north in Colombia and have no associated continental or arc fragments to the west or farther north along the trace of the Romeral Fault. Thus, although the blueschists might have originated during oblique closure of the Andean backarc basin, with the ages indicating onset of polarity reversal prior to the Aptian, it is also possible that the blueschists could have formed on the southwest side of the older, pre-reversal Inter-American Arc. If they did form on the southwest side of that arc, they must have been sheared into place during the intense southwest-northeast dextral slip that occurred in northern South America during the Cretaceous at the southeastern margin of the Caribbean plate.

Throughout the circum-Caribbean region, HP-LT metamorphism could have occurred at either the east-dipping subduction zone (prior to polarity reversal) or the west-dipping subduction zone (after polarity reversal), and we would expect significant unroofing to occur during the arc-polarity reversal event itself, because of the probability of transient doubly-vergent underthrusting beneath the arc. We conclude for the following reasons that the reversal occurred at, or at most a few million years before, 115–120 Ma:

1. With the possible exception of the rare Late Jurassic zircon ages from Cuba noted above (probably protolith ages) and a few slightly older K-Ar or Ar-Ar ages found in Andean terranes and in Guatemala, no HP-LT metamorphic rocks with ages older than ca. 120 Ma are known in the circum-Caribbean.

2. Potentially rapid burial and unroofing events are recorded in Cuba, Hispaniola, and Puerto Rico, where greenschist-grade metamorphic rocks were exhumed and then unconformably overlain by Albian limestones, and in Tobago, where the greenschist-grade North Coast Schist was intruded by the unmetamorphosed Tobago Volcanic Suite at ca. 105 Ma (Snoke et al., 2001). Part of the Villa de Cura Complex in Venezuela may also have been buried and exhumed at this time as metamorphic components of the Villa de Cura were intruded by unmetamorphosed 108–98 Ma ultrabasic rocks (Hebeda et al., 1984; Smith et al., 1999).

3. In *P-T-t* loops of HP-LT metamorphic rocks around the Caribbean, the 90–20 Ma interval is generally characterized by exhumation processes and not interrupted by burial or prograde metamorphic events, consistent with polarity reversal occurring before that time. Nowhere is peak HP-LT metamorphism recorded at ca. 85^Ma, as would be expected if subduction of the buoyant ca. 90 Ma Caribbean Large Igneous Province was the cause of polarity reversal, as argued by Burke (1988) and Kerr et al. (1998).

Numerous processes may lead to exhumation of HP rocks in the Great Arc, including arc-parallel stretching (mainly Late Cretaceous), obduction of forearc materials onto rifted continental margins (Late Cretaceous–Paleogene), low-angle detachment at intra-arc basins (Maastrichtian–Paleocene at Yucatán Basin), extensional detachment during creation of intra-arc pull-apart basins (early to middle Eocene at Grenada Basin), and possibly vertical extrusion along high-angle transpressive faults (Late Cretaceous in Andes terranes). All of these possibilities are discussed below.

APTIAN SUBDUCTION POLARITY REVERSAL: INTER-AMERICAN ARC BECOMES THE GREAT CARIBBEAN ARC

Constraints on Timing

Seven independent lines of evidence point to the Aptian as the time of the subduction polarity reversal that transformed the Inter-American Arc (southwest-facing) into the Great Caribbean Arc (northeast-facing). At no other time in the histories of the various parts of the Caribbean arc system was there such a profound change in depositional, magmatic, structural, and geochemical conditions (Fig. 6). Furthermore, the changes that occurred were very similar in many of the older arc fragments. This suggests that the various arc fragments once lay much closer together than they do today, an argument that itself points to the Early Cretaceous, when the gap between the Americas was still relatively small.

Onset of HP-LT Metamorphism

As shown in Table 1, circum-Caribbean HP metamorphic rocks indicate peak metamorphism dating back to the Aptian (119–112 Ma). The older age from the Escambray complex, Cuba (Table 1), appears to be exotic to the Inter-American Arc and Neocomian ages from the Motagua Fault Zone, and the northern Andes

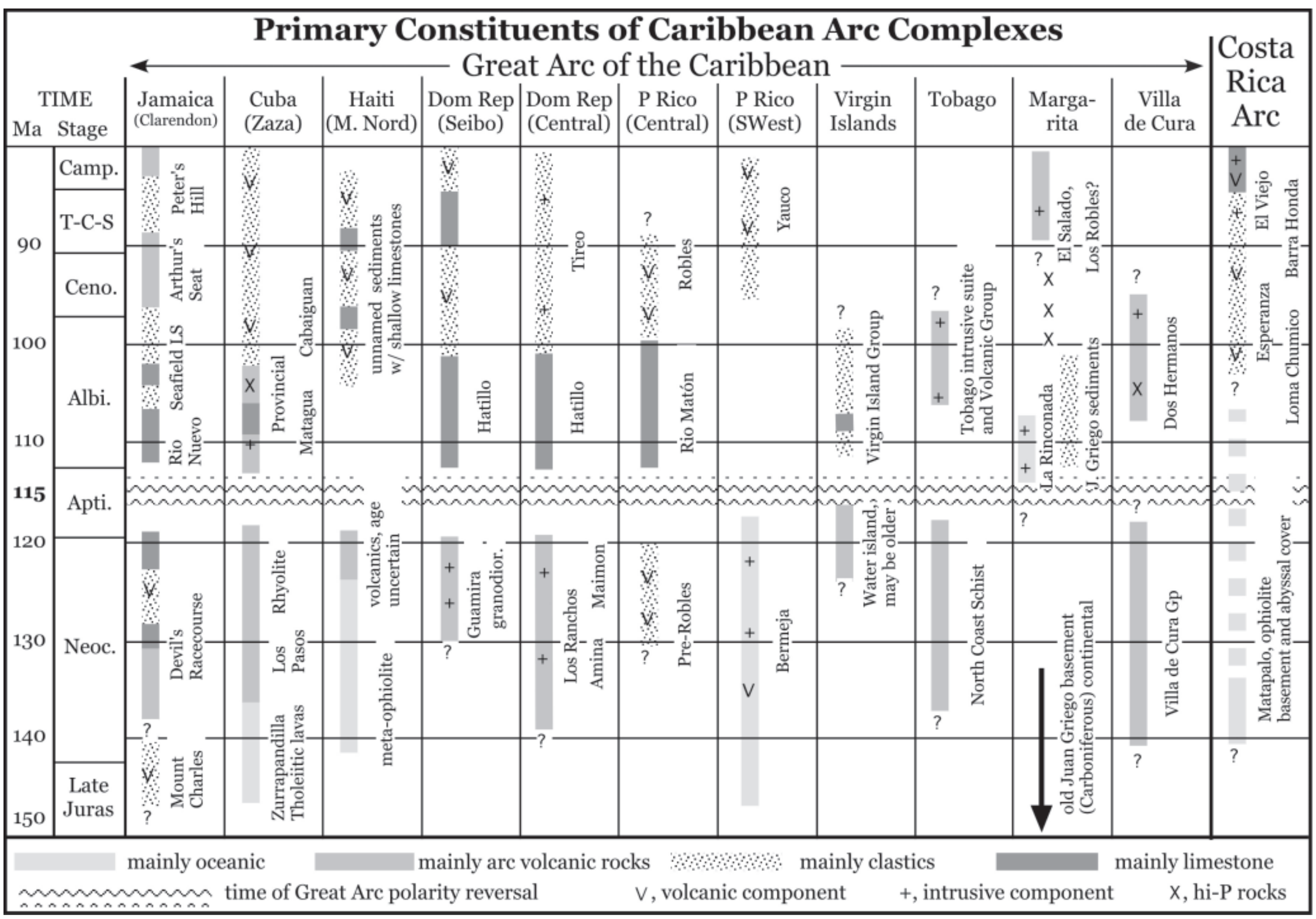

Figure 6. Main aspects of magmatic and stratigraphic development in Caribbean arcs. The most significant change during the Cretaceous occurred at ca. 120–115^Ma. Note lack of significant changes at 80–90^Ma, a time proposed for polarity reversal by other authors. Prior to the Aptian subduction polarity reversal and associated hiatus, nearly all Caribbean terranes show evidence of primitive island-arc magmatism, but it is not clear if these formed at a west-facing Inter-American Arc (requiring polarity reversal), or if they were dragged across the inter-American gap by transform motion. The complete lack of continental influence in the pre-Aptian stratigraphies of the arc pieces argues for an intra-oceanic arc (formed in the Inter-American gap), rather than an arc that was sheared away from Chortís. Camp.—Campanian; T-C-S—Turonian-Coniacian-Santonian; Ceno.—Cenomanian; Albi.—Albian; Apti.—Aptian; Neoc.—Neocomian; Juras—Jurassic; M. Nord—middle and north.

may be from HP-LT rocks that predate the polarity reversal and were emplaced along strike-slip faults at the flanks of the Caribbean plate. We interpret the Aptian onset of HP metamorphism in Cuba and Margarita as dating the inception of the southwest-dipping Benioff zone. Most of the Aptian and younger HP ages are from rocks on the northeast side of the Great Caribbean Arc; therefore, if the polarity reversal occurred any later than Aptian, it would have been necessary for HP rocks on the southwestern flank of the arc to pass beneath the higher temperature arc axis to be exhumed on the northeastern flank of the arc, which we think highly unlikely.

Hiatus in Volcano-Sedimentary History

In areas of the Great Caribbean Arc where pre-Albian arc rocks are present, a strong Aptian hiatus separates the older, usually metamorphosed rocks from little- or non-metamorphosed Albian and younger rocks (Fig. 6). Possible causes for this hiatus are (1) strong uplift in the wedge of arc material above the opposing Benioff zones during the transient arc polarity reversal, (2) tectonic unroofing due to arc-parallel extension (Draper, 2001), and (3) erosion at localized transpressional(?) uplifts along the Inter-American Arc prior to polarity reversal (or a combination of both).

Development of a Limestone Platform upon Parts of the Aptian Hiatus

In Cuba, Hispaniola, and Puerto Rico, the units beneath the Aptian hiatus are commonly metamorphosed to or near to greenschist grade and unconformably overlain by the unmetamorphosed Albian, shallow water, fossil-bearing Provincial, Hatillo, and Río Matón limestones, respectively. In Tobago, the greenschist-grade North Coast Schist (of PIA geochemical character; Jackson et al., 1988) is also deeply eroded and overlain by extrusive volcanics and intruded by plutons, both of the less-metamorphosed Tobago Volcanic Group, dated paleontologically and isotopically as middle Albian (Snoke et al., 2001; Snoke and Noble, 2001).

Shift in the Positions of Magmatic Axes

In at least the Hispaniolan and possibly the Cuban portions of the Great Caribbean Arc, the magmatic axis appears to have shifted southwest from the pre-Aptian to the Albian. In Hispaniola, we consider the Los Ranchos and Maimón Formations as pre-reversal arc and the Tireo Formation, which lies some 20–60 km southwest of Los Ranchos Formation (Bowin, 1966; Lewis and Draper, 1990), as post-reversal arc. In central Cuba (which we suggest represents only the forearc of the Great Arc; Paleocene opening of the Yucatán [intra-arc] Basin separated the Cuban forearc from the bulk of the Great Arc's magmatic axis beneath Cayman Ridge), the occurrence of the Los Pasos Formation (primitive, pre-reversal arc lavas) some 10–30 km northeast of the Cabaiguan Formation (evolved, post-reversal arc lavas) (Stanek, 2000) might suggest a southwesterly shift in the magmatic axis as well. The shift may also have occurred at the Aves Ridge portion of the arc, such that pre-reversal arc lithologies such as the La Desirade Complex (Mattinson et al., 1980) and the North Coast Schist of Tobago have occupied a forearc position since the reversal. The shift in the magmatic axis may also explain why HP-LT metamorphic rocks from the pre-reversal subduction zone have not clearly been recognized. The post-reversal arc may have buried and/or destroyed them.

Change in Arc Magmatic Chemistry

Magma chemistry is dramatically different above and below the Aptian discontinuity in some parts of the arc. Donnelly et al. (1990) and Lebrón and Perfit (1993) pointed out that in the Greater Antilles, pre-Albian arc magmas are generally PIA tholeiites, whereas Albian and younger magmas are mainly calc-alkaline (CA). The change from PIA to enriched CA magmas may pertain to differing sedimentary and/or crustal materials entering the Benioff zone before and after the reversal (Lebrón and Perfit, 1993; Pindell, 1993).

Emplacement of Nappes in Hispaniola

Draper et al. (1996) showed that the mylonitization and inverted metamorphism in the Maimón Schist and the development of penetrative fabrics in the Neocomian Los Ranchos Formation of central Hispaniola record a significant orogenic event, probably the northward thrusting of peridotites over these two units. Because the two units are overlain by the undeformed lower middle Albian Hatillo Limestone, the deformation must have occurred during the Aptian to early Albian. Draper et al. (1996) concluded that the timing of deformation coincides with the timing of the change in arc magmatism from PIA to CA, and that the orogenic event was related to arc-polarity reversal.

Geometric Simplicity

At 120 Ma (start of Aptian), the Inter-American Arc was very short and likely quite straight, bridging the juvenile gap between Mexico and Ecuador. By 80 Ma (Campanian), the arc would have been nearly 2000 km long and likely quite arcuate or cusped. We consider that reversal would have been far more complex, longer-lived, and comprised several significant sub-events once a long and arcuate arc geometry had been established. In the Campanian, it is difficult to pick out *any* regional arc event (Fig. 6), let alone one as complex as might be expected were the arc 2000 km long.

Cause of the Arc-Polarity Reversal

Much of the crust of the Caribbean plate is ~20 km thick (Case et al., 1990), and widespread basaltic extrusions (Caribbean basaltic plateau marked by seismic reflector B″) are believed to be at least partially responsible for the abnormal thickness. Burke et al. (1978) considered that this abnormal thickness would cause excessive buoyancy relative to "normal" oceanic crust. They, as well as Livaccari et al. (1981), Burke (1988), Kerr et al. (1998), and White et al. (1999) invoked choking of the west-facing (inter-American) trench by the arrival of this abnormally thick Caribbean crust as the cause for (Santonian–Campanian) polarity reversal. However, if the Caribbean plateau is 90^Ma, then it could not have driven our Aptian polarity reversal. But we are so confident in the Aptian age for polarity reversal (or onset of west-dipping subduction) that we doubt the plateau-trench choking hypothesis altogether. Large regions of Caribbean lithosphere *have* been subducted, such as the 900 km long Benioff zone imaged by seismic tomography beneath the Maracaibo region (van der Hilst and Mann, 1994), the "pre-subduction" area of which is shown on Figure 5.

Choking by very young (and therefore buoyant) crust at the Cordilleran trench is another possibility. However, seafloor spreading must also have been ongoing in the Proto-Caribbean Seaway at the time of the reversal, such that the age of crust on the east side of the arc may have been, if anything, younger than that entering the trench on the west. Another feasible hypothesis is that an east-facing intra-oceanic arc from farther out in the Pacific arrived at the west-facing Inter-American Arc and that the ensuing arc-arc collision was "won" by the east-facing arc, but so far parts of only a single pre-Albian magmatic axis have been recognized. Still another possibility is that east-dipping subduction drastically accelerated in the Aptian, but subduction *rate* is not often considered to be a strong control on tectonic style in the overriding plate (Jarrard, 1986).

Pindell (1993) noted the above timing discrepancy and recognized that Aptian arc polarity reversal was analogous to, and coeval with, the onset of compressive arc behavior (sensu Dewey, 1980, where trenchward advance of the overriding plate exceeds trench rollback, thereby increasing plate coupling and triggering backarc thrusting). Such compressive arc style is seen in Mexico (Sedlock et al., 1993), the Sevier fold-and-thrust belt of the western United States, and the West Peruvian Trough along the Andes (closure of backarc; Jaillard, 1994). This suggested a hemispheric-scale driving mechanism that spanned two continents on different plates and led to the proposal that the Aptian initiation of opening of the Equatorial Atlantic (Pindell, 1985b) and an Aptian doubling of spreading rate in the Central Atlantic (Klitgord and Schouten, 1986) caused the onset of compressive arc conditions. Such a profound change in the style of Cordilleran arcs over such large areas of two continents suggests to us that arrival of an oceanic plateau, or even a collection of plateaus, was not the driver of the Inter-American Arc polarity reversal. We favor the Aptian acceleration of Atlantic spreading as the dominant cause of the hemispheric scale onset of arc compression and reversal of the Inter-American Arc, which, upon reversal of its polarity, became the Great Caribbean Arc.

Our model, then, requires that the Caribbean Plateau basalts erupted onto a swath of Pacific-derived crust that was progressively engulfed by the American continents. In addition, Calvo and Bolz (1994) cite evidence for the Albian (ca. 100 Ma) onset of arc magmatism (calc-alkaline volcaniclastic turbidites) in Costa Rica. If this is true, then the Caribbean plate was already defined by plate boundaries in the Albian, at least three of which were subduction zones dipping Caribbean-ward at the time of the plateau basalt extrusion (Fig. 5). At the time of plateau basalt eruption, these subduction zones were relatively young, and subducted slabs did not penetrate deep enough to restrict plumes rising from the deep mantle. Kerr et al. (1998) propose that a subsidiary pulse of Caribbean basalt extrusion occurred at 72–78^Ma, which postdates the arc's collision with both the Yucatán and Colombian margins and thus must be well after reversal. Thus, we do not see any serious restraint on whether the main 90 Ma basalt pulse occurred before or after the reversal and, in our view, the extrusion of the Caribbean basalts is not an argument for or against the timing of the reversal. Central and southern Texas and the Oriente Basin of Ecuador also experienced basaltic magmatism between 90 and 100 Ma (Byerly, 1991; Barragán and Baby, 1999) that clearly had nothing to do with the Caribbean plate.

Paleogeographic Depiction of the Subduction-Polarity Reversal

The early breakup of western Pangea entailed the northwestward departure of North America from Africa–South America, but much of Mexico lagged behind North America by sinistral transform motion on one or more intra-continental transform faults such as the Texas Lineament and Trans-Mexican Volcanic Belt Lineament (Pindell and Dewey, 1982; Anderson and Schmidt, 1983). By the Late Jurassic, seafloor spreading in the Proto-Caribbean and Colombian Marginal Seaway was probably linked to the Pacific trench by a major transform fault passing along the south side of Chortís, because the stratigraphic sequences of Chortís and southern Mexico are very similar for the Early Cretaceous. Thus, one can infer that the original (Jurassic) east-dipping Cordilleran arc should have persisted for some time as the gap between the Americas grew, and hence there is a reason for the former existence of an east-dipping Benioff zone that initially spanned the widening gap between Chortís and the Andes (Fig. 3). Within the span of the Late Jurassic to Neocomian, however, motion at this Benioff zone may have become so oblique that an overlying magmatic arc was not developed. Relative plate motion at this boundary certainly had a strong sinistral component (Engebretson et al., 1985; Avé Lallemant and Oldow, 1988). Boreal Jurassic radiolarian cherts have been found in several Antillean islands (Montgomery et al., 1994), suggesting, in our view, southward terrane migration along the Mexico-Chortís margin prior to the subduction polarity reversal. Pre-Albian PIA rocks form "basement" throughout the Caribbean islands, but whether these rocks were extruded at an Inter-American Arc south of Mexico-Chortís formed above the east-dipping subduction zone, or were stripped from an intra-oceanic arc that lay outboard of the backarc basins of southern Mexico-Chortís and were carried southward along a mainly transform margin, is not clear. The primitive nature of the Antillean basement rocks suggests an intra-oceanic origin, which is more compatible with extrusion at an Inter-American Arc than with strike-slip removal from the Chortís continental block.

Thus, in the Early Cretaceous, there was either an elongating, highly sinistral, southwest-facing primitive arc system (Fig. 7A) or a sinistral shear system of unknown complexity with little or no Benioff zone (Fig. 7C) across the widening inter-American gap. This plate boundary would become the site of southwest-dipping subduction by the Aptian (Figs. 7B, 7D, and 8). If the circum-Caribbean pre-Albian primitive arc materials were carried into the Inter-American gap from outboard of Chortís by sinistral transcurrent motions along a complex shear system, then a northeastward-dipping subduction zone is not necessarily required, and hence the Aptian "event" earmarked in Figure 6 may have comprised the onset of southwest-dipping subduction at a preexisting transform zone rather than a true arc-polarity reversal (Fig. 7D). However, the general lack of quartz, continental sediments, or isotopic indicators of continental influence in the pre-Aptian primitive Caribbean arcs (e.g., Frost and Snoke, 1989; Donnelly et al., 1990) suggests that the Inter-American Arc was intra-oceanic.

Figures 7A and 7B schematically show the effects of polarity reversal at a sinistrally oblique trench for typical slab dips of 30° to 45°. In such a transpressive geometric model for reversal, axis-parallel extension along the arc should be an important process. The features of Figure 7B are generally recognized in the geology of the Caribbean arcs; namely, (1) the "new" arc axis can form southwest or west of the "old" arc (e.g., Cuba, Hispaniola); (2) the old arc, and even parts of the Pacific-derived Caribbean plate, can thus end up within the forearc of the new arc (e.g., the Los Pasos Formation relative to the Cabaiguan Formation in Cuba; the Los Ranchos and Maimón Formations relative to the Tireo Formation in Hispaniola; La Desirade and North Coast Schist terranes relative to Aves Ridge in the eastern Caribbean;

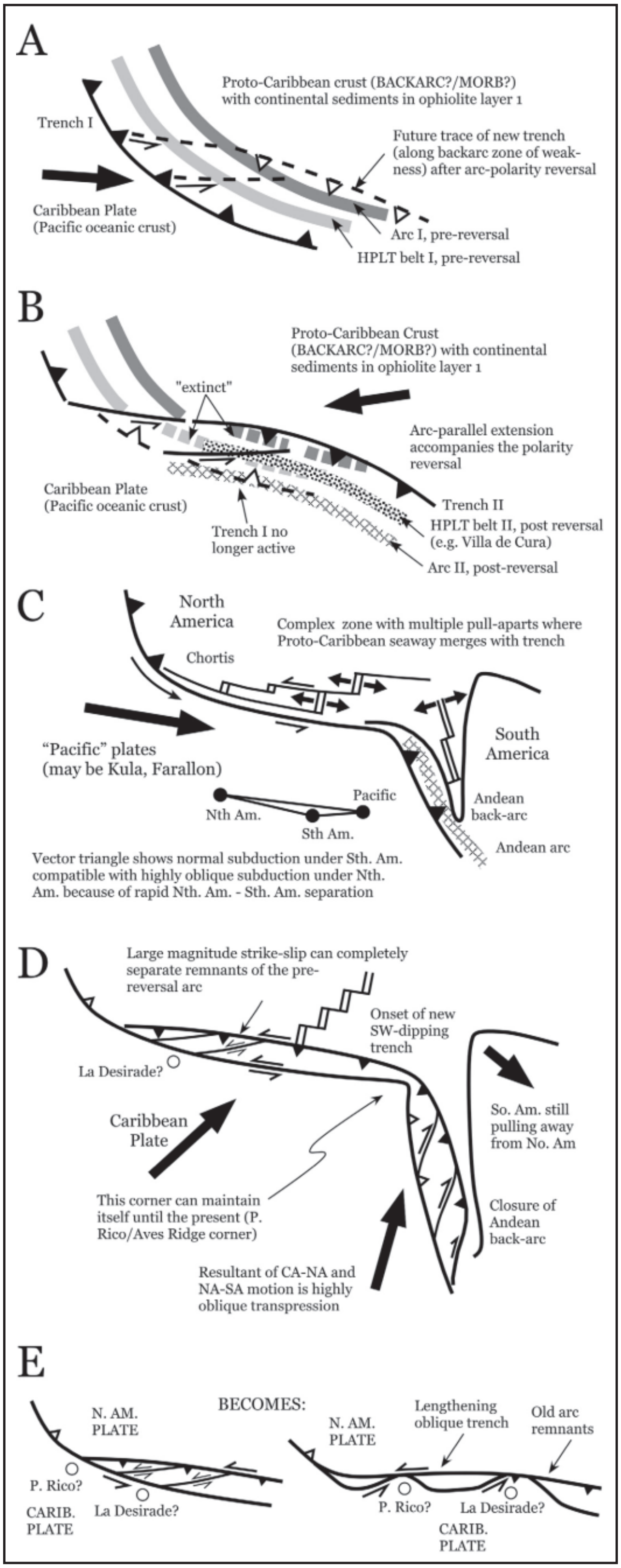

Figure 7. Models for the subduction-polarity reversal at the Inter-American Arc (North American end shown here; a mirror image could apply in the south). In (A) and (B), a southwest-facing primitive arc is replaced by a northeast-facing arc, with consequent reversal of the positions of high-pressure–low temperature (HP-LT) and high-temperature–low pressure (HT-LP) metamorphic zones. Note that we also show east-west–trending "cross-arc transform faults," which link the new Antilles trench to the continued east-dipping subduction zone in Mexico-Chortís and farther north. The degree of development of a preexisting arc is highly dependent on the direction and obliquity of "Pacific" relative plate motion. We show an extremely oblique case in (C) with little or no preexisting arc because the margin was strongly sinistral, with only a small component of subduction beneath the Proto-Caribbean, and also show how seafloor spreading in the Proto-Caribbean may merge with the Mexican trench, across one or more significant transforms south of Chortís. Parts of the new subduction zone may have nucleated along ridge segments and transforms of the Proto-Caribbean spreading ridge immediately adjacent to the original arc; thus, post-reversal forearc rocks (e.g., boninites of pre-Albian age and mid-oceanic-ridge basalt [MORB]-like basalt blocks in the northern Cuban ophiolite mélange noted by Kerr et al., 1999) may have a geochemistry consistent with a "backarc" origin or with melting due to spreading center subduction following polarity reversal. (D) shows a regional scale snapshot after subduction polarity reversal. Note that the transforms that link the two trenches must lengthen dramatically as Caribbean–American relative motion continues. One important consequence is that arc-parallel extension can be so severe that remnants of the pre-reversal arc can be strung out to the point where Pacific-origin crust of the Caribbean plate can be carried into the forearc position of the new arc as the Caribbean walls of the cross-arc faults become part of the hanging wall above the new trench (E). This may explain the apparent anomaly of Pacific-derived cherts (Montgomery et al., 1994) being present in the Antilles forearc in Hispaniola, Puerto Rico, and La Desirade. CA-NA—Caribbean–North America; NA-SA—North America–South America.

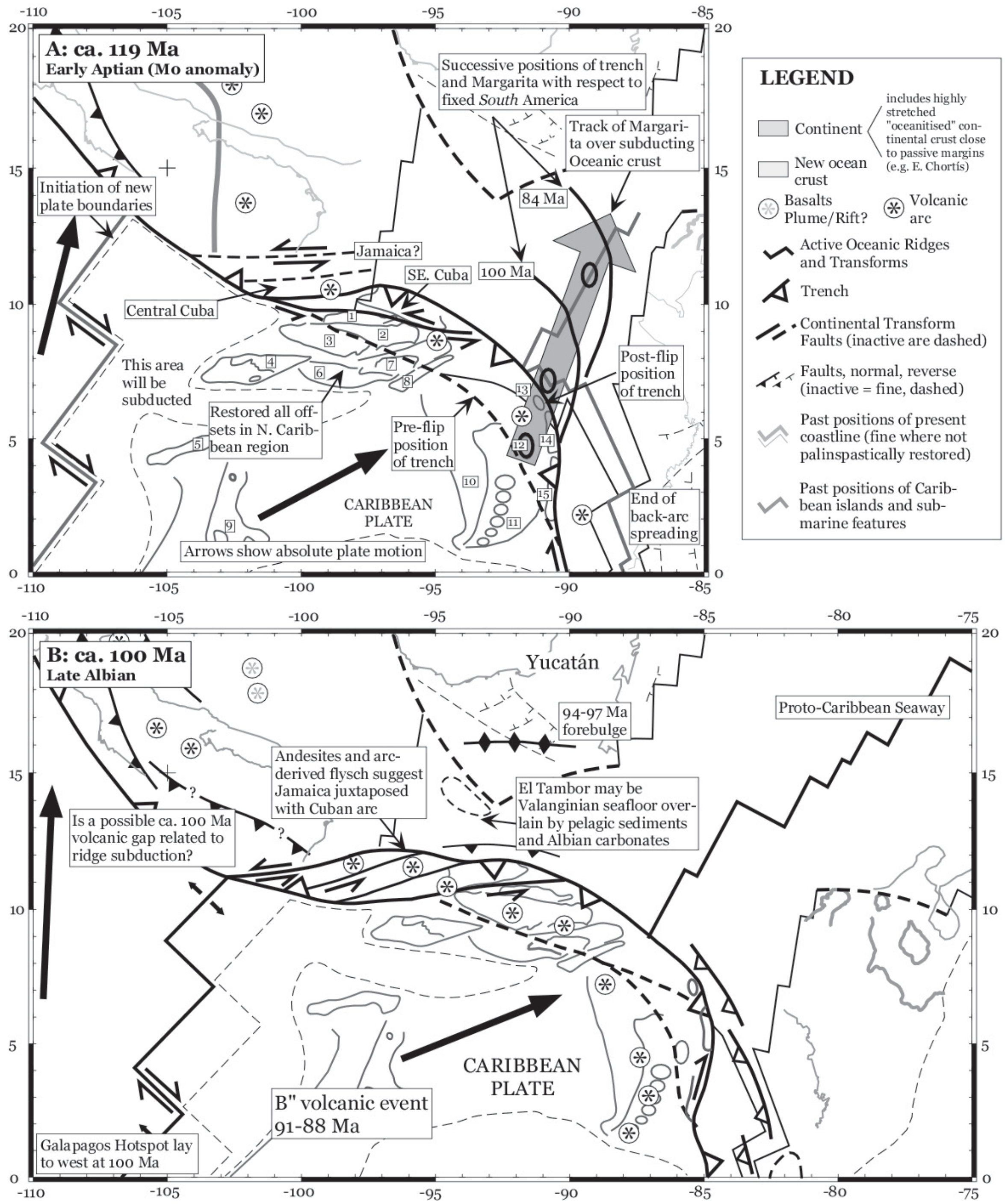

Figure 8. Detailed depiction of paleogeography of the subduction-polarity reversal, modified after Pindell and Kennan (2001b). (A) Inferred position of the Inter-American Arc immediately after the reversal. Note that the inter-American gap is quite narrow and we must restore large magnitudes of strike-slip in the northern Caribbean and arc-parallel rather than arc-normal extension in the Grenada Basin to allow the Caribbean plate to fit through. Key terranes are identified. We have also shown the modeled track of Margarita relative to South America and successive positions of the trench at 100^Ma and 84^Ma. The track passes close to spreading ridges active at 110^Ma, allowing the accretion of 110^Ma basalts prior to high-pressure–low-temperature metamorphism. (B) Modeled position of the Caribbean plate at 100^Ma. In the northern Caribbean, the Cuban terranes are lengthening east-west above the new southwest-dipping subduction zone. In the south, the modeled track of Margarita is consistent with slightly transpressive dextral motion relative to South America, allowing it to pass north of Guajira at 72^Ma. In the northwest, we suggest that the Caribbean plate was bounded by a ridge system, which may explain the "Icelandic" character of soon-to-be-erupted plateau basalts. This is also consistent with the lack of arc activity in Chortís and the lack of younger arc accretion in southern Mexico. Index to numbered localities: 1—Northern Cordillera, Dominican Republic; 2—eastern Cordillera; 3—central Cordillera; 4—central Haiti; 5—southern Haiti; 6—Muertos forearc; 7—Puerto Rico; 8—Virgin Islands; 9— Beata Ridge; 10—Aves Ridge; 11—Leeward Antilles; 12—Margarita (Juan Griego protolith); 13—Tobago; 14—Paria-Araya Peninsula; 15—Villa de Cura. M0—M0 oceanic magnetic anomaly.

Villa de Cura complex relative to Leeward Antilles Arc in the southern Caribbean; and (3) the transfer faults crossing from trench to trench may be the origin of faults of that orientation in Cuba (e.g., La Trocha Fault), and dextral ductile shear strains mapped in Margarita may also pertain to such faulting (but of opposite sense, being located at the southeast end of the arc).

Once southwestward dipping subduction had been initiated, Figure 8 shows how the American plates began to engulf the Caribbean lithosphere, the main points of which are (1) seafloor spreading continued in the Proto-Caribbean because the American plates continued to separate until ca. 85–90^Ma, after which the Proto-Caribbean ridge was effectively dead; (2) the Great Caribbean Arc must have lengthened to match this progressive separation, presumably by arc-parallel extension and by en echelon strike-slip faulting; (3) Albian and younger calc-alkaline magmatism continually intruded the arc as extension continued such that the arc began to acquire local paleogeographies that are recognizable today shortly after 100 Ma; (4) the western end of the Great Caribbean Arc began to override and subduct elements of the North American margin immediately after polarity reversal; (5) the eastern end of the arc began to override and subduct elements of the South American margin, including the Andean backarc basin, immediately after polarity reversal; (6) the arc went on to collide obliquely with the margins of the American continents in the Late Cretaceous. Only the central portion of the Great Arc (i.e., the Greater Antilles and the Aves Ridge–Leeward Antilles and their forearcs) would eventually fit through the gap between Yucatán and Guajira Peninsula.

NORTH AMERICA–GREAT CARIBBEAN ARC INTERACTIONS

Sinistral Oblique Collision between the Great Arc and Central America–Yucatán

At the North American end of the Great Arc, polarity reversal involved large-scale sinistral displacements along en echelon transfer faults that connected the Mexican trench to the Greater Antilles (Great Arc) trench. These faults lengthened the arc and allowed the Caribbean plate to enter the Proto-Caribbean realm (Figs. 7 and 8) and are probable precursors to the La Trocha, Pinar, Cauto, and other such faults crossing Cuba that were reactivated in the Maastrichtian to Eocene (Stanek et al., 2000). The faults also allowed fragments of the pre-reversal Inter-American Arc to become entirely separated (Fig. 7E), such that Caribbean plate rocks (parts of Puerto Rico and La Desiràde) or fragments of the pre-reversal forearc, including HP-LT rocks, could be pulled into the forearc of the younger west-dipping subduction zone.

The early history of the Jamaican portion of the Great Arc is murky at best. Today, Jamaica forms the eastern tip of Nicaragua Rise, but appears to be intra-oceanic; thus, a continent-oceanic arc boundary between Jamaica and Chortís likely occurs somewhere along Nicaragua Rise. From the inferred paleogeographies of Figures 5, 7, and 8, we deduce that Jamaica originated (Lower Devil's Racecourse Formation) southeast of Chortís and northwest of the Central Cuban terranes (but was still part of the Inter-American Arc), such that the sinistral transpressive shear noted above (Fig. 7) carried central Cuba east of Jamaica and carried Jamaica east of southern Chortís. Unfortunately, we do not know well enough when peak HP metamorphism was reached in the Blue Mountains of Jamaica, and thus it is difficult to define which subduction polarity produced the HP metamorphism. Draper (1986) suggested a single north-dipping Benioff Zone for Jamaica with the Blue Mountains HP rocks lying south of the arc axis, such that Upper Cretaceous arc magmatic intrusions in or near the Blue Mountains pertain to a steepening of that Benioff Zone. This might be feasible if the Blue Mountain HP metamorphism is Aptian or older, but if future geochronological work indicates a late Aptian or younger age for these rocks, as we expect will be the case, then the Blue Mountains HP suite probably originated from southwest-dipping subduction like those in Cuba, and was subsequently uplifted by faulting in the forearc. Transtensional collapse of the central Cuban terrane off of Jamaica during arc-parallel extension may be a possible cause of uplift. The Blue Mountains HP rocks might then have been intruded by magmas derived from rejuvenated *north-dipping* subduction of *Caribbean* crust beneath Chortís-Nicaragua Rise (Fig. 5D and 5E), *after* the Great Arc's Maastrichtian collision with Yucatán (see below). Part of the basis for this latter interpretation is that Jamaica remained volcanically active into the Eocene, long after collision with Yucatán (Pindell and Barrett, 1990).

The original context of the pre-Aptian HP-LT rocks south of the Motagua Fault is even less clear. If polarity reversal in this area is slightly older than in Cuba, then it is possible that they comprise part of the forearc of Jamaica and were emplaced ahead of Jamaica onto the Yucatán margin during the Late Cretaceous. In this case, they must have lain north of the Motagua Fault until Neogene time. Alternatively, if the Motagua HP-LT rocks were part of the Pacific-facing forearc prior to polarity reversal, axis-parallel extension within the Great Arc (Fig. 7E), separating Jamaica from southeastern Chortís, would allow them to lie west of Jamaica at time of collision with Yucatán, and be carried east by up to 500 km to their present position by Cenozoic sinistral motion on one of the strands of the Motagua Fault Zone.

Partial subduction of the attenuated southern margin of the continental Chortís Block (where Jurassic continental quartz-bearing strata are known; see lithologic column, Fig. 3), initiated by transpressive thrusting during and after polarity reversal, is an appealing origin for the quartzofeldspathic protoliths of the Escambray metamorphic complex of Cuba, as well as some of the HP blocks in the serpentinite mélanges of Las Villas. Las Villas mélanges also yield rare Grenville ages in their protolith (Somin and Millán, 1981, Renne et al., 1989; Grafe et al., 2001), again consistent with Chortís as an origin (e.g., Honduras; Manton, 1996). Four distinct nappes in the Escambray complex show northward ductile shearing, which may pertain to this early south-dipping subduction (Stanek, 2000). We suggest that thinned parts of the rifted margin of the Chortís Block, or quartzose sediments derived therefrom, were included in the downgoing Proto-Caribbean slab. HP-LT conditions were reached at Escambray and Las Villas Complex by 106 Ma (middle Albian).

The eastern Chortís margin may also have been the origin of the Isle of Youth (Isle of Pines) metamorphic complex, given that Somin and Millán (1981) have identified similarities between the protoliths of Escambray and Isle of Youth. The Barrovian metamorphism in the Isle of Youth indicates a quite different crustal setting to the higher-pressure metamorphic rocks of Escambray, possibly within a nappe pile at the western end of the arc. Other possibilities include an origin as a fragment of Chortís basement (if the metamorphism is old) or its passive margin cover (intercalated volcanics could be of proto-Caribbean origin or from the pre-reversal Inter-American Arc) or the passive margin cover of southernmost Yucatán. It could have been metamorphosed prior to polarity reversal, or within the root of the post-reversal Antilles arc during the middle Cretaceous, or during latest Cretaceous (in the case of a Yucatán origin). Distinguishing between these requires more accurate dating of protolith age and of peak metamorphism than is currently available; $^{40}Ar/^{39}Ar$ cooling ages (Garcia-Casco et al., 2001) simply constrain peak metamorphism to early Maastrichtian or older.

The central Cuban rocks were progressively uplifted through middle and Late Cretaceous. Given the oblique subduction history for this interval in Figures 5 and 8, this was likely due to arc-parallel stretching. At the same time, several km of post-reversal arc lavas had accumulated in southern central Cuba, which we interpret as pertaining to southward subduction of Proto-Caribbean crust. The primary plutonic-magmatic axis of the Cuban portion of the Great Arc was probably situated farther south in what is now the Cayman Ridge (the intra-arc Yucatán Basin had not yet opened; see below and also Pindell and Barrett, 1990), as suggested by the Cayman Ridge's far thicker crustal root (Case et al., 1990) relative to that of the Cuban "arc" rocks, beneath which lie the overthrust Bahamas carbonates. Thus, the "Central Cuban arc terrane," north of the Yucatán and Caúto basins and east of Pinar Fault (defining the boundary with Western Cuba), likely comprises mainly forearc elements of the Great Arc, where a number of "arc" intrusions and volcanic flows do occur, but where accreted sedimentary terranes, ophiolites, and HP and other metamorphic terranes are the rule. This interpretation is consistent with structural interpretations that the Cuban arc-related rocks comprise only relatively thin, obducted nappes above the underthrusted Bahamian carbonate section (Hempton and Barros, 1993; Draper and Barros, 1994).

Between 90 and 80^Ma, less-metamorphosed granitic pegmatites cut the contact between the Escambray HP rocks and the amphibolite grade base of the arc basement (Mabujina unit), indicating that significant uplift of the composite Escambray-Mabujina units had occurred by that time (Stanek et al., 2000; Grafe et al., 2001). The Campanian migration of the Great Arc toward the southern fringe of the Yucatán passive margin should have caused further shallowing of the subduction angle and a progressive southward shift of the magmatic axis. We suggest that this is why volcanism ceased in the central Cuban terrane by the Maastrichtian; if our reconstruction is correct, arc magmatism should have continued in the Yucatán Basin, a Maastrichtian–early Eocene intra-arc basin, and/or Cayman Ridge, as it did in the Oriente Province of Cuba where no significant intra-arc basin was present.

By the Campanian, north-vergent collision was imminent between the Great Arc and the southern Yucatán margin as marked by the drowning of the Cobán carbonate-evaporite platform beneath upward-deepening Campur Formation carbonates and the onset of Sépur clastic (including ophiolitic debris) flysch deposition (Rosenfeld, 1993). By the Maastrichtian, El Tambór and Santa Cruz ophiolites overthrust the Sépur foredeep basin, but it is not clear if these ophiolites represent the Great Arc's forearc basement or obducted Proto-Caribbean material; Giunta et al. (2002) suggest that both MORB and arc basalts are present. The Cuban part of the Great Arc was likely situated just east of Jamaica and avoided direct collision with the southern Yucatán margin such that it, but not Jamaica, continued to migrate toward the Bahamas thereafter. Thus, the Guatemalan ophiolites likely represent the Jamaican part of the Great Arc's forearc; new biostratigraphic assessments in Jamaica suggest that a primary orogenic pulse occurred in Jamaica during the Maastrichtian (Mitchell, 2003), matching the time of forearc emplacement onto northern Guatemala.

By 70 Ma (late Maastrichtian), the part of the Great Caribbean Arc that did not get blocked by southern Yucatán was passing through the narrowest point between the Americas, the gap between Yucatán and Guajira. Beyond this bottleneck, toward the northeast, the Proto-Caribbean margins became farther apart. However, a clear record of uninterrupted diachronous arc-continent collision exists along all Proto-Caribbean margins (Pindell and Barrett, 1990), and therefore, the arc must have undergone arc-parallel lengthening during further migration in order to accommodate the widening shape of the Proto-Caribbean. Arc-parallel extension was achieved at both ends of the Great Arc by opening of two intra-arc basins, the Yucatán and Grenada Basins, driven, we suggest, by trench rollback forces. Intra-arc rifting separated mainly forearc elements (Cuba in the northwest, and the composite Villa de Cura–Tobago terrane in the southeast) from the main Late Cretaceous magmatic axes. These forearc elements went on to collide with the American margins, which is why the American margins are today the sites of HP-LT and other Caribbean metamorphic complexes.

Evolution of the Yucatán Basin and Collision of the Cuban Forearc with Yucatán-Bahamas

Figure 9D shows the present structure of the Yucatán Basin and Cuba. Three distinct sets of faults dominate the region. Set 1 (Figs. 9A, 9B) comprises northeast- to east-northeast–striking extensional basement faults across much of the Yucatán Basin and Cayman Rise, which we believe record the north-northwest–south-southeast extension direction associated with generation of oceanic crust in Yucatán Basin. Set 2 (Fig. 9C) comprises high-angle block faults within arc-related crust along the Cayman Ridge that are probably younger than set 1 faults and that probably

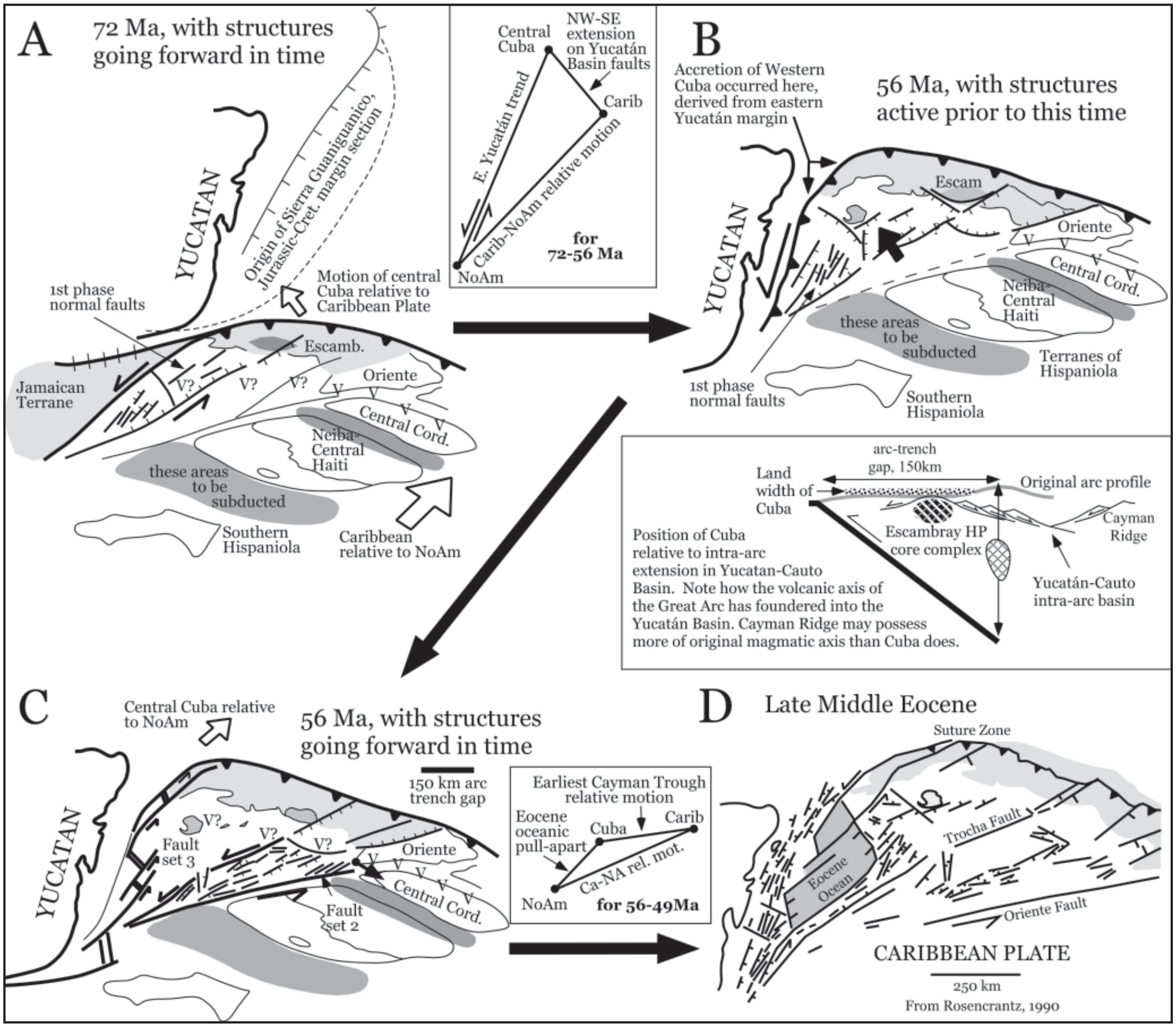

Figure 9. Three-stage model for opening of Yucatán Basin. (A) At 72^Ma, the Yucatán Basin had not opened, as the Great Arc had not yet passed through the Yucatán-Guajira bottleneck. (B) From 72 to 56^Ma, rollback of Proto-Caribbean oceanic crust (Fig. 10) sucked the northwest forearc flank of the Great Arc (i.e., Cuban portion) northwest with respect to the Caribbean plate, resulting in set 1 normal faults and oblique collision with Yucatán. (C) At ca. 56^Ma, tearing of the overthrust Proto-Caribbean slab along Yucatán (Fig. 10) caused the rollback direction to change from northwest to northeast, toward the Bahamas (slab was not yet torn there). We suggest the set 3 pull-apart formed above this tear, consistent with northeast-directed final suturing of Cuba with the Bahamas. (D) Late Eocene to present structure at the same scale as models A, B, and C. This model requires onset of Cayman Trough faulting by ca. 56 Ma; Rosencrantz et al. (1988) interpreted marine magnetic anomalies back to 50^Ma, and the 200-km-wide stretched, non-oceanic ends of the Trough suggest that motion began somewhat earlier. Ca-NA—Caribbean–North America; HP—high pressure; NoAm—North America; V—volcanism.

pertain to initial sinistral motion along the Cayman Trough transform. Set 3 (Fig. 9C), following Rosencrantz's (1990) interpretation, comprises faults defining a discrete north-northeast–striking pull-apart basin in the western Yucatán Basin. Northeast to east-northeast–striking faults crossing onshore Cuba (e.g., La Trocha, Cauto faults) correlate with, and probably belong to, fault set 1. These faults are mainly extensional, step down to the east in the onshore, and moved during the Maastrichtian to Eocene, judging from fault growth in flanking sediments (Iturralde-Vinent, 2000, personal commun.). Although untested by drilling as yet, the faults of the first set in the basin are probably of the same age. Rosencrantz (1990) cogently argues that set 3 (pull-apart basin) faults are early to middle Eocene and younger than the creation of the bulk of the Yucatán Basin. Set 3 faults transformed plate motion along the Pinar Fault into the Eocene Cuban thrust belt.

Given the general northeast migration of the Caribbean plate relative to North America during the early Paleogene, the north-northwest–south-southeast opening direction of the Yucatán Basin suggested by set 1 faults may initially seem anomalous but is actually required in the three-plate system: North America–Cuban forearc–Caribbean plate/Cayman Ridge (Fig. 9, vector triangle for 72–56^Ma, which defines motions for the 72–56 Ma interval). Note that the North America–Caribbean relative motion direction was more easterly than the trend of the eastern Yucatán Proto-Caribbean margin. Maastrichtian–Paleocene extension at set 1 faults and associated generation of quasi-oceanic crust in the Yucatán Basin allowed the Cuban forearc to migrate transpressively north-northeast along the Yucatán margin (Fig. 9B), while the rest of the Caribbean plate moved in a more eastward direction. We suggest that at least some set 1 faults were southeast-dipping low-angle detachments, such that the Cayman Ridge (arc) collapsed off the Escambray and Isle of Youth footwalls during the early opening of Yucatán Basin. Following Draper (2001), we suggest, therefore, that Escambray and Isle of Youth are extensional metamorphic core complexes as is suggested by their flat-lying and domal foliation patterns. The extension was driven by rollback of the Middle Jurassic Proto-Caribbean crust ahead of Cuba (Fig. 10), and we suggest that the 68 Ma $^{40}Ar/^{39}Ar$ age on Isle of Youth metamorphic rocks (Garcia-Casco et al., 2001) and

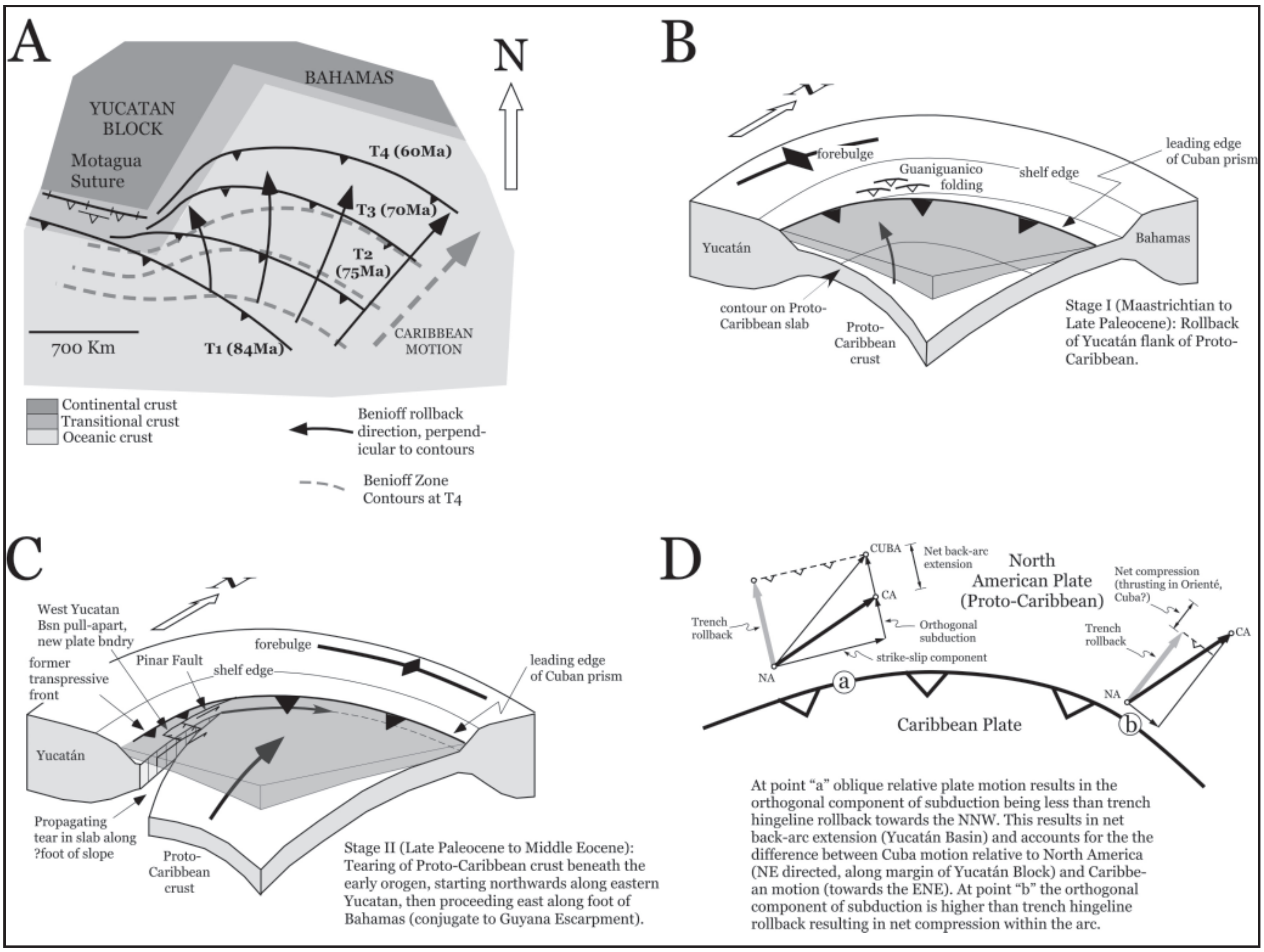

Figure 10. Block diagram of Proto-Caribbean rollback geometry beneath advancing Cuban terranes. (A) Schematic structure contours on subducting slab; rollback vectors relative to Yucatán and successive trench positions. Times T1–T4 are dated as ca. 84^Ma, 75^Ma, 70^Ma, and 60^Ma. (B) From 72 to 56^Ma, rollback toward Yucatán drove oblique thrusting in the Sierra Guaniguanico to the northwest of the opening Yucatán Basin. (C) Starting at ca. 56^Ma, rollback and extension orientations changed to southwest-northeast. We suggest this required tearing of the Proto-Caribbean slab southeast of Yucatán and was accompanied by a new oceanic pull-apart opening in the Yucatán Basin. Slab tearing led to rebound as the footwall unloaded, starting initially in Yucatán and migrating northeast and then east around the Cuban Orogen by the Eocene. (D) Vectors nets show that at very oblique margins (point a) rollback of the subducting plate can exceed the orthogonal component of subduction, resulting in extension within the forearc, in contrast to less oblique margins (point b) where the rollback of the subducting plate is unlikely to exceed the orthogonal component of subduction, resulting in continued compression within the forearc. CA—Caribbean; NA—North America.

the 60–70 Ma zircon fission track ages from Escambray (Stanek et al., 2000) date this tectonic unroofing.

By 56^Ma, set 3 faults (pull-apart) began to form in the western Yucatán Basin, following oblique collision between the Cuban forearc and the Yucatán margin. This oblique collision had imbricated the eastern Yucatán passive margin strata into the Sierra Guaniguanico fold-thrust belt (Pindell, 1985b; Hutson et al., 1998; Pszczólkowski, 1999). The set 3 pull-apart basin appears to have formed coevally (early to middle Eocene) with initial transform motion on the Oriente Fault of Cayman Trough (Fig. 8C; Rosencrantz et al., 1988), signaling a change in the regional kinematics. Thereafter, a new three-plate system, North America–Cuba/Yucatán Basin/Cayman Ridge–Caribbean plate, took over, the relative motions of which are shown in the lower vector triangle of Figure 9. The set 3 pull-apart (comprising Pinar Fault) allowed Cuba–Yucatán Basin to migrate toward eventual collision with the Bahamas, while the early Cayman Trough transform system became the site of Yucatán Basin–Caribbean relative plate motion. It is important to note that prior to final collision of Cuba with the Bahamas, the early Cayman Trough itself must have been moving north toward North America, following the Yucatán Basin plate. This also implies a component of north-directed compression in northern Central America and explains the more complex, disorganized structure of the easternmost and westernmost Cayman Trough. Following suturing of Cuba with the Bahamas, the Cayman Trough represents the relative motion between the Caribbean plate and North America.

Note that this two-stage model (Fig. 9) does *not* invoke synchronous backarc spreading (set 1 faults of Yucatán Basin) and arc-continent collision (Cuba–Bahamas). In this model, the composite Cuban forearc–Yucatán Basin terrane (second stage of Fig. 9) can move as an independent platelet between the Caribbean and North American plates, and the Cuba–Bahamas collision could ultimately have been driven by Caribbean–North America relative motion in the absence of a passive backarc spreading center. Thus, total contraction in the collision may have exceeded that which would be expected if Proto-Caribbean rollback were the only driving mechanism.

Seismic tomography data (van der Hilst, 1990) clearly image the Proto-Caribbean slab subducted beneath the Antilles and also clearly show that, to the east of the Dominican Republic, this slab has torn away from the North American plate and is sinking into the deep mantle. We expect that the overthrust Proto-Caribbean slab started to drop off into the mantle upon or shortly following collision of the Great Arc with the North American plate. In such a diachronous collision from southern Yucatán to the Bahamas, slab drop off must also have occurred diachronously, by progressive tearing: Maastrichtian at the Motagua Suture Zone, Paleocene along eastern Yucatán, and Eocene at the Bahamas. Thus, the cause for central Cuba's kinematic change from north-northwest to northeast at 56^Ma, relative to North America, was probably the cessation of the rollback driving mechanism upon Paleocene oblique collision along eastern Yucatán, because the subducting slab started to tear in that area.

We suggest that the set 3 pull-apart basin ("Eocene Ocean" in Fig. 9D) formed directly above the propagating Paleocene-Eocene tear between Yucatán (i.e., North America) and the Proto-Caribbean, such that further rollback would have been restricted to the eastern side of the tear and in a north-northeast direction. Thus, the Cuban forearc terrane would have been pulled, after 56^Ma, toward the Bahamas rather than toward Yucatán. Our two-stage model is supported by structural field studies (Gordon et al., 1997) that indicate northwest shortening and maximum stress during the Paleocene and north-south maximum stress due to sinistral transpression between the Pinar Fault and the offshore thrust belt during early to early middle Eocene.

As the tear in the slab propagated northward, we predict that isostatic rebound would occur in the North American margin to form a deep post-orogenic unconformity whose onset should young to the north. Such an unconformity does exist (e.g., middle Eocene hiatus in Cuba of Pszczólkowski, 1999), but it is difficult to demonstrate the northward-younging in Yucatán of its *onset*, because the erosional surface can remain subaerial for significant amounts of time thereafter. Rapid Paleocene–lower Eocene carbonate accumulation rates in the Bahamas ahead of the Cuban thrustbelt (Paulus, 1972) attest to Proto-Caribbean slab-pull and arrival of the Cuban terranes during that time, followed by a strong middle Eocene unconformity. Uplift surely occurred during imbrication of the Cuban nappes as they overrode the Bahamas margin, but the unconformity most certainly was severely enhanced by isostatic rebound due to slab drop off at the end of the orogeny. The fact that arc magmatism continued in the Oriente area of Cuba until ca. 45 Ma demonstrates that the tear had not propagated that far east until after that time.

Figure 10 explores how trench rollback initially drove north-westward extension (set 1 faults and seafloor spreading) in the Yucatán Basin. As the Great Arc encountered southern Yucatán, structural contours on the descending Proto-Caribbean slab would necessarily become warped as shown in Figure 10A (note contours for T4). As the obliquity of subduction direction at the trench increased, rollback velocity would begin to dominate the orthogonal rate at which Proto-Caribbean crust actually entered the trench (Fig. 10D). This would impart an extensional stress on the arc in a direction perpendicular to the Proto-Caribbean structural contours and parallel to the rollback direction (north-northwest). Thus, the leading edge of the arc was pulled toward eastern Yucatán while the rest of the Caribbean plate migrated to the north-northeast. Structural style at the plate *interface* along the eastern Yucatán Block was sinistrally transpressive. As the Cuban forearc terrane encountered the eastern Yucatán margin, the Yucatán margin strata were imbricated and accreted into the growing Cuban thrustbelt (Sierra Guaniguanico of western Cuba); structural imbrication and metamorphism there are both Paleocene (Gordon et al., 1997).

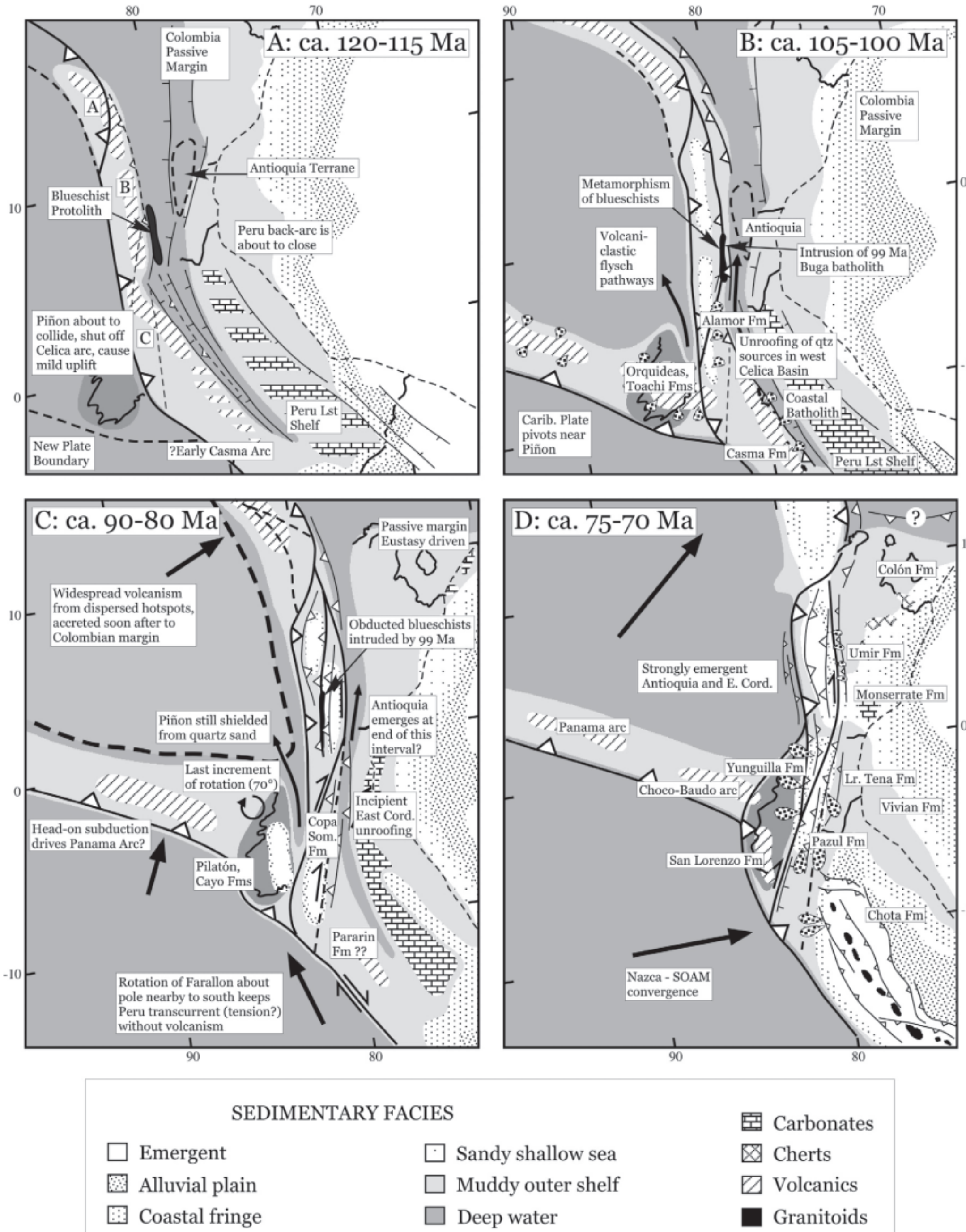

Figure 11. Snapshots of northern Central Andes development from 120 to 70^Ma, as discussed in the text. Positions of the trailing edge of the Caribbean plate off northern Peru are constrained by position of its leading edge off Yucatán, by equatorial paleomagnetic data from allochthonous Ecuadorian terranes, and by absence of arc volcanism farther north in the Ecuadorian and Colombian autochthon. Note the north-northeast–directed diachronous transpression of the former Andean backarc and northward migration of the Panama triple junction. Enhanced rate of uplift and erosion in the Andes from 84^Ma is driven by increased subduction of Caribbean crust beneath the Andes, related to the end of oceanic spreading between the Americas. Locations on (A): A—Margarita and Tobago portion of the pre-Aptian arc; B—arc fragments accreted to western Columbia; C—Amotape, Tahuín, and Chaucha terranes of northern Peru and southern Ecuador.

SOUTH AMERICA–GREAT CARIBBEAN ARC INTERACTIONS

Setting of Northern South America Prior to Caribbean Interaction: The "Andean Backarc Basin"

As with the northwest end of the Great Arc, the southeast end was also the site of HP-LT metamorphism dating back to at least the Aptian (Table 1). Such metamorphic rocks occur in Margarita, the Villa de Cura complex, and the Cordillera de la Costa of Venezuela, all of which lie on the eastern flank of the Great Arc of the Caribbean as reconstructed here. HP-LT rocks also occur along the Romeral–Peltetec Suture of Colombia and, although less easily interpreted, also probably formed along the eastern flank of the Great Arc. Again, unless we invoke keel-hauling to bring such old HP rocks under the arc from the western to eastern flank of the arc, the polarity reversal must be Aptian or older.

Figures 5 and 11 show the general south-to-north diachronous closure of the Andean backarc basin and arc-continent collision between the offshore arc (Great Arc) and the South American autochthon. One of the most important features to note in these maps is that, prior to the middle and Late Cretaceous, almost the entire Ecuadorian and Colombian margin is thought to have been essentially a passive margin, facing either the "Colombian Marginal Seaway" or the "Andean Backarc Basin." Both of these features were largely destroyed during proposed interaction with the Caribbean plate, and thus the key features that indicate their existence and nature warrant brief review:

- We define the Colombian Marginal Seaway as lying outboard of the Colombian passive margin. The arc to the west was intra-oceanic, founded on pre-reversal PIA rocks. The floor of the basin comprised entirely oceanic crust of Proto-Caribbean origin.
- We define the "Andean Backarc Basin" as that part of the Peruvian to southern Colombian margins separated from the passive margin by a narrower basin and with an arc founded on continental or transitional crust lying out to the west. The floor of the basin comprised highly stretched continental crust (central Peru) to transitional and possibly oceanic crust in northernmost Peru and Ecuador.
- There is strong paleogeographic evidence for the Colombian passive margin. Facies mapping in Colombia (Tectonic Analysis, 1998; Villamil, 1999) shows no sign of an active margin setting in the central Cordillera and areas to the east prior to the Late Cretaceous. There are no unambiguous remnants of a continental volcanic arc after the Late Jurassic, little or no volcanic material derived from an arc is found in the Magdalena Basin or areas to the east, and there is no indication prior to the Late Cretaceous of any west-derived sediment of arc or continent derivation.
- Plate reconstructions show that, prior to the Late Jurassic, Colombia (north of ~4°) was the conjugate margin to Chortís, and that the tail of Chortís (including our proposed protolith for central Cuba) extended at least this far south. Thus, the South American portion of the early "Intra-American Arc," above an east-dipping subduction zone and founded on stretched continental crust, must lie south of the latitude of Bogotá (see also maps in Jaillard et al., 1990).
- The lack of pre-Late Cretaceous active margin indicators in Colombia north of Bogotá and in Venezuela, and the fact that the Great Arc of the Caribbean forms the leading edge of the Caribbean plate, thus suggest that in South America the "Intra-American Arc," which rifted away from the passive margin, was the template for the Great Arc.
- In the Ecuadorian Cordillera Real, as in Colombia, the latest intrusive ages are Late Jurassic (Litherland et al., 1994, Noble et al., 1997) and only prior to Early to Middle Jurassic can volcanism with both extensional and subduction affinities be found in the Subandes and inner foreland basin (Romeuf et al., 1997).
- Volcanic components of the Cordillera show a geochemistry indicative of progressively more transitional crust to the west, and still farther west remnants of ophiolite (probable Early Cretaceous age) are found at ~2°S with no intervening Early Cretaceous arc. Thus, we suggest these are the remnants of "oceanic crust" formed in a backarc basin and that any arc of Early Cretaceous age lay farther to the west.
- To the west of these ophiolites, the "Chaucha Terrane" ("B" on Fig. 11A) comprises poorly exposed granite and metamorphic Paleozoic rocks with a Mesozoic clastic fringe on its eastern side ("Guamote Terrane" of Litherland et al., 1994). Farther west, middle Cretaceous and younger arc rocks are separated from this terrane by a second ophiolite suture. We propose that these continental fragments are the dismembered basement to the arc on the west side of the backarc basin. To the south, the Celica Volcanics in the Lancones Basin (e.g., Jaillard et al., 1999) are mainly andesites with a geochemical signature indicating underlying continental basement ("C" on Fig. 11A). These rocks are nearly in situ. Although separated from "stable" South America by significant north-south shear zones, paleomagnetic and other data (Mourier et al., 1988a, 1998b) indicate that they are not displaced significantly north of their latitude of origin.
- A pre-Albian "mafic arch" beneath the Coastal Batholith of western Peru is apparent from gravity data (Wilson, 1985). It lies immediately west of the "West Peruvian Trough" passive margin basin (Cobbing et al., 1981) and probably represents the highly stretched, but not truly oceanic, axis of the backarc basin. Farther south, rift basins with continental and basaltic fill, but no indication of oceanic crust, can be traced as far as northernmost Argentina, 500 km east of the arc (Salfity and Marquillas, 1994).
- The basement rocks of Margarita ("A" on Fig. 11A), where high-pressure rocks have both continental crust of Carboniferous age and oceanic crust of Albian age, are the final pieces required to reconstruct the Inter-American Arc. Dur-

ing the Late Jurassic, as South and North America separated, they must have lain south of the southern end of Chortís and the future Cuban terranes. They must have been north of the future Chaucha Terrane of Ecuador and thus are shown on our maps within the Inter-American Arc to the west of southern Colombia. This fragment of continental crust lay sufficiently far north along the Inter-American Arc that it was able to pass by the Guajira peninsula during subsequent motion (Fig. 8). The backarc basin and Colombian marginal seaway to the northeast are interpreted to be the original site of the oceanic protoliths, which were structurally juxtaposed with the continental crust before 90 Ma.

Following opening of the backarc basin, compression, with strong uplift and cooling of metamorphic terranes, dominated northern South America from the Aptian on, identical in age to the arc-polarity reversal farther north. Thus, we believe that these two events are genetically linked and that backarc basin closure, and accretion of fragments of arc terranes along Ecuador and western Colombia throughout the Late Cretaceous, was driven by relative motions of the Caribbean and South American plates. Because separation of South America from North America continued into at least the Campanian, relative motion was much more north-south oriented than in the northern Caribbean (see vector nests within Figs. 5A and 5B), consistent with the dextral transpressive tectonic style indicated by field studies (e.g., Litherland et al., 1994). During the Cenozoic, the arc continued to migrate along northern South America, all the way to Trinidad. In the sections below, we will trace this prolonged arc-continent interaction, looking first at the arc itself, and then examining the Andes of Ecuador–Colombia, and the Venezuela-Trinidad portions of northern South America, thus documenting the tectonic controls on basin development in northern South America.

The Caribbean Plate and the Eastern End of the Great Arc

Margarita, Tobago (Tobago terrane), and the Villa de Cura complex of Venezuela comprise forearc elements of the Great Arc whose metamorphism indicates burial to varying depths. In Margarita, two very different protoliths have reached HP conditions (Stöckhert et al., 1995; Maresch et al., 2000). The Juan Griego unit protolith had two components, Carboniferous gneisses and Aptian-Albian (*Heterohelix*-bearing) sedimentary rocks. Carboniferous gneisses are characteristic of the central Cordillera of Colombia and Ecuador, some of which must have been rifted from the South American autochthon in the Late Jurassic as the Andean backarc basin formed (magmatism ceased in Colombia's central Cordi llera remnant arc by 140 Ma; Irving, 1975). Backarc extension and spreading was presumably driven by subduction rollback of Pacific lithosphere; the basin persisted as marine during the Neocomian to at least the Albian. In contrast, the La Rinconada unit protolith was an oceanic MORB crystallized between 109 and 116 Ma (Stöckhert et al., 1995; Maresch et al., 2000). Thus, we suggest that both the Juan Griego unit and La Rinconada unit protoliths derive from the Andean backarc basin (Figs. 5, 8, and 11); if the protoliths came from different parts of that basin, then they were juxtaposed by faulting as they were taken down together to depths as great as 50 km. By 90^Ma, they had already begun to cool (Stöckhert et al., 1995; Maresch et al., 2000).

A broadly similar initial metamorphic history may apply to both the Villa de Cura complex and parts of the Cordillera de la Costa (Smith et al., 1999; Sisson et al., this volume, Chapter 3; Sorenson et al., this volume; Unger et al., this volume), but we will focus on the Margarita and other data to interpret some of the younger stages of Caribbean–South American interaction on the Caribbean side. By 86^Ma, as the arc converged obliquely with western Ecuador and Colombia, Margarita's HP suite had been elevated to ~25 km depth, as indicated by cross-cutting arc magmas (Stöckhert et al., 1995; Maresch et al., 2000). As with Cuba, arc-parallel extension is a viable mechanism for uplift, given the strong obliquity of subduction; alternatively, attempted subduction of a progressively thickening continental margin may have been the cause. Possible explanations for the intrusion of arc magmas into HP rocks at this depth include: (1) forearc magmatism, (2) widening of the forearc by addition of migrating terranes from further north along the plate edge, and (3) steepening of the subduction angle for a variety of possible reasons. Ductile dextral shear fabrics and a greenschist overprint were then imparted on both the HP suites and the 86 Ma arc magmatic rocks, indicating that Margarita was situated within the Late Cretaceous dextral transpressive plate boundary zone between the Caribbean and South American plates. Additional uplift of the HP rocks occurred during this shearing, possibly as vertically extruded wedges under transpression, and by 66 Ma as the Great Arc passed through the Yucatán-Guajira bottleneck, the Margarita Complex had been elevated to a depth of ~15 km.

Aruba presents a case that may also occur in other Leeward Antilles islands. The Aruba Batholith (83 Ma) intrudes the Aruba Lava Formation that has been interpreted as ca. 90 Ma Caribbean Plateau basalts (White et al., 1999), leading them to propose 85–88 Ma as the time of the arc polarity reversal. In that model, the Aruba Lava Formation (Caribbean Plateau basalts) choked an east-dipping subduction zone such that, after reversal, the basalts were situated above the new west-dipping Benioff zone, which produced new arc magmas (Aruba Batholith). If the Aruba Lava Formation does, in fact, represent the Caribbean basalt plateau, an alternative model consistent with Aptian polarity reversal and the paleogeography shown in Figure 8 would be that Aruba was situated within the Caribbean interior plateau province at 90^Ma, beyond the "reach" of the post-Aptian west-dipping Benioff zone, but strike-slip removal and incorporation of Caribbean hanging-wall elements into the Western Cordillera of Colombia during dextral oblique collision of the arc with northwest South America caused Aruba to move into the arc axis above the slab by 83^Ma. However, there is a prominent interval up to 100 m thick in the Aruba Lava Formation characterized by coarse conglomerates, paleosols, and accretionary lapilli tuffs clearly indicating subaerial conditions (Beets et al., 1984; A. Snoke, 2003, personal commun.). Thus, the Aruba Lava Formation may have an arc origin as opposed to being part of the Caribbean Plateau basalt province.

After passing through the Yucatán-Guajira bottleneck, in the Paleocene, the Great Arc must have undergone a significant arc-parallel lengthening in order to maintain oblique collision along both the northwestern and the southern Proto-Caribbean margins. Opening of the Yucatán Basin was examined above as one example of this process (Fig. 9), and opening of the Grenada intra-arc basin may be another. Pindell and Barrett (1990) reviewed structural, magnetic, magmatic, and kinematic arguments for a north-south growing dextral pull-apart model for Grenada Basin. Bird et al. (1993, 1999) claimed that magnetics do not support this model, and argued instead for arc-normal expansion of the basin. However, we caution against over-interpreting low-reliability low-latitude magnetic anomalies and note that east-west opening cannot account for early to middle Eocene emplacement of nappes onto the western Venezuelan margin (see section on Venezuela, below). It is possible, however, that the basin opened in two stages, the first one dominated by north-south dextral shear, thereby creating the very sharp western wall of the basin (former strike-slip fault) and explaining the kinematic requirement for forearc obduction in Maracaibo, followed by a period of more southeastward-directed opening and mafic intrusion in the basin itself.

The important point here is that rock units in Margarita cooled very rapidly between 50 and 55 Ma (early Eocene). In the north-south opening model for Grenada Basin, which pulls the Leeward Antilles arc fragments out of the position now occupied by the basin (Pindell and Barrett, 1990), Margarita lies very near the southern rifted margin of the basin (Fig. 5). Thus, Margarita likely formed the footwall to an asymmetric detachment at the site of Grenada Basin. If so, the cooling ages of 50–55 Ma in Margarita may record the time of opening of Grenada Basin. This period matches the time of peak foredeep subsidence in northeastern Maracaibo Basin as allochthonous Caribbean nappes were emplaced southeastward in the Falcón region (Bockmeulen et al., 1983). Therefore, as with central Cuba, rollback of Proto-Caribbean lithosphere toward the continental margin appears to have caused intra-arc extension in the Great Arc, thereby creating Grenada Basin. Margarita was analogous to the south flank of central Cuba during this Paleogene phase of arc lengthening; it was part of the Great Arc's forearc, which was sucked by trench rollback from under the flank of the Great Arc and toward the Proto-Caribbean passive margin with which it eventually collided. Once the Margarita basement units had been tectonically unroofed in the middle Eocene, subsequent dextral-oblique Caribbean–South American collision caused further progressive uplift and cooling of most allochthonous southern Caribbean rocks.

Late Cretaceous and Paleogene of Peru, Ecuador, and Colombia

The lack of younger faulting or volcanism strongly suggests that the Caribbean plate had reached its present size or larger by ca. 90 Ma (Driscoll and Diebold, 1999). Its northern edge was situated south of the Yucatán Block prior to the Campanian (and would collide with southern Mexico during the Maastrichtian), and thus the southern, trailing, edge of the Caribbean plate must have been situated in the vicinity of northern Peru at this time (Figs. 5, 11). As the Caribbean plate then migrated north, the southern edge of the plate also migrated northward along western South America, reaching its present position at 4°N (southern tip of Panama-Baúdo arc) by the Eocene. Following the Early Cretaceous passive margin stage, deformation in the northern Central Andes was therefore first driven by Caribbean interactions and only later, after the Panama Arc had migrated north of any given point, by interaction with the Farallon and Nazca plates. North of 4°N, Andean deformation is still controlled by Caribbean interactions.

There is clear evidence for significant tectonic changes that are probably directly related to the Aptian polarity inversion event in the Great Arc. At ca. 120–115 Ma (Fig. 11A), volcanism ceased in the Celica Arc of northernmost Peru and southern Ecuador. Subduction of the newly-isolated Caribbean plate beneath the active arc along the backarc basin farther north also ceased, and subduction of "Pacific" (Phoenix, Proto-Farallon) plates began beneath the new Panama–Costa Rica arc (initiated in the Albian; Calvo and Bolz, 1994). In Ecuador, the extinct arc (Jaillard et al., 1996), associated plutons (115 Ma; Kennerley, 1980), forearc, and underlying continental basement (Amotape-Tahuín; Aspden et al., 1995) were uplifted and eroded prior to deposition of southwest-derived volcaniclastic flysch (Alamór Fm.; Jaillard et al., 1996). At this time, the oceanic plateau basement of the Piñon Terrane (123 Ma; Reynaud et al., 1999) of western Ecuador lay close to the paleoequator (Roperch et al., 1987), consistent with a position at the trailing edge of the Caribbean plate. The presence of large mafic clasts in basal Talara Basin sediments in Peru (Pecora et al., 1999) indicates that until the Paleocene, the Piñon Terrane still lay 300–500 km south of its present position. The newly formed Panama–Costa Rica Trench at the trailing edge of the Caribbean plate now accommodated most of the eastward convergence of the "Pacific" oceanic plates. Caribbean–South American relative motion was near north-northeast–directed dextral transpression. To the south of the Panama triple junction, volcanism and associated plutonism continued in Peru (Casma Group; Cobbing et al., 1981) during the Albian.

The Caribbean–South America plate boundary remained transpressional, causing closure of the Andean backarc basin and metamorphism of blueschists, which are found today along the Romeral-Peltetec Suture—the west flank of the Colombian central Cordillera and Ecuadorian Cordillera Real (Aspden and McCourt, 1986). Basalts were thrust eastward out of the backarc axis and over Albian limestones on the Antioquia Block, burying the western flank of the Colombian central Cordillera to at least 10–15 km depth, sufficient to drive greenschist-type metamorphism and to reset K-Ar ages. In this model, we propose that the Antioquia Block lay at least 300–400 km south of its present position at 120–100^Ma, close to present day Ecuador, because nowhere in northern or central Colombia have mid-Cretaceous synorogenic deposits or metamorphism been identified. Northward motion of the Antioquia Block (along the Palestina-Otú fault system) to its present position in the central Cordillera can be constrained to the time between backarc basin closure (110–120 Ma) and the late Eocene deposition of the Chorro Group in the Middle Magdalena Basin (overlap assemblage).

Dextral transpression continued during the Albian-Cenomanian (Fig. 11B and 11C), but as separation between the Americas slowed by 84^Ma, relative motion of the Caribbean with respect to South America became more east-west. Initially, this resulted in a higher component of contraction across the closed former backarc basin, and may be the cause of the consistent 85–65 Ma peak in K-Ar cooling ages in the Cordillera Real of Ecuador (Litherland et al., 1994) and the central Cordillera of Colombia (McCourt et al., 1984). From the Maastrichtian on, the east-west component of Caribbean-Colombia convergence was taken up by the establishment of east-dipping subduction of the Caribbean beneath Colombia (Fig. 11D). Seismic tomography data (van der Hilst and Mann, 1994) clearly show an enormous area of the Caribbean plate, which was subsequently subducted eastward beneath Colombia. This reorganization of subduction initiated the accretion of young Caribbean oceanic plateau basalts to the Colombian western Cordillera.

In contrast, in northernmost Colombia, the original Caribbean–over–South America vergence was maintained and the Ruma metamorphic belt was emplaced by the Paleocene onto the northwest Guajira margin, driving latest Maastrichtian–Paleocene foredeep subsidence in the Cesar Basin (Molina Formation). An important river flowed north through this basin, providing immature clastic sediments to the trench ahead of the Caribbean plate. These clastic sediments were then accreted into the Caribbean accretionary prism and later emplaced onto the Falcón portion of the Venezuelan margin (Matatére Formation).

Other aspects of the regional geology also strongly support the picture of south to north diachronous transpression indicated in Figure 11. Inception of the Andean foredeep (indicated by passage of a "peripheral bulge unconformity" and markedly increased rates of subsidence) youngs from south to north. The West Peruvian Trough may have been incorporated into the foredeep as early as the Cenomanian in Peru (Jaillard, 1993, 1994), the Putumayo–Llanos Basin during the Turonian–Santonian, and the César Basin (northernmost Colombia) during the Maastrichtian (Tectonic Analysis, 1998, unpublished report). As the west-east component of relative plate motion increased, underthrusting of the Colombian margin by the Caribbean plate caused Maastrichtian emergence of the northward-migrating Antioquia Terrane, from which west-derived clastic sediments were shed into the Middle Magdalena Valley for the first time (Cimmarona-Umir Formations; Villamil, 1999). Whereas the early Andean foreland was terrestrial to the south of Vaupes Arch by the Maastrichtian (Lower Tena Fm. in Ecuador; Chota Fm. in Peru; Mourier et al., 1988c), it was still open marine in the north (Colón Formation; Villamil, 1999).

Interaction of the trailing edge of the Caribbean plate (Panama Arc) with South America was also clearly diachronous, leaving an unmistakable imprint on the forearc basins and the Andean arc system. The San Lorenzo intra-oceanic volcanic arc was accreted to the western Piñon Terrane during the latest Cretaceous and/or early Paleocene, and the Piñon Terrane was then accreted to northernmost Peru and southern Ecuador. The Macuchi Arc was accreted to central Ecuador and southern Colombia during the Eocene. Uplift of the extinct Early Cretaceous Celica Arc and forearc basement blocks led to forearc limestones giving way to volcaniclastic sediments in the Celica-Lancones area (Copa Sombrero Fm.; Jaillard et al., 1999). Maastrichtian continent-derived siliciclastic turbidites overlap the accreted terranes of the western Cordillera (Yunguilla Fm.; Henderson, 1979; Jaillard et al., 1999; Reynaud et al., 1999; Kerr et al., 2002), while red beds derived from the Cordillera Real were deposited over marine carbonate and shale sequences in the foreland (Jaillard, 1997). Farther west, the Piñon terrane was the site of mostly pelagic sedimentation, with minor local volcaniclastic components, during the Late Cretaceous (Calentura, Cayo, Pilatón Formations; Reynaud et al., 1999) and had not yet been accreted. All of these accreted terranes lie west of the Pallatanga Fault, site of still active dextral faulting (5–7 mm/yr; Trenkamp et al., 2002) and were probably accreted at least 200 km south of their present site, in the area of present-day northern Peru.

Once the Panama triple junction passed to the north, subduction of the Nazca plate beneath South America (which occurred at roughly 5 times the rate of Caribbean subduction) resulted in the reestablishment of andesitic arc volcanism by earliest Eocene in southern Ecuador and middle to late Eocene in southern Colombia. Unmetamorphosed stitching plutons of Eocene age clearly indicate that accretion of basalts and deep-sea sediments had ceased in the Western Cordillera (Kerr et al., 1997). Subsequent northward migration of the Piñon Terrane was driven by oblique subduction of the Nazca plate beneath Ecuador, and its onset is probably indicated by initiation of rapid subsidence in the Talara Basin (northern Peru) during the Eocene. The tail of the terrane is marked by progressively younger onset of subsidence northward into the Progreso and Guayaquil basins of Ecuador (Jaillard et al., 1996).

Paleogene Underthrusting of Caribbean Crust beneath Colombia and the Development of "Negative Flexure" Basins

Progressive early Andean uplift continued into the Paleogene, producing a regional unconformity that affected the central Cordillera, the Lower, Middle, and Upper Magdalena basins, the Putumayo Basin, and parts of the Llanos Basin. The style of uplift can be inferred from the erosional unconformity in the Magdalena Valley; Cretaceous strata beneath the unconformity generally form a deeply eroded, east-dipping homocline (eroded section of ~5 km in the west, diminishing eastward; Pindell et al., 1998). The broad homocline suggests a flexural origin rather than a typical thrust-belt behind a foreland basin. We propose that this unconformity was produced by the obduction and telescoping of the Colombian crust westward onto the Caribbean plate (Figs. 12 and 13). In order for this to be the case, westward drift of Colombia would need to exceed the rollback velocity of the Caribbean Benioff zone. To the east of the area of unconformity, a broad sag-like basin is recorded by up to 300 m of Paleocene and early Eocene deposits (Fig. 13).

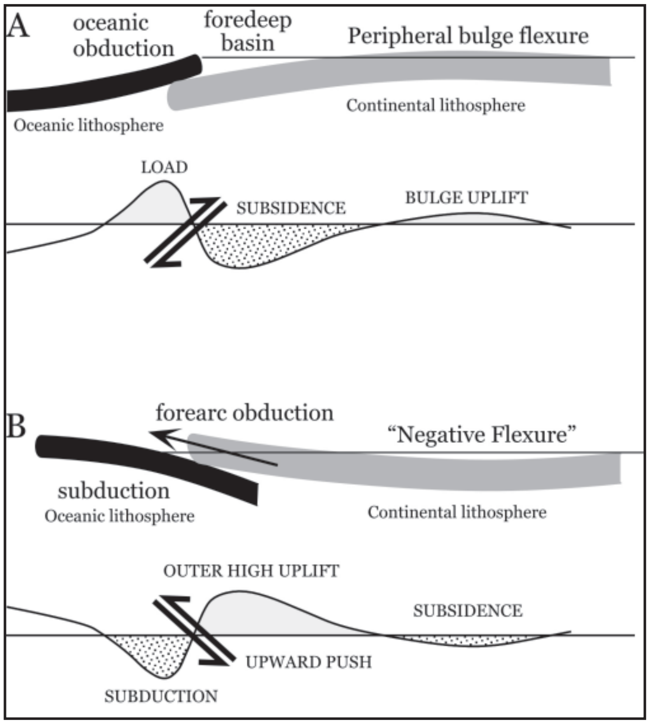

Figure 12. Contrasts between flexural foredeep basins and the "negative flexure" model proposed for the Paleogene development of Colombia. (A) Flexural foredeep basins reflect subsidence adjacent to a load (e.g., a foreland fold-thrust belt) placed on an elastic lithospheric plate. (B) In contrast, "negative flexure" basins can form as a result of the onset of obduction of continental crust onto oceanic crust to form a new subduction zone (i.e., in cases where the overriding plate moves toward the trench faster than the subducting plate rolls back). In such cases, outer highs are flexurally uplifted by the subducting oceanic slab while the interior will sag. Because this is an elastic flexural process, the wavelength of the sag basin will be comparable to the width of flexural forebulges, typically 200–300 km. Where subduction is sufficiently fast, arc development will eventually overprint the records of these early synsubduction basins. However, the early Paleogene sections in Colombia and eastern Venezuela–Trinidad appear to be two cases where subduction was sufficiently slow that the record of the "negative flexure" can be seen.

We coin the term "negative flexure basin" to describe this depocenter, and suggest that such basins may be a standard feature of the onset of subduction beneath formerly passive continental margins, whereby the outer edge of the margin is uplifted anywhere up to 10 km due to underthrusting of the oceanic slab.

The early Paleogene strata in the Colombian negative flexural basin predate significant thrusting, orogen growth, and arc volcanism within the Andes, most of which are of Oligocene and younger age, the palinspastic effects of which have been removed in Figure 13. The Paleogene stratigraphy of this area overlies a post-foredeep unconformity (top of the Maastrichtian foreland) and comprises basal fluvial sandstones (lower Socha, Barco, lower Regadera Formations) overlain by lacustrine shales (upper Socha, upper Regadera Formations). Marine incursion into this basin ("Lake Socha") was intermittent, over a sill at the rising forebulge in the Caribbean region (produced by southeast-directed overthrusting by leading Caribbean nappes). Ongoing uplift and erosion of the central Cordillera eventually resulted in filling of this basin by the early Eocene and deposition of the regressive lower Mirador sandstones.

The negative flexure basin was short-lived. It appears that rollback of Caribbean lithosphere allowed a relaxation of the earlier uplift, as Colombia's advance over the Caribbean plate slowed from ~20 mm/yr to 10mm/yr in the late middle and late Eocene (Pindell et al., 1998). This led to subsidence of the Colombian hanging wall and unflexing of the South American lithosphere, renewed sedimentation at the previously erosive margins of the negative flexure basin (Chorro Group onlaps westward onto central Cordillera), and rebound of the core of the negative flexure, producing an erosional unconformity at the top of the negative flexure fill (a second, higher, "Eocene unconformity") and the retreat of sedimentation northward toward modern Lake Maracaibo.

Continued thrusting of Colombia over the downgoing Caribbean plate cooled the base of the Colombian lithosphere over the following 10 m.y., driving further regional subsidence and renewed deposition, namely the upper Mirador and equivalent transgressive sandstones, and the eventual re-establishment of widespread lacustrine and intermittently marine conditions in a younger basin ("Lake Concentración"; Pindell et al., 1998). The rate and magnitude of this younger pre-Andean subsidence, as indicated by stratal thicknesses, are consistent with known rates of conductive cooling (Tectonic Analysis, 1998, unpublished report).

Paleogene Underthrusting of Proto-Caribbean Crust beneath Eastern Venezuela and Trinidad

Concurrently, to the east in the Paleocene to Eocene, several tens of km of plate convergence occurred between the Americas across the Proto-Caribbean that must have been taken up at structures between the Bahamas and northern South America. Caribbean seismic tomography (van der Hilst, 1990) and regional stratigraphy suggested to Pindell et al. (1991, 1998) and Pindell and Kennan (2001a, 2001b) that this convergence accumulated as the eastern Venezuela–Trinidadian continental crust overthrust the Proto-Caribbean oceanic lithosphere and caused considerable uplift and erosion (Figs. 14 and 15). Recent and ongoing field studies by the authors are confirming and expanding upon long-lived concepts (Guppy, 1911; Hedberg, 1937; Higgs, 2000) of a Paleogene northern outer high along eastern Venezuela and Trinidad (the "North Trinidad Basement High" of Pindell and Kennan, 2002) that provided erosional detritus toward the south. In the northernmost Serranía del Interior, erosion of probable Paleocene age cut down locally to the Barremian level of the Barranquín Formation (Vierbuchen, 1984), and in the subsurface of the Caroni Basin and northern flank of the Central Range of Trinidad, erosion of probable Eocene-Oligocene age cut variably to Eocene, Paleocene, and mid-Cretaceous levels. A second negative flexural trough formed south of this outer high, but unlike the Paleogene tough

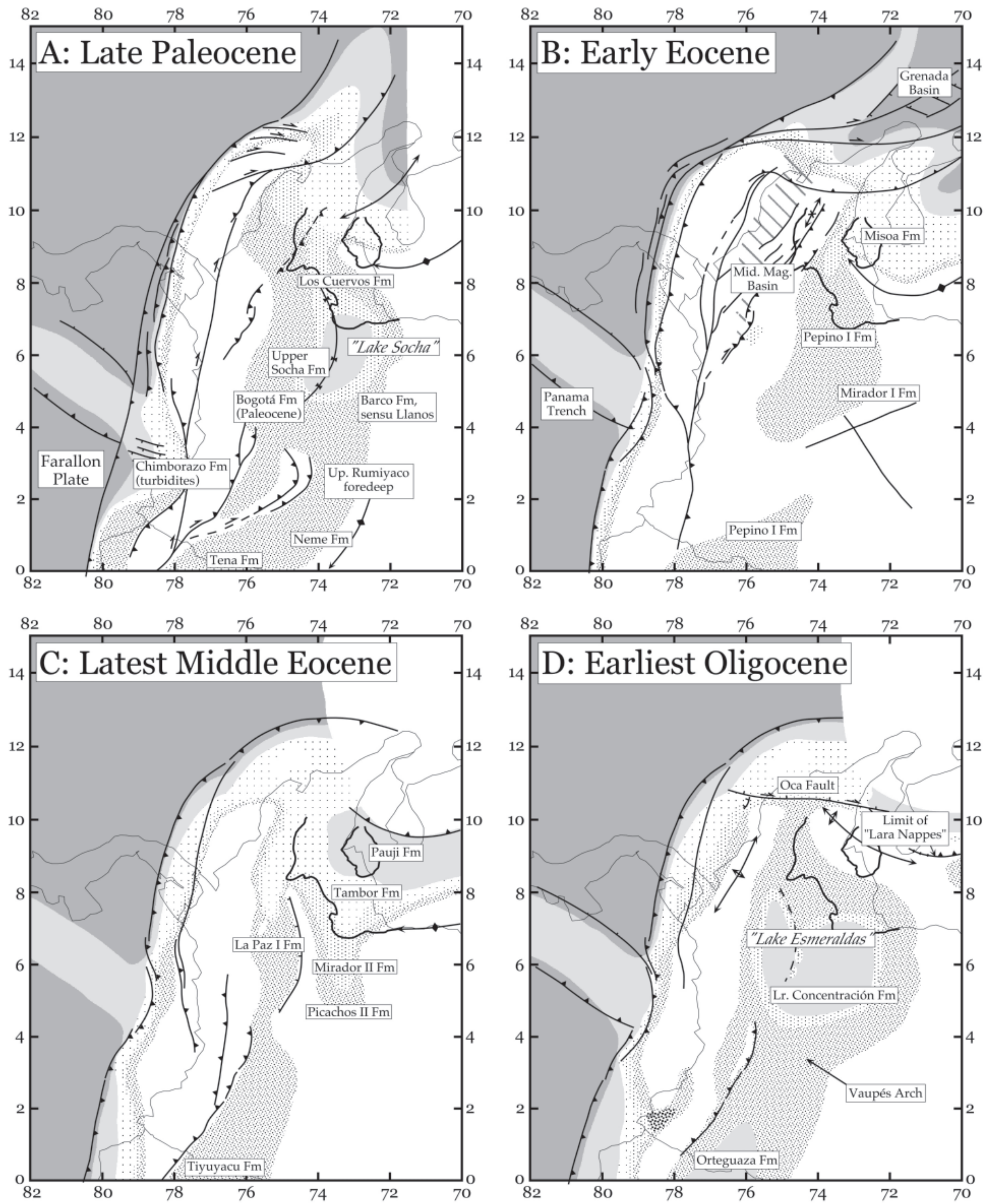

Figure 13. Paleogeographic evolution of Colombian "negative flexure" basins, derived partly from Pindell et al. (1998). (A) Maastrichtian-Paleocene underthrusting resulted in uplift of the central Cordillera, eastward tilting, and erosion, which is recorded in the Middle Magdalena Basin and development of a negative flexure basin in the eastern Cordillera and Llanos region farther east, without associated large-magnitude overthrusting. (B) Ongoing uplift and erosion of the central Cordillera eventually resulted in regressive sandstones filling of this basin by the early Eocene. (C) Slowing subduction and sinking of the Caribbean plate resulted in "unflexing" of the basin, reducing accommodation space and resulting in a subtle overlying unconformity. (D) Continued subsidence of the Caribbean slab, and cooling of the base of the South American lithosphere, resulted in long-wavelength thermal subsidence and development of a wide lacustrine basin in the absence of associated thrusting. Facies patterns as in Figure 11

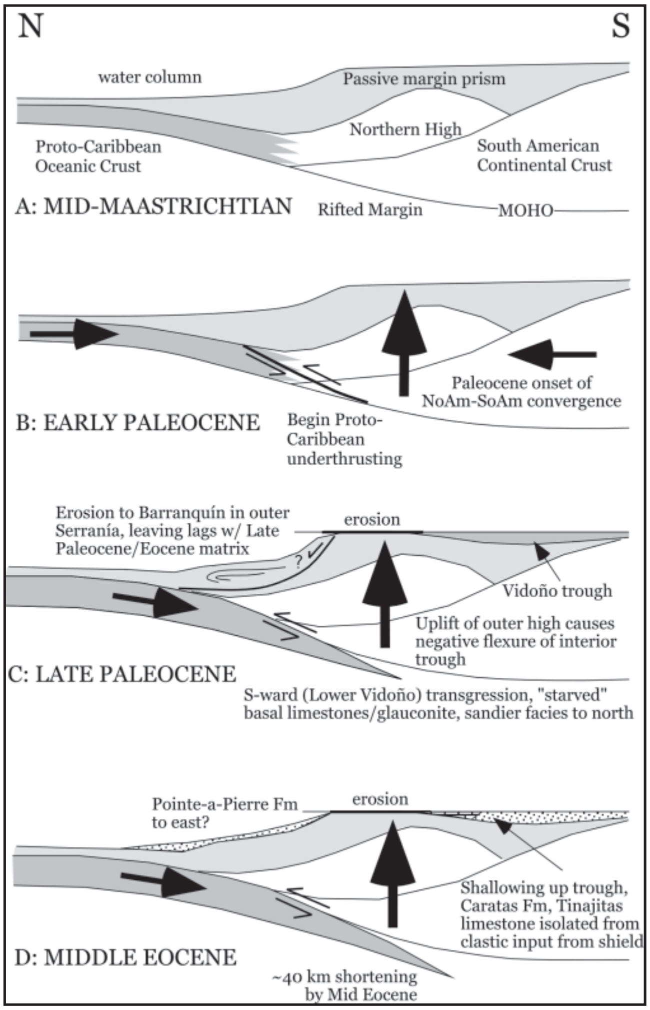

Figure 14. Hypothetical cross sections of the onset of subduction beneath northern South America. As in Colombia, onset of subduction led to the uplift of an outer high, which was a source area for sediments deposited in the "negative flexure" basin to the south.

of Colombia, this trough continued as a marine depocenter that remained connected to the ocean in its eastern and probably its western end. The northern flank of this trough in Paleocene–early Oligocene time received north-derived "orogenic" fan-delta slope sediments including sandy mudstones and turbidites (Vidoño [northern facies], Chaudière [northern facies], Pointe-a-Pierre, and San Fernando [northern facies] Formations), whereas the southern flank was starved of clastics: the Paleocene of Venezuela is characterized by authigenic glauconitic greensands and carbonates (southern Vidoño, possibly Soldado), and the Paleocene-Eocene of Trinidad lacks sand altogether (Lizard Springs and Navet Formations). In Venezuela, the Caratas Formation sands filled much of the trough in the Eocene, due to its proximity to the Shield. In the Oligocene, this basin configuration was changed by drowning of the northern outer high due to the arrival by then of the Caribbean prism's foredeep basin, as the Caribbean plate came ever closer to eastern Venezuela and Trinidad (Fig. 15).

A large volume of mid-Cretaceous hydrocarbon source rocks was probably eroded from the northern outer high. This pre-Caribbean uplift and unroofing of Cretaceous section may explain why fission track studies of northerly Barranquín Formation strata (Locke and Garver, this volume) show depositional rather than Neogene, synorogenic ages (lack of sufficient burial due to prior erosion). Because the total amount of Proto-Caribbean subduction at this margin is very small, ~70 km in eastern Venezuela–Trinidad and ~150 km in western Venezuela (Pindell et al., 1998), a subduction-related magmatic arc has not formed along the margin. Therefore, the record of this marine negative flexural basin has been preserved.

Cenozoic Dextral-Oblique Caribbean–South America Collision in Venezuela and Trinidad

Although the northern Andean terranes of northwest South America are currently being extruded northward (relative to the Guyana Shield) onto the Caribbean plate at the South Caribbean foldbelt (Fig. 1; Mann and Burke, 1984; van der Hilst and Mann, 1994), this plate boundary geometry has developed only since middle Oligocene time (Dewey and Pindell, 1985; Pindell et al., 1998). Prior to the Oligocene, from Santa Marta Massif eastward, initial Caribbean–South America interaction was southward vergent, such that oceanic material was emplaced by dextral transpression onto the former South American margin (Fig. 2). This emplacement of allochthonous material is well recorded by the eastwardly-younging Caribbean foredeep basin (Pindell, 1985a; Dewey and Pindell, 1986; Pindell et al., 1988), which started in the Paleocene at César Basin, Colombia, and continued until the middle Miocene in eastern Venezuela–Trinidad. The onset of Andean extrusion onto the Caribbean plate coincides with a westward acceleration of South America across the mantle (Pindell et al., 1998), and we believe that coupling between the North Andean and Caribbean lithospheres and collision of buoyant parts of the Panama Arc with western Colombia have combined to cause strong, roughly east-west contraction in the eastern Cordillera of Colombia at roughly the same azimuth as Caribbean–South American relative motion. This "Andean" development has involved a foldbelt polarity reversal, probably eastward-younging from offshore western Colombia, which accommodated continued convergence, after arc collision, between the Caribbean and South America (Fig. 5). The Oligocene-Recent extrusion of the northern Andes has taken advantage of the free-face provided by these early north-vergent structures. Foredeep sections have formed in the basinal areas adjacent to the zone of strong Andean strain, which are critical to hydrocarbon maturation history in these basins.

Thus, dextral oblique collision between the Caribbean and South American crusts, with local complexities, characterizes the Paleocene to middle Miocene of the southern Caribbean margin (Pindell et al., 1988; Audemard and Serrano, 2001). After the Great Caribbean Arc rounded the Guajira corner, eastward-

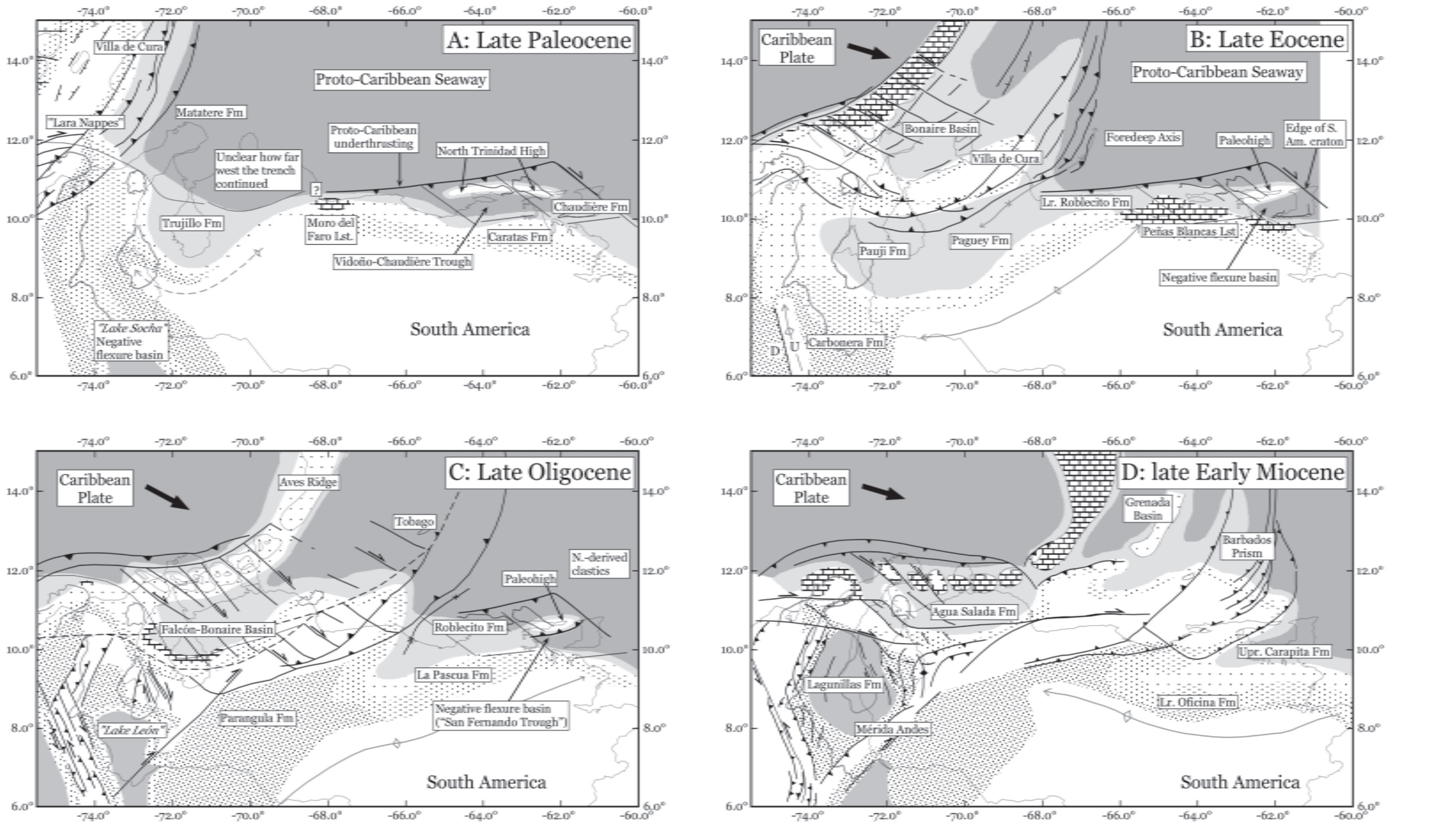

Figure 15. Caribbean–Venezuela/Trinidad arc–continent interactions, simplified from Pindell et al. (1998). (A) By late Paleocene, the leading edge of the Caribbean plate was driving foredeep subsidence in the Maracaibo area. In the east, a negative-flexure basin (coined here the "Vidoño–Chaudière Trough") was the consequence of onset of subduction of Proto-Caribbean lithosphere beneath northern South America. (B) By late Eocene, foredeep subsidence had spread to central Venezuela. Migration of the associated flexural forebulge into eastern Venezuela provided a source for sediments deposited in the negative flexure trough. D and U are down and up sides of faults, respectively. (C) By late Oligocene, the Caribbean foldbelt reached central Venezuela and the foredeep had encroached upon eastern Venezuela. Basement-involved faulting may have accentuated uplift of the North Trinidad Basement High, such that coarse clastics were deposited southward into what we term here the "San Fernando Trough." Note the restored coastline of Northern Range of Trinidad, showing the extent of subsequent transpressive deformations. (D) By late early Miocene, the Proto-Caribbean subduction zone and associated outer high were inactive and incorporated into the Caribbean foredeep. Facies patterns as in Figure 11.

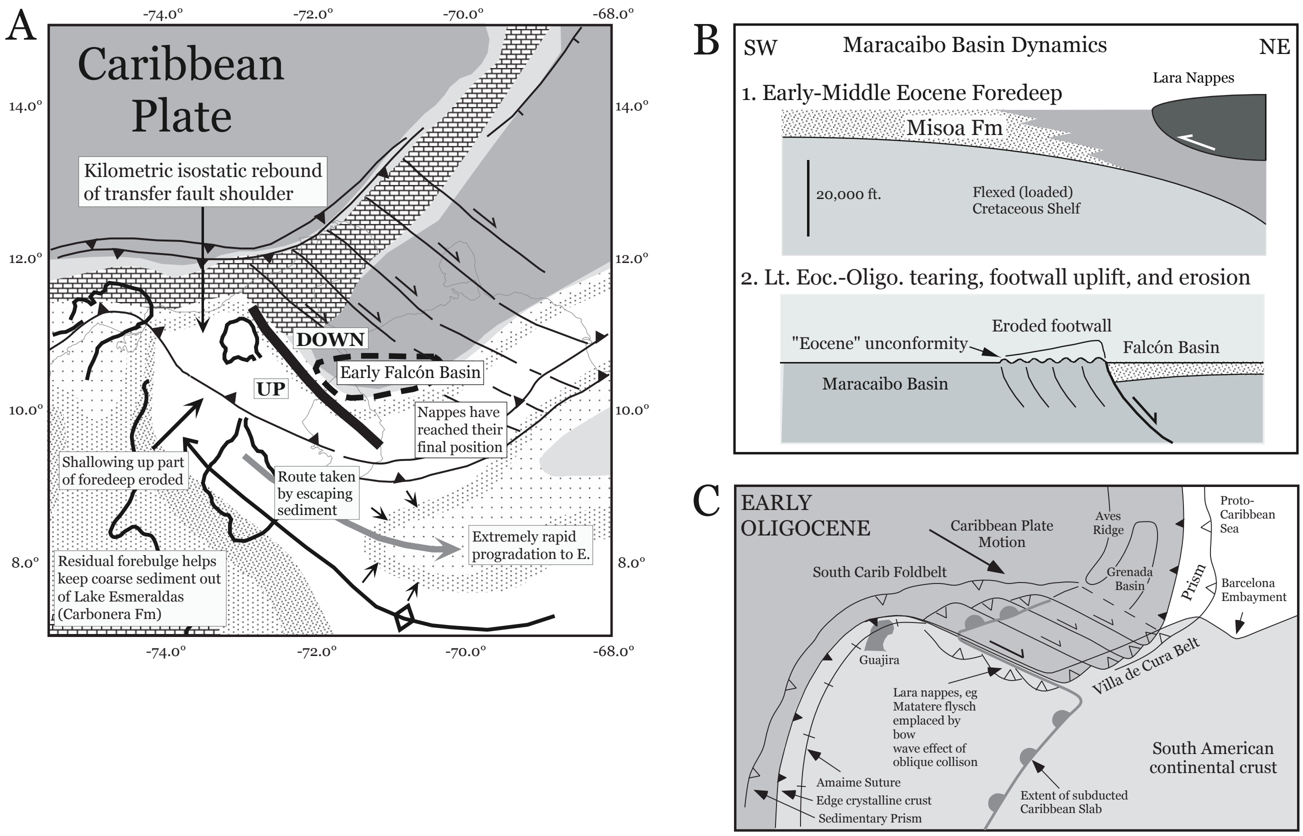

Figure 16. (A) Middle Oligocene reconstruction of the Maracaibo-Falcón area. Southeast-directed thrusting stalled over the thicker continental crust of the Maracaibo Basin but continued over the transitional to oceanic crust northeast of the Mesozoic Maracaibo Transform (see Figs. 4 and 5). (B) Tearing of the downgoing plate led to flexural rebound in the northeast Maracaibo Basin and onset of subsidence in the Falcón Basin. (C) Overview of the complex, lengthening transfer zone that developed between areas of subducting Caribbean plate (northern Colombia) and overriding Caribbean plate (central Venezuela). Facies patterns as in Figure 11.

younging foredeep development from the Maracaibo to Maturín basins, as well as eastward-younging cooling ages of obducted and metamorphosed parautochthonous rocks involved in the collision (Foland et al., 1992), attest to continuous, west-to-east collision along the plate interface. Rounding of Guajira Peninsula and initial collision in western Venezuela was assisted by arc-parallel lengthening as the Grenada Basin opened (Figs. 5 and 15). The north-south opening model of Grenada Basin suggests that the Leeward Antilles Arc originated east of, and is equivalent to, the southern half of Aves Ridge (Pindell and Barrett, 1990). In all our paleogeographic-kinematic modeling, there is insufficient space to bring the Great Arc through the Yucatán-Guajira bottleneck if the Leeward Antilles islands originated along strike of, rather than in front of, the southern Aves Ridge.

It is not clear how far west the Paleogene Proto-Caribbean Trench of eastern Venezuela had developed as Caribbean–South America collision ensued (Fig. 15). It is possible that north-south contraction between the Americas at the longitude of western Venezuela was achieved by downwarping of Proto-Caribbean crust beneath the Caribbean Arc, such that there was no kinematic need for a trench in the west (Pindell et al., 1998). In the east, however, this mechanism is not viable in the Paleogene, because the Caribbean plate was too far west, and seismic tomography and Paleogene stratigraphy support the existence of a Proto-Caribbean Trench there. Thus, for at least the eastern portion of this margin, the Caribbean plate has obliquely overthrust the preexisting trace of the Proto-Caribbean Trench. To the east of Trinidad, toward the North America–South America pole of rotation for this boundary, contraction has been so small, and the mid-Atlantic Ridge was so near, that the eastward continuation of the Proto-Caribbean Trench for much of Cenozoic time has not been identified clearly (Pindell and Kennan, 2001a, 2001b).

Within the Caribbean oblique collisional history, a secondary development occurred in the northern Maracaibo–Falcón region, where the Trujillo Embayment existed in the original passive margin (Fig. 4). As the Lara Nappes were emplaced toward the southeast across this embayment in the Eocene-Oligocene, the emplacement direction was parallel to the western side of the embayment (Fig. 16).

Thus, the thrust belt north of the Maracaibo Basin is a transpressional lateral ramp rather than a south-directed thrust front, but a significant load was still imparted on northern Maracaibo Basin. However, shortening progressed so far to the southeast toward central Venezuela that the autochthonous South American lithosphere was forced to tear, northeastern side down, to accommodate the continued shortening. The effect of this was that the western side of the tear rebounded isostatically, leading to several km of late Eocene–early Oligocene uplift and erosion in northern Maracaibo Basin. This event most certainly arrested hydrocarbon maturation that had been in progress in northern Maracaibo Basin due to early and middle Eocene foredeep deposition. In addition, tearing of the autochthon probably facilitated the Oligocene emplacement of basalts in the Falcón Basin section (Muessig, 1978).

Another secondary development was the late middle Miocene change in Caribbean–South American relative motion direction (Pindell, 1994; Pindell et al., 1998). Prior to 12^Ma, Caribbean–South America motion was directed east-southeast, whereas relative motion since 12 Ma has been directed slightly north of east; GPS studies confirm this for the present day (Weber et al., 2001). Figure 17 shows three "snapshots" of this late Cenozoic evolution in the eastern Venezuela–Trinidad area, showing: (1) end of oblique collision, (2) onset of transcurrent plate boundary development, and (3) post-4 Ma dextral transpression across the region.

In the Miocene, until ca. 12^Ma, dextral oblique thrusting in the Serranía del Interior of Venezuela and the Nariva Fold Belt of Trinidad was accompanied by foredeep loading in Maturín and Southern basins. However, after the 12 Ma relative motion change, late Miocene sediments in the Maturín and Southern basins step northward and bury deeper thrust structures with little or no evidence for further thrusting in the subsurface thrust wedge. Between 12 and 4^Ma, east-west motion between the Caribbean and South American plates appears to have become strongly partitioned along the Cariaco–El Pilar fault zones, stepping southward across the late Miocene to Recent Gulf of Paria pull-apart basin, along which there is evidence for subtle collapse of the middle Miocene allochthons back to the north. Farther east, along the Central Range and south to the Point Radix–Darien

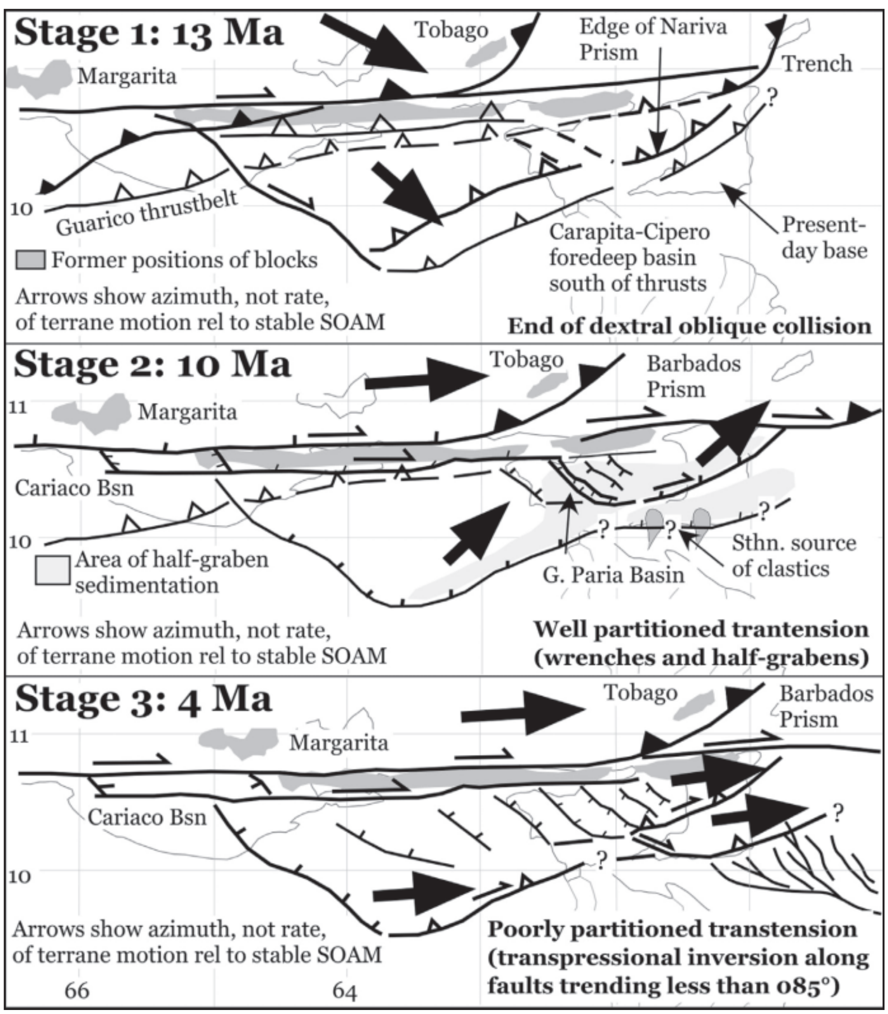

Figure 17. Proposed map view evolution of eastern Venezuela and Trinidad since 12^Ma, modified after Pindell and Kennan (2001a). Stage 1 shows the end of oblique collision of the Caribbean plate. During Stage 2, transtensional movements between the Caribbean and South American (SOAM on figure) plates were well partitioned into strike-slip at the Coche–North Coast Fault and Gulf of Paria pull-apart basin and northward listric extension along the half-graben faults in the Maturín and the South Trinidad basins. Finally, in Stage 3, the Caribbean and South American plates become more strongly coupled, possibly because the deep root of the Antilles arc could not override the North Trinidad High of South American lithosphere. As a result, both contractional and strike-slip faulting, with associated pull-apart formation, spread southeast from their late Miocene location.

Ridge Fault Zone, displacement was transpressional because reactivated middle Miocene thrust structures were oriented slightly counterclockwise of late Miocene relative plate motion.

A third secondary development is seen at ca. 4^Ma, documented by the onset of shortening in the Southern Basin of Trinidad, enhanced transpressional uplift in the Central Range, onset of significant subsidence in the Columbus Channel and eastern offshore areas (drowning of late Miocene deltas by Pliocene shelf sediments), enhanced rates of pull-apart formation in the Gulf of Paria and of transpression in the eastern offshore, and initiation of extensional collapse (Wood, 2000) of the foredeep basin fill toward the Atlantic south of the Darien Ridge (Pindell and Kennan, 2001a). The strong partitioning of strain appears to have broken down, and little or no significant strike-slip on the North Coast Fault Zone can be recognized since 4 Ma (Tectonic Analysis, 2002, unpublished seismic interpretation). The Caribbean plate appears to have become more strongly coupled with deeper basement of South America, possibly because the deep root of the Antilles volcanic arc (Margarita Ridge) began to encounter the northern limit of the North Trinidad Basement High (Pindell and Kennan, 2002).

SUMMARY DEPICTION OF TECTONIC STYLES OF SELECT ARC-CONTINENT INTERACTIONS

Figure 18A–N shows a series of schematic maps and cross sections summarizing proposed and possible mechanisms for arc-continent interaction between elements of the Caribbean plate and the Americas and outlining the context of key HP-LT occurrences.

At ca. 120^Ma, west-dipping subduction of Proto-Caribbean oceanic crust was initiated by subduction polarity reversal beneath the Great Arc. We envision that the new Great Arc trench was connected to the ongoing, east-dipping western Chortís trench by an ever-lengthening transfer zone of arc-parallel stretching and through-going, propeller-shaped, sinistral shear zones (Fig. 18A). The cross section (Fig. 18B) shows the sutured subduction or strike-slip zone between Pacific-derived (Caribbean plate) crust and the former Inter-American Arc, the through-going sinistral shear zones, and newly-initiated oblique subduction of thinned Chortís passive margin beneath the Great Arc. The protolith for the Escambray Massif, Cuba, was probably Jurassic quartzose sandstone and carbonate from part of the Chortís passive margin. Protoliths of other Caribbean HP complexes include (1) Proto-Caribbean oceanic crust (e.g., Río San Juan Complex, Dominican Republic); (2) fragments of the former northeastern fringe of the Inter-American Arc that entered the trench immediately upon reversal, or sometime later by subduction erosion of the Great Arc hanging wall (e.g., Villa de Cura Complex); and (3) continental blocks of the Inter-American Arc that may have been rifted from the Americas during Neocomian intra-arc spreading (e.g., Juan Griego Unit, Margarita). Although Sisson et al. (2003) suggested that the southern Motagua HP-LT rocks, which formed from oceanic crust and give Ar-Ar ages of 116–125^Ma, may pertain to an early collision between Chortís and Mexico, two other possible origins worth noting are (1) Proto-Caribbean or Inter-American Arc crust that was subducted at 120^Ma, like other Caribbean HP complexes (although this does not explain apparent ages at the old end of Sisson et al.'s [2003] age range); and (2) Pacific crust (like that at Baja California; Baldwin and Harrison, 1989) subducted at the west-facing trench near the southern end of Chortís shortly prior to subduction polarity reversal (ca. 130 Ma), such that initial exhumation (toward the end of the 116–125 Ma interval) may directly relate to arc-parallel stretching within the lengthening sinistral transfer zone between the Chortís and Great Arc trenches during the subduction polarity reversal. Such an origin may best explain the possible Jurassic HP components in Cuba's Escambray complex (Maresch et al., 2003), as well as possible 118 Ma components in the Northern Serpentinite Mélange (García-Casco et al., 2002).

In the Late Cretaceous, the Great Arc migrated northeast over Proto-Caribbean oceanic crust toward sinistral-oblique collision with Yucatán at ca. 70 Ma (Fig. 18C). The south Motagua HP-LT rocks may have formed the hanging wall of the trench at this time in the two possible origins for them suggested above, but not if they were formed by Chortís-Mexico collision. The ophiolites and mélanges later emplaced onto the Yucatán margin are interpreted to form part of the accretionary wedge above subducting Cretaceous Proto-Caribbean oceanic crust (Fig. 18D). Proto-Caribbean oceanic crust also likely formed the protolith for the younger, northern Motagua HP-LT rocks, and may only have been subducted shortly before Maastrichtian collision. This Late Cretaceous collision event exhumed HP-LT rocks and thrust less-metamorphosed ophiolites and ophiolite mélange (Santa Cruz, Cuchumantanes) onto the Yucatán margin, driving subsidence of the Maastrichtian Sepúr foredeep (Fig. 18E). Highly oblique collision between the Great Arc and Yucatán may have allowed slivers of older HP-LT rocks to be left behind north of a "Proto-Motagua Fault" and south of the younger HP-LT belt, while most were carried farther east (i.e., Jamaican and Cuban HP rocks). Alternatively, the southern Motagua HP-LT rocks may have been emplaced after passage of the Great Arc to the east, with the southern Yucatán margin forming the northwestern side of the sinistral shear zones that link the Chortís and Caribbean trenches. Relative motions between the Great Arc, the easternmost end of Chortís, and southern Yucatán during the Late Cretaceous are likely to have been complex. Our maps suggest that sinistral transtension is likely, and may play a role exhuming the younger HP-LT rocks in Yucatán. Finally, during the middle to late Cenozoic (Fig. 18F), strike-slip on the Motagua Fault, linked to opening of the Cayman Trough, juxtaposed the Chortís Block and southern Yucatán, and the older and younger HP-LT suites reached their current relative positions. We consider that south-directed thrusting in this highly transpressive zone may have placed the older, more southerly HP-LT suite onto the Chortís Block, because we are confident that prior to this time Chortís lay well west of the Great Arc–Yucatán suture zone (Pindell and Dewey, 1982), and we are not convinced that a former oceanic basin ever existed between Chortís and southwest Mexico that may have produced these rocks. The trace of the neotectonic Motagua Fault appears

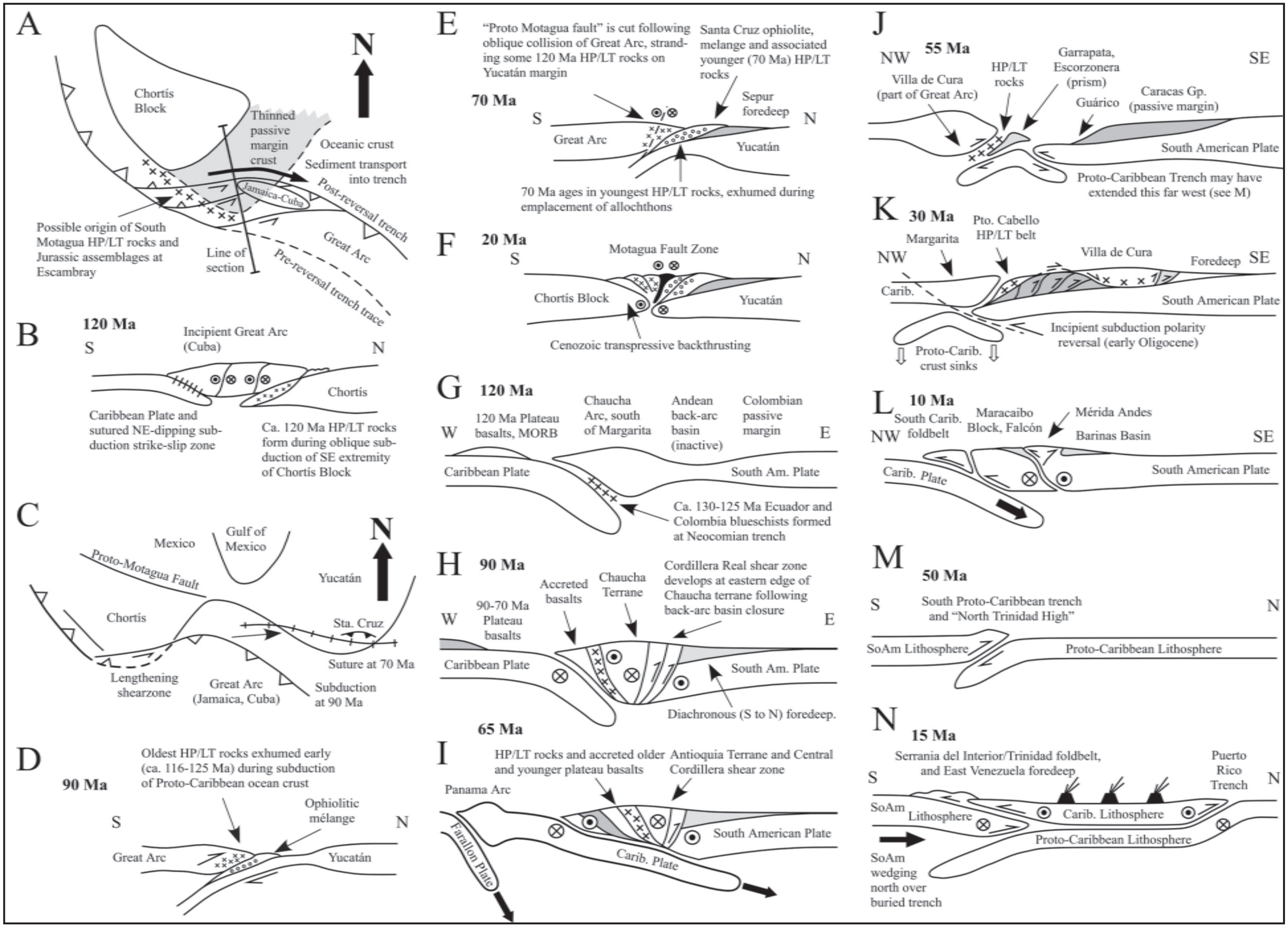

Figure 18. Schematic maps and cross sections, not necessarily to scale, illustrating the proposed and possible mechanisms of formation and emplacement of allochthons and associated high-pressure–low-temperature (HP-LT) rocks on the margins of North and South America. MORB—mid-oceanic-ridge basalt; SoAm—South America.

to follow the boundary between older (ca. 120 Ma) and younger (70 Ma) HP-LT belts, suggesting that prior to the ?Miocene (age of emplacement onto Chortís?), the older HP-LT suite lay some distance west of its present position.

A similar scenario, with dextral strike-slip along the southeast flanks of the Caribbean plate linking southwest-dipping subduction beneath the Great Arc with ongoing northeast-dipping subduction beneath the Central Andes, appears to best explain the age and location of HP-LT rocks and associated structures in Ecuador and Colombia. A schematic cross section of the northern Andes at ca. 120 Ma (Fig. 18G), located south of the paleoposition of Margarita at approximately the latitude of present-day Ecuador, shows key Andean elements at the time of subduction polarity reversal farther north. The Chaucha Terrane (rifted blocks of Carboniferous to Jurassic continental crust) is interpreted as the basement of an arc formed due to subduction of "Pacific" oceanic crust beneath South America prior to 120^Ma, and there is no evidence of arc activity from 120 to 60 Ma north of Peru. The blueschists in Ecuador and Colombia are older than those in Cuba and Margarita and lie west or northwest of the Chaucha (pre-reversal) arc and forearc fragments, suggesting that they formed in the "Pacific" subduction zone prior to subduction polarity reversal. Continued opening of the Proto-Caribbean Seaway after the subduction polarity reversal resulted in highly oblique dextral relative motion between the Caribbean and South America, in contrast to the head-on subduction beneath the Great Arc farther north. The oblique closure of the Andean backarc basin formed the Cordillera Real of Ecuador and the Central Cordillera of Colombia, and inactive fragments of the former Inter-American Arc were stranded in Ecuador (Chaucha Terrane, Fig. 18H). Some slivers of the passive margin were transported several hundred km north of their original position in Ecuador (e.g., Antioquia Terrane of central Cordillera, Colombia). As in the north, this zone of transpressive shear deformation must have lengthened northward with time, and interaction between the leading edge of the Caribbean must have younged from south to north. Several lines of evidence support this conclusion. Exhumation of parts of the shear zone in central Ecuador is pre-Campanian (cooling ages, first appearance of "orogenic" clastic sediments), but the first west-derived, "orogenic" sediments are of Maastrichtian age in the Middle Magdalena valley. Also, unconformities and sequence boundaries related to northward migration of the Caribbean foredeep are Coniacian-Santonian age in southern Colombia, but are of late Maastrichtian age in northernmost Colombia.

A ca. 70–65 Ma section through Colombia (Fig. 18I), slightly north of that shown in Figure 18H, shows no fragments of the former arc. Instead, blueschists, serpentinites, and fragments of ca. 120 Ma plateau basalts, derived from the Caribbean plate, are juxtaposed directly against the central Cordillera–Antioquia Terrane along the line of the present-day Romeral Fault. Remnants of the arc may be present in the subsurface of the Lower Magdalena Basin, and arc or forearc remnants outcrop on the Guajira Peninsula and farther east. Campanian termination of Proto-Caribbean seafloor spreading (Pindell et al., 1988) caused Caribbean–South America relative motion to change from northeast to east, triggering, in the west, renewed low-angle, non-volcanic, subduction of Caribbean crust beneath South America. This led to the accretion of Caribbean Plateau Basalts in the Western Cordillera and development of an accretionary prism of offscraped Caribbean plate strata (San Jacinto Belt). The Panama Arc, which marks the trailing edge of the Caribbean plate, still lay to the southwest at this time, but began to underthrust the Colombian margin during the Paleogene, driving eastward thrusting in and east of the Central Cordillera at the latitude of the Upper Magdalena Valley (Chusma belt), and leading to the establishment of a volcanic arc in southern Colombia that was produced by subduction of Farallon crust south of the Panama Arc–Andean Trench triple junction.

By Paleogene time, the leading edge of the Caribbean plate had rounded the Guajira Peninsula of Colombia. The dextral-oblique opening of the Grenada backarc basin (Pindell and Barrett, 1990) allowed the Caribbean arc and forearc to be thrust southeastward onto the western Venezuelan margin. Figure 18J shows northern Venezuela during the early Eocene, shortly before obduction of the Villa de Cura allochthon. The Eocene Garrapata (and possibly the Los Cajones Member of Guárico and the Escorzonera Formations) allochthonous rocks (prism) have an arc-orogenic character and are interpreted to have formed ahead of the Caribbean arc and forearc. The bulk of the quartz-turbiditic Guárico Formation is shown in a slope and rise position north of older passive margin section, and the Caracas Group is shown as pre-Cretaceous passive margin section overlying extended Paleozoic continental crust. As noted earlier, it is not clear if the Proto-Caribbean Trench along northern South America, well-developed to the east (see below), had propagated this far west. If it had, then the Caribbean prism must have overridden the trace of the Proto-Caribbean trench, as opposed to a simple passive margin, before emplacement onto the Venezuelan margin, and the hanging wall of the Proto-Caribbean trench may have contributed to the Garrapata and other orogenic facies. During the late Eocene and through Oligocene time, the Villa de Cura allochthon (partly HP-LT metamorphosed arc and MORB rocks) was emplaced over the Caracas Group, which was strongly deformed, forming a duplex beneath a major Caribbean sole thrust marked by HP-LT rocks (Fig. 18K). The primitive (PIA) character of parts of the Villa de Cura Group suggests to us a partly pre-polarity reversal (i.e., older than 120 Ma) age for protoliths, in keeping with the ages of other PIA rocks in the Great Arc. The northern portion of this sole thrust could be interpreted to coincide with the coastal Puerto Cabello HP-LT rocks. The Garrapata-Escorzonera-?Cajones prism, the Guárico Formation, and the Upper Cretaceous cover of the passive margin (Querecual and Mucaria Formations) were detached from the underlying, mostly older Caracas Group and thrust ahead of the Villa de Cura allochthon, driving subsidence in the La Pascua–Roblecito foredeep. Although Smith et al. (1999) proposed a western Colombian derivation of Paleozoic continental components in the Puerto Cabello zone, another alternative origin to consider is the basement beneath the Caracas Group. Because the Caracas Group was itself detached from its basement, the Caribbean sole thrust may have been able to incorporate Paleozoic basement material, perhaps leading to a complex mix of lithologies and protoliths seen in the Puerto Cabello complex (Avé Lallemant and Sisson, 1993). Caribbean plate oceanic and arc rocks of Margarita and the Leeward Antilles lay to the north of the Puerto Cabello area.

Toward the end of or after emplacement of the Caribbean allochthons in northern Colombia and western Venezuela, the accreted arc material and elements of South American basement began to be backthrusted onto the Caribbean plate (Fig. 18K and 18L; note section L is drawn somewhat west of K to include Maracaibo Basin). Northward thrusting is limited in central Venezuela, north of the Morón Fault, but was strongly enhanced in the Maracaibo area by northward tectonic escape of the triangular Maracaibo Block between the Santa-Marta-Bucaramanga and Boconó faults.

Figure 18M summarizes eastern Venezuela–Trinidad during the Paleogene, where a Proto-Caribbean Trench developed prior to the arrival of the Caribbean plate from the west to accommodate north-south convergence between North and South America, producing an outer high in the Paleogene. Subduction was minor (<150 km) and slow and did not produce a volcanic arc on northern South America. During the Miocene to Recent (Fig. 18N), the Caribbean plate migrated into the line of section from the west, burying the northern edge of South American continental basement and driving southeast-directed thrusting in Venezuela (Serranía del Interior Oriental) and Trinidad (Naparima belt). Continued convergence between North and South America has enhanced the apparent "wedging" of South America between the Caribbean and Proto-Caribbean plates. Volcanism on the Caribbean plate has been confined to those areas where a significant asthenospheric wedge existed above the subducting Proto-Caribbean plate, and thus, volcanism has probably been shut off diachronously from west to east during the oblique Caribbean–South America collision.

CONCLUSIONS

The American continental margins of the Caribbean region formed during Jurassic continental breakup of Pangea and originally faced the Proto-Caribbean Seaway in which seafloor spreading resulted from separation of North and South America. These margins were later overridden diachronously by allochthonous arc and oceanic complexes that were part of the Caribbean plate or its accretionary complex. The stratigraphic and structural development of the American margins was strongly controlled by tectonic events prior to and including these arc-continent interactions.

Caribbean lithosphere originated in the Pacific and was progressively engulfed between the Americas after Aptian time by the inception of a west-dipping Benioff zone in the widening gap between the Chortís Block and the northern Andes. This trench is now marked by the Motagua (Guatemalan) and Cuban sutures of southern North America, and by the Ruma (Guajira) and Villa de Cura (Venezuela) nappes of northern South America, and persists today as the Lesser Antilles subduction zone. Most Caribbean HP metamorphic complexes originated at this east-facing Aptian and younger subduction zone. Arc-polarity reversal at an earlier (Neocomian) west-facing Inter-American Arc is implied, but the site of this "arc" may have been dominated by sinistral transform motions between two oceanic plates rather than by significant amounts of east-dipping subduction. Geochronologic data and metamorphic relations of the HP complexes, as well as the arc magmatic and structural histories of today's pieces of the Great Arc, all support a roughly 120 Ma age for this polarity reversal event, with uplift and exhumation well under way prior to 90^Ma. Several HP complexes along the Ecuadorian and southern Colombian Andes, and two Caribbean HP complexes (south Motagua, Guatemala, and parts of Escambray, Cuba) possess components that pre-date the Aptian; this suggests that they originated along the west-facing Inter-American Arc (like those in Baja California; Baldwin and Harrison, 1989) prior to polarity-reversal, but were then intimately involved with the arc-polarity reversal and part of the subsequent migration of the Great Arc. Processes that we can tie to progressive exhumation of HP rocks in the Great Caribbean Arc include arc-parallel stretching, obduction of forearc materials onto continental rifted margins (e.g., north Motagua), and low-angle detachment at intra-arc basins (Yucatán and Grenada Basins, final cooling at Escambray and Margarita). Counterflow within the subduction complex is also possible. A number of Caribbean HP complexes were considerably uplifted by 90^Ma, and we consider that arc-parallel stretching (Avé Lallemant and Guth, 1990) by various processes at the highly oblique northwestern and southeastern portions of the Great Arc subduction zone (as opposed to the more head-on, northeastern part of the Great Arc) was a major factor in the progressive uplift histories.

Given the clear Aptian age (ca. 120 Ma) for the onset of polarity reversal, which was achieved by the Albian, the 90 Ma Caribbean Basalt Plateau could not have played a role in triggering the polarity reversal. Rather, the Caribbean basalts were extruded onto preexisting Caribbean crust after polarity reversal and during engulfment by the Americas. Polarity reversal was probably triggered, instead, by the westward acceleration of the Americas, thereby causing strong compression at the entire Cordilleran arc system. Oceanic plateau basalts were able to erupt onto the Caribbean Basin because that lithosphere remained almost fixed relative to a deep mantle reference frame, and the bounding subduction zones were young and did not root deeply enough into the mantle by 90 Ma to block the paths of plumes to the surface.

Evidence increasingly suggests that amagmatic, Paleogene subduction of Proto-Caribbean lithosphere occurred beneath eastern Venezuela and Trinidad, ahead of the migrating Caribbean plate, which accommodated the well-documented Cenozoic convergence between North and South America. Subduction of Caribbean crust beneath western Colombia also began in Maastrichtian time. In both settings, strong uplift occurred in the hanging walls close to the trench as subduction began, while shallow subaqueous troughs formed landward of these highs in the apparent absence of bounding thrusts. We propose a new class of basin to describe this condition, the "negative flexure basin," produced as a consequence of flexure during the onset of subduction at a preexisting passive margin. The early Paleogene sections of Colombia and eastern Venezuela–Trinidad provide epicontinental and pericontinental examples, respectively.

Finally, stratigraphic and structural style was altered dramatically in eastern Venezuela and Trinidad at ca. 12^Ma, when Caribbean–South America relative motion changed from east-southeast–directed to east-directed. Since 4^Ma, increased coupling between the Lesser Antilles Arc and deep South American basement has resulted in less partitioning of interplate shear and the *onset* of transpressive shortening in southern Trinidad, coeval with pull-apart basin formation in the Gulf of Paria, where little recent thrusting is recorded.

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