

Late Cretaceous to Miocene tectonic reconstruction of the northwestern Caribbean

- regional analysis of Cuban geology

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Ernst-Moritz-Arndt-Universität Greifswald

vorgelegt von

Max Sommer

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in Mannheim

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Dekan: Prof. Dr. Klaus Fesser (Universität Greifswald)

.....

1. Gutachter : Prof. Dr. Martin Meschede (Universität Greifswald)

.....

2. Gutachter: Prof. em. Dr. Wolfgang Frisch (Universität Tübingen)

.....

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Abstract

The Caribbean is a geologically complex region with several different plate boundary interactions. Geodynamic reconstructions of the northwestern Caribbean region have been particularly controversial in terms of the number of arcs, subduction polarity, and timing of collision. This thesis develops a refined tectonic reconstruction for the northwestern Caribbean based on a review of geological data of Cuba and a regional analysis within the northwestern Caribbean context.

With regard to plausibility, significant emphasis was put on the degree and quality of visualization. Three crustal sections across key areas in western, central, and eastern Cuba have been constructed in order to conduct an evolutionary interpretation in three dimensions.

Western and central Cuba constitute an orogenic belt resulting from the collision of a mid- to Late Cretaceous volcanic arc – the "Great Caribbean Arc" – with the southern paleomargin of North America. The collision process apparently started in the Campanian, but major north- to northeast-directed thrusting processes at the southern Bahamas margin culminated during the Paleocene.

A Continuous southwest-dipping polarity of the "Great Caribbean Arc", at least from the Aptian-Albian, can be inferred from (1) its Late Cretaceous approach towards the North American margin, (2) the magnitude of top to the north directed tectonic transport in the Cuba orogenic belt, and (3) the internal structures of the metamorphic fore-arc assemblages and their evolution on the north side of the arc.

An Early Cretaceous southwest-dipping origin of the "Great Caribbean Arc" along the northern fringe of the Chortís Block appears to be in all probability. This concept provides a link between (1) middle Late Cretaceous collision processes along the Matagua suture zone, (2) the Turonian termination of "Great Caribbean Arc"-activity on Jamaica, and (3) the late Campanian onset of collision in the Cuba orogenic belt.

The collision of the "Great Caribbean Arc" with the Bahamas margin hampered relative northward motion of the Caribbean Plate from the late Campanian onward. Continued northward push finally resulted in the commencement of north-dipping subduction. Late Cretaceous commencement of north-dipping subduction was accompanied by superposition of oceanic crust and large-scale north-directed gravity sliding on the upper plate, as documented by ophiolitic slide-masses and Maastrichtian olistostromes in eastern Cuba (Nipe – Cristal and Moa Baracoa ophiolite massifs) as well as on Jamaica (ophiolites of the Bath-Dunrobin Complex) and the southern peninsula of Hispaniola.

Progress of north-dipping subduction was responsible for the emergence of a Paleocene to Middle Eocene volcanic arc which spanned the northwestern Caribbean along the southern boundary of the Yucatán Basin while the Chortís Block and the

Nicaragua Rise were still in a paleoposition to the south of the Maya Block. North-dipping subduction and the associated volcanic arc isolated the Yucatán Basin from its original affiliation to the Caribbean Plate.

Relative northward motion of the Caribbean Plate and activity of the Paleogene volcanic arc stopped after the Eocene arrival of thickened oceanic crust of the Caribbean Large Igneous Province at the north-dipping subduction zone.

After the late Early Eocene commencement of spreading at the Mid-Cayman Rise, North America – Caribbean relative motion was taken up along the sinistral Oriente Fault with estimated amounts of 800 to 1000 km offset since the Middle Eocene. This transform margin dismembered the northwestern Caribbean extend of the Paleocene to Middle Eocene volcanic arc. Its eastern bend was uncoupled in the course of this process and is probably represented by the Aves Ridge.

South-central Hispaniola can be restored to a Middle Eocene position to the south of eastern Cuba, which accounts for an approximate Cenozoic displacement of 200 to 300 km. Therefore, most of the western prolongation of the Oriente Fault must be accommodated at the northern bounding-faults of the southern peninsula of Hispaniola.

The proposed synthesis is in clear accordance with the paradigm of plate tectonics, corroborating its capability to incorporate even a complex region like the Caribbean.

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1 Introduction

Just a glance at a bathymetric map of the Caribbean region reveals a premonition of the fascinating dynamics of the Earth's lithosphere. Looking at both, the Caribbean and its tectonic counterpart in the region of the Scotia Sea is even more suggestive. At least from a mobilistic perspective, the impression of some energetic process and quasi-fluid behavior of oceanic crust becomes apparent, as if the South American Continent would displace bow waves by its westward drift.

However, the Caribbean is a geologically complex region with several different plate boundary interactions - and the devil is in the details. Up to recent time, some authors are convinced that the paradigm of plate tectonics cannot incorporate the Caribbean region (Morris et al. 1990) or at least that the development of the Caribbean did not involve diachronous processes, major plate migration, changes in direction of plate migration, and rotations of continental blocks or parts of volcanic arcs (James 2006).

Mobilist views of the tectonic development of the Caribbean Plate assume significant amounts of eastward migration relative to the Americas, but differ with regard to the total amounts of relative eastward migration and whether to suppose a "Pacific" or "intra-American" location of origin. Essential differences are concerning the concepts of active margin developments.

Geodynamic reconstructions of the northwestern Caribbean region have been particularly controversial in terms of the number of arcs, subduction polarity, and timing of collision. Current models partly contradict or neglect geological facts on the regional and sub-regional scale.

The aim of this thesis is a review of the fundamental geological data of Cuba and a regional analysis within the northwestern Caribbean context. As a synthesis, an evolutionary interpretation is aspired, which represents a refined tectonic reconstruction for the northwestern Caribbean region and which perhaps allows for some essential inferences for the evolution of the Caribbean Plate.

2 Methodology

This thesis is mainly based on a review of literature. An attempt was made to survey the current state of geological data about Cuba with regard to geodynamic relevance. For the purpose of this thesis, the focus needed to be expanded to geological relations between Cuba, Hispaniola, Jamaica and a general overview of the Caribbean region.

Outcrop observations were gathered during two field campaigns on Cuba in 2003 and 2005. These observations were fundamental to the understanding of literature and many geological features, e.g. the deformational styles and the structure of the fold and thrust belt of central Cuba (Fig. 22), or the structural position of ophiolite massifs in eastern Cuba (Fig. 25; Appendix, VI A171).

Field observations and the evaluation of literature were accompanied by a careful study of the Cuban geological cartography 1 : 250.000 which is mostly based on mapping results from the 1970's and 1980's (Puscharovskiy et al. 1988). The map series also contains documentations of type sections and boreholes.

The compilation of data from literature, outcrop observations and the geological map series are concentrated in the "overview" chapters 3 and 4. These chapters are projected as a plain listing of facts, largely free from any evolutionary interpretations. This may in parts take its toll in terms of reduced readability, but this was accepted for the benefit of a reasonable organization of the thesis.

In order to recapitulate the compilation of geological data from Cuba, three crustal sections across key areas on Cuba have been constructed. The section lines, presented in chapter 5, are located in western, central, and eastern Cuba respectively (Fig. 9). Structural and lithostratigraphic information from the Cuban geological map series 1 : 250.000 are summarized in an accessory figure for each section.

The crustal sections are a basic requirement for conducting an evolutionary interpretation in three dimensions. Understanding the fundamentals of current geodynamic concepts of the Caribbean Plate is a further one. Chapter 6 discusses these fundamentals and defines the basis for the detailed Late Cretaceous to Miocene tectonic reconstructions of the northwestern Caribbean which are proposed in chapter 7.

With regard to plausibility, significant emphasis was put on the degree and quality of visualization. All essential data and interpretations were tried to be expressed in images.

Digital elevation models (DEMs) permit various visualisations of topographic and bathymetric features and allow for a vivid approach to a study area or region. Three easily accessible datasets of global coverage were used for that purpose: Topographic information was taken from SRTM-3 data (spatial resolution: 3 arc seconds, ~90 m; Farr & Kobrick 2000; Rosen et al. 2000; <ftp://e0mss21u.ecs.nasa.gov/srtm>) and from GTOPO30 data respectively (spatial resolution: 30 arc seconds, ~ 925 m; <ftp://edcftp.cr.usgs.gov>). Bathymetric features were visualized from the ETOPO2 "Smith/Sandwell" data base with a horizontal grid spacing of 2 arc minutes (Smith & Sandwell 1997). A DEM with increased detail of bathymetric information within the coastal regions of Cuba was computed from a combination of ETOPO2 with isobaths and sounding data from the Cuban geological map series 1 : 250.000. All DEM datasets used for this thesis are included on a supplementary data carrier.

3 Geological overview of the Caribbean region

The Caribbean region displays a variety of plate boundary interactions including subduction in Central America and the Lesser Antilles, strike-slip motions on the northern and southern boundaries, and sea floor spreading in the Cayman Trough. The crustal types in the Caribbean basically involve (1) pre-Mesozoic continental blocks, (2) accretionary crust formed during the Mesozoic and Cenozoic, (3) Jurassic - Early Cretaceous oceanic crust, and (4) a mid-Cretaceous oceanic plateau comprising major parts of the central Caribbean.

The purpose of this chapter is to provide a framework for the focus on Cuban geology and the geodynamic conclusions drawn for the northwestern portion of the Caribbean.

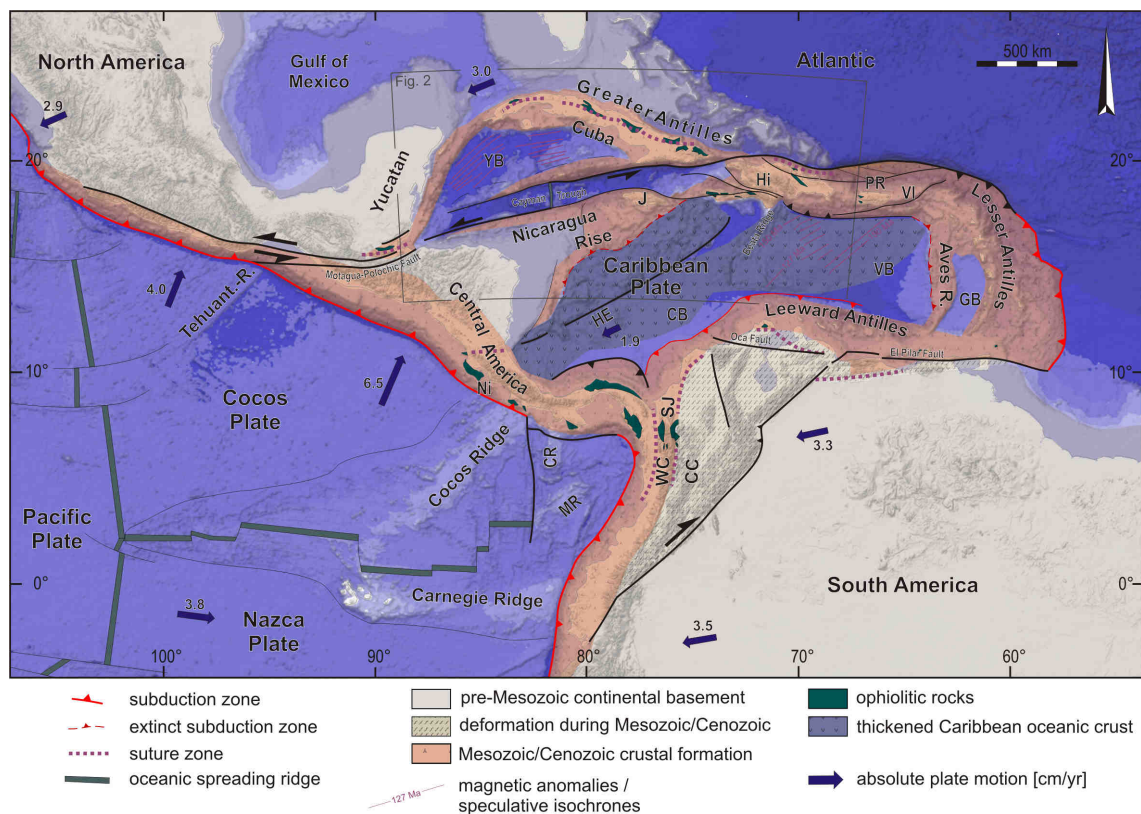


Fig. 1: Tectonic overview of the Caribbean Plate and its neighboring regions (modified from Meschede & Frisch 1998; Pindell & Barrett 1990). Bathymetric data was taken from ETOPO2 (Smith & Sandwell 1997), Topographic data taken from GTOPO30. CB = Colombian Basin; CC = Central Cordillera; CR = Coiba Ridge; GB = Grenada Basin; HE = Hess Escarpment; Hi = Hispaniola; J = Jamaica; Malpelo Ridge; Ni = Nicoya; PR = Puerto Rico; San Jacinto Belt; VB = Venezuelan Basin; VI = Virgin Islands; WC = Western Cordillera; YB = Yucatán Basin.

3.1 Active margins of the Caribbean Plate

The northern Caribbean Plate boundary consists of a zone of left-lateral strike-slip deformation extending from the Motagua-Polochic Fault System, the Cayman Trough (chapter 3.3.3) into a multibranch fault system along Hispaniola (chapter 3.3.2), Puerto Rico and the Virgin Islands (Fig. 1). Correspondingly, the Caribbean Plate is bounded by right-lateral strike-slip fault zones towards South America, namely by the Oca and El Pilar Fault Zones (Fig. 1). The North and South American plates are moving to the west with respect to the hotspot reference frame at current velocities of 3.05 cm yr^{-1} and 3.30 cm yr^{-1} . The Caribbean Plate has a smaller motion vector of 1.88 cm yr^{-1} in a western direction (DeMets et al. 1990). Therefore, relative eastward displacement of the Caribbean with respect to the Americas sums up to about $1.2 - 1.5 \text{ cm yr}^{-1}$ (DeMets et al. 1990), which corresponds to the current spreading rate in the Cayman Trough (Rosencrantz et al. 1988). Presuming similar plate motion vectors throughout the Cenozoic, these movements result in 800-1000 km of relative lateral displacement of the Caribbean Plate with respect to North America. This is in rough agreement with most of the estimated amounts of sinistral displacement along the northern Caribbean Plate boundary (e.g. Mann & Burke 1984; Rosencrantz & Sclater 1986; Rosencrantz et al. 1988; Mann et al. 1990; Pindell & Barrett 1990; Meschede 1998).

To the west and to the east, the Caribbean Plate is bounded by active subduction zones. Atlantic Ocean crust is subducted beneath the eastern margin of the Caribbean, producing the Lesser Antilles oceanic island arc system. The presently active volcanic centers are only dating back to early Miocene (Donnelly 1989). Early Cretaceous to Paleogene volcanic arc development of the Greater Antilles and examples of Cretaceous igneous activity from the Lesser Antilles (Donnelly 1989) are proof for a complex but long lived active subduction history on the eastern margin of the Caribbean.

At the southwestern margin of the Caribbean, eastward motions of the Pacific and Cocos Plates result in subduction of these plates beneath Costa Rica - Panamá - Arc. A primitive antecessor of this arc already existed in the Albian, as inferred from volcanoclastic sediments from the Nicoya Complex in Costa Rica and from drillings in Nicaragua (Calvo & Bolz 1994; Meschede & Frisch 1998).

3.2 Caribbean crust and the structure of the plate interior

The main part of the Caribbean Plate is essentially oceanic, whereas vast areas have an increased thickness of 10-15 km (Fig. 2) and a smoother upper reflector (horizon B'') than typical oceanic crust (e.g. Burke et al. 1978; Diebold et al. 1981; Holcombe et al. 1990; Donnelly 1994; Mauffret & Leroy 1997). From the geophysical character and DSDP results it was concluded that these areas represent a large flood basalt province (Fig. 3). Extensive basaltic magmatism was apparently active at $\sim 90 \text{ Ma}$ (Sinton et al. 1998) although some biostratigraphic indications and some K-Ar data yield a range of

magmatic activity from as early as Aptian-Albian to as young as Santonian, and even early Campanian (Donnelly et al. 1990a; Donnelly 1994; Weidmann 1978; Diebold et al. 1999).

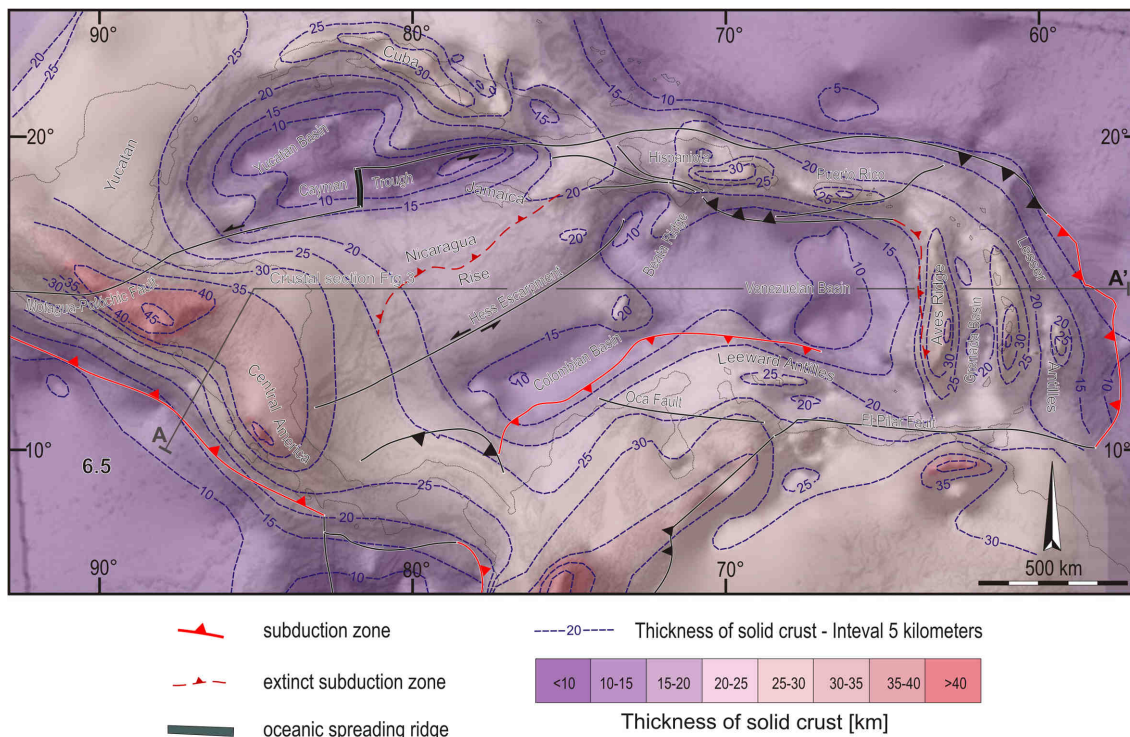


Fig. 2: DEM representation of the thickness of solid crust in the Caribbean region (contours and further punctual values used for processing taken from Case & MacDonald 1990). The central Caribbean consists mainly of anomalously thick oceanic crust (lower Nicaraguan Rise, Colombian Basin, Venezuelan Basin; Fig. 3). Only the Cayman Trough and major parts of the Yucatán Basin have crustal thicknesses typical of oceanic crust. Thicknesses of the continental provinces (compare Fig.1) vary from 20 to 45 km.

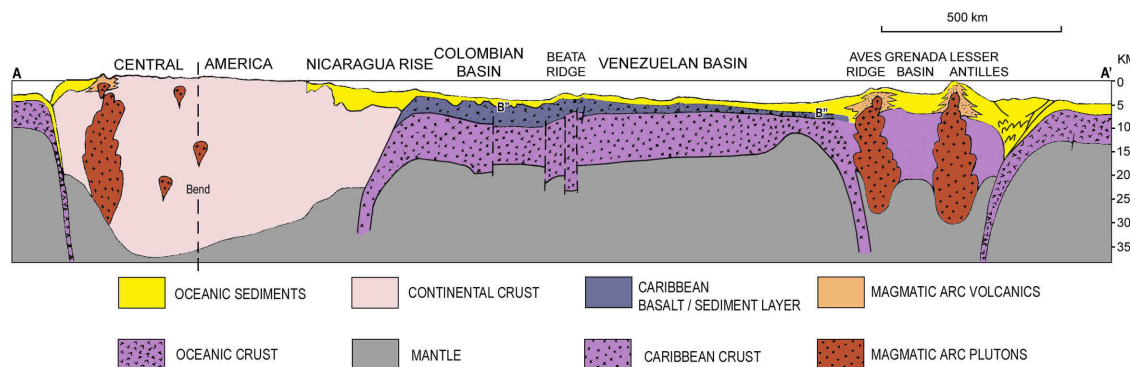


Fig. 3: Schematic east-west section across the Caribbean Plate (modified from Case & MacDonald 1990 in Donnelly 1994). The Caribbean basalt / sediment layer is the inferred Cretaceous flood basalt forming the upper part of Caribbean crust; the sediment portion of this layer is conjectured. The position of the section line is plotted in Fig. 2.

Several ridges subdivide the interior of the Caribbean Plate into different basins (Fig. 1). These features of the plate interior are: The Nicaragua Rise, the Colombian Basin, the Beata Ridge, the Venezuelan Basin, the Aves Ridge and the Grenada Basin.

3.2.1 Venezuelan Basin

The Venezuelan Basin in the eastern Caribbean is the deepest and largest of the Caribbean basins. Gosh et al. (1984) fitted a NE-SW-trending pattern of magnetic anomalies found in the Venezuelan Basin (Fig. 1, 4) to a seafloor spreading sequence of Jurassic to early Cretaceous age. The pattern of the magnetic anomalies in the Venezuelan Basin also parallels one set of a conjugate fault system and may also reflect a dyke swarm following this fault set (Donnelly 1994).

The greatest depths in the Venezuelan Basin occur at its northern and southern margins where it approaches the island arc and continental margins, respectively. These boundaries show evidence of young subduction beneath the adjacent plates (Ladd et al. 1990).

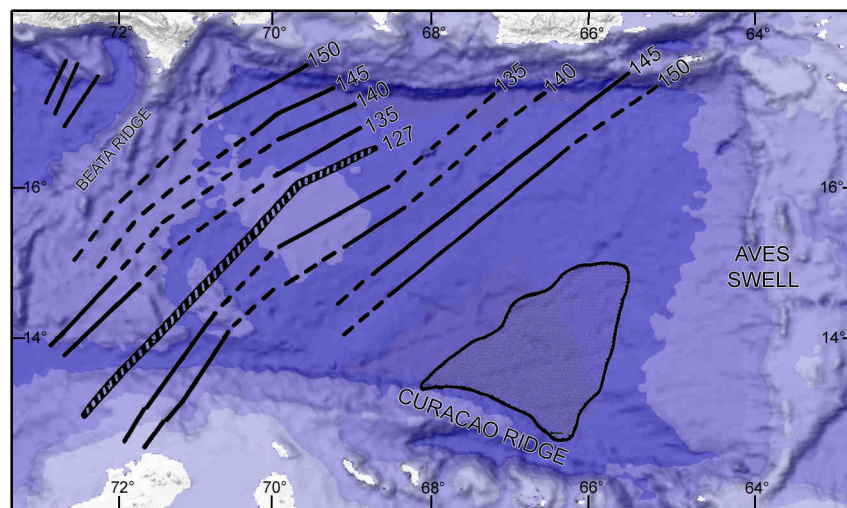


Fig. 4: Map showing correlation of magnetic anomalies and proposed isochrones, in Ma, of crustal Formation in the Venezuelan Basin (redrawn from Gosh et al. 1984). Extend of the Cretaceous flood basalt province is inferred from the rough-smooth horizon B" boundary (Diebold et al. 1981; Holcombe et al. 1990; compare Fig. 5). Bathymetric data was taken from ETOPO2 (Smith & Sandwell 1997). The interval of the blue shading is 1 km.

Geophysical data divides the Venezuelan Basin into a northwestern section where horizon B" is a characteristically smooth surface, and a southeastern section where horizon B" is an irregular surface similar to the top of igneous oceanic crust in the Atlantic Ocean (Biju-Duval et al. 1978; Diebold et al. 1981). Seismic velocity results by Diebold et al. (1981) indicate that smooth B" to the NW is underlain by normal oceanic

crust (Fig. 3) at depths corresponding more or less to those of rough horizon B" to the SE. Hence, the extent of the inferred Cretaceous flood basalt province in the Venezuelan Basin may correspond to that geophysical boundary (Holcombe et al. 1990, Driscoll & Diebold 1999; Fig. 4, 5).

3.2.2 Aves Ridge and Grenada Basin

On its eastern margin, the Venezuelan Basin is bounded by the linear N-S-trending western flank of the Aves Ridge. The Aves Ridge separates the Grenada Basin from the Venezuelan Basin (Fig. 1). From its geophysical characteristics and from rock samples, Bouysse (1988) and Holcombe et al. (1990) conclude that the Aves Ridge is primarily a Late Cretaceous to Paleogene island arc. Donnelly (1989) suggests that Venezuelan Basin crust had been subducted beneath the Aves Ridge. Most authors however, interpret the Aves Ridge as an extinct east-facing arc (Holcombe et al. 1990; Pindell & Barrett 1990).

Likewise, the origin of the Grenada Basin is also undefined. Earlier models suggest that the Grenada Basin represents fore-arc crust which was isolated by an eastward jump of the subduction zone at the beginning of the Eocene (Kearey 1974; Bouysse & Martin 1979). According to younger models, the Grenada Basin formed by back-arc spreading, which split the Aves Ridge from the Lesser Antilles (Bouysse 1988; Pindell & Barrett 1990; Holcombe et al. 1990; Bird et al. 1999).

All existing theories concerning the origins of the Aves Ridge and the Grenada Basin encounter certain contradictions (Holcombe et al. 1990; James 2006). The subduction-jump theories, however, encounter most serious contradictions, last but not least because Late Cretaceous and Paleocene arc activity is accounted for the Aves Ridge as well as the southern branch of the Lesser Antilles (Santamaría & Schubert 1974). This may also be relevant to the east-facing Aves Ridge hypotheses (chapter 7.3).

3.2.3 Beata Ridge

The Beata Ridge is a structural high accentuated by horst and graben structures extending to the south of Hispaniola (Fig. 1). A normal faulted escarpment faces the Colombian Basin to the northwest. The ridge shallows gradually to its faulted eastern boundary towards the Venezuelan Basin.

Interpretations of seismic reflection data are consistent in the conclusion that the Beata Ridge was formed as a morphological feature after horizon B" time, that is after the emplacement of the middle-/ Late Cretaceous flood basalt complex (Fig. 5). Holcombe et al. (1990) conclude that the Beata Ridge evolved episodically or continually over an extended period of Late Cretaceous - Early Cenozoic time, while Driscoll & Diebold (1999) infer that regional uplift of the Beata Ridge was caused by collapse of the footwall block soon after magmatic emplacement of the flood basalts.

Samples recovered from the ridge range in age between 80 and 75 Ma and are very similar to the inferred flood basalts from other parts of the Caribbean. Some of the

^{40}Ar - ^{39}Ar datings, however, yielded surprisingly young ages of around 55 Ma (Révillon et al. 2000).

According to James (2004, 2006) the Beata Ridge represents a Late Jurassic - Early Cretaceous spreading axis related to the Pangean rift in Middle America.

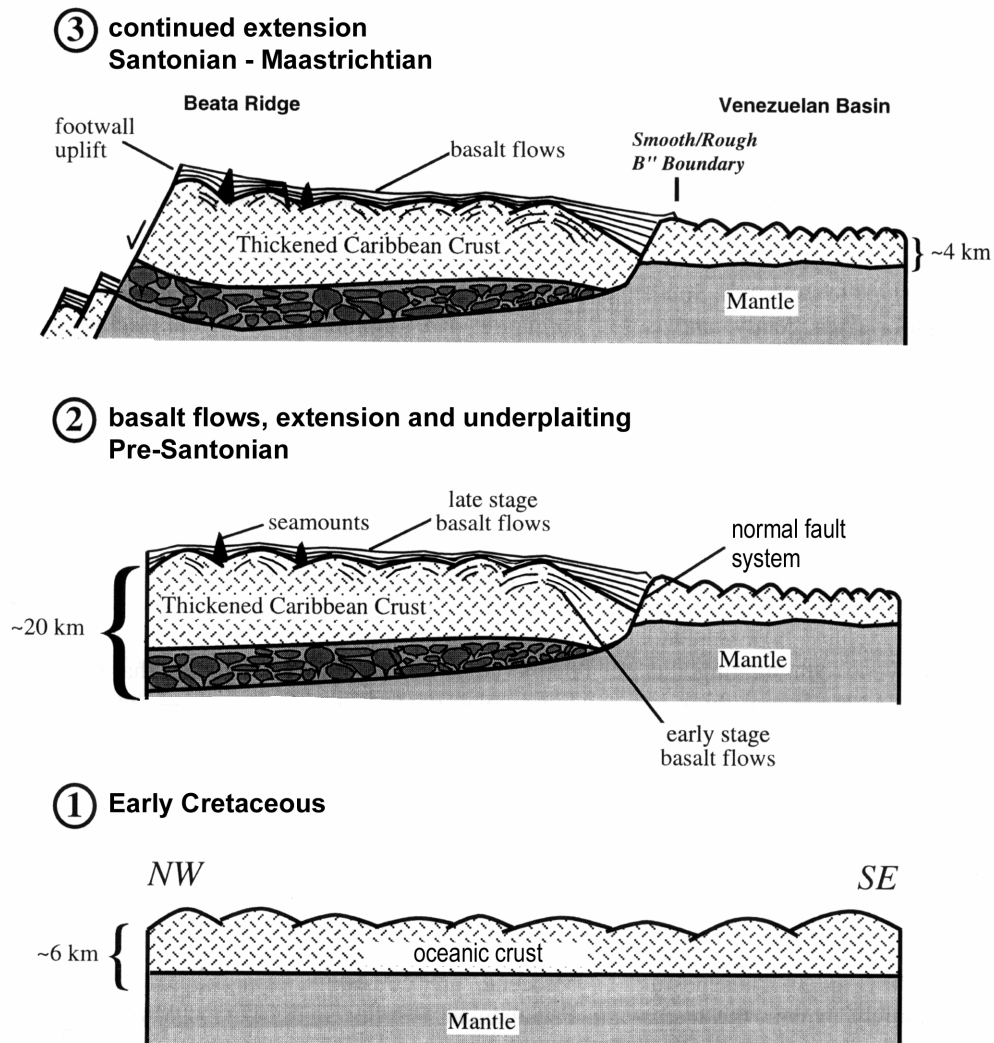


Fig. 5: Schematic illustration of the development of the Caribbean crust (Driscoll & Diebold 1999). (1) Oceanic crust formed by seafloor spreading in Late Jurassic - Early Cretaceous time. (2) Eruptions of basaltic flows are attributed to at least two episodes of magmatic activity. The early stage was more localized with steeper dipping reflectors; the late stage is regionally more extensive. The southeastern limit of the basalt province in the Venezuelan Basin is inferred from the rough-smooth horizon B'' boundary and its coincidence with: (a) an abrupt rise in Moho level, (b) a normal fault system. (3) Formation of the Beata Ridge as a morphological feature is attributed to extensional tectonics in the Late Cretaceous.

3.2.4 Colombian Basin

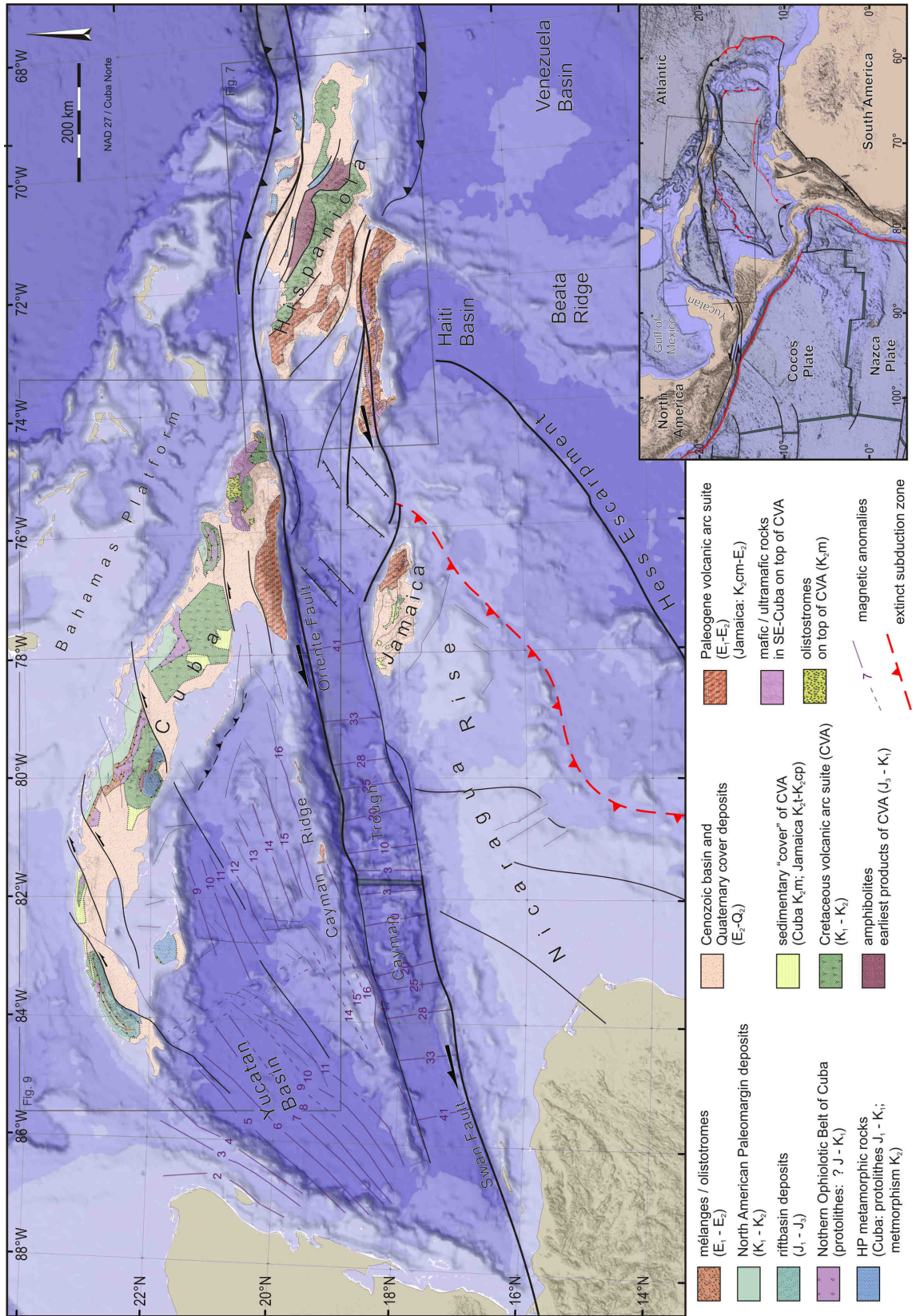
The Colombian Basin extends to the southwest between the sharp western escarpment of the Beata Ridge and a deformed belt on the Caribbean side of Panamá and Costa Rica. On the northwest it is separated from the lower Nicaragua Rise by the Hess Escarpment (Fig.1). Large submarine fans dominate the sea-bottom morphology of the eastern half of the basin.

Seismic refraction results show that the entire crustal section is thicker in the Colombian Basin than in the Venezuelan Basin, but instead of a very smooth horizon-B" reflector, the top of the crust reveals regional ridges and basins (Ewing et al. 1960). However, at least the western Colombian Basin seems to be underlain by a large oceanic plateau related to Late Cretaceous Caribbean basalt flows (Fig. 3, Bowland & Rosencrantz 1988).

3.3 Northwestern Caribbean

The northwestern Caribbean region encompasses a suture zone marking the collision between the Cretaceous Greater Antilles volcanic arc (CVA) and the southern edge of North America (Yucatán and Bahamas). Subduction in this portion of the CVA dates back at least to the Aptian and ceased at about 70 Ma in the Campanian (Stanek et al. 2006). Folding and thrusting of the Bahamas carbonate sequences and the ages of associated olistostromes constrain the timing of the final collision as Paleogene, no younger than early Late Eocene (Iturralde-Vinent 1996a). Southeastern Cuba and south-central Hispaniola enclose remnants of renewed Paleogene volcanic arc activity (e.g. Gyarmati 1983; Cobiella et al. 1984; Draper et al. 1994). On Jamaica, Cretaceous arc activity spans the Barremian to Turonian and a renewed episode lasted from the Campanian to the Early Eocene (Lewis & Draper 1990; Mitchell 2003).

Fig. 6: (Next page) Tectonic overview of the northwestern Caribbean. Geological units are generalized from Puscharovskiy et al. (1988), Lewis & Draper (1990). On Hispaniola and Jamaica, the pattern of the Paleogene magmatic arc suite includes presumed back-arc and fore-arc deposits. In southeastern Cuba the same pattern is restricted to Paleogene – Early Eocene units of direct arc origin; on Jamaica the same pattern includes Campanian to Early Eocene volcanic arc units. The numbered magnetic anomaly sequence of the Yucatán Basin (Hall & Yeung 1982) does not imply time sequence or correlation with numbered magnetic lineations recognized worldwide. The numbers of the proposed magnetic isochrones in the Cayman Trough (Rosencrantz et al. 1988 in Müller et al. 1999) reflect approximate ages given in millions of years (Cande & Kent 1995). The main fault system of the northern Caribbean Plate boundary is reproduced from Calais et al. (1998). Bathymetric data is taken from ETOPO2.



At least since the Oligocene, a left-lateral transform system dissects the northwestern Caribbean. The eastern part of the Greater Antilles arc was displaced eastward together with the present day Caribbean Plate, leaving behind the oceanic Yucatán Basin as a fault-bounded triangular tectonic unit (Pindell et al. 2005).

The following section gives an overview on the primary elements of the northwestern Caribbean (Fig. 6), except of the island of Cuba, which is outlined in chapter 4.

3.3.1 Nicaragua Rise, Chortís Block and Jamaica

The Nicaraguan Rise extends northeastwards from Honduras and Nicaragua to Jamaica. To its northern edge it is bounded by the Cayman Trough and along its southern margin by the northeast-southwest trending Hess Escarpment (Fig.1).

The shallow northwestern portion of the Nicaragua Rise appears to be the offshore extension of the continental structure of the Chortís Block, which has a presumed Precambrian / Paleozoic basement. The overlying sedimentary units consist of Mesozoic shallow marine and terrigenous deposits. Volcanic rocks are very limited and only some Cretaceous and Tertiary granite plutons occur. Cenozoic deposits, however, consist mostly of volcanic rocks with some intercalated volcanoclastic rocks and red beds (Gordon 1990).

From west to east the crust of the upper Nicaraguan Rise thins and becomes more magmatic arc in character (Arden 1975), as exposed on Jamaica (see below). The boundary zones between the plausibly continental and the Cretaceous - Paleogene volcanic arc portions of the Nicaragua Rise are enigmatic.

The deeper southern portion of the Nicaraguan Rise is an elevated extension of the Colombian Basin. The continuation of the horizon B" northwest of the Hess Escarpment was demonstrated at DSDP Site 152 (Donnelly 1994). The boundary between the oceanic and the continental / magmatic arc portion of the rise probably has the nature of an extinct subduction zone (Fig.1, 2; chapter 7.3).

Jamaica

The easterly uplifted tip of the Nicaraguan Rise is emerged on the island of Jamaica, which in its eastern part rises up to 2,225 m in altitude. Most of the island displays a plateau of Late Eocene - Miocene limestone. Northwest-trending graben structures separate main structural blocks. Erosional inliers in each block expose different Cretaceous and early Paleogene rocks.

The oldest rocks in western and central Jamaica consist of a Barremian - Turonian volcanic arc succession ("older volcanic sequence", Devils Race Course Formation, Athurs Seat Formation, Mitchell 2003, 2006). Tholeiitic lavas in the lower part of the section are succeeded by a calc-alkaline suite in the upper Albian to Turonian section (Jackson 1987). Minor occurrences of amphibolites and high-pressure, low to medium temperature schists are believed to have formed in an Early Cretaceous subduction

zone (chapter 6.5.2, chapter 7.2). K-Ar mineral-ages from similar rocks in southeastern Jamaica yielded Campanian and Early Eocene ages, but these are ascribed to thermal resetting (Lewis et al. 1973; Draper 1986; Lewis & Draper 1990; Mitchell 2003).

The older volcanic arc sequence is overlain with an angular unconformity by the late Turonian - Campanian Crofts Synthem which mainly consists of deep-water sediments (mudstones, turbiditic sandstones, and shales; Mitchell 2003, 2006). A regional shallowing event in the Middle Campanian is followed by rapid subsidence in the Late Campanian (Mitchell 2003).

The Croft Synthem is overlain by volcanoclastics and limestones of the Maastrichtian - Early Eocene Kellits Synthem with a well-constrained unconformity at its base. The top of the synthem is represented by dacitic ignimbrites. Renewed magmatic activity (chapter 7.3) is also evident with the intrusion of granodiorites in the eastern part of central Jamaica (Above Rocks granodiorite, 63 ± 5 , Cubb & Burke 1963).

The Blue Mountain Block of eastern Jamaica already records renewed andesitic volcanic activity from the Campanian onward. Granitoid intrusive rocks are evidently of latest Cretaceous age (Lewis & Draper 1990), the oldest volcanics apparently date back to the early Campanian (Wadge et al. 1982; chapter 7.3). The Paleocene to Early Eocene Wagwater Group, more or less contemporaneous to the upper part of the Kellits Synthem in central Jamaica, comprises volcanic rocks throughout the entire section. This volcanic activity ceased by early Middle Eocene (Lewis & Draper 1990; chapter 7.3) followed by the transition to island wide limestone sequences.

Outcrops of an ophiolitic suite occur in the Bath-Dunrobin Complex on the southern flank of the Blue Mountains. Cherts assumingly interbedded within the pillowed basalts indicate a Turonian-Coniacian age (Montgomery & Pessagno 1999). Deep-water mudstones and cherty limestones overlying the ophiolites were deposited during the Campanian to early Maastrichtian. Similarities to the occurrences of mafic rocks in southeastern Cuba (Sierra de Nipe-Cristal and Moa-Baracoa Massifs) and the southern peninsula of Hispaniola are accentuated by massive polymictic conglomerates of Campanian-Maastrichtian age, which overlay the ophiolitic suite (Krijnen & Lee Chin 1978; chapter 7.3).

3.3.2 Hispaniola

Hispaniola is situated in a crosspoint where the left-lateral northern Caribbean Plate boundary, i.e. the boundary faults of the Cayman Trough, converges with the main axis of the Greater Antilles and splits into a multibranching fault system (Fig. 6). The island exhibits a magmatic arc structure (CVA) that evolved from at least early Cretaceous time. Strike-slip faulting has begun by the mid-Tertiary, resulting in an E-W to NW-SE trending system of lineaments, which define several geological provinces. Each fault bounded block displays a distinct geological history and some accumulated thick basin sequences which cover the basement units. Following Mann et al. (1991), the geological Provinces of Hispaniola are generally referred to as tectonostratigraphic

terrane. However, it is obvious that the basement blocks of Hispaniola are somehow compatible in a common geological history of the northern Caribbean Plate boundary, even if some intervening intermediate facies are lacking due to strike-slip tectonics.

The geological provinces along the northeastern coast are essentially of a fore-arc tectonic origin. Basement rocks in the Cordillera Septentrional and of the Samana peninsula (Fig. 7) consist of heterogeneous assemblages of igneous and metamorphic rocks, such as blueschist - eclogite mélanges, gabbroic intrusives and serpentinites, greenschist-blueschist facies rocks, marbles and gneisses (Samana metamorphic terrane and Rio San Juan / Puerto Plata / Pedro Garcia disrupted terrane). A late Cretaceous age of high-pressure metamorphism has been suggested for an eclogite from the Samana peninsula (Joyce 1991) and Bowin & Nagle (1982) reported a Maastrichtian age for a basaltic flow from the Cordillera Septentrional. Sedimentary rocks following the cessation of subduction developed at least from the Late Eocene onward (Draper et al. 1994).

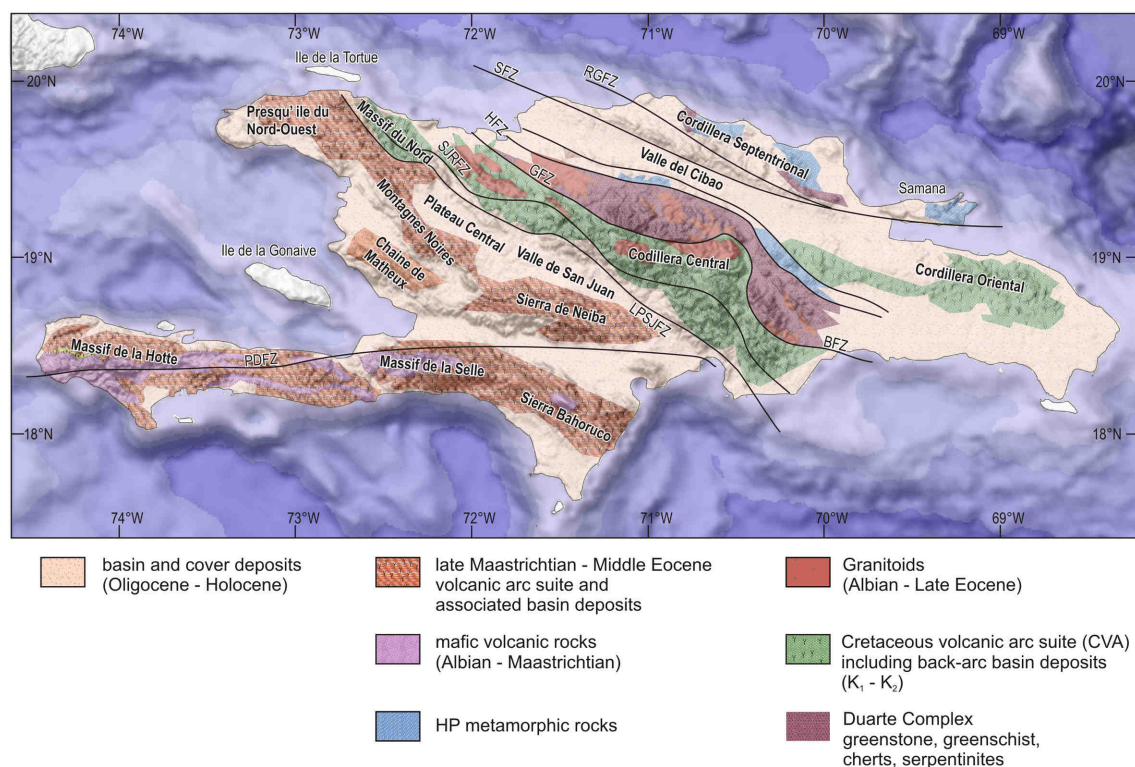


Fig. 7: Simplified geological map of Hispaniola and names of physiographic provinces (compiled from Lewis & Draper 1990; Draper et al. 1994). Bathymetric data is taken from ETOPO2. BFZ = Bonafo fault zone; GFZ = La Guacera fault zone; HFZ = Hispaniola fault zone; LPSJFZ = Los Pozos - San Juan fault zone; PDFZ = Presqu'île del Sud fault zone; RGFZ = Rio Grande fault zone; SJRFZ = San José - Restauración fault zone.

Inbound of the northeastern coastal area the presence of a magmatic arc basement is apparent from outcrops in the Cordillera Oriental (Fig. 7). The El Seibo unit and the Los Ranchos Formation mainly consist of about 1 km of bimodal low-grade metavolcanic rocks, which have characteristics of a primitive island arc series (Donnelly et al. 1990a,

Kesler et al. 1991). U-Pb ages for zircons from the Los Ranchos Formation show, that it was emplaced during the Aptian-Albian transition (Kesler et al. 2005). This series is (presumably truncated and unconformably) overlain by carbonates containing a rich Aptian-Albian fauna (Hatillo Limestone, Bourdon 1985 in Draper et al. 1994), which in turn are overlain by a further thick volcano-sedimentary sequence probably reaching as long as early Maastrichtian. A trachyte sample has been dated at 65.7 ± 3.3 Ma (Lewis & Draper 1990) and a granodiorite intrusion yielded a Santonian age (Draper et al. 1994).

The Duarte Complex along the northern flank of the Cordillera Central exposes regionally metamorphosed, massive and locally volcanic rocks of mafic to ultramafic composition. Geochemical analyses suggest that the Duarte rocks have similar signatures to those of Pacific seamounts. They may represent the very earliest volcanic products of the Hispaniola island arc (Lewis & Draper 1990). Several granitoid batholiths and stocks intrude the Duarte Complex. These date between the Albian and late Eocene (Draper et al. 1994).

The Hispaniola Fault Zone constitutes the high angle northern bounding fault of the Duarte Complex. To the north of the Hispaniola Fault Zone the Duarte Complex is flanked by a narrow belt of northeast-verging foliated meta-volcanic and meta-sedimentary rocks (Tortue-Amina-Maimon metamorphic terrane).

Across the southern bounding fault, the Duarte Complex is flanked by broad belt of Late Cretaceous volcanic and clastic rocks extending throughout the central mountain chain of Hispaniola to the Massif du Nord. The rocks of this belt are attributed to a formation above the magmatic arc (Tireo stratigraphic terrane). Several unfoliated granitoids intruded up to the Late Eocene (Draper et al. 1994). Corresponding back-arc deposits parallel the volcanic series in a further fault bounded zone to the south (Trois Rivières-Peralta stratigraphic terrane), which consists of Coniacian - early Late Eocene turbiditic sandstones, siltstones, limestone, chert and tuffs (Draper et al. 1994).

The basement character of the zone (or terrane) to the southeast of the Trois Rivières-Peralta terrane is ill-defined. Large areas are covered by Oligocene - Pliocene basin deposits. Some Late Cretaceous intermediate volcanic rocks intruded by small Late Cretaceous to Early Eocene plutons are reported from the Presqu'île du Nord-Ouest (Kesler 1971, Cheillett et al. 1978; see Draper et al. 1994) and some Late Cretaceous pelagic limestone occur in the Montagnes Noires (Draper et al. 1994). Furthermore, most of the Montagnes Noires, Chaîne de Matheux and the Sierra de Neiba consist of Eocene limestones and argillites intercalated with andesitic breccias, tuffs and rare basalts (Perodin Formation, Butterlin 1957, 1960 in Lewis & Draper 1990). The Perodin Formation also contains andesite dikes and sills, dacites, and quartz-diorites; hence Butterlin (1960) compared it with the El Cobre Formation in the Sierra Maestra of Cuba but its general description also exhibit similarities to the back-arc character of the Bayamo – San Luis basin successions (chapter 7.3).

The three major mountain ranges on the southern peninsula of Hispaniola, the Massif de la Hotte, Massif de la Selle, and the Sierra de Bahoruco, expose thick tholeiitic sequences with oceanic characteristics. Age data from this basaltic sequence in the Massif de la Selle (Dumisseau Formation, Maurrasse et al. 1979) range from Albian to Maastrichtian (Sayeed et al. 1978; Bellon et al. 1985). The mafic volcanics are associated with tightly folded late Coniacian to Campanian - Maastrichtian pelagic limestones and cherts (Macaya Formation, Calmus & Vila 1988). Both units are unconformably overlain by heterogeneous detrital formations containing Danian olistostromes (Calmus & Vila 1988). Scattered outcrops of alkalic volcanic rocks are interbedded in the overlying Paleocene to Middle Eocene section of the Massif de la Hotte (Calmus 1987; see Draper et al. 1994; Calmus & Vila 1988).

Maurrasse et al. (1979) suggested that the mafic volcanics of the southern peninsula of Hispaniola represent an uplifted piece of the thickened Caribbean oceanic crust. Calmus & Vila (1988, p. 65) ascribe the Danian olistostromes to "a phase of northward motion of the Macaya unit upon the lower plate, which consist of the volcanic unit and the Riviere Glace Formation." Nevertheless, the occurrence of oceanic massifs associated with Maastrichtian - Danian olistostromes and conglomerates resemble the situations of the Bath-Dunrobin Complex in Jamaica and the Sierra de Nipe-Cristal / Moa-Baracoa Massifs in southeastern Cuba (chapter 7.3).

3.3.3 Cayman Trough

The Cayman Trough, extending ~ 1400 km eastwards from the Gulf of Honduras to the Windward Passage, represents a prominent feature of the left-lateral strike-slip zone forming the northern Caribbean Plate boundary (Fig. 1, 6). At only 120 km in width and 5 km depth, it comprises a floor of thin oceanic crust (Fig. 2) bounded by steep walls towards continental / magmatic arc crust of the Cayman Ridge to the north and the Nicaraguan Rise to the south. The north bounding Oriente and the south bounding Swan Island Transform Faults are connected by the perpendicular oriented Mid-Cayman Rise, which is the site of seafloor spreading (Holcombe et al. 1973; Edgar et al. 1990). Magnetic anomalies show lineations on both sides of the axis and partial anomaly sequences have been interpreted (Fig. 8; Rosencrantz et al. 1988; Leroy et al. 2000). From depth, heat flow, and geomagnetic studies Rosencrantz et al. (1988) suggest an overall rate of opening of about 1.5 cm yr^{-1} since 26 Ma and 3.0 cm yr^{-1} prior to 26 Ma. From these results it is further argued, that the trough opened at least since the Early Eocene (45–50 Ma, Rosencrantz et al. 1988; 49 Ma, Leroy et al. 2000).

The minimum rates given by Rosencrantz et al. (1988) summarize to at least 775 km of spreading at the Mid-Cayman Rise. Pindell & Barrett (1990) estimate a total sinistral offset of 1050 to 1100 km related to the Cayman Trough, accounting to additional extension of arc-related or continental crust at the western and eastern ends of the trough. However, estimations on the amount of offset along the Cayman Trough are controversial, because the accommodation of about 1000 km sinistral Cenozoic offset is not evident in the eastern and western prolongations of the Cayman Trough. At least

for the western prolongation along the Polochic-, the Motagua-, and the Guayape Faults, Gordon & Avé Lallement (1995) calculated a total of 725 km of distributed movement on cryptic strike-slip faults across the entire Chortís Block to account for remaining offset. Hitherto reconstructions for the alignment of eastern Cuba and Hispaniola only account for a Cenozoic displacement of ~ 200 - 300 km (e.g. Meyerhoff 1966 in Pindell & Barrett 1990, Iturralde-Vinent & MacPhee 1999). Indeed, the Oriente Transform Fault splits eastwards into a multibranch fault system. Hence, the Cenozoic displacement is split up and most of it could be accommodated south of the Cordillera Central of Hispaniola (cf. Pindell & Dewey 1982; chapter 7.5).

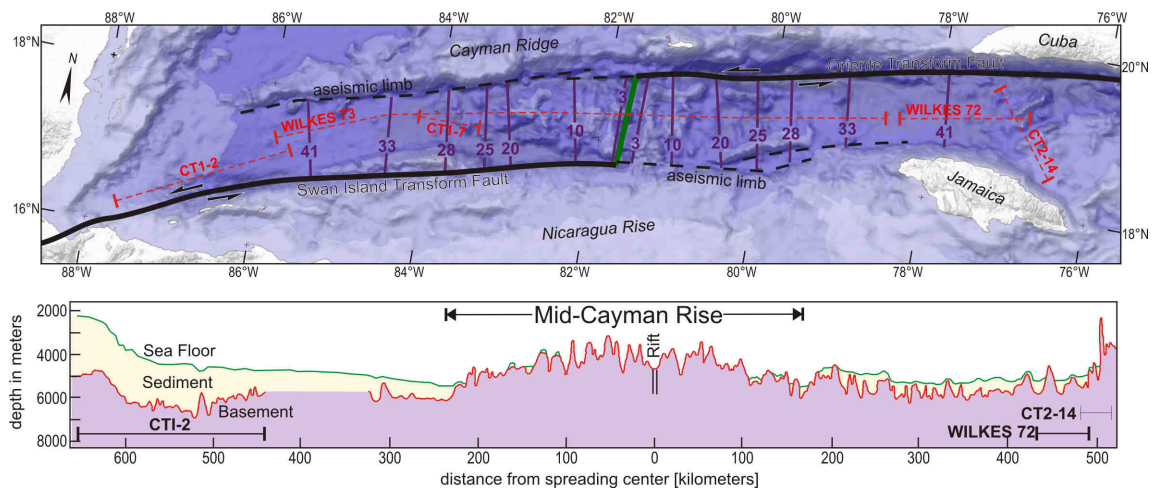


Fig. 8: Bathymetric map and seismic profile of the Cayman Trough (compiled from Rosencrantz & Sclater 1986 in Edgar et al. 1990; Rosencrantz et al. 1988 in Müller et al. 1999). Dashed red lines in the map show the locations of the seismic lines combined in the profile. Locations of proposed magnetic isochrones (Rosencrantz et al. 1988 in Müller et al. 1999) are shown as purple lines. Approximate ages of the isochrones are given in Ma (Cande & Kent 1995).

3.3.4 Yucatán Basin and Cayman Ridge

The Yucatán Basin extends between the margins of the Yucatán Platform and Cuba. It is separated into a deeper northwestern part (-4000 to -4600 m) containing the Yucatán Plain and a shallower, topographically heterogeneous southeastern part (-2000 to -3500 m). The latter merges with the Cayman Ridge on the north side of the Cayman Trough. Both parts are separated by a northeast to east-northeast trending series of lineaments, continued on Cuba as the La Trocha Fault (Fig. 6). The margin on the western side of the Yucatán Basin, against the Yucatán Platform, is steep and highly faulted and has the appearance of a transform margin. The northeastern margin dips beneath the Cuban crust along a sediment filled depression.

Crustal thickness of the basin is only slightly greater than typical for oceanic crust, whereas the southeastern part and the Cayman Ridge show a crustal thickness of up to 20 km (Fig. 2; Ewing et al. 1960; Case & MacDonald 1990).

Hall & Yeung (1982) tentatively identified a series of magnetic anomalies, trending northeast in the western basin, and east-northeast in its eastern part, separated and offset along a fracture zone (Fig. 6). According to their interpretation, the anomalies would have been generated on one side of a spreading center to the south of the present basin. A definite correlation to the geomagnetic timescale is not identified for the anomalies of the Yucatán Basin, but speculations on a Late Cretaceous age were supported by the interpretation of heat flow values (Epp et al. 1970; Erickson et al. 1972). Rosencrantz (1990) concluded that the oldest crust in the eastern Yucatán Basin is at least Late Cretaceous and could be Aptian-Albian or even Late Jurassic in age (chapter 7.4).

Geophysical studies as well as dredged and drilled samples point to a Paleogene volcanic arc origin of the Cayman Ridge. The ridge consists of up to 20-km-thick crust (Ewing et al. 1960). Volcanics and granodiorites of Maastrichtian and Paleocene ages have been dredged from the north wall of the Cayman Trough (Perfit & Heezen 1978; Lewis et al. 2005) and Sigurdsson et al. (1997) found extensive volcanoclastic deposits and rhyolitic volcanic ash layers in a lower to Middle Eocene section from the central Cayman Rise, whereas ash layers also occurred in the overlying pelagic deposits. Rosencrantz (1990, 1996) identified a distinct seismic unit thinning away from the southern crest of the ridge, presumably representing the arc volcanics. At least topographically the Cayman Ridge represents an evident continuation of the Paleocene arc remnants in the Oriente province on southeastern Cuba. However, Lewis et al. (2005) suggest a closer relationship to the northern margin of the Chortís Block, based on a distinct continental affinity of the Cayman Ridge granitoids.

4 Geological overview of Cuba

The geological record of western and central Cuba differs significantly from that of southeastern Cuba, which divides the island into two broad geological provinces.

4.1 Western and central Cuba

Western and central Cuba constitute an orogenic belt resulting from the collision of a mid- to Late Cretaceous volcanic arc (CVA) and its associated subduction-accretion complex with the Late Jurassic to Late Cretaceous sedimentary rocks of the North American Paleomargin. The collision process apparently started in the Campanian (e.g. Stanek et al. 2006). Major north- to northeast-directed thrusting processes culminated during the Paleocene and Eocene, resulting in

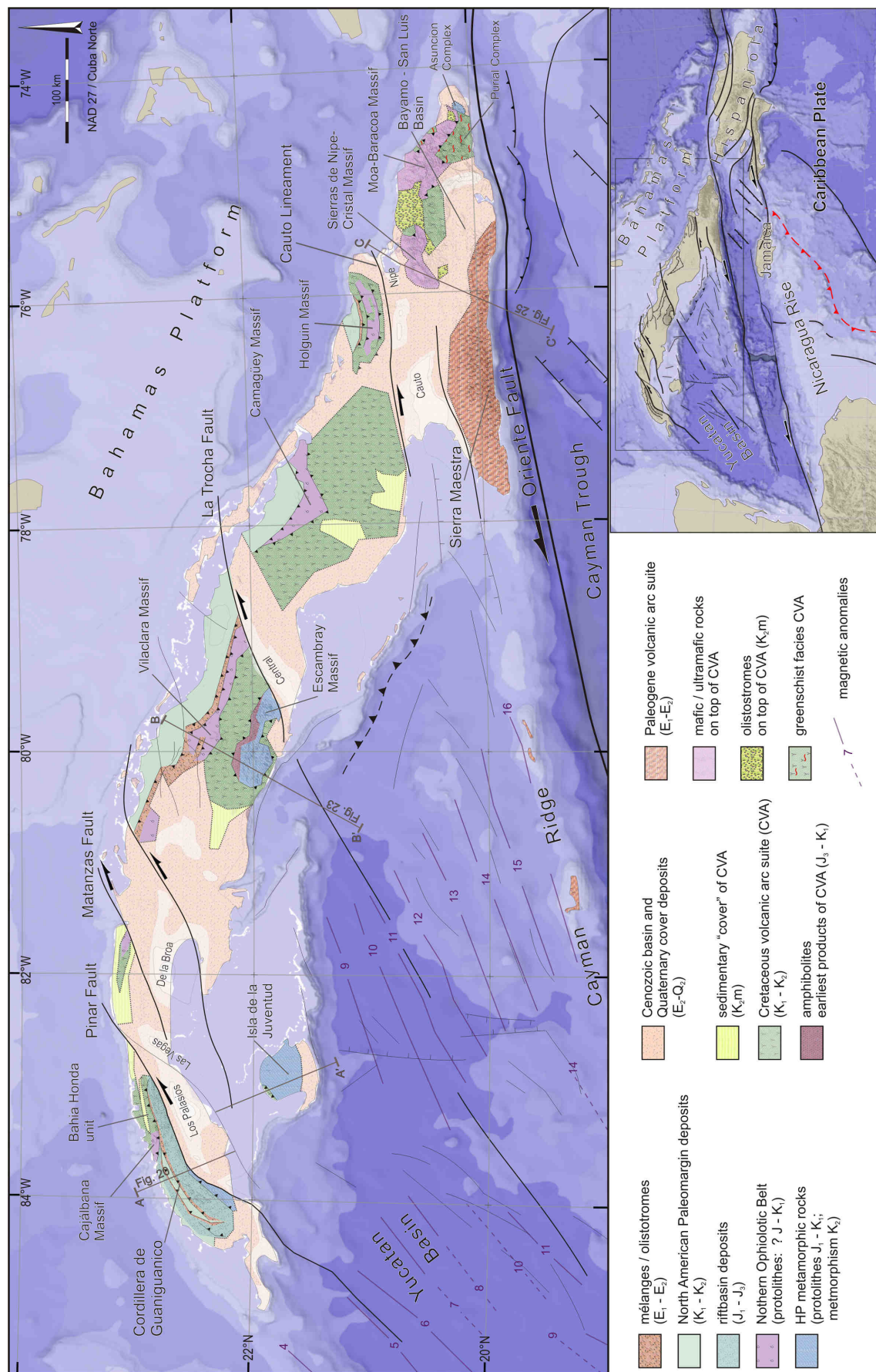
- tectonic transport of the CVA units onto the southern shoulder of the Bahamas Platform,
- formation of foreland basins on the North American Paleomargin and piggyback basins on the CVA,
- obduction and intense shearing of ophiolites (Northern Ophiolite Belt),
- uplift and exhumation of metamorphic rocks comprising the former subduction-accretion complex,
- formation of a north to northeastward-verging fold and thrust belt.

Late-tectonic molding of the thrust belt and the Paleogene development of foreland and piggyback basins were associated with the formation of northeast striking left-lateral transform faults dissecting the Cuba orogenic belt (e.g. Pinar-, Matanzas-, La Trocha-, Cauto Faults; Fig. 9).

4.1.1 Riftbasin and paleomargin deposits

Mesozoic passive continental margin deposits are exposed in a fold and thrust belt along the northern coast of western and central Cuba. The oldest sequences of Jurassic age are represented by riftbasin deposits coherent with the breakup of Pangea and the opening of the Proto-Caribbean (e.g. Pszczółkowski 1999, Stanek 2000; chapter 6.4).

Fig. 9: (Next page) Geological and tectonic overview of Cuba. Geological units are generalized from Puscharovskiy et al. (1988, 1989). The numbered magnetic anomaly sequence of the Yucatán Basin (Hall & Yeung 1982) does not imply time sequence or correlation with numbered magnetic lineations recognized worldwide. The bathymetric data is taken from ETOPO2 and Puscharovskiy et al. (1988).



The thrust nappes of the Cordillera Guaniguanico in westernmost Cuba contain the largest exposure of Jurassic rocks, whereas the passive margin sequences in central Cuba are dominated by stacked Late Jurassic to Cretaceous continental rise, slope, and platform units of the southern extension of the Bahamas Block (Pardo 1975, Hatten et al. 1988).

Cordillera Guaniguanico

The Cordillera Guaniguanico consists of several thrust nappes (Fig. 20) that are subdivided into distinct structural facies zones. The top of the nappe pile is represented by the Bahía Honda unit which contains ophiolites and remnants of the Cretaceous volcanic arc (chapter 4.1.3, chapter 5.1). The underlying nappes consist of up to 3500 m thick sediments from the continental rise east of Yucatán. From south to north the following structural facies zones are distinguished (Fig. 10): the Cangre belt along the northern side of the Pinar Fault, the Sierra de los Organos, Southern Rosario, and Northern Rosario belt.

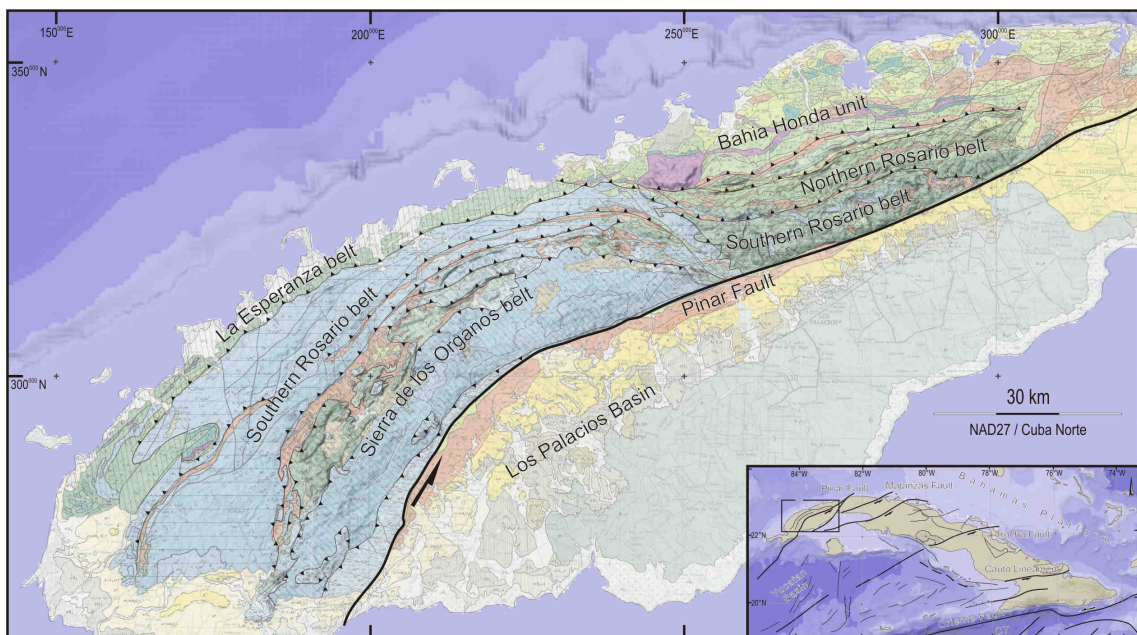


Fig. 10: Tectonic map and structural facies zones (belts) of the Corillera Guaniguanico in western Cuba (modified from Puscharovskiy et al. 1988, Pszczólkowski 1999).

The San Cayetano Formation (Appendix, I A5, I A26.) encompasses the oldest deposits in the Cordillera Guaniguanico. It comprises Early and Middle Jurassic deltaic sequences of at least 1000 m thickness. Bimodal volcanic und subvolcanic rocks are intercalated in the upper section. Geochemical characteristics of the El Sábolo Fomation basalts from the Northern Rosario belt are consistent with a formation at a continental rift margin (Kerr et al. 1999, chapter 6.4).

The Cangre Belt consists mainly of high pressure / low temperature metamorphic equivalents of the San Cayetano Formation (Millán & Somin 1976). Moreover, the

protoliths of the Escambray and Isla de la Juventud metamorphic units may correlate with these deltaic riftbasin deposits (see below, Millán & Myczynski 1978).

The San Cayetano Formation is overlain by a Late Jurassic to Cenomanian sequence of mainly deep-water limestones (Pszczółkowski 1999). The trend towards deeper-water carbonates peaks in Middle Cretaceous, when radiolarian cherts are frequent (Cobiella-Reguera 2000). Rare tuff beds occur throughout the entire Aptian to Cenomanian section in the Sierra del Rosario (Cobiella-Reguera 2000).

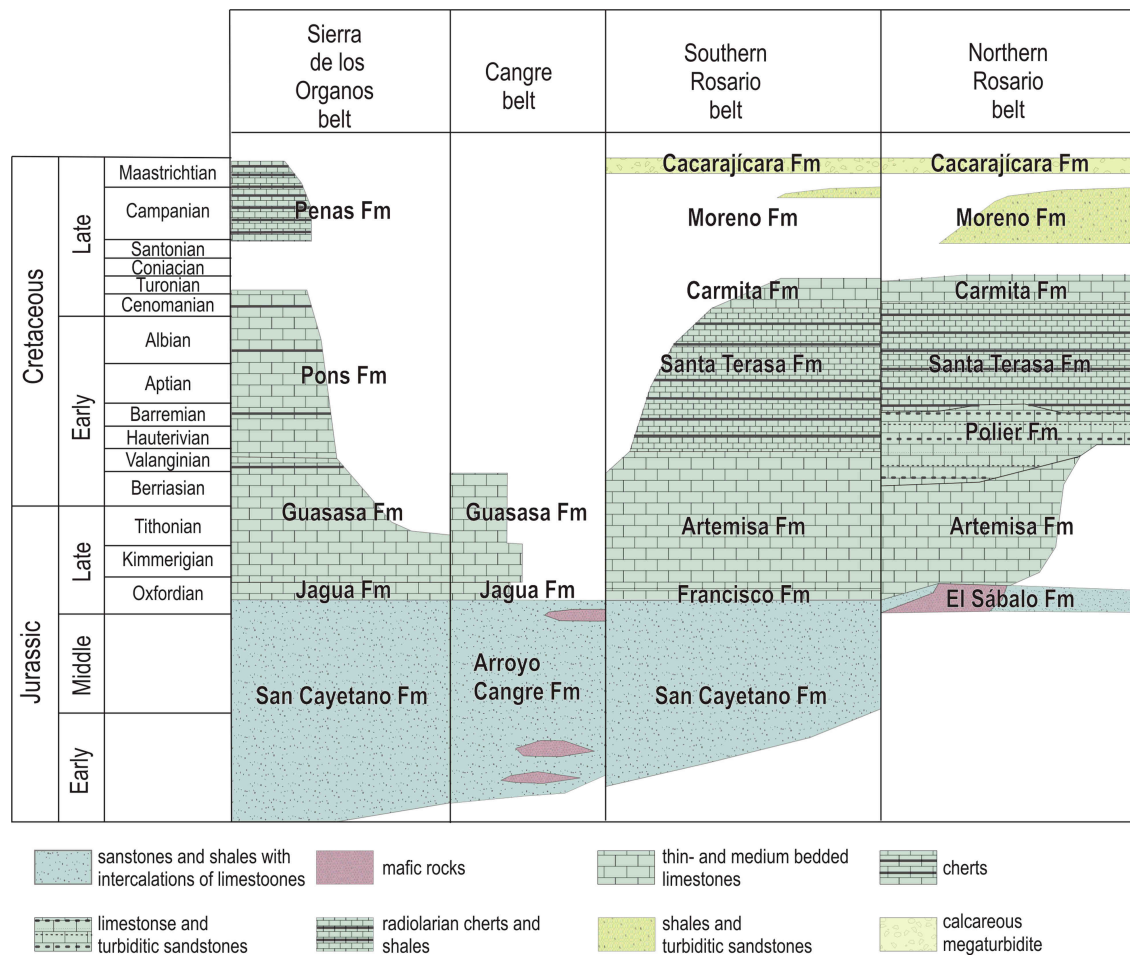


Fig. 11: Generalized lithostratigraphic scheme of the Cordillera Guaniguanico (from Pszczółkowski 1999).

The origin of a regionally extensive Turonian-Santonian hiatus in the pelagic sequence is unknown, but there is evidence for strong erosion at the base of the Cacarajícara Formation – an up to 450m thick Cretaceous - Tertiary boundary megaturbidite unit (Pszczółkowski 1986, Cobiella-Reguera 2000; Kiyokawa et al. 2002; Appendix, I A33). Pindell (1994) ascribes the hiatus to “flexural arching” and erosion in the fore-arc region of the approaching CVA.

However, there are some limited outcrops of Campanian and Maastrichtian pelagic limestones, namely the Peñas Formation in the Sierra de los Organos and the Moreno Formation in the Sierra del Rosario (Fig. 11). The latter contains abundant turbiditic influx of volcanoclastic material from the approaching CVA to the south (Pszczółkowski 1999).

Central Cuba

The stacked passive margin sequences in the thrust belt of central Cuba are subdivided into zones, reflecting different intensities of Paleogene deformation and different facies zones of the Late Jurassic to Cretaceous southern extension of the Bahamas Block (Fig. 12; Fig. 22).

The Cayo Coco and Remedios zones consist of several 1000 m of platform deposits. The oldest rocks penetrated by drill core sections are interbedded dolomites and anhydrites of the Late Jurassic Punta Alegre Formation. This formation crops out only as blocks in salt domes in Camgüey and in northwestern Matanzas (Puscharovskiy et al. 1988; Lewis & Draper 1990). The overlying Cretaceous section is continuous and dominated by shallow-water limestones, dolomites and some anhydrites. Pelagic shales and marls were deposited in the Cayo Coco zone during Late Cretaceous. Therefore, Pardo (1975) suggested that the Late Cretaceous carbonate banks might have had a bathymetry similar to the present-day Bahamas Banks, in which shallow-water platforms are separated by deeper-water tongues.

The Maastrichtian limestones are fragmental and in places coarsely conglomeratic (Pardo 1975). They are overlain by several hundred meters of marls and limestones, which represent the Paleogene foreland basin and post collisional deposits (chapter 4.1.5).

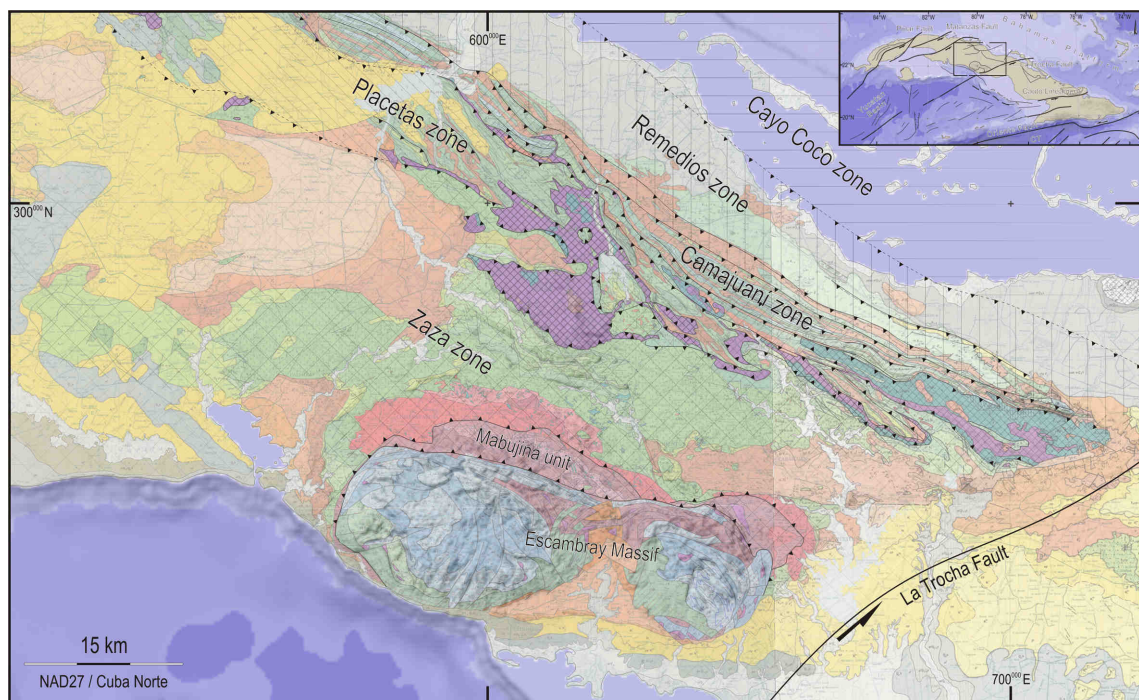


Fig. 12: Tectonic map and structural facies zones of central Cuba (modified from Puscharovskiy et al. 1988; Draper & Barros 1994).

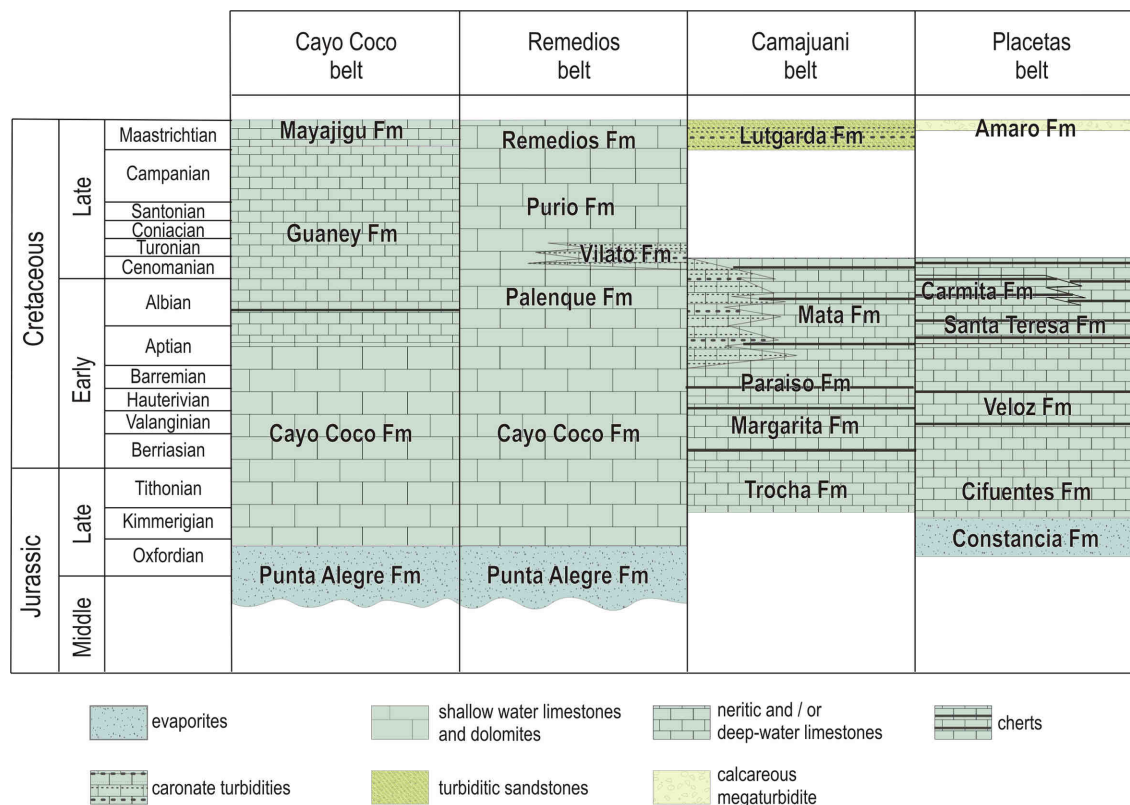


Fig.13: Generalized lithostratigraphic scheme of the Mesozoic North American Paleomargin rocks in central Cuba (from Cobiella-Reguera 2000).

The Camajuani and Placetas zones essentially represent the deep-water-facies extension to the south of the Cayo Coco and Remedios zones. The thickness of Camajuani and Remedios Mesozoic deposits is only one-fifth of the time-equivalent shallow-water carbonates of the Cayo Coco – Remedios zones and the sequences are very similar to the coeval deposits in western Cuba. The rocks of these zones span a time range of Kimmeridgian to Maastrichtian and include pelagic limestones, calcareous turbidites, radiolarian cherts and some sandstone turbidites.

As in the paleomargin units of the Cordillera Guaniguanico, a Late Cretaceous (Coniacian – Campanian) hiatus also occurs in Camajuani – Placetas zones and a Cretaceous - Tertiary boundary megaturbidite unit rests unconformably on the pelagic sequences of the Placetas zone (Amaro Formation, Pszczółkowski 1986). The Maastrichtian Lutgarda Formation of the Camajuani zone records influx of shallow-water coarse grained material from the north as well as influx of fine-grained igneous material from the approaching CVA to the south (Pardo 1975).

4.1.2 Northern Ophiolite Belt

The Northern Ophiolite Belt is an ophiolite-bearing mélangé like assemblage (Appendix, II A46, II A71). It constitutes an almost continuous body transported from the south over the North American Paleomargin (Pardo-Echarte 1996, Cobiella-

Reguera 2005). Its outcrops stretch along the northern half of almost the entire island of Cuba (Fig. 9).

Tectonized ultramafic rocks and gabbros floating in a serpentinite matrix are the most common lithologies. Tectonically embedded Mesozoic volcanic and sedimentary rocks are composed of basalts, hyaloclastites, cherts, limestones, shales and other rocks (e.g. Knipper & Cabrera 1974; Iturralde-Vinent 1990, 1994, 1996b; Cobiella-Reguera 1984, 2005). Inclusions of high-pressure subduction-derived metamorphic rocks (eclogites and blueschists) have also been found within the ophiolite mélangé (Iturralde-Vinent 1996b, Millán Trujillo 1996), which in general bears many lithological and tectonic similarities to a typical subduction accretion assemblage (Kerr et al. 1999).

Several of the sedimentary samples yielded Tithonian, Hauterivian – Barremian and Aptian – Albian fossils (Llanes Castro et al. 1998; Andó et al. 1996; Iturralde-Vinent 1990). A maximum K-Ar age of 160 ± 24 Ma was obtained from anorthosites sampled in Camgüey (Somin & Millán 1981; Iturralde-Vinent et al. 1996).

Geochemically, the basalts of the NOB are grossly characterized as oceanic tholeiites (e.g. Iturralde-Vinent 1996b). Some samples have been classified as intraplate lavas, boninites, island arc tholeiites or even back-arc basalts (Fonseca et al. 1990; Iturralde-Vinent 1996b; Kerr et al. 1999). Based on the intraplate characteristics of some samples from the Bahía Honda and the Matanzas regions, Kerr et al. (1999) consider a Pacific origin of these rocks. Their tectonic position, however, points to an origin from an oceanic realm which was located between the northwestern edge of the CVA and the Bahamas Platform, not to the south of the CVA (chapter 7.1).

4.1.3 Cretaceous volcanic arc

Cretaceous volcanic arc units occupy large parts of the Cuban basement below the Cenozoic deposits. Often referred to as “Zaza zone” or “volcanic arc terrane”, these units have an allochthonous structural position, as they are thrust from the south upon the Northern Ophiolite Belt and the North American continental paleomargin. In westernmost Cuba, outcrops of the CVA are confined to the Bahía Honda unit, which occupies the highest structural position of the Cordillera Guaniguanico nappe pile (see below, chapter 5.1).

Principle components and development

The earliest metamorphosed products of the initial stage of the CVA, and thus constraining the minimum age for the beginning of subduction, may be represented by the Mabujina amphibolites exposed along the northern edge of the Escambray Massif (chapter 4.1.4; Millán & Somin 1985; Stanek 2000). Associated gneisses and metavolcanics belong to the lower part of the actual CVA (Somin & Millán 1977; Boyanov et al. 1975; Millán Trujillo 1996; Stanek et al. 2006). The age of the protolith of the Mabujina Formation has been estimated paleobotanically as Late Jurassic to Early Cretaceous (Dublan et al. 1988). Pb-Pb zircon ages from gneisses gave 90-130 Ma

(Bibikova et al. 1988 in Stanek et al. 2006), and SHRIMP zircon dating has yielded 132 Ma (Rojas-Agramonte et al. 2005a). Geochemically and isotopically, the igneous protoliths of the Mabujina unit are interpreted as island-arc rocks derived from a depleted mantle source with sedimentary contamination (Blein et al. 2003).

The Cretaceous arc suite broadly consists of a two stage volcanic development: A primitive pre-Aptian stage is followed by a mainly calc-alkaline Aptian to Campanian phase (Iturralde-Vinent 1996c, 1998; Díaz de Villalvilla et al. 1997; Kerr et al. 1999; Stanek 2000). This development is also reflected by the grouping of granitoid intrusives, which crop out over large areas in the Camagüey area. A presumably Early Cretaceous suite of low K₂O granitoids is succeeded by K₂O-rich granitoids in the Middle Cretaceous and early Late Cretaceous (Stanek 2000).

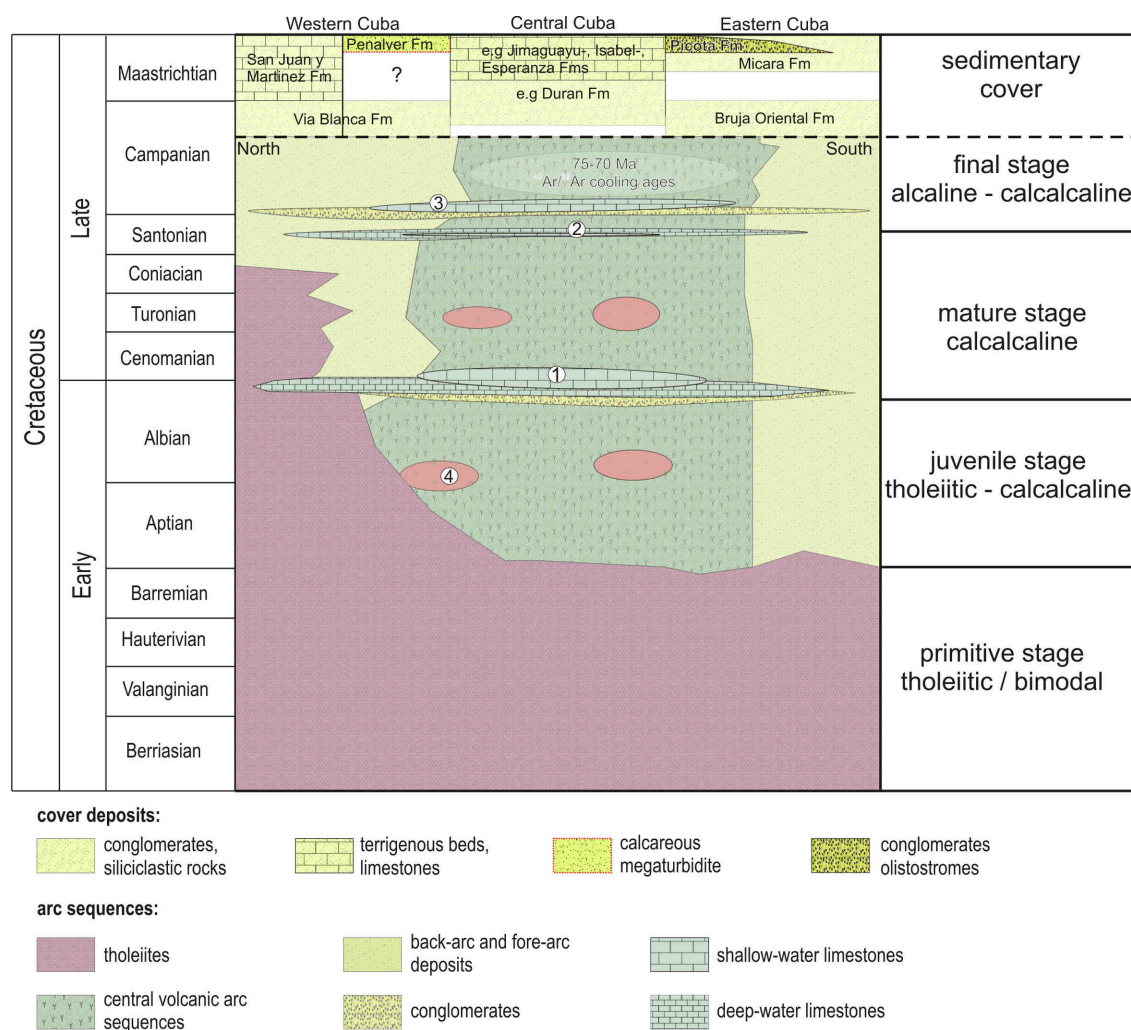


Fig. 14: Generalized scheme of the principle components of the Cretaceous volcanic arc in Cuba (from Iturralde-Vinent 1996c). 1. late Albian limestones with *Tepeyacia corrugate*; 2. Santonian limestones with *Durania curasavica*; 3. Campanian limestones with *Barrettia monilifera*; 4. intrusion events.

A distinct stratigraphic level in the lower part of the calc-alkaline section of the volcanic arc consists of Albian to early Cenomanian limestone horizons. In some places, the

Albian to early Cenomanian limestones occur unconformably with a basal conglomerate. Similar horizons are found throughout the entire Greater Antilles (Hatillo Limestone on Hispaniola, Bourdon 1985 in Draper et al. 1994; Río Matón Limestone on Puerto Rico, Kerdraon 1985 in Stanek 2000) and they approximately correlate with the change in the geochemical signature of the arc. This is consulted by some authors to support the scenario of an Aptian - Albian flip in subduction polarity (e.g. Lewis & Draper 1994; Pindell 1994; Lebrón and Perfit 1994; Draper et al. 1996; Cobiella-Reguera 2000; Pindell & Kennan 2001; Pindell et al. 2006). Stanek (2000) points to a correlation of the Albian to Cenomanian change of the CVA with the extrusion of the Caribbean flood basalts.

Onset of uplift and erosion constrains the minimum age for the termination of subduction and related volcanic activity in the Cretaceous volcanic arc. Rapid uplift and erosion of the arc during the Campanian is indicated by Ar-Ar cooling ages from volcanic and intrusive rocks of the Camagüey area (Hall et al. 2004) and by Campanian-Maastrichtian terrigenous cover beds, which even crosscut eroded granitoid intrusives (Monos, Durán, Yáquimo, and Sirvén Formations; Cobiella-Reguera 2000).

Further indications for a Campanian cessation of volcanic arc activity comes from the structural and pressure-temperature-time evolution of the Escambray metamorphic complex (chapter 4.1.4; Fig. 15). In western Cuba, post-volcanic cover beds of the arc consist of Campanian to Maastrichtian limestones and conglomerates (Vía Blanca and San Juan y Martínez Formations, Piotrowski 1987, Puscharovskiy et al. 1988). In contrast to its widespread occurrence on the continental margin units, a Cretaceous - Tertiary boundary megaturbidite overlays only the western parts of the Cretaceous volcanic arc, i.e. in the Bahía Honda unit and the Habana - Matanzas area (Peñalver Formation, Takayama et al. 2000).

Bahía Honda unit

The Bahía Honda unit represents a controversial component of CVA affinity. It occupies the highest structural position in the Cordillera Guaniguanico nappe pile. Internally, its upper sheets are composed of ultramafic rocks, mafic lavas and gabbros. The lower sheets are composed of intensely deformed mafic tuffs interstratified with mafic and intermediate lavas, siliceous slates, laminated limestones, and radiolarites (e.g. Encrucijada Formation, see Appendix, III A29). Therefore the Bahía Honda unit seems to represent an overturned oceanic – volcanic arc sequence. First biostratigraphic dating of the Encrucijada Formation yielded ages between Aptian to Cenomanian (Aiello & Chiari 1995).

Stanek (2000) concludes that the entire Bahía Honda unit may represent slide masses and olistostrome deposits, which slid from the collision front of the CVA to the foreland basin during the latest Maastrichtian. Thus, the Bahía Honda rocks would have been in an allochthonous position on top of the continental margin before the Paleogene thrust

tectonics commenced (chapter 4.1.5; chapter 7.2). According to Stanek (2000) this conclusion can be drawn from the following facts:

- Exploration wells encountered Maastrichtian conglomerates below the Encrucijada Formation;
- The entire unit contains high-energetic deposits ranging from breccias, olistostromes to mafic and ultramafic olistoplaques;
- The absence of contact metamorphism between magmatic, volcanic bodies and their surrounding sedimentary rocks prove the contacts to be strictly sedimentary or tectonic;
- Compared to standard volcanic arc sequences, the sequence of deposition is inverted;
- Internal deformation and thrusting of the units seem to have occurred after the deposition of the inverted sequence.

4.1.4 Metamorphic complexes

Two tectonic windows within the Cretaceous volcanic arc suite expose high-pressure metamorphic rocks (Escambray Massif, Isla de la Juventud; Fig. 9). The dome shaped metamorphic complexes are tectonically overridden by the CVA units from the south (Millán-Trujillo 1997; Grafe et al. 2001).

The metasedimentary sequences of the Escambray Massif are surrounded and separated by tectonic contacts from a series of amphibolites, metagabbros and gneisses of the Mabujina Complex (Somin & Millán 1976 in Stanek et al. 2006) which represent the basement of the CVA (Somin & Millán 1981 in Stanek et al. 2006; Hatten et al. 1988). The shear zone below the Mabujina unit is the main detachment between the high-pressure metamorphic rocks and the overlying arc complex (Stanek et al. 2006; Fig. 23).

The dominant rock types of the Escambray Complex are monotonous carbonate- and quartz-mica schists. On the basis of rare remnants of fossils the protolith ages have been suggested to be Late Jurassic to Early Cretaceous, comparable to the stratigraphic profile in western Cuba (Khudoley & Meyerhoff 1971; Millán & Myczynski 1978; Somin et al. 1992 in Stanek et al. 2006). A second group comprises dark marbles with tectonic slivers of metagabbro, greenschist, sulfide bodies and serpentinite. Additionally, various bodies of up to several tens of meters in size of eclogite, blueschist, garnetiferous mica schist, serpentinite and metaquartzite form mélange-like zones within the carbonate- and quartz-mica schists (Stanek et al. 2006).

On the basis of structural data and pressure-temperature-time evolution Stanek et al. (2006) subdivide the metamorphic complex into three units which form a top to north directed stack of nappes. The tectonically lowermost nappe, called Pitajones unit, never exceeded high-grade greenschist-facies conditions. In contrast, the overlying

nappes, the Gavilanes and Yayabo units, record clear evidence of HP/LT metamorphic conditions (Schneider et al. 2004; Stanek et al. 2006).

The Mabujina unit and the neighbouring unit of the Escambray metamorphic complex have been intruded by pegmatites, which did not undergo ductile deformation as their amphibolite-facies host rocks (Stanek et al. 2006). Thus, the 88-80 Ma intrusion ages of the pegmatites (Grafe et al. 2001) mark the minimum age of HP-metamorphism in metamorphic rocks and the ductile deformation of the arc basement (Stanek et al. 2006).

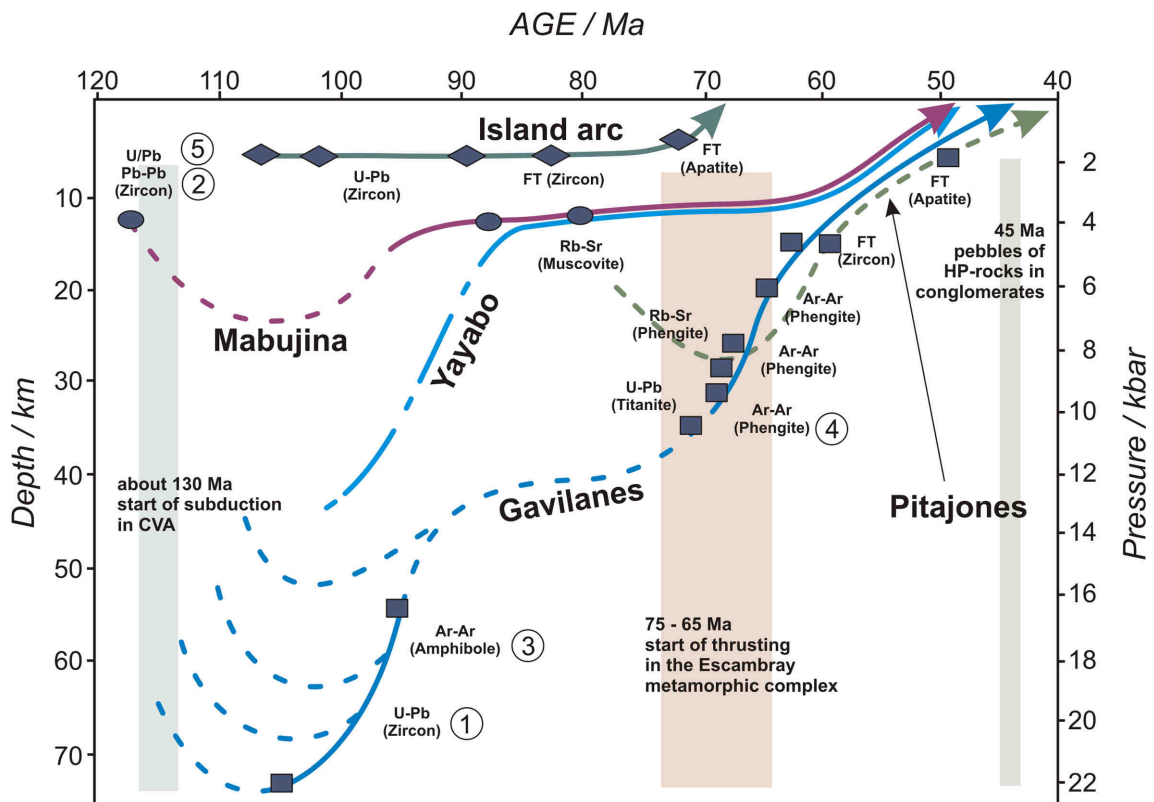


Fig. 15: Pressure – temperature paths of the nappe units in the Escambray metamorphic complex and the Cretaceous volcanic arc units in central Cuba (compiled by Stanek et al. 2006 based on data by Grafe 2001, as well as Hatten et al. 1988 [=1], Bibikova et al. 1988 [=2], Renne in Draper & Nagle 1991 [=3], Schneider et al. 2004 [=4], Rojas-Agramonte et al. 2005a [=5]).

Stanek et al. (2006) compiled pressure-temperature-time paths for each nappe in the Escambray Massif and the CVA units in central Cuba (Fig. 15). Based on this data several conclusions are suggested (Stanek et al. 2006):

- The high-pressure metamorphic rocks represent the subduction-accretion complex associated to the CVA.
- Subduction began at least as far back as Aptian / Albian, and magmatic activity of the arc is indicated until at least 80 Ma.

- The Mabujina unit (CVA basement) and the Yayabo unit of the subducting plate were already welded together by 90 Ma.
- Rapid cooling started at ~75 Ma, most probably by underthrusting of sediments of the Bahamas margin and contemporaneous uplift. At this time, the CVA was already being deeply eroded.
- The sedimentary sequences of the Bahamas margin which were involved in the down-going slab at ~75 Ma, led to a shallowing of the subduction angle and a thickening of the accretionary wedge. The volcanic front was shifted away from the arc axis and the magmatism in the arc ceased. The subducted continental margin sediments underwent a high greenschist-facies metamorphic overprint and are now represented by the Pitajones unit.
- Thrusting of the metamorphic subduction-accretion complex and the CVA onto the southern margin of the Bahamas margin began at about 65 Ma when further cooling was assisted by extensional tectonic unroofing (Pindell et al. 2005). Buoyancy forces and slab rebound helped to uplift and exhume the metamorphic complex. In the sedimentary basins surrounding the Escambray complex, the first pebbles of HP-metamorphic rocks occur at about 45 Ma (Kantchev et al. 1976).

The geological record of metasedimentary rocks of the Isla de la Juventud is very similar to that of the Escambray Complex. Petrographic and biostratigraphic evidence also point to an affinity of the protoliths to the stratigraphic profile in western Cuba (Khudoley & Meyerhoff 1971; Millán 1975; Somin & Millán 1977; Millán & Myczyński 1978). Structural observations, metamorphic mineral assemblages, as well as K-Ar data from metamorphic micas and from post-metamorphic intrusions advise to an evolution analogue to the metamorphic rocks of the Escambray Massif (Stanek 2000). The tectonically overlying CVA units are exposed in the Sabana Grande Zone at the northwestern fringe of the island (Fig. 9; Fig. 20).

4.1.5 Paleogene foreland and piggyback basins

Paleogene foreland basins related to the evolution of the Cuba orogenic belt developed on the Mesozoic paleomargin deposits in the front of the extinct CVA, whereas contemporaneous piggyback basins evolved above the allochthonous thrust units of the arc. Parts of the foreland basins were eventually overthrust by the advancing tectonic prism of the orogenic belt ([Iturralde-Vinent, 1998 #344]; Saura et al. 2008).

From about the Early Eocene onward, northeast striking transform faults (e.g. Pinar-, Matanzas-, La Trocha-, Cauto Faults) started to dissect the previously compressive structure. Normal offsets of up to several thousand meters induce the development of transtensional basins to the south of the major faults. Simultaneously, deformed basement rocks exhumed on the northwestern shoulders of the transform faults (Puscharovskiy et al. 1989; Gordon et al. 1997; Iturralde-Vinent 1995, 1998; Saura et al. 2008; Fig. 9).

Foreland basins

Paleogene foreland basin sequences of several hundred meters of thickness occur in the central Cuban fold and thrust belt and in the Cordillera Guaniguanico.

In the **Cordillera Guaniguanico**, the pre-tectonic hemipelagic limestones (Ancón Formation, Fig. 16) have an age span from late Early Paleocene to earliest Eocene (Puscharovskiy et al. 1988; Puszczółkowski 1978, 1999; Bralower & Iturralde-Vinent 1997; Iturralde-Vinent 1998). The overlying latest Paleocene to Early Eocene Manacas Formation is considered to represent syn-tectonic deposits. Sandstones of its basal Pica Pica Member record an increased influx of detritus from the CVA and from Jurassic siliciclastic rocks.

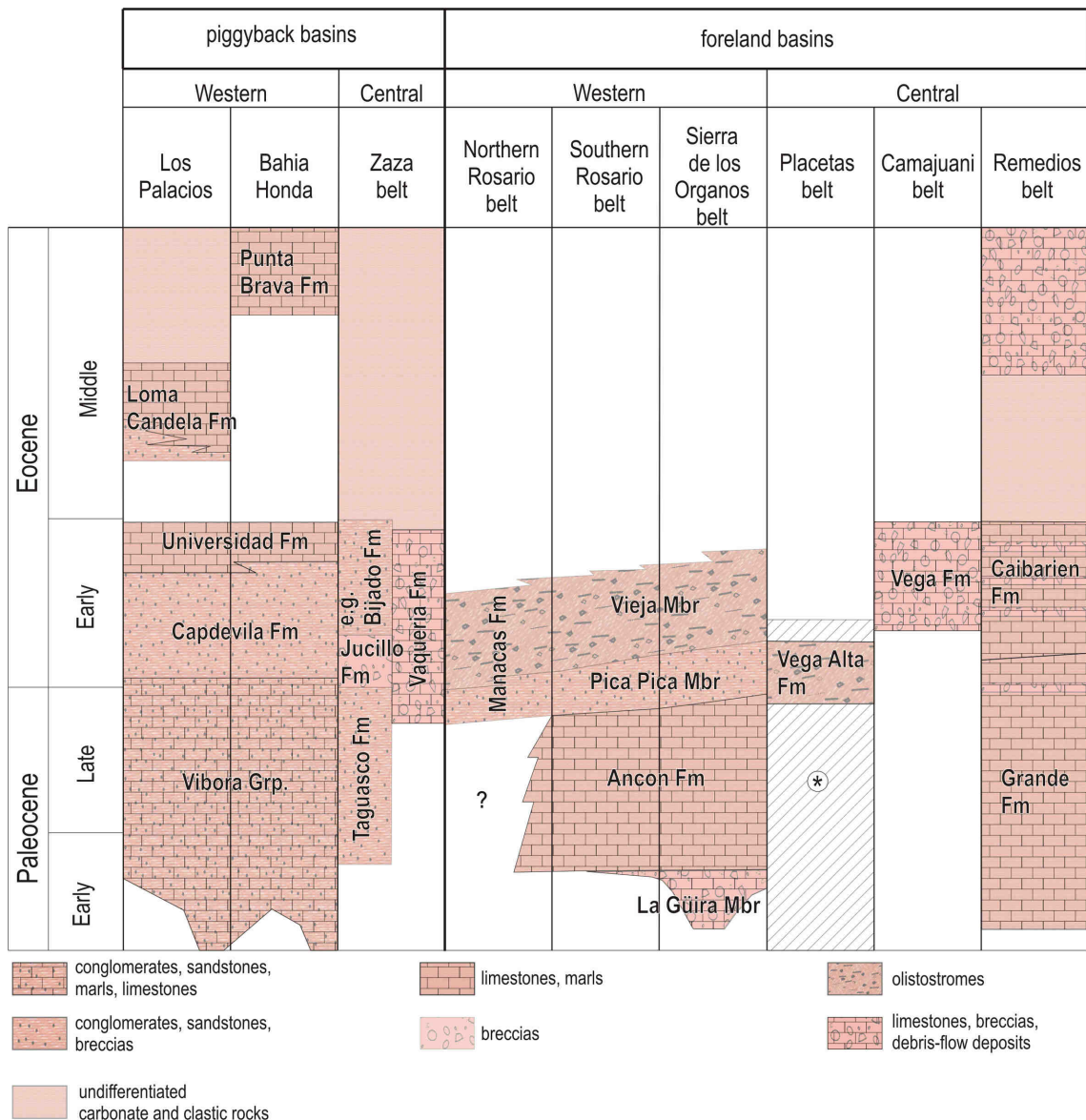


Fig. 16: Generalized lithostratigraphic scheme of the Paleogene foreland and piggyback basins in western and central Cuba (from Puscharovskiy et al. 1988; Bralower & Iturralde-Vinent 1997; Iturralde-Vinent 1998; Puszczółkowski 1999).

The Vieja Member, possibly representing olistostromes, contains abundant blocks and phacoids of serpentinites, gabbros, basalts and Mesozoic limestones, thus constraining the final collision between the CVA and the North American Paleomargin to a latest Paleocene to Early Eocene interval (Pszczółkowski 1978; Bralower & Iturralde-Vinent 1997).

According to Iturralde-Vinent (1998), the syn-tectonic deposits in **central Cuba** would span a considerably longer period, whereas the olistostromes would get progressively younger from south to north. Thus, syn-tectonic deposits with clastic material from the ophiolites and the CVA would be of Paleocene age in the Placetas belt, of Late Paleocene to Early Eocene age in the Camajuaní belt, and of late Middle Eocene age in the Remedios belt. The Cayo Coco belt lacks Paleocene sediments and olistostromes, presumably because it represented a forebulge to the foreland basin (Iturralde-Vinent 1998).

The cessation of syn-tectonic deposition in the Cuba orogenic belt apparently becomes younger towards the east. Its end is constrained to the Early Eocene in western Cuba (Pszczółkowski 1978; Bralower & Iturralde-Vinent 1997) to the Middle Eocene in the region between La Habana and Ciego de Avila (Kantchev et al. 1976), and to the Late Eocene in the region between Camagüey and Holguín (Nagy et al. 1983; Iturralde-Vinent et al. 1981, 1986). This apparent trisection resembles the Eocene establishment of the left-lateral strike-slip faults. With regard to the different ages in the cessation of syn-tectonic deposition, the Pinar- and the La Trocha Faults seem to represent major limitations of individual tectonic blocks.

Overall, latest Eocene deposits discordantly cover the deformed foreland basin deposits, marking the orogen-wide termination of folding and thrusting (Iturralde-Vinent 1978, 1995).

Large proportions of the Paleocene to Eocene foreland basin were eventually overridden by the advancing thrust units from the south. In the Cordillera Guianiguanico, the pre- and syn-orogenic foreland basin deposits are generally foliated and sheared with great intensity. They occur between each of the main superposed nappe units. Due to the *mélange*-type deformation of these units, an original olistostrome or debris flow character of the Vieja Member is not always obvious (Appendix, IV A31). The same applies to the Vega Alta Formation in the Placetas belt in central Cuba, but in the fold and thrust belt of central Cuba, the intensity of deformation decreases from southwest to northeast. Only a few kilometres to the northeast of the Vega Alta outcrops, the Early Paleocene deposits in the Camajuaní belt (Vega Formation) are hardly altered by shear deformation and clearly recognizable as debris flow deposits (Appendix, IV A75).

Piggyback basins

Paleocene to Late Eocene syn-tectonic basins evolved on the allochthonous thrust units of the CVA and the Northern Ophiolite Belt. In many locations the piggyback basin rocks rest with an erosive hiatus on the late Campanian to Maastrichtian cover beds of the extinct CVA.

The 'cover beds' of the CVA generally constitute a first transgressive cycle, beginning with sandstones and conglomerates. Clastic components derive from the erosion of the CVA but also encompass ophiolitic material. By the late Maastrichtian, shallow-water limestones and marls are predominant, but facies and thickness of the 'cover beds' display many lateral variations (Iturralde-Vinent 1998).

The Paleocene to Late Eocene piggyback basin successions are dominated by sandstones, conglomerates, marls, and limestones (e.g. Appendix, V A27). Iturralde-Vinent (1998) describes lateral transitions between deep-water and shallow-water limestones and paleo-soils. From the Middle Eocene onward, transtensional basins associated to the strike-slip faults – from west to east: Cuenca Los Palacios, Cuenca Las Vegas, Cuenca de la Broa, Cuenca Central, Cuenca Cauto (Fig.9) - accumulate several thousand meters of sedimentary infill without any major unconformities (Puscharovskiy et al. 1989; Saura et al. 2008).

In the Holguin area, the Paleocene and Early to Middle Eocene piggyback deposits contain volcanoclastics and tuff horizons (Appendix, VIII A143), grading over into the back-arc basin of the Paleogene volcanic arc to the south (chapter 4.2.1; chapter 7.3).

4.2 Eastern Cuba

The geology of eastern Cuba, i.e. of the Oriente Block southeast of the Cauto – Nipe Basin, is characterized by two major components:

- The Paleogene volcanic arc complex of the Sierra Maestra (PVA), and
- the Nipe – Cristal / Moa Baracoa ophiolite massifs which rest on top of greenschist facies CVA rocks.

Further components are the metamorphic rocks of the Asunción Complex and Cenozoic basin succession, apparently including Paleogene back-arc deposits of the PVA.

Immediately offshore to the south, the Oriente Block is cut by the Oriente Transform Fault with a steep topographic gradient, ranging from 1974 m (Pico Turquino) to -6642 m in the Oriente deep (Magáz García 1989 in Rojas-Agramonte et al. 2005b; ETOPO2, Smith & Sandwell 1997). The Oriente Fault represents the north bounding fault system of the left-lateral strike-slip zone forming the northern Caribbean Plate boundary (see chapter 3.3.3).

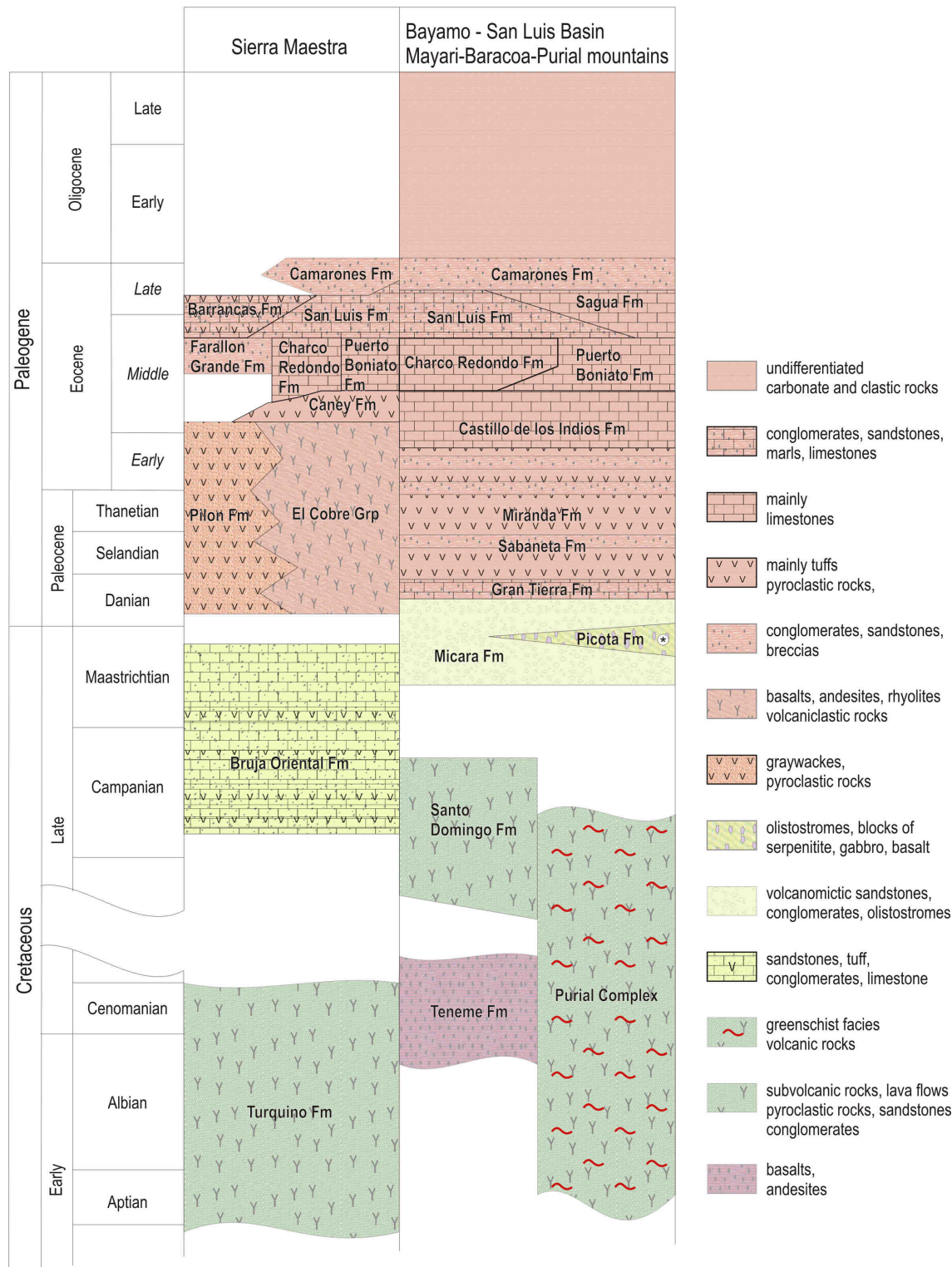


Fig. 17: Generalized lithostratigraphic scheme of the Sierra Maestra, the Mayarí-Baracoa -Sierra del Purial Mountains and the Bayamo – San Luis Basin, eastern Cuba. (from Puscharovskiy et al. 1988; Iturralde-Vinent et al. 2006 and from Rojas-Agramonte et al. 2004, after Linares et al. 1985; García Delgado & Torres Silva 1997; Méndez Calderón 1997; Remane et al. 2002). The Picota Formation (*) is associated to the Nipe-Cristall / Moa Baracoa ophiolite massifs (chapter 4.2.2).

4.2.1 Paleogene volcanic arc

Paleogene volcanic arc rocks essentially crop out in the Sierra Maestra, comprising the southern portion of the Oriente Block. Minor outcrops of the unconformably underlying CVA occur along the southern coastline of the Sierra Maestra (Puscharovskiy et al. 1988). The CVA is represented by the Aptian to Cenomanian Turquino Formation (Fig. 17), consisting of different types of volcanic and subvolcanic rocks, tuffogenic sandstones and limestones. With a stratigraphic hiatus, the Campanian to Maastichtian Bruja Oriental Formation unconformably lies on top, consisting of sandstones, conglomerates, limestones and some tuffs. The Bruja Oriental Formation corresponds to CVA cover beds of western and central Cuba.

The unconformably overlying more than 4000 m of Paleogene volcanic arc successions make up the major part of the Sierra Maestra. The PVA rocks are represented by the basaltic – andesitic and sometimes rhyolitic sequences of the El Cobre Group (Méndez Calderón 1997), the Pilón Formation with greywackes and pyroclastic rocks, and the Caney Formation, characterized by pyroclastic and sedimentary rocks, agglomerates and lava flows (Fig. 17; Iturralde-Vinent 1996d).

The sequences of the El Cobre Group are intruded by a large number of bodies and plutons of gabbro, diorite, tonalite, granodiorite and granite composition. Broadly consistent with previous age dating (Iturralde-Vinent 1996d; Rodriguez Crombet et al. 1997; Kysar et al. 1998; Mattiotti-Kysar 1999), the granitoid massifs yielded intrusion ages between 60 Ma and 48 Ma on recent $^{207}\text{Pb}/^{206}\text{Pb}$ SHRIMP analyses of single zircons (Rojas-Agramonte et al. 2004).

Tuffs and pyroclastic rocks make up large proportions of the Paleocene basin deposits that accumulated to the north of the Sierra Maestra (Bayamo – San Luis Basin and Mayari-Baracoa-Purial mountains, Fig.17).

The PVA is overlain by Middle Eocene shallow-water limestones of the Charco Redondo Formation and its deeper-water equivalents, the Puerto Boniato Formation (Fig. 17). A weak late Eocene remnant volcanism is documented by some layers of the terrigenous Barrancas Formation (Cobiella-Reguera 1988). Granitoid clasts in conglomerates of the Camarones Formation indicate that the PVA was already uplifted and eroded to the upper level of its intrusive bodies by the Late Eocene.

4.2.2 Nipe – Cristal / Moa Baracoa ophiolite massifs

North and east of the Sierra Maestra, the Sierra de Nipe – Cristal and Moa Baracoa Massifs expose two large bodies of mafic and ultramafic rocks, i.e. mainly of serpentized harzburgite, gabbros, and dolerite (Draper & Barros 1994).

The ophiolitic massifs of eastern Cuba differ from those in western and central Cuba in terms of their significantly lower degree of deformation (Appendix, VI A171; c.f. II A46, II A71) and their exceptional structural position: The Nipe – Cristal / Moa Baracoa Massifs rest on top of the CVA.

The underlying CVA is exposed primarily in the Purial Complex south of the Moa Baracoa Massif and at some localities within the Moa-Baracoa Mountains (Fig. 9). The Purial Complex consists of greenschist facies sedimentary, volcano-sedimentary, and plutonic rocks (e.g. Iturralde-Vinent et al. 2006). The protolith ages of these rocks are poorly constrained and may altogether span the Aptian / Albian to Campanian interval (Millán Trujillo 1996; Iturralde-Vinent et al. 2006). Besides the ophiolitic massifs, the Purial Complex is unconformably overlain by non-metamorphosed Danian and younger rocks (Fig. 17; see below), which suggests a Maastrichtian age of the greenschist facies metamorphism (Iturralde-Vinent et al. 2006).

In between the Nipe - Cristal and the Moa Baracoa Massifs, the underlying CVA is largely represented by non-metamorphic volcanoclastic and sedimentary rocks of the late Coniacian to mid-Campanian Santo Domingo Formation (Fig. 17; Iturralde-Vinent et al. 2006). According to Quintas (1987, 1988), the rocks of the Santo Domingo Formation are locally overthrust by metamorphic rocks assigned to the Purial Complex.

The age of the ophiolitic rocks in eastern Cuba is unknown, but the time and mode of their emplacement as allochthonous massifs on top of the CVA can be narrowed down to a certain degree. The ophiolitic sheets are essentially associated with late Maastrichtian to early Danian olistostromes, conglomerates and debris flow deposits of the La Picota Formation (e.g. Knipper & Cabrera 1974; Cobiella 1978; Iturralde-Vinent 1996b; Appendix, VI A171).

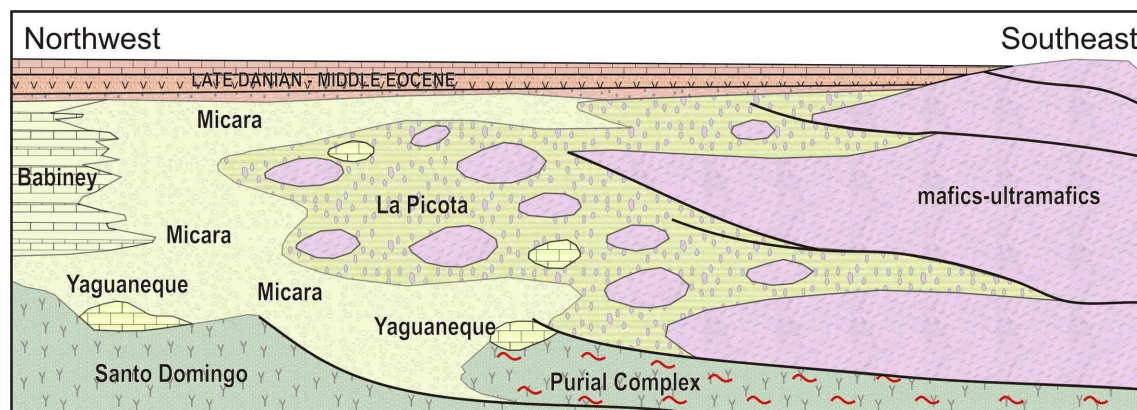


Fig. 18: Schematic cross section of the late Maastrichtian to early Danian Mícará – La Picota basin, depicting the relationships between the Mícará-, La Picota Formations, the ophiolite sheets, and the underlying Cretaceous volcanic arc (from Iturralde-Vinent et al. 2006, after Cobiella 1974, 1978).

The La Picota Formation occurs as lenticular intercalations within the Mícará Formation. The latter is composed of polymict sandstones and shales with frequent intercalations of conglomerates and debris flow deposits (Appendix, VI A168, VI A169). The clastic components encompass volcanogenic, serpentinite and biogenic material, whereas serpentinites are absent at the base of the section and increases towards the top. The olistostromes of the La Picota Formation are dominated by clastic components

of the ophiolitic suite (Cobiella 1974, 1978; Iturralde-Vinent et al. 2006, Rabehl 2006). Laterally, the proportion of olistostromes increases towards the southeast and the general lithological zonation as well as some structural hints suggest that the source of the allochthonous debris and ophiolites was located to the south of the Mícará – La Picota basin (Cobiella 1974, 1978; Iturralde-Vinent et al. 2006). Both formations display no deformations that would point to severe thrust and nappe tectonics (Appendix VI A171, VI A168, VI A169) as manifested in the Northern Ophiolite Belt. Merely some drag folds have been described, which support the presumed direction of emplacement.

Cobiella (1974, 1978) and Iturralde-Vinent et al. (2006) summarize the relationships between the Mícará, La Picota Formations, the ophiolite sheets, and the underlying CVA (Fig. 18), which suggest gravitational sliding from a southern location into a marine basin as the most probable mode of emplacement of the eastern Cuban ophiolites. The triggering process of this late Maastrichtian – early Danian event is enigmatic, but a close relationship to the mafic volcanics of the southern peninsula of Hispaniola (chapter 3.3.2) and the Bath-Dunrobin Complex in Jamaica (chapter 3.3.1) is incidental (chapter 7.3).

The rocks of the Mícará Formation grade upward into late Danian marls, limestones of the Gran Tierra Formation (Fig. 17), which still contains some debris flow deposits. The overlying Sabaneta Formation is dominated by tuffaceous intercalations, which document the emergence of the Paleocene to Middle Eocene volcanic arc.

4.2.3 Asunción Complex

At the easternmost end of the Purial Complex, the so called Asunción Complex exposes metamorphic Jurassic to Cretaceous rocks in an area of only a few square kilometers. The structurally lowermost rocks are black phyllites and slates with some marbles and metamafites (Sierra Verde Formation, Somin & Millán 1981; Cobiella-Reguera 1983; Millán et al. 1985). Late Jurassic – Early Cretaceous fossils have been found in some of the phyllites and slates (Millán et al. 1985). Dark marbles and calcareous schists (La Asunción Formation) of assumed Late Jurassic age rest tectonically on the Sierra Verde Formation (Cobiella-Reguera 1983, Millán et al. 1985). On the basis of their lithology, the rocks of the Asunción Complex have been correlated with the Escambray region, thus a passive continental margin origin has been suggested. The tectonic relationship with the neighbouring Purial Complex and the eastern Cuban mafic /ultramafic massifs or even a possible link to metamorphic units in northern Hispaniola is not clear (Cobiella-Reguera 1983; Cobiella et al. 1984; Millán et al. 1985; Iturralde-Vinent 1996a, Iturralde-Vinent et al. 2006).

5 Crustal sections across key areas on Cuba

The evolutionary interpretation of a given set of geological observations must be conducted in three dimensions, i.e. in map view as well as in cross section (Pindell et al. 2006). Therefore, several adequately spaced cross sections of the orogenic belt are an essential requirement of any palinspastic and paleogeographic reconstruction.

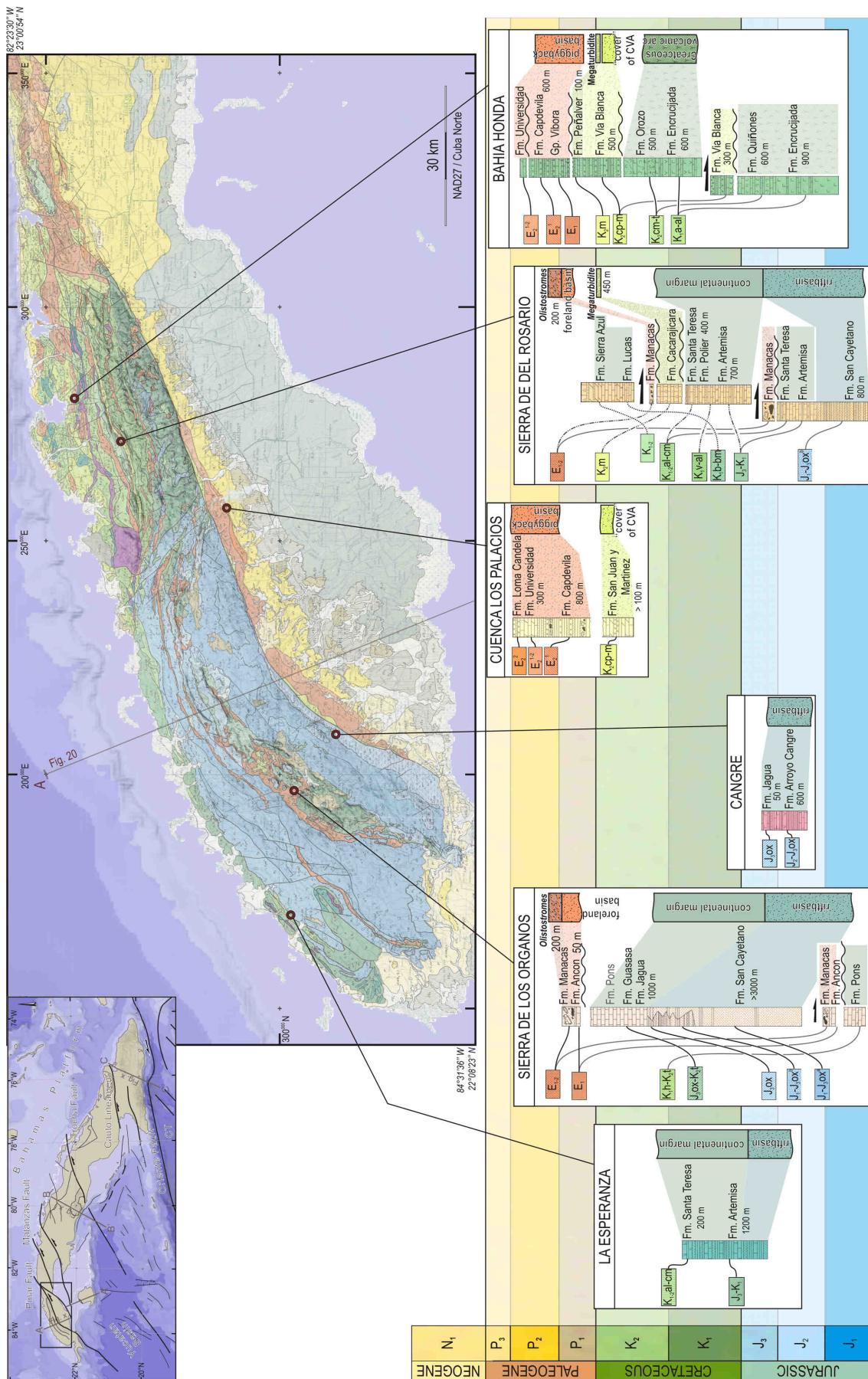
5.1 Section across western Cuba

Early considerations about the nappe structure of the Cordillera Guaniguanico (Rigassi-Studer 1963; Hatten 1967; Piotrowska 1978; chapter 4.1.1) have been largely confirmed by the results of deep wells and seismic profiles (López Rivera et al. 1987; Saura et al. 2008; Fig. 19). The lowest units encountered in depths of more than 3000 m are Jurassic terrigenous rocks, similar to those of the San Cayetano Formation, as well as contemporaneous limestones and evaporites. The nappe structure becomes obvious from recurrences of the stratigraphic section, oftentimes separated by sheared olistostromes of Early Eocene age (Fig. 20). Remnants of the overthrust CVA in the uppermost thrust sheet, (Bahía Honda unit, chapter 4.1.3), the intensity of deformation and the partial metamorphism of the Guaniguanico sedimentary rocks, suggest that the Cordillera Guaniguanico units may have been entirely overridden by CVA units during the Paleocene.

The thin-skinned thrust sheets were emplaced north- and northwestward and later folded into a gentle anticline.

The Pinar Fault abruptly truncates the anticline at its southern limb (Fig. 20). Fault slip analysis revealed a polyphase compressional and tensional sinistral strike-slip along the Pinar Fault during the latest Early Eocene to Miocene (Gordon et al. 1997; Appendix, X A124, X A125).

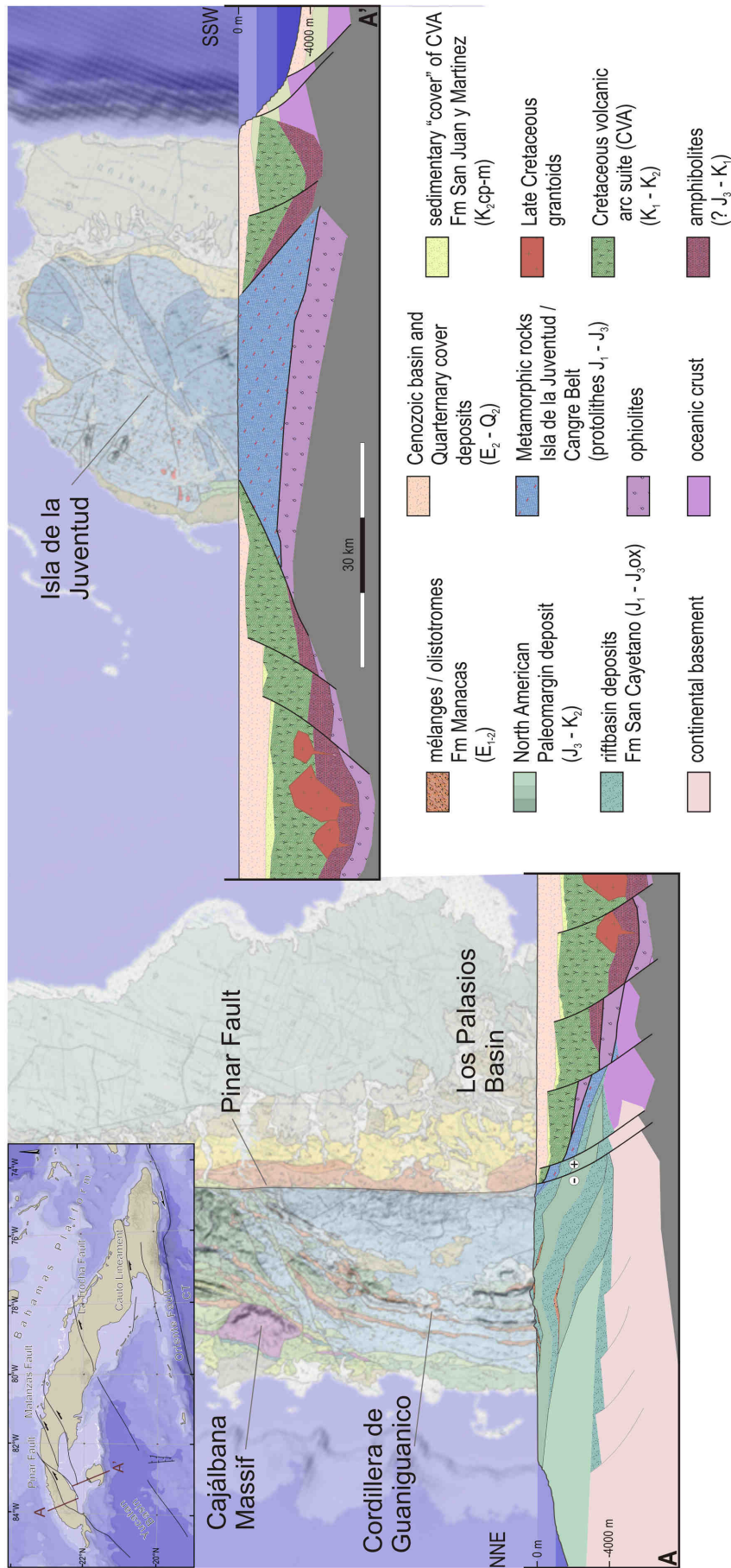
Fig. 19: (Next page) Geological map and type sections of western Cuba (compiled from Puscharovskiy et al. 1988). A-A' represents the line of the constructed cross section. Topographic and bathymetric data was taken from SRTM, ETOPO2 and Puscharovskiy et al. (1988).



The basement of the syn- to post-collisional Paleogene and Neogene sedimentary rocks of the Los Palacios Basin (chapter 4.1.5; Fig. 9) evidently consist of CVA units. These are exposed in the Savana Grande zone, tectonically overlying metamorphic Jurassic sequences (Puscharovskiy et al. 1988) at the northwestern coast of the Isla de la Juventud. Geophysical data indicate a continuation of CVA rocks below the Gulf of Batabano and the Los Palacios Basin (Bush & Sherbakova 1986). Maastrichtian cover deposits of the CVA are exposed in a narrow belt south of the Pinar Fault to the northwest of Pinar del Río and have been encountered in deep drillings (San Juan y Martínez Formation, Puscharovskiy et al. 1988, Fig. 19).

Following Stanek et al. (2006), the metamorphic fold and thrust structure exposed on the Isla de la Juventud may have been part of the Cretaceous subduction-accretion complex of the Greater Antilles Arc. The subduction-accretion complex has been completely overridden by the arc from the south. The symmetry of structures in the cross section suits this theory (Fig. 19). Other workers assume that the metamorphic complexes may be underlain by a piece of Caribbean continental crust, which was accreted to the Cretaceous volcanic arc from the south (e.g. Iturralde-Vinent 1994; Iturralde-Vinent & MacPhee 1999, chapter 6.5.2, chapter 7.1).

Fig. 20: (Next page) Section across western Cuba, i.e. the metamorphic dome exposed on the Isla de la Juventud, the Los Palacios Basin, and the thrust nappes of the Cordillera Guaniguanico mainly consisting of Jurassic to Paleogene rift basin, continental margin and foreland basin deposits. For type section reference see Fig. 19. For lithostratigraphic reference see chapter 4.1. For outcrop documentations see Appendix: 1) deformational style Cordillera Guaniguanico units (I A5, I A10, I A26); 2) Late Paleocene – Early Eocene sheared Olistostromes of the Cordillera Guaniguanico (IV A31); 3) Early Eocene piggyback basin deposits of the Los Palacios Basin (V A27); 4) sense of shear indicators from the Pinar Fault Zone (X A124, X A125).



5.2 Section across central Cuba

The northern half of central Cuba consists of Mesozoic deep water-, slope-, and platform deposits of the Bahamas continental margin as well as of partly olistostromal Paleogene foreland basin deposits (Fig. 21; chapter 4.1.1, chapter 4.1.5). The Camajuaní – Placetas rocks (deep water and slope) are folded and sheared much more tightly than those of the Cayo Coco – Remedios zones (platform). The intensity of folding and thrusting decreases towards the north (Fig. 22).

The Camajuaní – Placetas zones are overlain by thrust nappes of the Northern Ophiolite Belt (chapter 4.1.2). In some places the ophiolite nappes are notably thin-skinned and the underlying Paleomargin rocks are exposed in tectonic windows (Fig. 22, 23).

The ophiolites represent a *mélange* like assemblage, consisting of tectonized ultramafic and mafic rocks floating in a serpentinite matrix (Appendix II A46, II A71).

To the south of the NOB, the overlaying CVA rocks form a megasyncline, exposing the lower, pre-Aptian part of the volcanic sequence as well as the Middle to Late Cretaceous volcanoclastic and igneous suites (Pardo 1975; chapter 4.1.3).

South of the allochthonous island arc rocks, the Escambray metamorphic complex is exposed in a tectonic window (chapter 4.1.4).

The allochthonous Mabujina amphibolites at the top of the Escambray complex (Fig. 23) represent the metamorphosed basement of the CVA (Iturralde-Vinent 1994; Kerr et al. 1999; Stanek et al. 2006). The shear zone at the base of the Mabujina unit acted as the main detachment between the overlaying arc complex and the underlying Jurassic – Early Cretaceous metamorphic sequences. The symmetry of the large scale nappe pile of the Escambray metamorphic complex (Fig. 23) as well as the older generations of its internal structures (D1 and D2), which are ascribed to subduction and collision, indicate top to the north direction of tectonic transport (Stanek et al. 2006). Younger structures encompass sinistral strike-slip faults. Only the youngest structures, ascribed to exhumation, indicate top to the south direction of tectonic movement (Stanek et al. 2006).

The exhumation and uplift of the Escambray metamorphic complex in the early Paleogene was accompanied by the formation of flanking sedimentary basins (e.g. Trinidad Basin, Fig. 21).

Fig. 21: (Next page) Geological map and type sections of central Cuba (compiled from Puscharovskiy et al. 1988). B-B' represents the lines of the constructed cross section. Topographic and bathymetric data was taken from SRTM, ETOPO2 and Puscharovskiy et al. (1988).

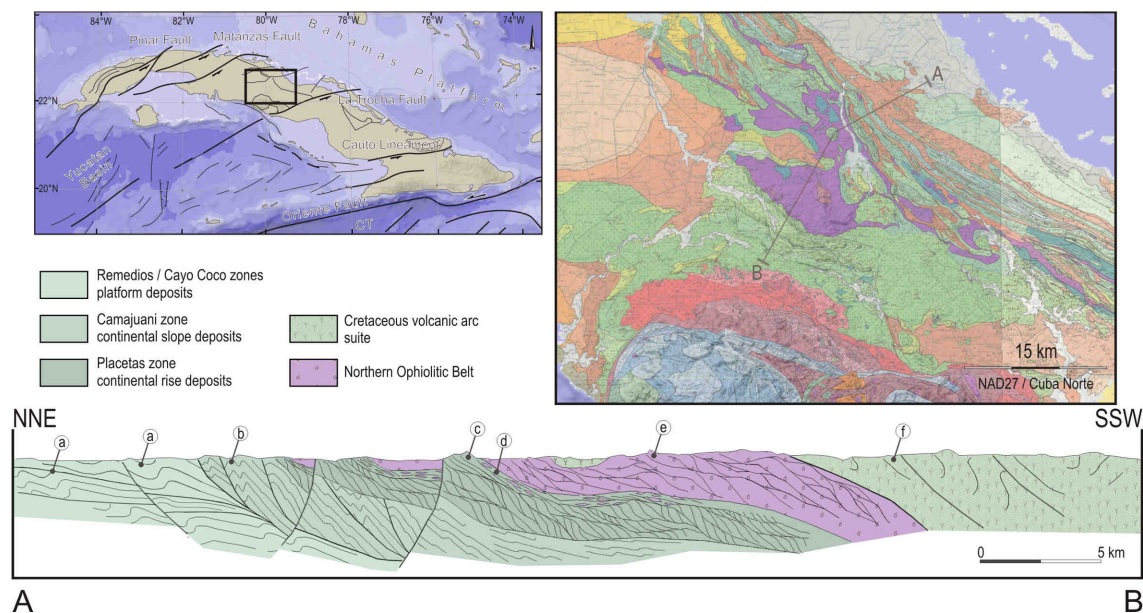
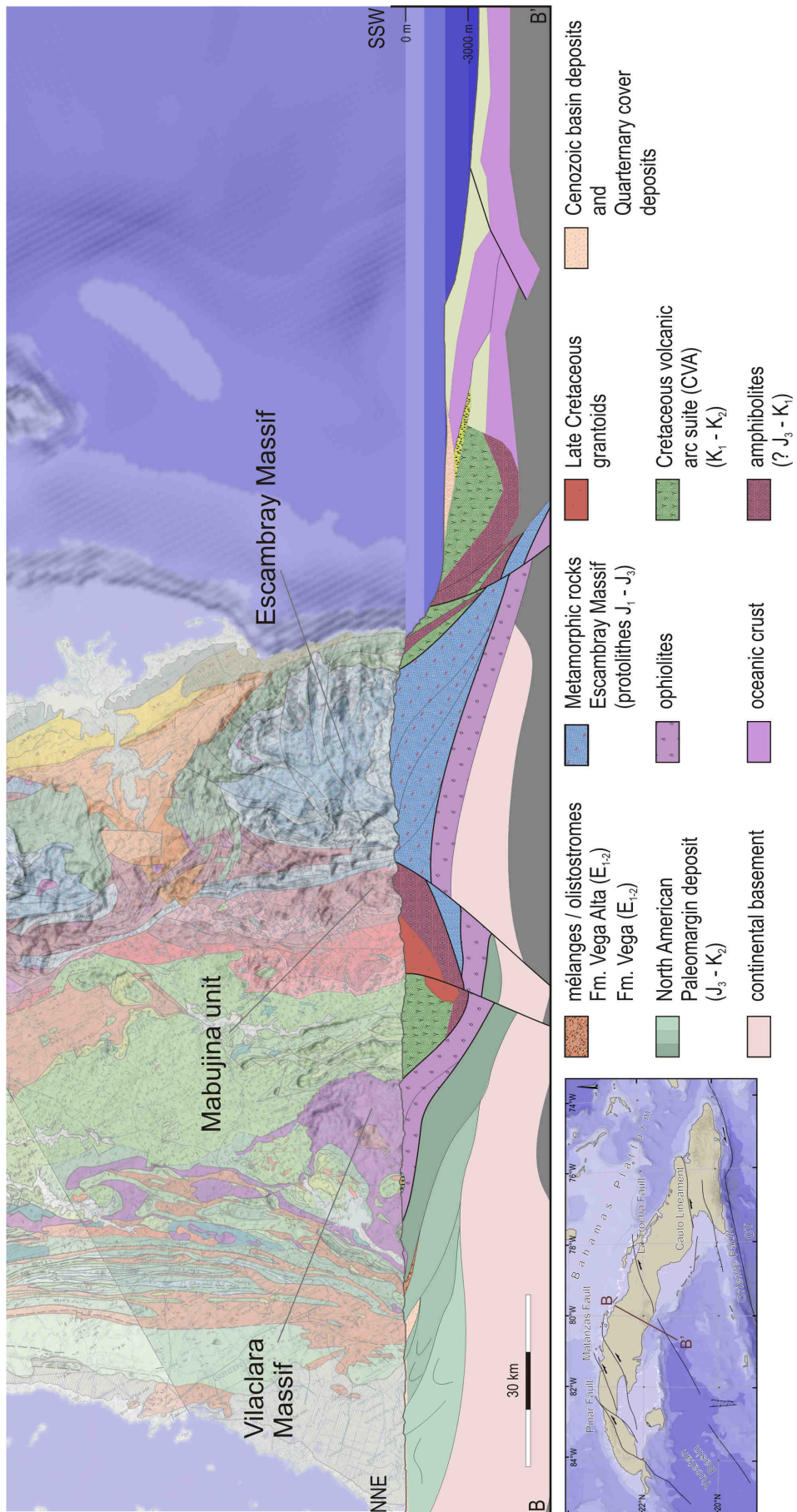


Fig. 22: Deformational styles of geological units in the fold and thrust belt of central Cuba as observed in outcrops along the section line. The Camajuaní – Placetás rocks (**b** – Appendix, I A65, I A76, I A129) are folded and sheared much more tightly than those of the Cayo Coco – Remedios zones (**a** – Appendix, I A86, I A131). The ophiolites are represented by mélanges (**d** – Appendix, I A67) and assemblages of tectonized ultramafic and mafic rocks floating in a serpentinite matrix (**e** – Appendix, II A71). Tectonic windows within the NOB expose the underlying Paleomargin rocks (**c** – Appendix, I A65). The overlying Cretaceous volcanic arc rocks to the south are moderately folded and sheared (**f** – Appendix, III A99), forming a megasyncline between the NOB and the Escambray Massif.

Fig. 23: (Next page) Section across central Cuba, i.e. the metamorphic dome exposed in the Escambray Massif, the Cretaceous volcanic arc units exposed in a megasyncline to its north, the Northern Ophiolite Belt, and the thrust nappes of the continental margin and foreland basin deposits. For type section reference see Fig. 21. For lithostratigraphic reference see chapter 4.1. For outcrop documentations see Appendix: 1) deformational style of continental margin units (I A65, I A76, I A86, I A129, I A131); 2) Late Paleocene – Early Eocene olistostromes and debris flow deposits of the central Cuban foreland basin (IV A75); 3) deformational style of the northern ophiolites (II A71); 4) exemplary lithologies of the Cretaceous volcanic arc suite (III A92, III A99, III A101).



5.3 Section across eastern Cuba

The basement of eastern Cuba consists of CVA units, as exposed by the Turquino Formation at the southern margin of the Sierra Maestra, several locations in the Mayari-Baracoa Mountains, and by the greenschist facies equivalents in the Sierra del Purial (Fig. 17, 24, 25). The greenschist facies CVA units of the Purial Complex seem to be thrust over the non-metamorphic CVA rocks from a southern direction (Quintas 1987, 1988; Fig. 25).

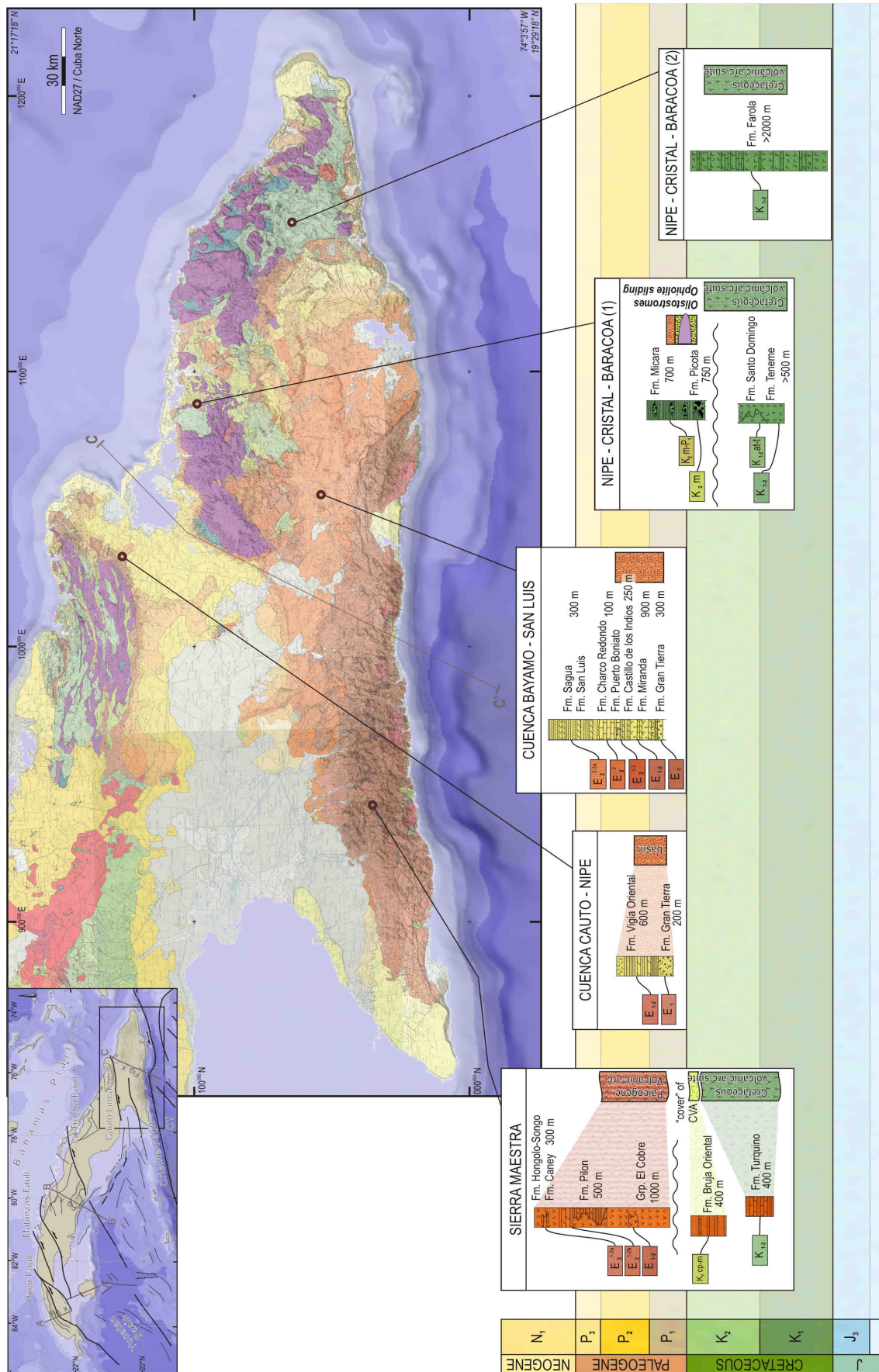
The main part of a subduction-accretion complex must be presumed deeply buried below the Cretaceous volcanic arc suite (Fig. 25), but a small proportion may be exposed by the Asunción Complex (chapter 4.2.3).

A distinctive feature of the geology of eastern Cuba is represented by sheets of mafic and ultramafic rocks which rest on top of the CVA units (Fig. 18, 24, 25), and which are associated with late Maastrichtian to early Danian olistostromes, conglomerates and debris flow deposits (La Picota and Micara Formations, 4.2.2 Nipe-Cristal / Moa Baracoa ophiolite massifs; Appendix, VI A171). Gravitational sliding from a southern location has been proposed as the most probable mode of emplacement of the eastern Cuban ophiolites (e.g. Cobiella 1974, 1978; Iturralde et al. 2006; chapter 4.2.2). Therefore, these rocks most likely underlay large proportions of the Bayamo San Luis Basin (Fig. 25).

A further unique characteristic of the geology of eastern Cuba is represented by the Paleogene volcanic arc complex of the Sierra Maestra. The Paleocene to early Eocene volcano-sedimentary sequences of the Bayamo San Luis Basin (e.g. Miranda Formation; Fig. 17, Appendix, VIII A154) apparently evolved in a corresponding back-arc basin, which extended northward at least as far as the present day Holguin Massif (Vigia Formation, Appendix, VIII A143). Its eastward extension may reach out far beyond the present day Cauto-Nipe Basin onto the Cayman Ridge (Sigurdsson et al. 1997; chapter 3.3.4)

At the southern coast, the PVA unconformably overlays Campanian-Maastrichtian cover deposits of the CVA (Bruja Oriental Formation, chapter 4.2.1). Below the Bayamo San Luis Basin to the north, the PVA units presumably overlay the sheets of the eastern Cuban ophiolites - a stratigraphic contact which can not be observed at the southern coast due to late Maastrichtian – early Danian uplift and erosion before the commencement of the PVA (chapter 7.3).

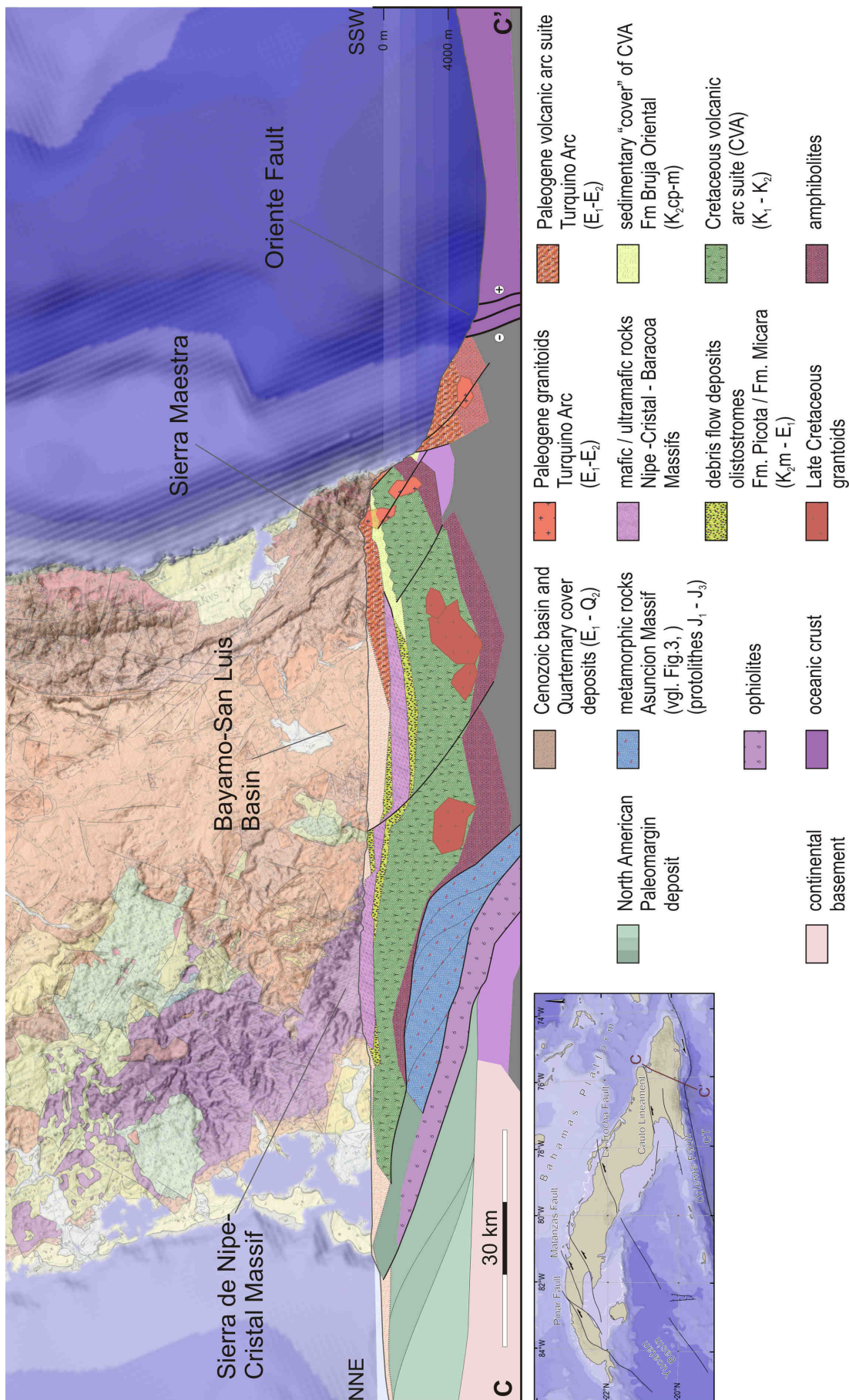
Fig. 24: (Next page) Geological map and type sections of eastern Cuba (compiled from Puscharovskiy et al. 1988). C-C' represents the line of the constructed cross section. Topographic and bathymetric data were taken from SRTM, ETOPO2 and Puscharovskiy et al. (1988).



The rocks of the Sierra Maestra are folded with increasing intensity from south to north, apparently deformed by a collision from the south postdating the Paleogene volcanic arc activity (chapter 7.5).

The Oriente Fault abruptly truncates the PVA complex at the southern margin of the Sierra Maestra (Fig. 25). Rojas-Agramonte et al. (2005b, 2006) distinguish two generations of east-striking faults along the southern coast. The fault systems seem to postdate the folds and are apparently related to the evolution of the Cayman Trough and the Oriente transform fault. The older generation is represented by normal faults of at least Miocene age. The younger generation shows sinistral displacement. A conjugated set of northwest-striking faults, some of them with sinistral displacement, and dextral northeast-oriented strike-slip faults occur over the entire mountain range and may correlate with the strike-slip movements along the Oriente Fault (Rojas-Agramonte et al. 2005b, 2006).

Fig. 25: (Next page) Section across eastern Cuba, i.e. the northern wall of the Cayman Trough and the Sierra Maestra, the Bayamo – San Luis Basin, and the Sierra de Nipe – Cristal Massif consisting of mafic and ultramafic rocks. For type section reference see Fig. 24. For lithostratigraphic reference see chapter 4.2. For outcrop documentations see Appendix: 1) sheets of eastern Cuban ophiolites and associated olistostromes (VI A171); 2) deformational style of Paleogene volcanic arc rocks of the Sierra Maestra (VII A161); 3) exemplary lithologies of the Paleogene back-arc and the cover deposits of the arc (VIII A143, VIII A154, IX A156, IX A157, IX A163, IX A187).



6 Geodynamic setting

This chapter discusses the fundamentals of current Jurassic to Early Cretaceous reconstructions of the Caribbean. It defines the basis for detailed reconstructions of the northwestern Caribbean which are proposed in chapter 7.

6.1 Tectonic hypotheses and touchstones

Morris et al. (1990) give a comprehensive overview of the early tectonic hypotheses of the Caribbean. Accordingly, it was Suess (1885, 1909) who published the first geological synthesis of the Caribbean region. By the 1960's, the reconstructions involved the following mechanisms:

- Contraction (Schuchert 1932, 1935; Willis 1932; Waters & Hedberg 1939; Senn 1940; Meyerhoff 1946, 1954; Meyerhoff & Meyerhoff 1972);
- Mantle convection (Vening Meinesz et al. 1934; Rutten 1935; Hess 1938; Hess & Maxwell 1953, Eardley 1954; Barr 1963; MacGillavry 1970);
- Earth expansion (Carey 1958);
- Continental drift (Issac del Corral 1940; North 1965);
- Oceanization (Judoley & Furrázola-Bermúdez 1971).

Furthermore, Morris et al. (1990) distinguish six basic concepts and ideas from the plate-tectonic era up to the mid 1980's. These are summarized as follows:

- "The Caribbean Plate predates the onset of seafloor spreading in Jurassic time" (Nafe & Drake 1969, in Morris et al. 1990, p.434);
- "The Caribbean Plate is not of Pacific origin; all available evidence indicates its Tethys affinity." (Aubouin et al. 1982, p. 755);
- "...the Caribbean Plate was inserted from the Pacific ...(Mattson 1969; Malfait & Dinkelman 1972). Later versions [of this concept] generally [have] a spreading center extending from the East Pacific Rise to the Mid-Atlantic Ridge until Late Cretaceous time..." (Morris et al. 1990, p.434);
- "Mooney (1980) proposed .. that, after 80 Ma, a..Venezuela Basin spreading center remained. Between 65 and 55 Ma, South America overrode that spreading center. The magmatic anomalies in the Venezuelan Basin today were interpreted to be remnant on the northern flank of that postulated spreading center." (Morris et al. 1990, p.434).
- Salvador (1986) is cited to have presented a different model of Pacific insertion (Morris et al. 1990, p.435).

- Anderson & Schmidt's (1983) spreading model for the region involves megashears (Morris et al. 1990, p.435).

Pindell (1994) emphasizes a distinction between models which generate the lithosphere of the Caribbean Plate...

- between the Americas (Salvador & Green 1980; Anderson & Schmidt 1983; Klitgord & Schouten 1986; Donnelly 1989; James 1990),
- in the Pacific (Dickinson & Coney 1980; Pindell & Dewey 1982; Duncan & Hargraves 1984; Pindell & Barrett 1990; Pindell 1990).

Finally, Morris et al. (1990), Pindell (1994) and Stanek (2000) list some problems of the cited models and further touchstones for any reconstruction of the Caribbean Plate. Thus, tectonic reconstructions of the Caribbean Plate are demanded to integrate several aspects:

- Rifting history
 - Reconstruction of Pangea, restoration of extension along passive margins, restoration of transcurrent / convergent faulting in the Americas, restoration of accreted arc terranes (Pindell 1994);
 - Opening of the Atlantic, the Gulf of Mexico (Pindell 1994), timing and dynamics of Proto-Caribbean rifting, and the formation of passive continental margins (Stanek 2000);
 - Motions of the Americas (Pindell 1994).
- Volcanic arc developments
 - Beginning of subduction between the Pacific / Atlantic plates and the formation of a primitive island arc (Stanek 2000);
 - Periods and polarities of subduction (Pindell 1994) and the role of a mid-Cretaceous unconformity in many of the Caribbean units (Stanek 2000).
- Collision processes
 - Timing and vergence of closure of former oceans / basins along the arc-continent suture zones (Pindell 1994; Stanek 2000);
 - Processes and timing of circum-Caribbean collisions during the Late Cretaceous and early Paleogene (Pindell 1994);
 - Links between these processes and an inferred eastward migration of the Caribbean Plate (Pindell 1994).

Further aspects concern:

- Reconfiguration of the Caribbean Plate during the early Cenozoic and the formation of east-west striking transform faults (Stanek 2000);

- Origin and age of the Cayman Trough (Morris et al. 1990);
- Histories of the Grenada Basin and the Yucatán Basin (Pindell 1994);
- Presence of Paleozoic and older rocks in Cuba (Morris et al. 1990);
- Origin and role of the Nicaragua Rise (Morris et al. 1990).

6.2 Plate-kinematic framework

In simple cases, restorations of current oceanic basins can be achieved by realigning coeval magnetic anomalies along fracture zones. In fact, magnetic anomalies of the Caribbean oceanic crust have been proposed for the Venezuelan, and more tentatively for the Yucatán Basin (chapter 3.2.1, chapter 3.3.4). However, a closure of the Caribbean realm can not be reconstructed by simply realigning the pairs of coeval magnetic anomalies. Large portions of the Caribbean oceanic crust have been altered and thickened by extensive basaltic magmatism during the Late Cretaceous (chapter 3.2). Much of the geological record has been destroyed by subduction. Moreover, transform fault displacements indicate, that the Caribbean Plate may not have formed in its present day relative position.

In contrast, the history of seafloor spreading of the Atlantic Ocean is well documented from magnetic anomaly and fracture zone data (e.g. Bullard et al. 1965; Heirtzler et al. 1968; Le Pichon & Fox 1971; Sclater et al. 1977; Müller et al. 1999). Therefore, former relative positions of North and South America have been determined mainly by finite difference solutions for the three plate system of South America, Africa and North America (Fig. 26; Ladd 1976; Sclater et al. 1977; Pindell & Dewey 1982; Pindell et al. 1988).

The rates of relative motion between North and South America define an era of separation from the Early Jurassic to the Aptian and slow convergence since the Middle Eocene (Fig. 26, Pindell 1994). These relative motions provide the plate-kinematic framework for the evolution of the Caribbean region and define the size and shape of the Caribbean realm at stages through time.

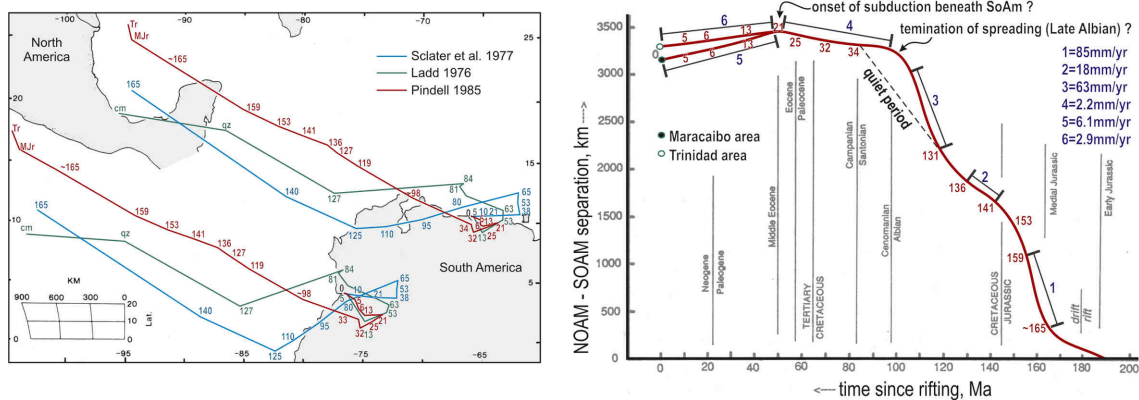


Fig. 26: Post-Triassic relative motion histories for two points (10.5°N, 66°W and 4°N, 77°W) along South America, with respect to North America, from three data sets (Sclater et al. 1977; Ladd 1976; Pindell 1985b; after Pindell & Barrett 1990). The chart on the right side shows rates of relative motion for North-South America from the vector of Pindell (1985b) in Pindell (1994). The relative positions through time were computed from vector addition of the Atlantic rotation parameters in the three-plate system North America – Africa – South America and by anomaly identification. The vector of Sclater et al. (1977) is based on rotation parameters defined by age-depth curve in the ocean basins. The numerals indicate age in millions of years.

6.3 Western Pangea reconstructions

Triassic – Early Jurassic reconstructions of western Pangea show, that the Caribbean region did not exist at this time (Fig. 27). Therefore, these reconstructions are the natural starting point for tectonic models of the Caribbean.

Tightness of continental fit in the Triassic – Early Jurassic reconstructions has improved significantly over the last decades (Fig. 27; e.g. Bullard et al. 1965; Le Pichon & Fox 1971; Van der Voo & French 1974; Pindell 1985a; Ross & Scotese 1988; Pindell et al. 2006). Mappings of marine magnetic anomalies and fracture zone traces in the Atlantic oceans allow the reconstruction of circum-Atlantic continents to accuracies of better than 50 km error for most anomalies (Klitgord & Schouten 1986; Pindell et al. 1988; Müller et al. 1999; Pindell et al. 2006). Further inferences i.a. come from

- onshore determinations of paleo-inclination trough time (Van der Voo & French 1974, Van der Voo et al. 1976),
- realignments of marginal offsets from opposing continental margins (Le Pichon & Fox 1971),
- realignment of the Demerara Rise of northeastern Brazil and the Guinea Plateau of Mauritania and Liberia (Masclé et al. 1986),

- considerations of strike-slip, rotations of continental blocks as well as intracontinental extension and shortening (e.g. Pindell 1985a; Dunbar & Sawyer 1986).

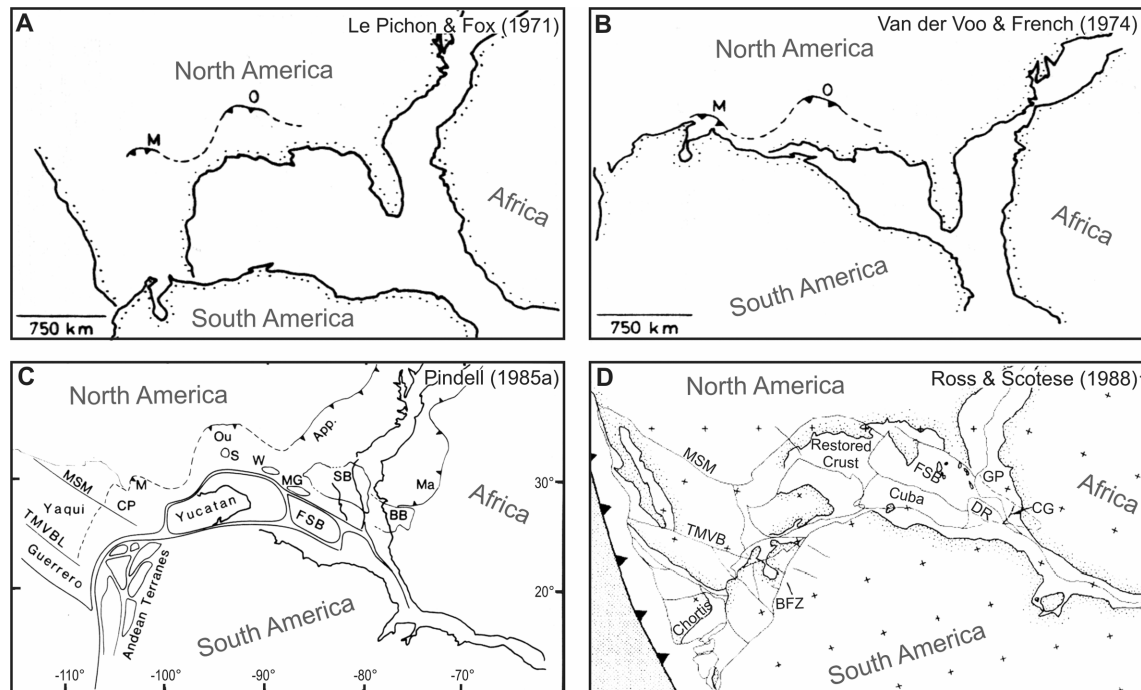


Fig. 27: (A) Mesozoic continental reconstruction (Le Pichon & Fox 1971 in Burke et al. 1984) based on alignment of opposing marginal fracture zones. (B) Mesozoic continental reconstruction (Van der Voo & French 1974 in Burke et al. 1984), based on paleomagnetic data from circum-Atlantic continents (C) Triassic-Early Jurassic reconstruction of circum-Atlantic continents (Pindell 1985a, in Pindell & Barrett 1990), based on paleomagnetism, fracture zones, and considerations of strike-slip, rotations of continental blocks as well as intracontinental extension and shortening. App = Appalachians; BB = Bove Basin; CP = Coahuila Peninsula; FSB = Florida Straits Block; M = Marathons; Ma = Mauritanides; MG = Florida Middle Grounds; MSM = Mojave-Sonora Megashear; Ou = Ouachitas; S = Sabine Uplift; SB = Suwannee Basin; TMVBL = Trans-Mexican Volcanic Belt; W = Wiggins Arch. (D) Middle Jurassic reconstruction after Ross & Scotese (1988), modified after Pindell (1985a). MSM = Mojave-Sonora Megashear; TMVB = Trans-Mexican Volcanic Belt; BFZ = Bocono Fault Zone; FSB = Florida Straits Block; DR = Demerara Rise; GP = Guinea Plateau; CG = Cat Gap.

Tight fitting reconstructions (e.g. Pindell 1985a, Ross & Scotese 1988) place the Bahamas-type basement of Cuba adjacent to the northeastern margin of South America. The southeastern limit of the Bahamas Platform overlaps with Africa, which is ascribed to a probable volcanic origin in relation to Atlantic seafloor spreading. The formation of this basement would therefore postdate the opening of the central Atlantic (e.g. Ross & Scotese 1988, Pindell et al. 2006). Some authors also infer an eastward

transform migration of south Florida and the Bahamas during the opening of the Gulf of Mexico (Pindell 1985a, Pindell et al. 1988; Pindell & Kennan 2001).

A tight fit between the Americas and the Maya Block is achieved by restoring the volume of stretched crust in the Gulf of Mexico (Dunbar & Sawyer 1986), a clockwise rotation of the Maya Block by 40°-50° from its present position (Pindell & Dewey 1982; Pindell 1985a; Molina-Garza et al. 1992; Pindell & Kennan 2001; Steiner et al. 2005) and the breaking of Mexico along several major strike-slip faults (e.g. Pindell & Dewey 1982).

Ross & Scotese (1988) place the Chortís Block west of the Guerrero Block of Mexico based on westward extrapolation of the strike-slip offset along northern Caribbean Plate boundary (chapter 3.1, chapter 7.5).

6.4 Breakup of Pangea and the Caribbean spreading system

During the Middle Jurassic, North America rifted from Africa and South America, opening the Central Atlantic (Klitgord & Schouten 1986). All reconstructions assume a propagation of the Central Atlantic rift along transform boundaries into the Gulf of Mexico (Fig. 28). Some place a transform boundary along the northern Bahamas, (e.g. Burke et al. 1984; Mann 1999), some to the south (e.g. Meschede & Frisch 1998), while others place transform boundaries to the north and to the south of the Florida Straits Block (e.g. Pindell 1994; Ross & Scotese 1988; Stanek 2000).

Middle Jurassic salt deposits in Gulf of Mexico point to regional attenuation and subsidence of continental crust (Buffler et al. 1980; Salvador 1997). During the Middle Jurassic the salt deposits were overlain by deep water clastics. Buffler & Sawyer (1985) conclude that generation of oceanic crust in the central Gulf of Mexico commenced during the Middle Jurassic. The youngest ocean floor in the gulf is assumed to have formed around 145 -140 Ma (Burke et al. 1984; Ross & Scotese 1988).

Many authors propose that rifting between the Maya Block and South America did not begin until the Gulf of Mexico was completely formed (Klitgord et al. 1984; Ross & Scotese 1988; Pindell et al. 1988).

However, the San Cayetano Formation in the Cordillera Guaniguanico in western Cuba (chapter 4.1.1) indicate, that Early and Middle Jurassic rift basin deposits south of the Maya Block developed contemporaneously to those in the Gulf of Mexico (Hutson et al. 1998, Pszczółkowski 1999; Stanek 2000). The Oxfordian transition to marine carbonates in the Cordillera Guaniguanico is accompanied by bimodal volcanic rocks and the El Sábolo Formation basalts (Kerr et al. 1999).

Moreover, a maximum K-Ar age of 160 ± 24 Ma was obtained from anorthosites sampled in the Northern Ophiolite Belt in Cuba (Somin & Millán 1981; Iturralde-Vinent et al. 1996). The tectonic position of the NOB and its age data (chapter 4.1.2) suggest that the northern ophiolites of Cuba most probably evolved from Jurassic to Early

Cretaceous ocean floor which occupied an oceanic realm south of the Bahamas rift/passive continental margin. Remnants of this ocean floor were later obducted during the collision of the northeast-facing CVA with the southwestern margin of the Bahamas Platform (chapter 7.1, chapter 7.2).

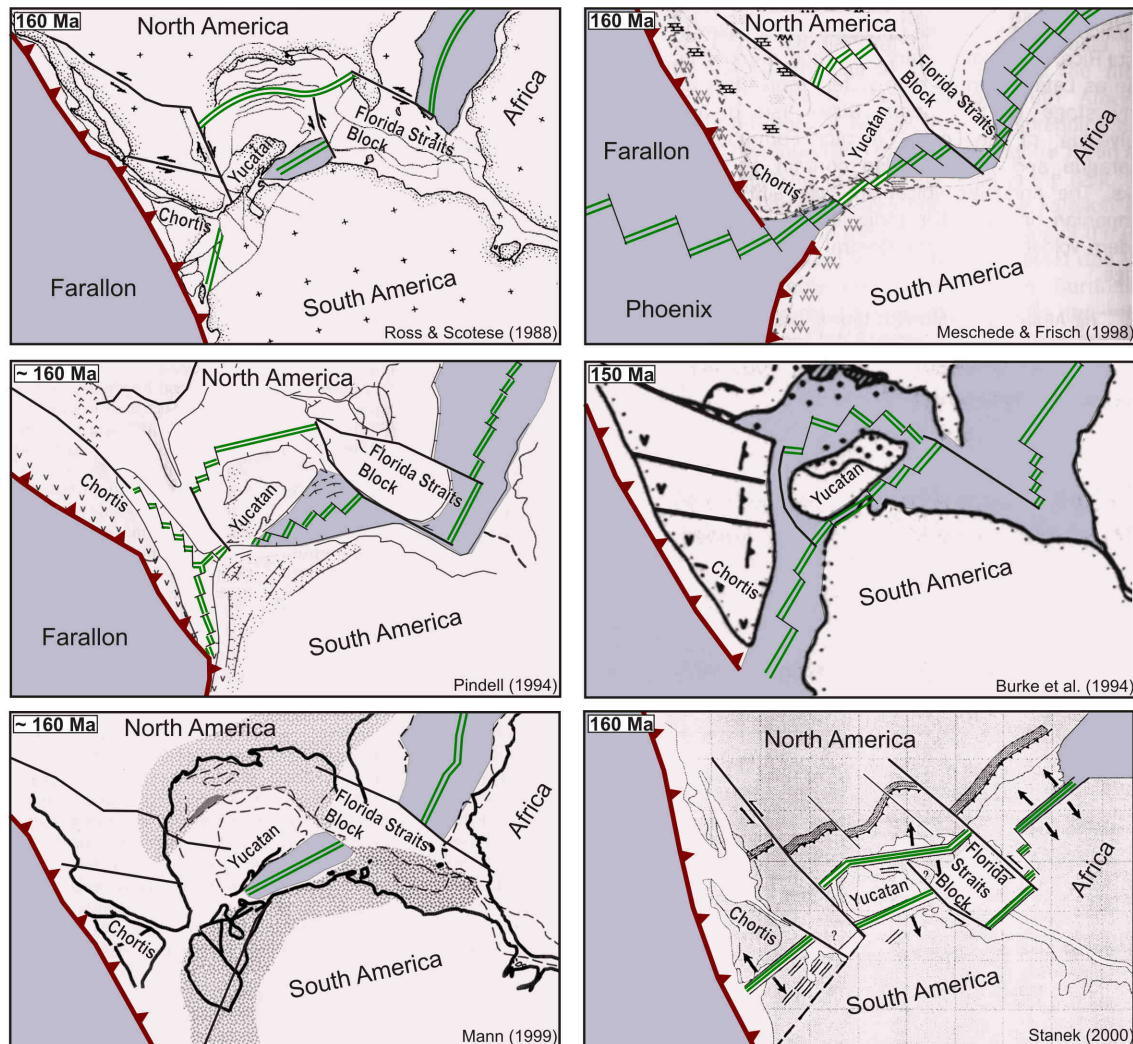


Fig. 28: Comparison of Callovian/Oxfordian reconstructions of the Caribbean region (compiled from Pindell 1994; Burke et al. 1984; Ross & Scotese 1988; Meschede & Frisch 1998; Mann 1999; Stanek 2000).

The above stated evidence is consistent with a simultaneous formation of Jurassic continental rifts and ocean floor at two spreading centers: One in the Gulf of Mexico, the other located between the Maya / Florida Straits Blocks and South America (Caribbean spreading system).

A westward continuation of the Caribbean spreading system is interpreted differently (Fig. 28). Some models argue in favor of an eastward dipping Jurassic subduction system along the entire western coast of Central America (Fig. 28), either truncating the Caribbean spreading system to the west (e.g. Ross & Scotese 1988; Mann 1999; Stanek 2000), or connecting it to back-arc spreading centers of central Mexico and/or offshore Colombia (e.g. Pindell 1994, Pindell & Kennan 2001).

Gosh et al. (1984) proposed a model for magnetic anomalies in the Venezuelan Basin (Fig. 4; chapter 3.2.1). They assume, that these anomalies would trace Jurassic to Early Cretaceous rifting much more westerly from the present day position of the Venezuelan Basin, thus in a westward continuation of the Caribbean spreading system into the Farallon-Phoenix rift. Interpretations of the magnetic anomalies in the Venezuelan Basin, however, are controversial. Other interpretations relate them to a dyke swarm or to a linear deformation pattern (Donnelly 1994).

Frisch et al. (1992) determined a paleomagnetic drift path for the Central American ophiolites. Accordingly, these ophiolites evolved from Jurassic ocean floor generated in a westerly equatorial position between the North and South American continental blocks. Therefore, the formation of the Venezuelan Basin oceanic crust must also be located in the Caribbean spreading system, somewhat more to the east than it was assumed by Gosh et al. (1984). Nevertheless, the results of Frisch et al. (1992) and the interpretations by Meschede & Frisch (1998) support the idea of a westward continuation of the Caribbean spreading system into the Farallon-Phoenix rift.

6.5 Inception of the Caribbean Plate

During the Early Cretaceous, Proto-Caribbean oceanic crust continued to form as South and North America separated (Fig. 26, 29).

Passive margins extended along northern South America and the southern margin of Chortís (Ross & Scotese 1988). Passive margin sequences of the eastern Maya Block and the southern Bahamas Platform are preserved within the Cordillera Guaniguanico and the thrust belt of central Cuba (chapter 4.1.1).

According to the interpretation of Gosh et al. (1984), seafloor spreading in the Venezuelan Basin ceased at ~127 Ma, but the rates of relative motion indicate rapid separation of the Americas until ~100 Ma (Fig. 26; Pindell 1985b).

6.5.1 Onset of subduction along the northeastern Caribbean margin

An early indication for the beginning of subduction along the northeastern Caribbean realm is represented by the Mabujina amphibolites on Cuba. The protoliths of the Mabujina unit are interpreted as primitive volcanic arc rocks and may be as old as 130 Ma (Millán & Somin 1985; Stanek 2000; Blein et al. 2003; Bibikova et al. 1988 in Stanek et al. 2006; Rojas-Agramonte et al. 2005). Subduction was active at least as far back as Aptian, as indicated by the magmatic development in central Cuba (chapter 4.1.3; Iturralde-Vinent 1996c, 1998; Díaz de Villalvilla et al. 1997; Kerr et al. 1999; Stanek 2000) and structural/metamorphic data from the Escambray Massif (chapter 4.1.4; Stanek et al. 2006).

On Jamaica, the oldest primitive arc volcanics are of Barremian age (~ 115 Ma; Mitchell 2003, 2006) and on Hispaniola, primitive island arc volcanics are of Aptian-Albian age (El Seibo unit, Los Ranchos Formation; Donnelly et al. 1990a, Kesler et al.

1991; Kesler et al. 2005). The metamorphosed mafic rocks of the Duarte Complex may represent even earlier products of volcanic arc activity on Hispaniola. The mafic rocks of the Duarte Complex have been intruded by Albian granitoid stocks (Draper et al. 1994), so that their age might correspond to the protolith ages of the Mabujina amphibolites.

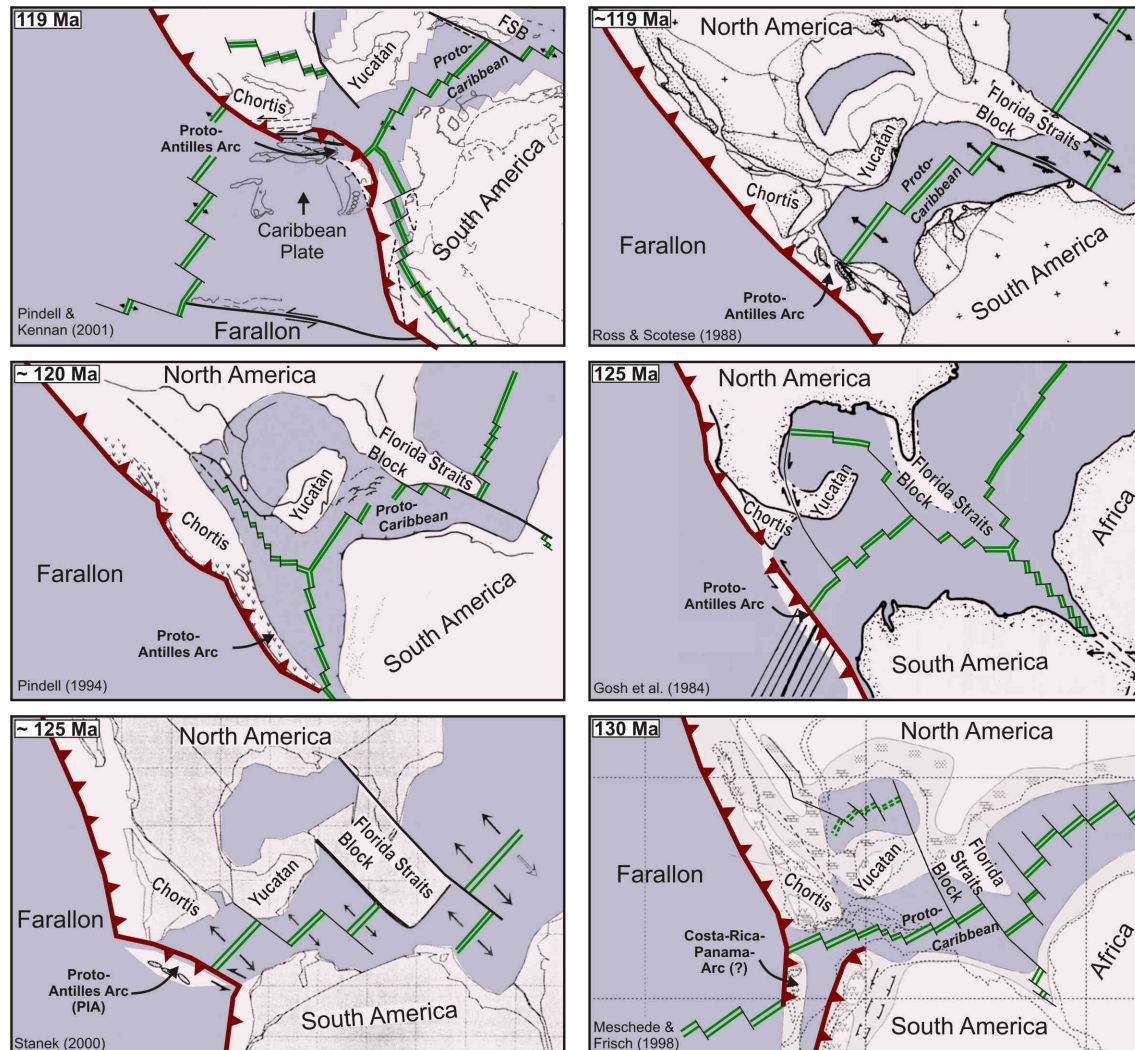


Fig. 29: Comparison of Early Cretaceous reconstructions of the Caribbean region (compiled from Gosh et al. 1984; Pindell 1994; Ross & Scotese 1988; Meschede & Frisch 1998; Stanek 2000; Pindell & Kennan 2001).

In summary, the magmatic histories of Jamaica, Cuba and Hispaniola point to a beginning of subduction and the formation of a common Greater Antilles Arc antecessor at sometime during the Early Cretaceous.

6.5.2 Concepts of the northeastern Caribbean margin

The vast majority of mobilistic plate tectonic models concur with a Meso- and Cenozoic eastward migration of the northeastern Caribbean margin relative to the Americas up to its presentday position at the Lesser Antilles Arc. During the last decades the debate was often times tapered by the controversy about whether the Caribbean is of Pacific

or intra-American origin. However, supporters of both sides share common aspects and the proposed concepts are obviously more differentiated. Thus, there are several hypotheses about the Cretaceous development of the northeast-east bounding arc of the Caribbean, often referred to as the “Great Caribbean Arc” (GCA).

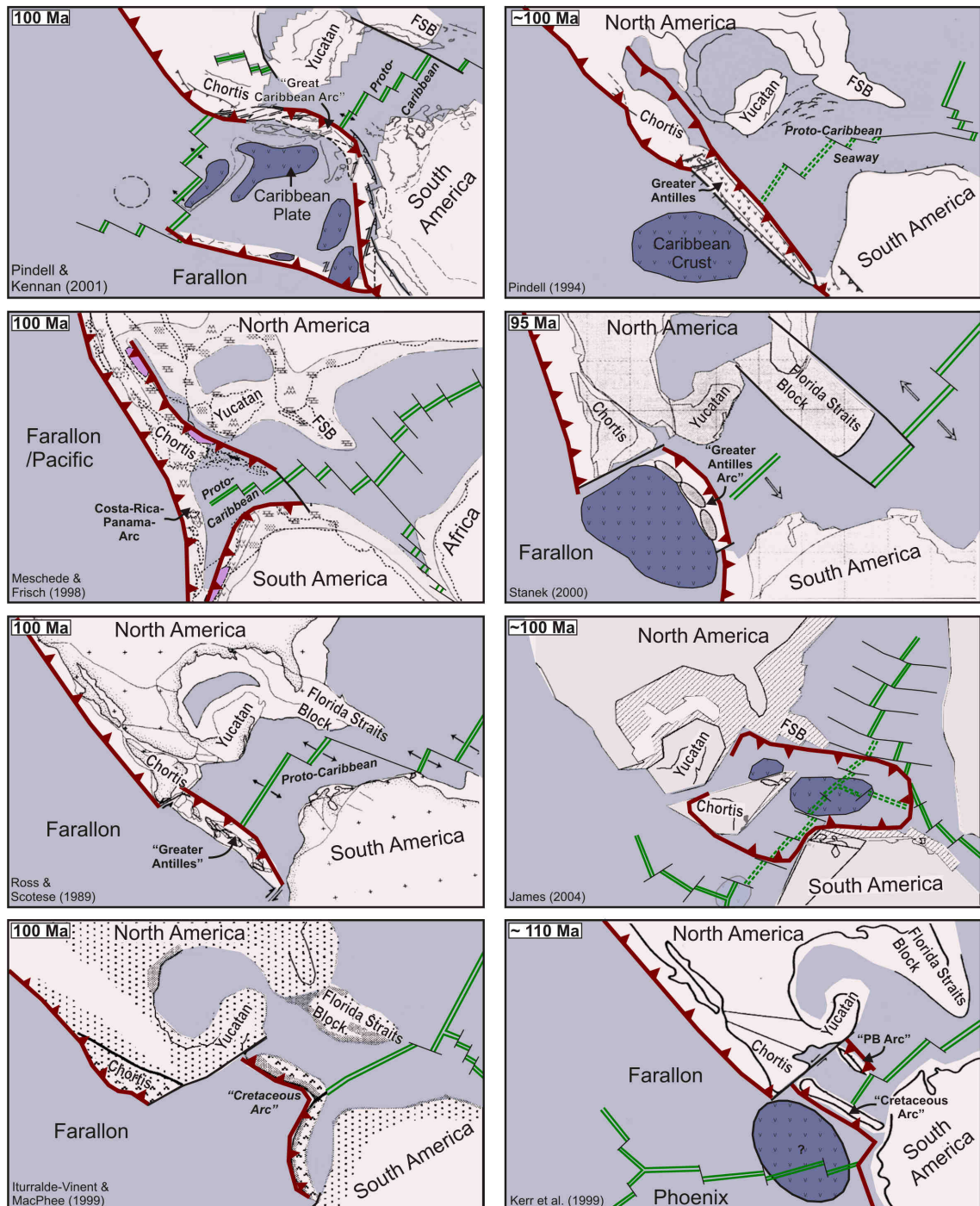


Fig. 30: Comparison of Albian reconstructions of the Caribbean region (compiled from Pindell 1994; Ross & Scotese 1988; Meschede & Frisch 1998; Iturralde-Vinent & MacPhee 1999; Stanek 2000; Pindell & Kennan 2001; James 2004).

With regard to neutral terminology and regional geology, in previous chapters the northwestern portion of the GCA was referred to as “Cretaceous volcanic arc” (CVA). The term “Great Caribbean Arc” is referring to the coherency of the entire northeastern and eastern active margin of the Caribbean Plate, of which former elements may range into the Netherlands- and Venezuelan Antilles along the northern edge of South America.

“Cordilleran” starting points of the “Great Caribbean Arc”

Most interpretations place the Early Cretaceous “Proto-Caribbean Arc” along the western flank of the Americas (Fig. 29). Four different types of models with this common starting point are distinguishable:

- 1) One group of models starts with an **east-dipping “Cordilleran” subduction zone** and assumes a **Campanian polarity reversal** driven by the arrival of the buoyant Caribbean Large Igneous Province (CLIP) along the western fore-arc (e.g. Burke 1988; Duncan & Hargraves 1984, Kerr et al. 1999; Fig. 31). According to these models, the **CLIP** (chapter 3.2) **was extruded** onto Pacific-derived Caribbean lithosphere as it passed **over the Galapagos hotspot**.

Kerr et al. (1999) extend this concept by introducing a second Early Cretaceous arc, a north-facing “primitive boninite arc” which was located near the southern edge of Yucatán (Fig. 30).

Contradictions to these models have been summarized by Meschede (1998) and Pindell et al. (2006). Two major contradictions concerning the Galapagos origin of the CLIP are depicted here:

- The oldest rocks which are undoubtedly related to the Galapagos hotspot are 15-20 Ma old (Christie et al. 1992, Protti et al. 1994), but the most extensive basaltic magmatism in the Caribbean was apparently active at ~90 Ma (Sinton et al. 1998).
- At present day the Caribbean Plate is moving towards the west with respect to the hotspot reference frame (DeMets et al. 1990). Its slower westward motion compared to the Americas is responsible for the relative eastward movements along the northern and southern boundaries of the Caribbean Plate. Extrapolating these motion vectors to the past increases the distance between the Galapagos hotspot and the Caribbean Plate, rather than letting it pass over the hotspot.

Further difficulties of these models arise from the evolution of the Costa Rica – Panamá - Arc (chapter 6.5.3) and constraints coming from the paleomagnetic drift path of the Central American ophiolites (Frisch et al. 1992).

The assumed Campanian polarity reversal due to collision of the CLIP is objected by evidence from the northwestern Caribbean, according to which the CLIP did not

arrive at the southwestern fringe of the GCA until about the middle Eocene when the northern Caribbean transform margin was established (chapter 7.5).

- 2) Another group of models also starts with an **east-dipping “Cordilleran” subduction zone** and assumes an **Aptian-Albian arc reversal** (e.g. Pindell 1994; Ross & Scotese 1988; Pindell & Kennan 2001, Pindell et al. 2006; Fig. 29, 30).

The assumed pre-Aptian east-dipping polarity of the GCA is mainly based on geochemical characteristics of some basalt samples from the Northern Ophiolite Belt (chapter 4.1.2), according to which they may have formed in a back-arc basin (Fonseca et al. 1990; Iturralde-Vinent 1996b; Kerr et al. 1999).

The following main arguments are requested for an Aptian-Albian reversal of polarity:

- Albian to Cenomanian limestone horizons occur in the volcanic arc sections throughout the Greater Antilles, sometimes unconformably with a basal conglomerate. These roughly correlate to a change in the geochemical signature of the arc from the primitive pre-Aptian stage to the mainly calc-alkaline Aptian to Cenomanian phase (chapter 4.1.3);
- there are mainly Aptian-Albian age data for HP-LT metamorphism;
- there are indications for pre-Aptian deformations in the Maimon and Los Ranchos Formations on Hispaniola.

However, these arguments do not procure the definitive inference of an Aptian-Albian arc reversal:

- The presumed back-arc origin of some NOB basalts is based on geochemical characteristics of a few basalt samples and lacks back up by any other findings concerning tectonics or regional geology.
- Unconformities and a change from primitive to calc-alkaline characteristics may be found in the history of any volcanic arc. Besides a polarity reversal, these effects can be attributed to changes of any other parameter in the subduction / volcanic arc system.
- Structural/metamorphic data from the Escambray Massif are consistent with continuous southwest-dipping subduction from at least Aptian-Albian to Cenomanian time (e.g. Stanek et al. 2006; chapter 7.1), but do not indicate a polarity reversal during the early history of the arc.

- 3) A third type of model with a “Cordilleran” starting point is proposed by Gosh et al. (1984) and Stanek (2000), who assume a **west-dipping subduction right from the inception of the GCA** (Fig. 29).

- 4) The fourth type is proposed by Iturralde-Vinent & MacPhee (1999), according to which the **GCA** and its subsequent “Paleogene Arc” **never faced west- and southwest-dipping subduction** until the early Eocene, when the Lesser Antilles Arc came into existence (Fig. 30, 33, 36, 39). This model fundamentally conflicts

the clear evidence from the Cuba orogenic belt for a northeast-facing Cretaceous volcanic arc (chapter 7.1).

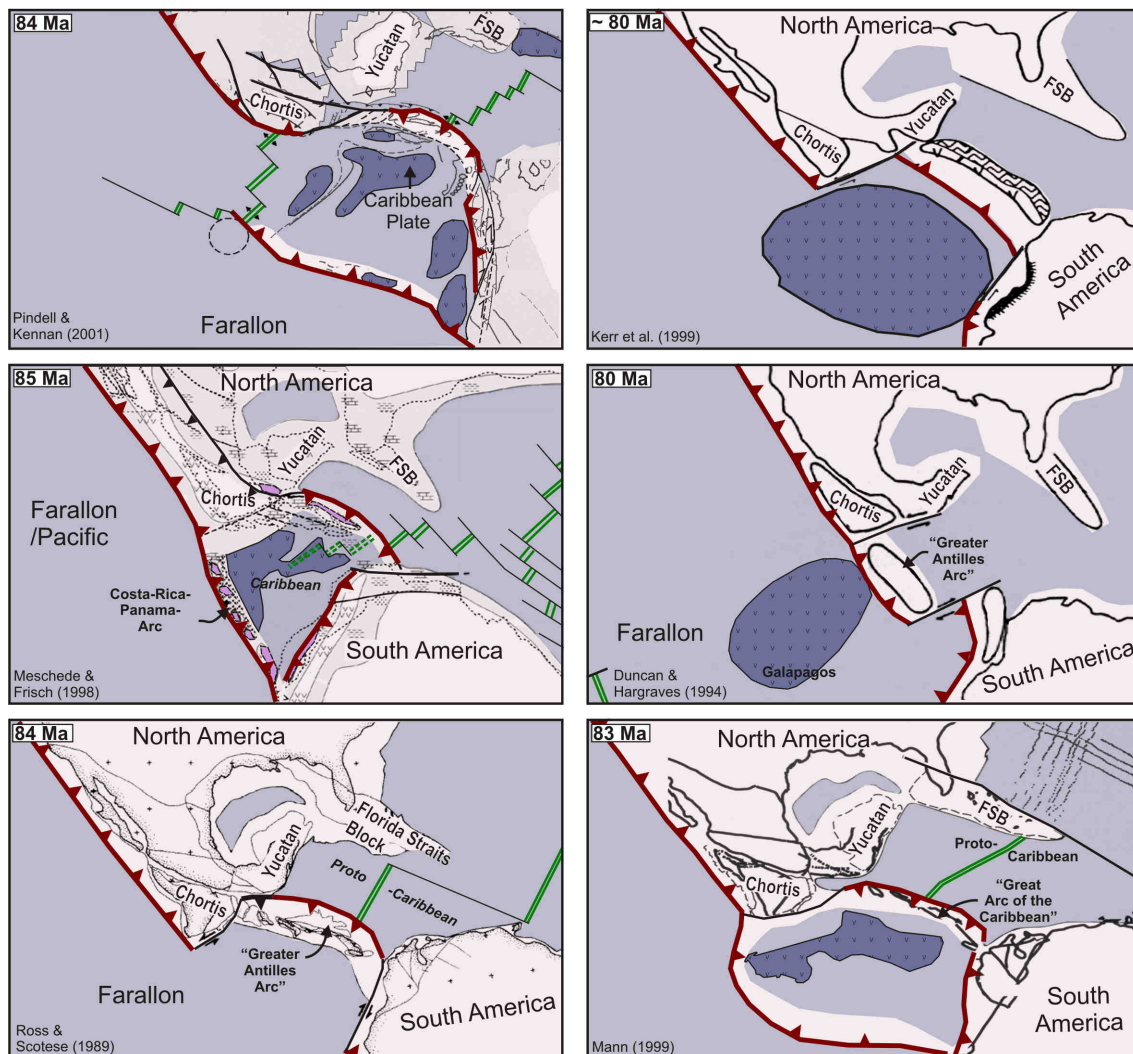


Fig. 31: Comparison of Santonian / early Campanian reconstructions of the Caribbean region (compiled from Duncan & Hargraves 1984; Ross & Scotese 1988; Meschede & Frisch 1998; Kerr et al. 1999; Mann 1999; Pindell & Kennan 2001).

Northern Chortís inception of the “Great Caribbean Arc”

Donnelly (1989) and Meschede & Frisch (1998) propose an alternative model to the “Cordilleran”-type reconstructions. Instead of an inception of the GCA along the western flank of the Americas, the early arc was initiated by southwest-dipping subduction between the northern fringe of the Chortís Block and South America. It may have had a northwestern continental continuation along the ophiolite complexes of Guatemala, along the northeastern front of the Juarez complex in southern Mexico, and into the ophiolite complexes at the northeastern margin of the Guerrero Block in Central Mexico (Fig. 30; Burkart 1994; Tardy et al. 1994; Freydier et al. 1996, 1997; Meschede & Frisch 1998). The continental continuation of the arc was later cut off by the Campanian collision of the Chortís and Maya Blocks, while the intra-oceanic eastern part of the GCA was approaching the Bahamas Platform (Fig. 31).

6.5.3 Southwestern Caribbean margin

Remnants of early magmatism in the Costa Rica - Panamá - Arc are only preserved in middle to early Late Cretaceous clastic erosional products. However, most workers conclude, that these remnants indicate at least an Albian age for the onset of subduction at the Costa Rica - Panamá - Arc (Calvo & Bolz 1994; Ehrlich et al. 1996; Meschede & Frisch 1998; Pindell & Kennan 2001; Calvo 2003; Flores et al. 2005; Pindell et al. 2006).

Hardly any of the hitherto Caribbean models integrate an island arc at this time at the southwestern boundary of the Caribbean Plate. The middle Cretaceous onset of subduction at the southwestern boundary of the Caribbean is crucial to all models with a "Cordilleran" starting point for the GCA (chapter 6.5.2). Furthermore, the middle Cretaceous predecessor of the Costa Rica - Panamá - Arc raises another objection against the "Galapagos Plateau" concept for the origin of the CLIP (e.g. Burke 1988; Duncan & Hargraves 1984; Kerr et al. 1999; chapter 6.5.2). If this concept would be correct, then the Galapagos hot spot would have had to migrate westward across the trace of the southwestern Caribbean subduction zone / plate boundary to get to its present position west of the Caribbean Plate (Pindell et al. 2006).

The models of Meschede & Frisch (1998) and Pindell & Kennan (2001) account, however in different ways, for Albian subduction along the southwestern Caribbean margin.

As Meschede & Frisch (1998) interpret the GCA to be initiated at the northern fringe of the Chortís Block (chapter 6.5.2), their Albian reconstruction shows a "Cordilleran"-type Costa Rica - Panamá - Arc between the southwestern edge of the Chortís Block and South America (Fig. 30).

Pindell & Kennan (2001) assume a "Cordilleran" starting point of the GCA and therefore place the predecessor of the Costa Rica - Panamá - Arc farther out into the Pacific constituting the southern margin of the Caribbean Plate in the Albian. Its western plate boundary is defined by the hypothetical Kula-Caribbean spreading center that connects Northern Costa Rica and the Chortís Block (Fig. 30).

Both models may have their strengths and weaknesses. The model by Meschede & Frisch (1998) for example does not leave much room for the shape of the Caribbean Plate during the Cretaceous, especially when subducted crust is extracted from beneath South America and from the Paleogene volcanic arc subduction zone in the northwestern Caribbean (chapter 7.3).

On the other hand, Pindell & Kennan (2001) neglect the appealing concept of a northwestern continuation of the GCA into the ophiolite complexes at the northeastern margin of the Guerrero Block in Central Mexico (Fig. 30; Burkart 1994; Tardy et al. 1994; Freyrier et al. 1996, 1997; Meschede & Frisch 1998), while their argumentation in favor of an east-dipping "Cordilleran" onset of the GCA is not fully conclusive (chapter 6.5.2).

7 Tectonic reconstruction of the northwestern Caribbean

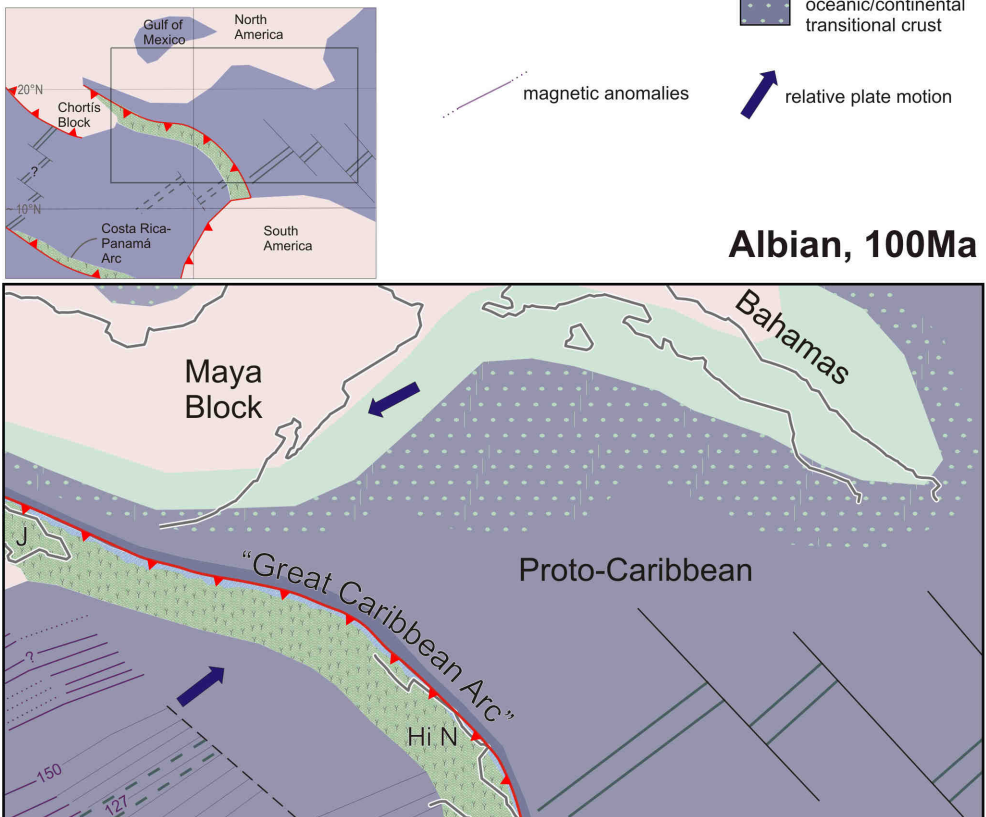
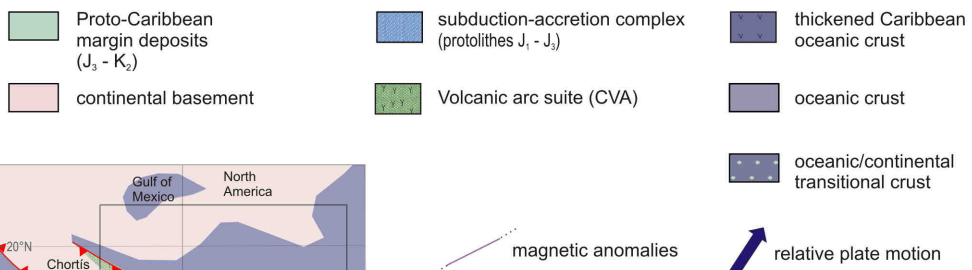
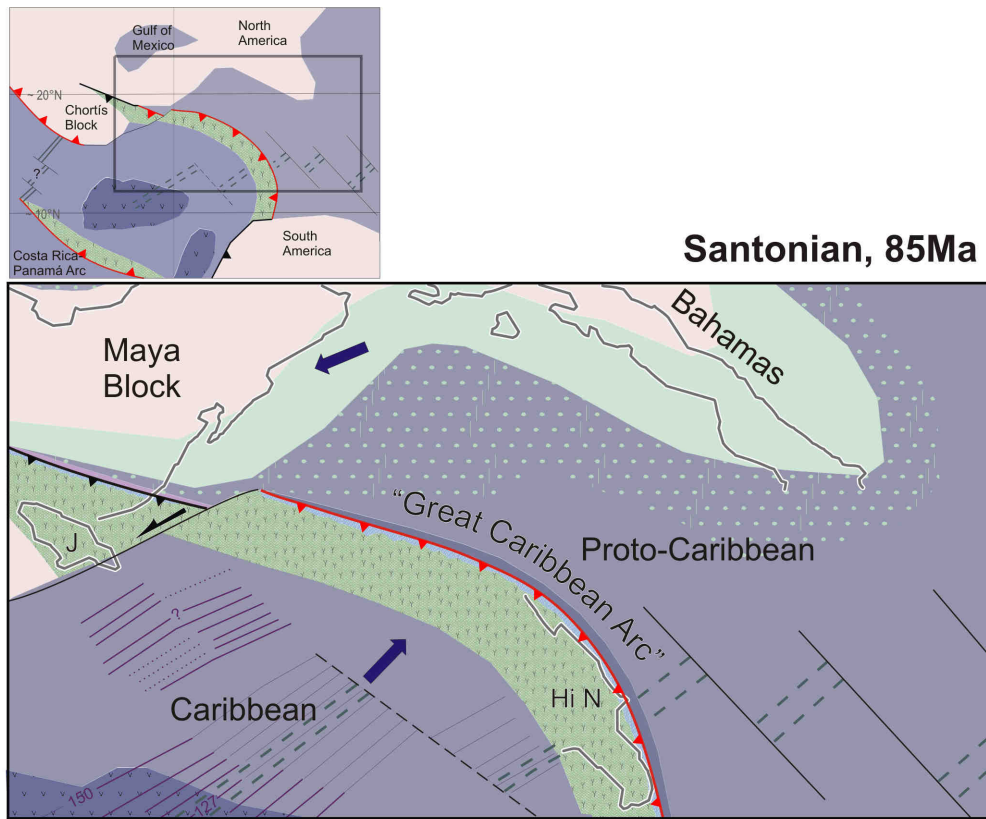
This chapter focuses on detailed Late Cretaceous to early Miocene tectonic reconstructions for the northwestern Caribbean. The proposed reconstructions are based on the regional geological data compiled in chapters 3 and 4 as well as the constructed cross sections presented in chapter 5. With regard to plausibility, it was attempted to embrace a high level of detail in the proposed reconstructions, in map view as well as in cross section.

7.1 Polarity of the “Great Caribbean Arc”

The Albian starting point of the tectonic reconstruction of the northwestern Caribbean (Fig. 32) requires southwest-dipping subduction along the GCA in order to satisfy the following aspects:

- A southwest-dipping trench must have existed in order for the arc to be able to approach the American margin.
- Furthermore, a southwest-dipping polarity is required in order to explain the magnitude of top to the north directed tectonic transport in the Cuba orogenic belt. It is difficult to see how low angle thrust nappes would propagate for great distances onto to southern shoulder of the Bahamas Platform (chapter 7.2) if it would be the upper plate of the subduction system, as suggested by Iturralde-Vinent (1994, 1996a) and Iturralde-Vinent & MacPhee (1999).
- Metamorphic fore-arc assemblages on Hispaniola are situated along the north side of the GCA range (Fig. 7; chapter 3.3.2). From their internal structures and tectonic position, there is little doubt about the fact that the metamorphic fore-arc assemblages on Cuba (Escambray Massif, Isla de la Juventud) also evolved on the north side of the arc, but were later overridden by arc units from the south (4.1.4 Metamorphic complexes; Fig. 23).

Fig. 32: (Next page) Albian and Santonian tectonic reconstructions of the northwestern Caribbean. Hi N = northern Hispaniola; J = Jamaica.



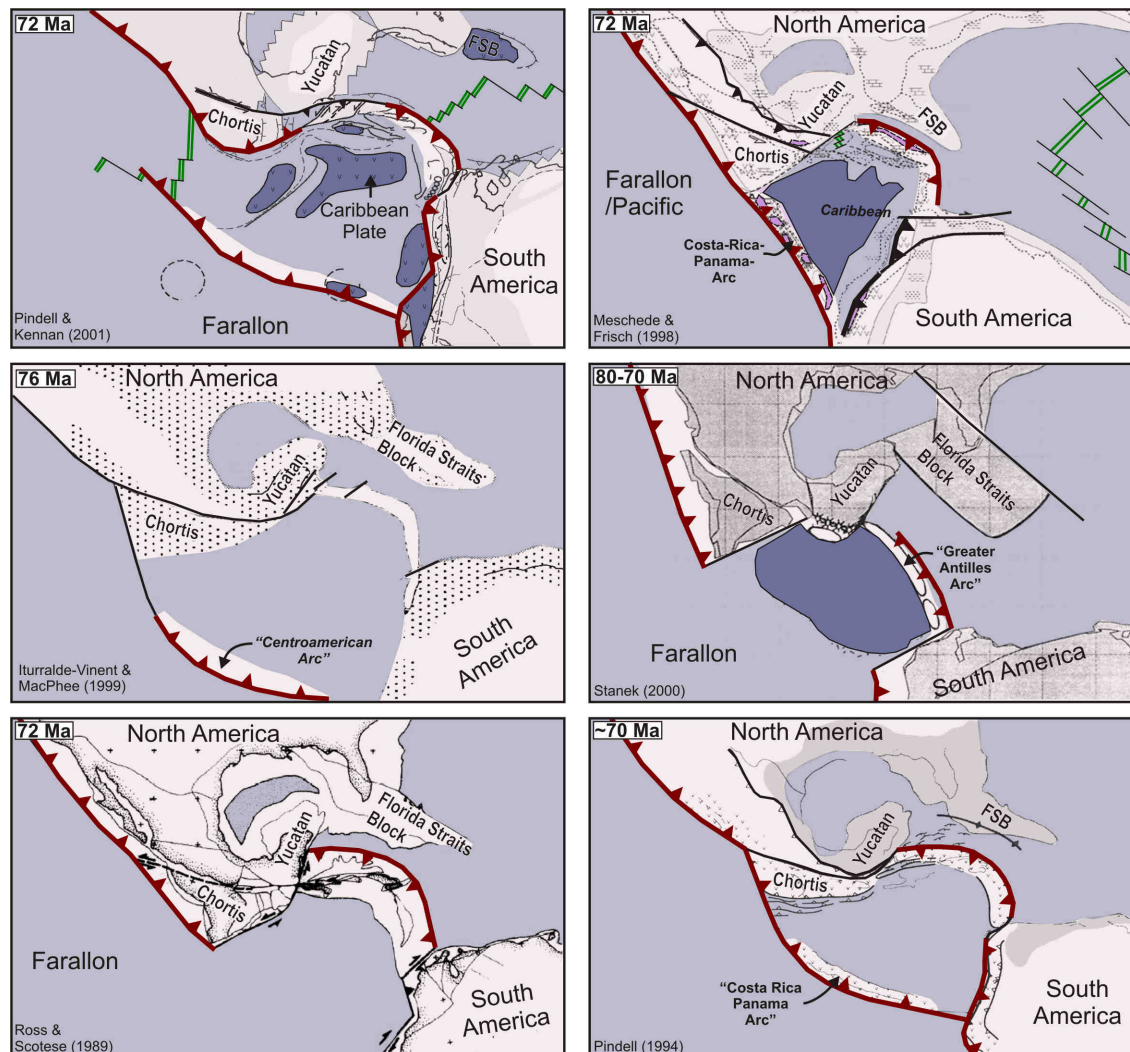


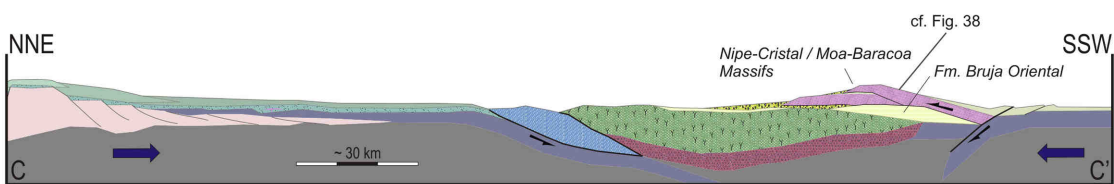
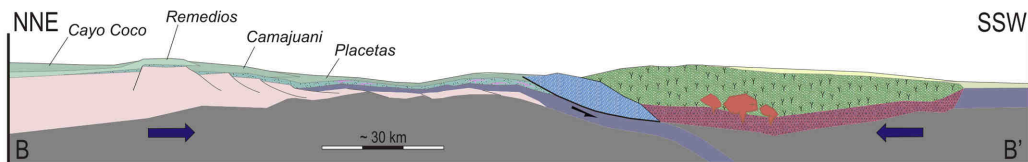
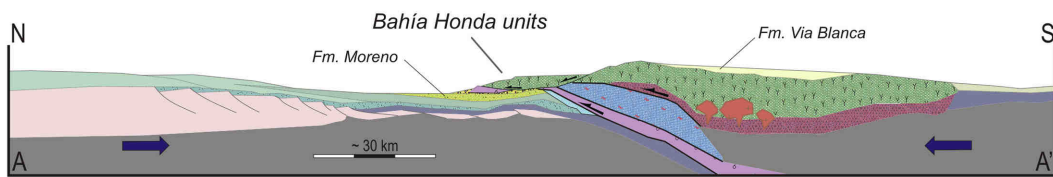
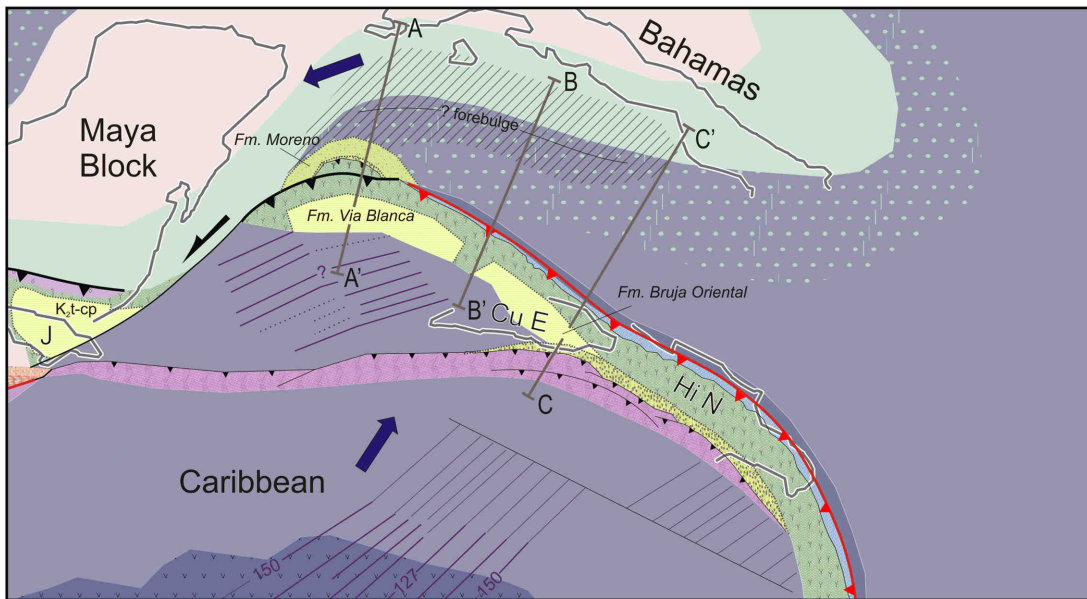
Fig. 33: Comparison of Campanian / late Campanian reconstructions of the Caribbean region (compiled from Pindell 1994; Meschede & Frisch 1998; Ross & Scotese 1988; Iturralde-Vinent & MacPhee 1999; Stanek 2000; Pindell & Kennan 2001).

7.2 Collision of the “Great Caribbean Arc”

The Albian reconstruction (Fig. 32) adopts the “northern Chortís” concept for the GCA (chapter 6.5.2). This situation suits the evidence from Jamaica, where the Turonian termination of the “older volcanic sequence” (chapter 3.3.1; Mitchell 2003, 2006) may correlate with the onset of collision between the continental part of the GCA along the northern Chortís Block and the Maya Blocks. Hence, the middle Campanian shallowing event in the overlying Crofts Synthem may have been triggered by the collision, while the rapid subsidence documented by the late Campanian section would correlate to the tearing of the remaining intra-oceanic part of the arc (Fig. 32).

Fig. 34: (Next page) Late Campanian tectonic reconstruction of the northwestern Caribbean. Cu E = eastern Cuba; Hi N = northern Hispaniola; J = Jamaica.

Late Campanian, 72Ma



- | | | |
|--|---|---|
| hemipelagic deposits
turbidites, volcanoclastic debris
Fm. Moreno | Cretaceous volcanic arc suite (CVA)
(K ₁ - K ₂) | sedimentary "cover" of CVA |
| North American
Paleomargin deposits
(J ₃ - K ₂) | granitoids | Nipe-Cristal / Moa-Baracoa
oceanic rocks |
| riftbasin deposits
(J ₁ -J ₃) | amphibolites
earliest products of CVA (J ₃ - K ₁) | olistostromes |
| continental basement | subduction-accretion complex
(protoliths J ₁ - J ₃) | relative plate motion |
| thickened Caribbean
oceanic crust | ophiolite obduction
(protoliths: ? J - K ₁) | magnetic anomalies |
| oceanic crust | oceanic/continental
transitional crust | |
| lack of Campanian
deposits | | |

Due to the latest Paleocene to Eocene age of syn-orogenic foreland deposits (Fig. 16), many workers assumed that the collision process between the arc and the Bahamas Platform did not start before the Paleocene (e.g. Bralower & Iturralde-Vinent 1997, Kerr et al. 1999). However, numerous indications point to the fact, that the GCA was actually approaching the southernmost extension of thinned continental crust of the Bahamas Platform by the late Campanian (Fig. 34):

As indicated by Ar-Ar cooling ages, magmatism in the Cuban portion of the GCA ended during the Campanian (Fig. 14; Hall et al. 2004). Uplift and erosion of the arc is approved by pressure-temperature-time paths from the Escambray Massif (Fig. 15; Stanek et al. 2006) and by Campanian-Maastrichtian terrigenous cover beds which locally crosscut granitoid massifs (Fig. 14; chapter 4.1.3; chapter 4.1.5; Cobiella-Reguera 2000). Furthermore, uplift and erosion of the arc can be inferred by the Campanian Moreno Formation in the Northern Rosario belt of western Cuba (Fig. 11). The Moreno Formation is the oldest passive margin unit that contains abundant volcanoclastic material eroded from the approaching GCA to the south (Fig. 34; Pszczółkowski 1999).

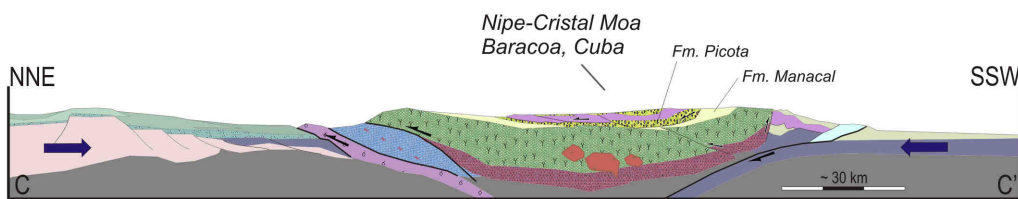
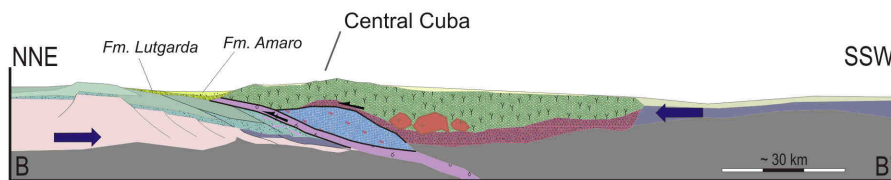
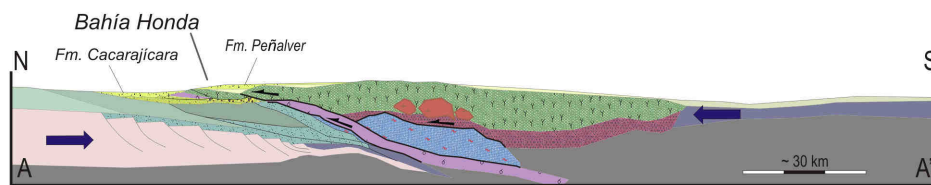
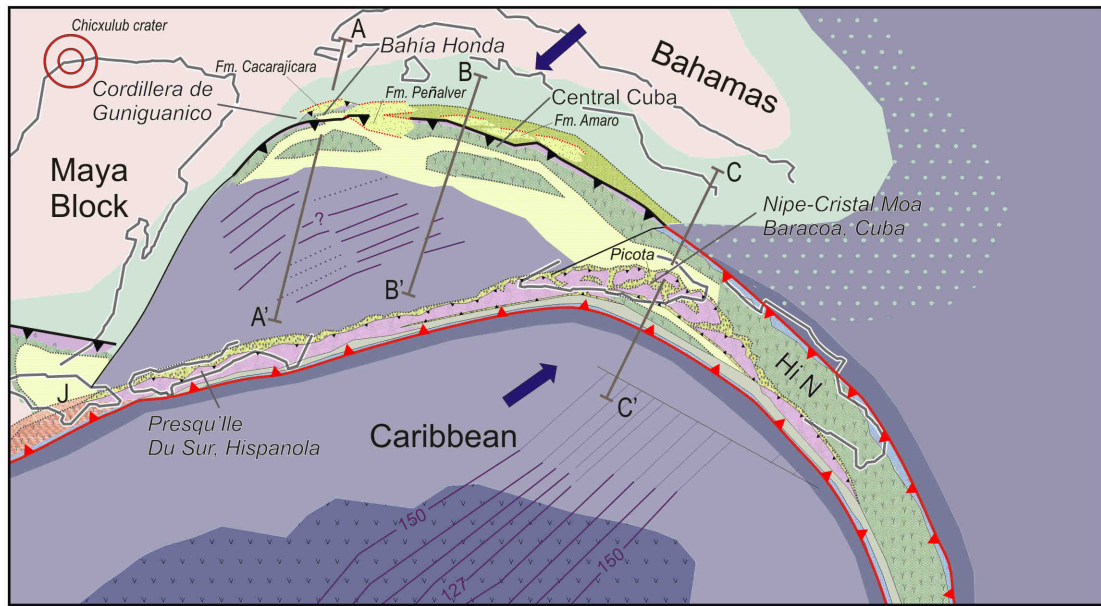
It seems probable, that the GCA was torn off its western continuation along the Matagua Suture Zone at some time during the middle Late Cretaceous (collision of the Chortís and Maya Blocks; Fig. 32), several million years before the late Campanian / early Maastrichtian obduction of the Matagua ophiolites (Donnelly et al. 1990b). Hence, the GCA may have been located somewhere east of the southern half of the Maya Block during the late Campanian onset of collision with the southernmost fringe of the Bahamas Platform (Fig. 34). A far southward extension of pre-Campanian passive margin deposits in the Cuba orogenic belt is confirmed by the interpretation of $^{40}\text{Ar}/^{39}\text{Ar}$ dating of single muscovite grains from the San Cayetano Formation (Hutson et al. 1998). Regarding the late Campanian onset of collision at the approximate position at the southern half of the Maya Block, an estimated amount of 250 km of shortening since the Maastrichtin in the western Cuban thrust belt (Saura et al. 2008), or even estimations of 450 km of shortening for the central Cuban passive margin deposits (Hempton & Barros 1993) appear realistic.

Stanek et al. (2006) attribute the late Campanian cessation of magmatism, simultaneous onset of uplift, cooling and erosion to underthrusting of sediments of the Bahamas margin, thickening of the accretionary wedge and shallowing of the subduction angle.

The Lutgarda Formation of the Camajuaní zone (Fig. 13) and the Bahía Honda units which apparently slid from the collision front of the arc (Stanek et al. 2000, Saura et al. 2008) are indicative for the emergence of a foreland depression during the Maastrichtian (Fig. 35).

Fig. 35: (Next page) Late Maastrichtian tectonic reconstruction of the northwestern Caribbean. Hi N = northern Hispaniola; J = Jamaica.

Late Maastrichtian, 65Ma



- | | | |
|--|--|--|
| carbonate turbidites, turbiditic sandstone
Fm. Lutgarda | megaturbidite
Fm. Cacarajicara, Fm. Peñalver; Fm. Amaro | sedimentary "cover" of CVA
Fm. San Juan y Martinez, Fm. Vía Blanca, Fm. Manacal |
| hemipelagic turbidites, volcaniclastic debris,
Fm. Moreno (K ₂ cp) | Cretaceous volcanic arc suite (CVA)
(K ₁ - K ₂) | Nipe-Cristal / Moa-Baracoa oceanic rocks |
| North American Paleomargin deposits
(J ₃ - K ₂) | grantoids | olistostromes,
Fm. Picota, SE Cuba |
| riftbasin deposits
(J ₁ -J ₂) | amphibolites
earliest products of CVA (J ₃ - K ₁) | Wagwater volcanics,
Jamaica |
| continental basement | subduction-accretion complex
(protolithes J ₁ - J ₃) | magnetic anomalies |
| thickened Caribbean oceanic crust | ophiolite obduction
(protolithes: ? J - K ₁) | relative plate motion |
| oceanic crust | | |

Some authors correlate the widespread Campanian stratigraphic gaps in the southern Bahamas passive margin sequences (Fig. 11, 13) to the early onset of collision (e.g. Pindell 1994; Stanek 2000). However, the stratigraphic gaps may also represent the effect of erosion at the base of the Cretaceous – Tertiary boundary megaturbidite (Pszczółkowski 1986, Cobiella-Reguera 2000, Takayama et al. 2000; Tada et al. 2002), that might have propagated eastward through the channel of the emerging foreland depression, parallel to the collision front (Cacarajícara, Peñalver, Amaro Formations; Fig. 35). The Cretaceous – Tertiary boundary megaturbidite is widespread on the continental margin units. The only arc units covered by the megaturbidite are those of the Bahía Honda complex and the Habana – Matanzas area. Hence, the arc units of the Havana – Matanzas area might also represent a Maastrichtian slide mass into the emerging foreland basin, analogue to the Bahía Honda units.

According to the P-T history of the Escambray Massif (Fig. 15), actual thrusting of the metamorphic subduction-accretionary complex and the extinct arc onto the southern margin of the Bahamas Platform begun at about 65 Ma (Fig. 35) when further cooling of the Escambray and Isla de la Juventud metamorphic units was assisted by extensional tectonic unroofing (Pindell et al. 2005, Stanek et al. 2006).

Consequently, the Paleogene development of the Cuban foreland basins including their late Paleocene to Eocene olistostromes and debris flow deposits (Fig. 16; chapter 4.1.5) is an expression of the final collision in the Cuba orogenic belt (Fig. 37, 40) and does not mark the onset collision. The late tectonic implementation of the Cuba orogenic belt is characterized by

- the final thrusting episodes in the nappe piles of the Cordillera Guaniguanico and central Cuba;
- the obduction of the Northern Ophiolite Belt. The NOB originates from the lower plate oceanic realm south of the Bahamas Platform which was subducted beneath GCA from the north (Fig. 32). It may also encompass some primitive arc components from the basement of the GCA, but a Pacific origin as considered by Kerr et al. (1999) can be ruled out from the general tectonic frameset;
- the exhumation of the metamorphic subduction-accretion complex on the Isla de la Juventud and in the Escambray Massif. Paleogene uplift and exhumation of the metamorphic subduction-accretionary complex may have been triggered by buoyancy forces and slab rebound after large portions of the arc have been thrust over from the south (Fig. 37, 40, 42; Stanek et al. 2006). Flanking sedimentary basins of the Isla de la Juventud dome and the Escambray Massif record the first pebbles of HP-metamorphic rocks at about 45 Ma (Kantshev et al. 1976, Stanek 2000).

At least from Early Eocene time a significant change started to affect the entire northern Caribbean region. Relative motion of the Caribbean shifted to a more eastward direction with respect to North America. As a result, major northeast trending sinistral strike slip faults started to crosscut the compressive structures in the Cuba

orogenic belt (Pinar, Matanzas, La Trocha, Cauto Faults). North America – Caribbean relative motion began to be increasingly taken up at the site of the sinistral Oriente Fault and the Cayman Trough (chapter 7.5). Regarding the probable Early Eocene opening of the Cayman Trough (chapter 3.3.3; Rosencrantz et al. 1988, Leroy et al. 2000), the initiation of the major strike slip faults in the Cuban belt may even predate the Early Eocene.

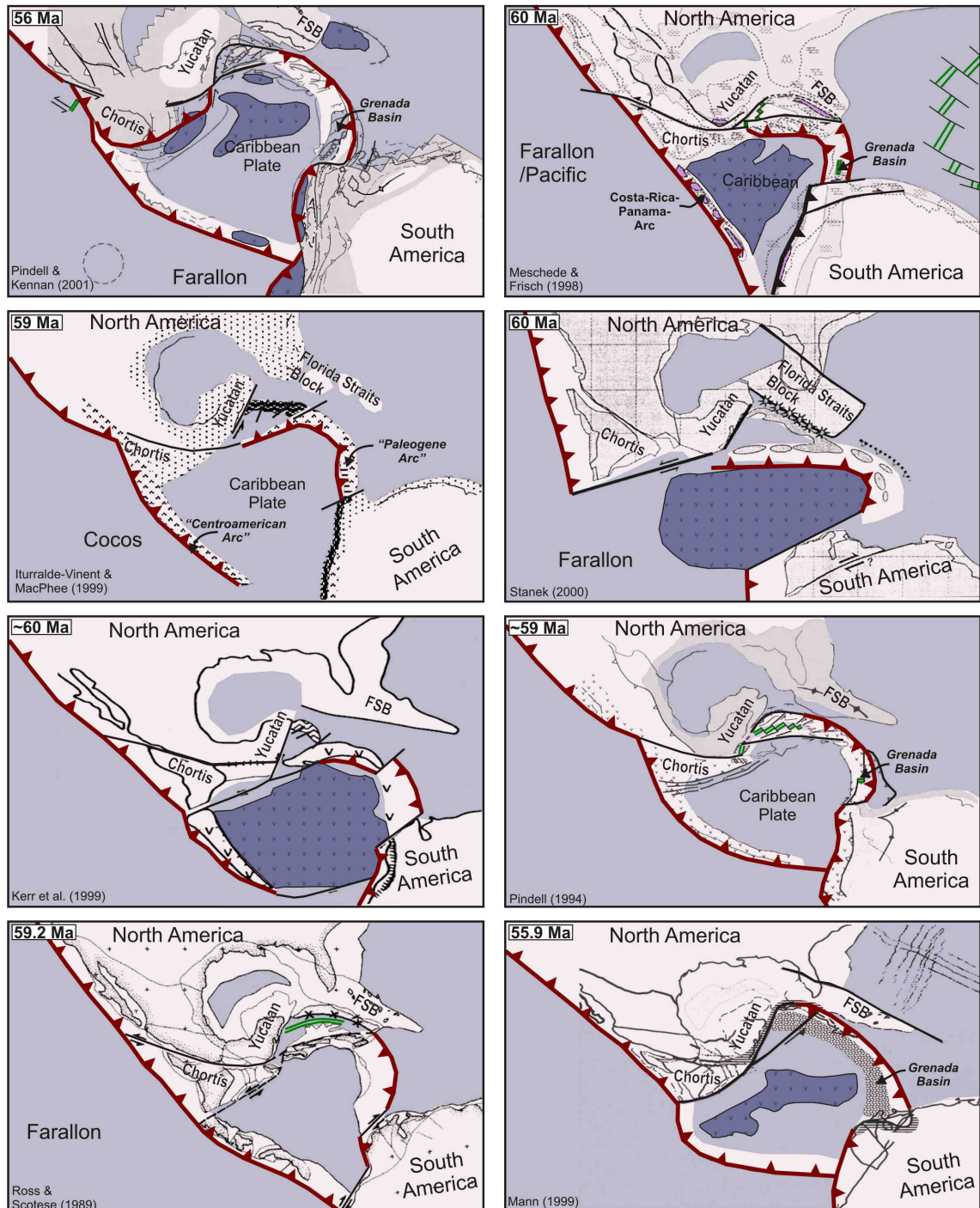


Fig. 36: Comparison of Paleocene reconstructions of the Caribbean region (compiled from Pindell 1994; Iturralde-Vinent & MacPhee 1999; Ross & Scotese 1988; Meschede & Frisch 1998; Kerr et al. 1999; Mann 1999; Stanek 2000; Pindell & Kennan 2001).

7.3 Origin of the Paleogene volcanic arc

Some distinctive aspects of regional geology to the south of the Cuban orogenic belt and the understanding of their causal connection are absolutely vital to a comprehensive hypothesis of the latest Cretaceous and Paleogene tectonic development of the northwestern Caribbean:

- In eastern Cuba, sheets of mafic and ultramafic rocks rest on top of the GCA units (chapter 4.2.2, Fig. 24, 25). These are associated with late Maastrichtian to early Danian olistostromes, conglomerates and debris flow deposits (La Picota and Micara Formations, Fig. 18, Appendix, VI A171, VI A168, VI A169). Gravitational sliding from a southern location has been proposed as the most probable mode of emplacement of the eastern Cuban ophiolites (Cobiella 1974, 1978; Iturralde-Vinent et al. 2006). Maastrichtian greenschist facies metamorphism of GCA units in southeastern Cuba (Purial Complex; chapter 4.2.2; Iturralde-Vinent et al. 2006) seems to be directly related to the cause of the gravitational thrust sheet emplacement. A close relationship to the mafic volcanics of the southern peninsula of Hispaniola (chapter 3.3.2) and the Bath-Dunrobin Complex in Jamaica (3.3.1) is incidental.
- Several million years after the Campanian cessation of magmatic activity in the Cuban portion of the GCA, the Sierra Maestra records renewed Paleocene – Middle Eocene volcanic arc activity (chapter 4.2.1). To the north, the Paleogene arc volcanics thin out into basin successions of distinct back-arc basin character (Bayamo – San Luis and the Cauto – Nipe Basins). Therefore, the Sierra Maestra seems to comprise the northern fringe of a Paleogene volcanic arc, whose associated fore-arc region was located to the south and has been truncated sharply by the left-lateral Oriente Fault (Fig. 24, 25). The Cayman Ridge (chapter 3.3.4) may represent the western continuation of the Paleogene arc remnants of the Sierra Maestra.
- The bisection of magmatic arc activity constitutes another significant parallelism between eastern Cuba and Jamaica, whereas on Jamaica, the “older volcanic sequence” ends during the Turonian and renewed magmatic activity already starts during the Campanian (chapter 3.3.1). However, Paleogene volcanism of both regions ceases in the Middle Eocene.
- On Hispaniola, a time-based delimitation of Cretaceous GCA activity and the activity of a distinctive Paleogene arc is not obvious. Nevertheless, the spatial distribution of Cretaceous and Paleogene magmatic arc units fits the concept of a Late Cretaceous cessation of GCA activity and the new emergence of a south-facing magmatic arc during the Paleogene: The GCA fore-arc units along the northeastern coast do not record subduction any younger than Late Cretaceous. Paleogene volcanics are confined to the southwestern portion of the island and the volcano-sedimentary sequences of the Montagne Noir, Chaîne de Matheux, Sierra

de Neiba (chapter 3.3.2, Fig.7) are apparently closely related to those of the Bayamo – San Luis Basin and the Sierra Maestra in eastern Cuba.

Many authors who assume west– southwest-dipping subduction for the Late Cretaceous GCA do not support the concept of north-facing Paleogene subduction to explain the above stated aspects (e.g. Pindell 1994; Ross & Scotese 1988; Pindell & Kennan 2001, Pindell et al. 2006; Fig. 36; 39). In the concepts of Pindell & Kennan (2001) and Pindell et al. (2006), the Paleogene magmatism in eastern Cuba and Hispaniola is attributed to a late episode of GCA activity, which would have been merely dormant in Cuba since the Campanian. Accordingly, the Paleogene Cayman Ridge would represent a part of the GCA which was separated by latest Maastrichtian – Paleocene opening of the “intra-arc” Yucatán Basin.

However, it is very difficult to see how Paleocene volcanic activity to the south of the Yucatán Basin could have been induced by subduction beneath the northern edge of the GCA, while there was apparently no oceanic crust left to be subducted and the GCA was actually thrust onto the southern Bahamas margin. Assuming that there would have been Paleogene subduction beneath the northern edge of the GCA, it is just as difficult to comprehend how this scenario would facilitate the opening of the Yucatán Basin to the north of the Paleogene volcanic axis. Furthermore, these reconstructions fail to explain late Maastrichtian to early Danian north-directed emplacement of mafic and ultramafic rocks in eastern Cuba, the southern peninsula of Hispaniola and in eastern Jamaica.

Other authors tentatively highlight the necessity of Paleogene north-dipping subduction in order to explain the geological history of the northwestern Caribbean (e.g. Meschede & Frisch 1998; Stanek 2000; Fig. 36; 39), while some even elevate north-dipping subduction to an exclusive concept, neglecting all evidence for a south-facing Cretaceous GCA (Iturralde-Vinent & MacPhee 1999; Kerr et al. 1999; Fig. 36; 39).

Figures 34, 35, 37 comprise a hypotheses that embraces the main aspects of the plate tectonics and links the previously noted aspects to a common process.

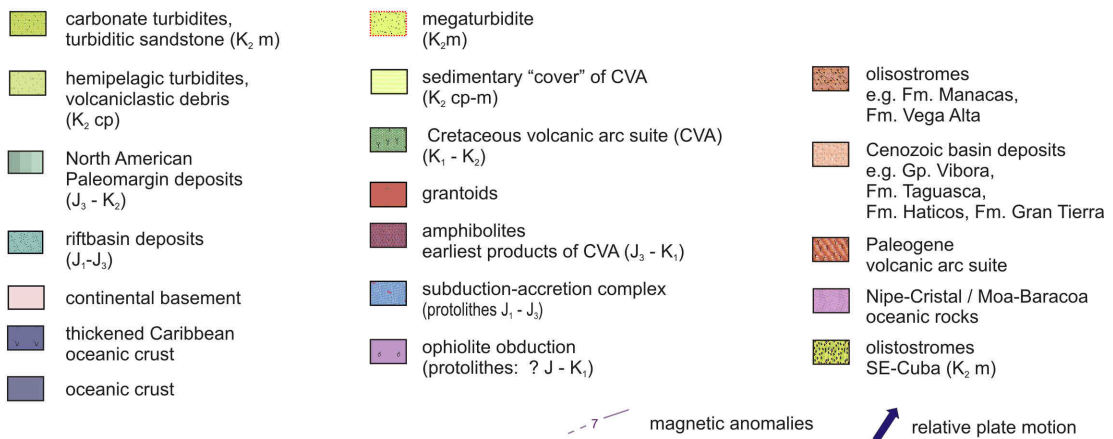
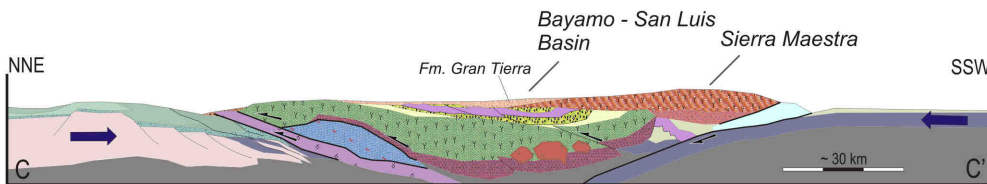
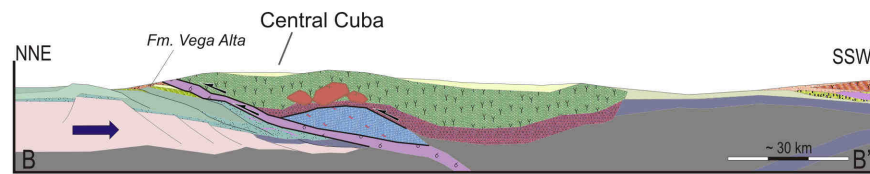
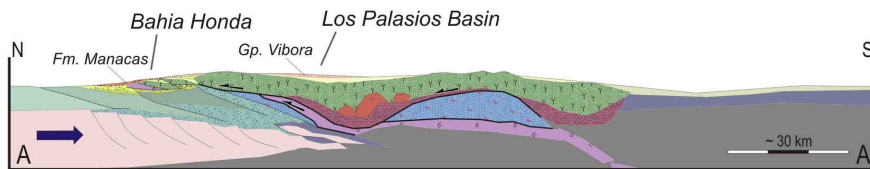
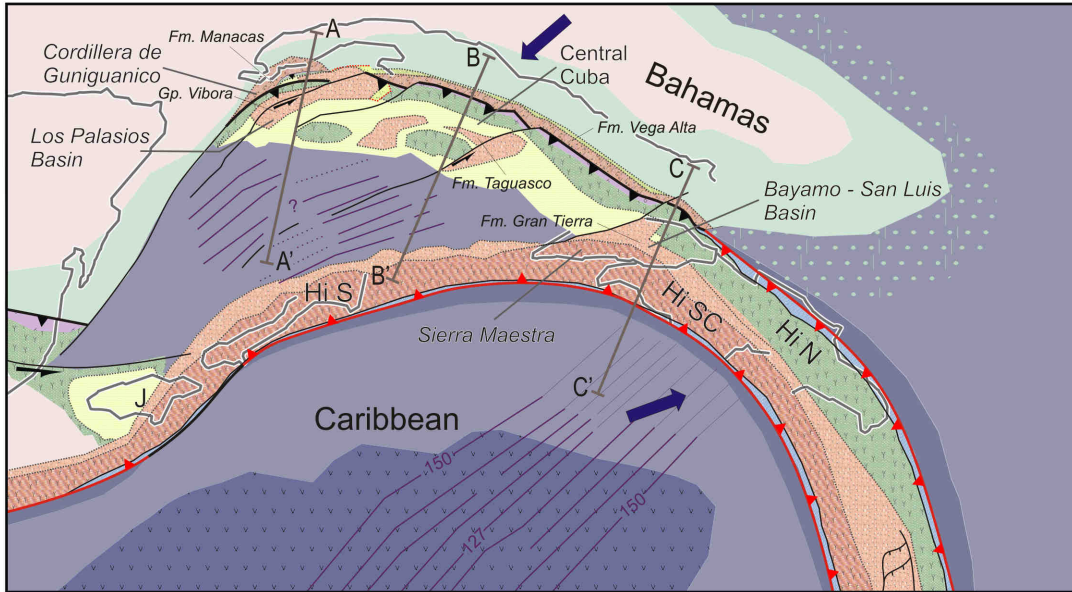
The southern margin of Chortís may have faced north - northeast-directed subduction from the late Campanian onward (e.g. Pindell & Kennen 2001; Fig. 33). Impediment of northward relative motion of the Caribbean Plate already affected this area subsequently to the collision between the Chortís and Maya Blocks which put an end to the GCA activity on Jamaica (“older volcanic sequence”, chapter 3.3.1; Mitchell 2003, 2006). The resulting north to northeast-directed subduction along the southern margin of Chortís is represented by Campanian – Maastrichtian granitoid intrusion and andesitic volcanism in the Kellits Synthem and the Blue Mountain Block on Jamaica (Mitchell 2003, 2006).

Fig. 37: (Next page) Paleocene tectonic reconstruction of the northwestern Caribbean.

Hi N = northern Hispaniola; J = Jamaica; Hi S = southern peninsula of Hispaniola;

Hi SC = south-central Hispaniola.

Paleocene, 55Ma



A key problem of with the apparent north-directed gravity emplacement of the eastern Cuban ophiolites is the origin of the large scale uplift that is necessary to emplace oceanic rocks onto the southern fringe of the GCA. Nevertheless, the same phenomenon affected the southern peninsula of Hispaniola and is also traceable on the southern flank of the Blue Mountains Block of eastern Jamaica.

The relatively abrupt late Maastrichtian – early Danian emplacement of the ophiolite was evidently coincident with the emergence of Paleocene volcanic activity to the south of the Cuba orogenic belt. Assuming that the Paleogene arc was triggered by north-dipping subduction, the emplacement would be coincident with the initiation of north-dipping subduction at a convergent boundary to the south of the colliding GCA and the Yucatán Basin.

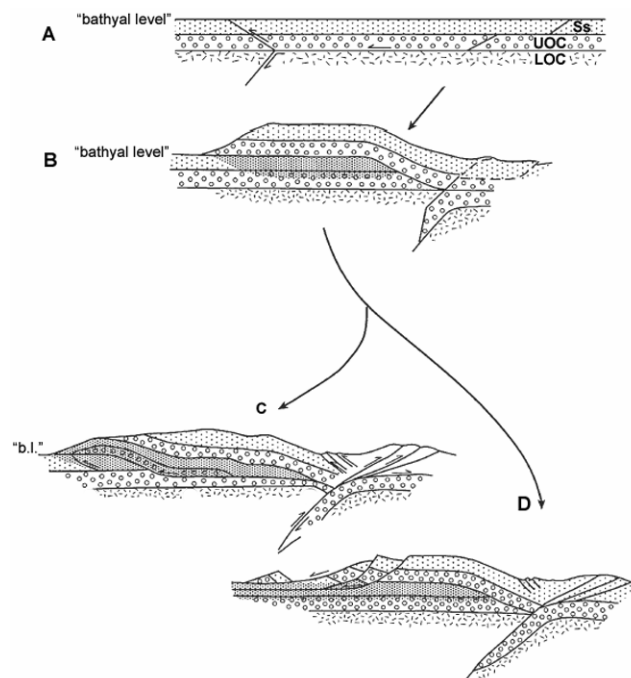


Fig. 38: Schematic diagrams of “flake” emplacement at the commencement of subduction, based on Gretener (1981) in Bradshaw (2004). **(A)** Initial compressive stress induces subhorizontal shear between upper and lower portions of oceanic crust. **(B)** Superposition of flake is resulting in high pore-water pressure zone (stippled). **(C)** Gravity induced spreading of the pile concentrated in the high pore-water pressure zone leads to thrusting to the left while the right is buttressed by initial subduction convergence. **(D)** High pore-water pressure leads to large-scale gravity sliding with thinning and subsidence of the upper plate. Ss = sedimentary succession; UOC = upper ocean crust; LOC = lower ocean crust.

The Northland Allochthon, which comprises oceanic rocks on top of the autochthonous continental crust of the Northland peninsular of New Zealand (e.g. Ballance & Spörl 1979, Bradshaw 2004), may represent an analogue of the eastern Cuban ophiolites. A

recent model for the origin of the Northland Allochthon is based on the concept of “flake” emplacement at the commencement of subduction (Bradshaw 2004; Gretener 1981; Fig. 38). Accordingly, initial compressive stress induces subhorizontal shear between upper and lower portions of oceanic crust. The superposition of “flakes” of oceanic crust is resulting in zones of high pore-water pressure in the sediments trapped below the overthrust slap. Thus, a situation is created that is highly conducive to further displacement, either by gravity induced spreading of the pile concentrated in the high pore-water pressure zone (Fig. 38c), or by large-scale gravity sliding with thinning and load-induced subsidence of the upper plate (Fig. 38d). Contemporary basin sequences of the upper plate contain olistostromes and other material from the advancing sheets. Triggers for a major collapse may have been the growing accretionary complex after the onset of an active subduction zone and/or a change of direction in the convergence vector (Fig. 35). In the cases of the Northland Allochthon as well as the northwestern Caribbean, the base of the overthrust slab would be on the upper plate of the emerging subduction system and unlikely to undergo significant further shortening.

Regarding the positions of ophiolitic rocks in the Blue Mountains of Jamaica, eastern Cuba, the southern peninsula of Hispaniola and their association with Late Maastrichtian – Early Danian olistostromes, a similar process probably affected the northwestern Caribbean realm before the commencement of the Paleogene volcanic arc. Progression of the collision process between the GCA and the Bahamas margin hampered the Caribbean Plate in its northward relative motion with respect to North America. Continued northward push triggered initial compressive stress and “flake” emplacement (Fig. 34) which finally resulted in the commencement of north-dipping subduction (Fig. 35) and the emergence of the PVA (Fig. 37).

After the impediment of northward relative motion of the Caribbean Plate and the resulting “flake” emplacements had affected the entire northwestern Caribbean, the new Paleogene volcanic arc spanned the region from the southern margin of the Chortís Block, along the southern boundary of the Yucatán Basin to the southeastern tip of Cuba and the southern half of Hispaniola (Fig. 37). Cenozoic sinistral strike slip along the northern Caribbean Plate boundary and the relative eastward migration of the Nicaragua Rise and the Chortís Block later dismembered the original extent of the PVA (chapter 7.5). The eastern bend of the PVA was uncoupled in the course of this process and is probably represented by the Aves Ridge (chapter 3.2.2).

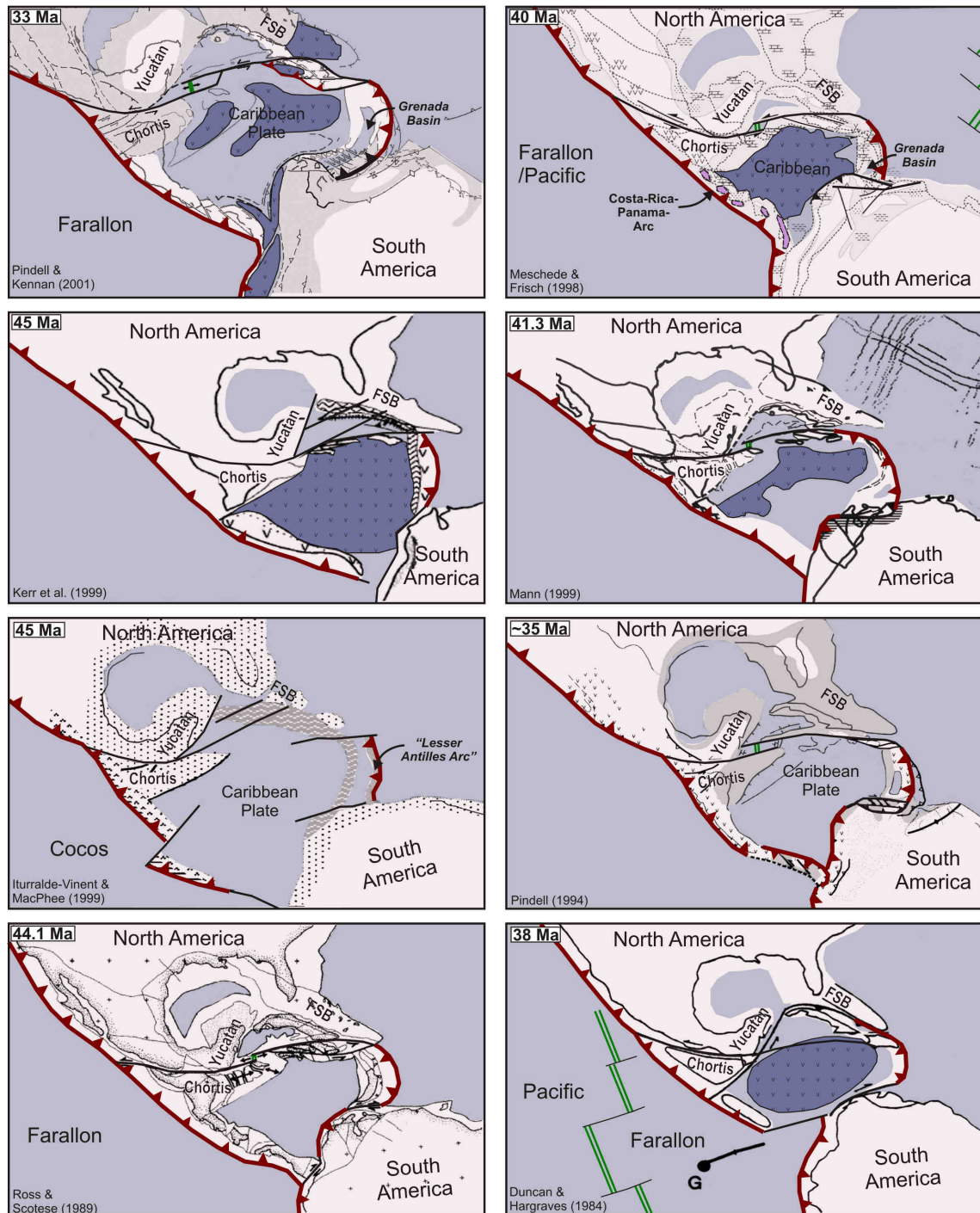


Fig. 39: Comparison of Eocene reconstructions of the Caribbean region (compiled from & Duncan & Hargraves 1984; Pindell 1994; Iturralde-Vinent & MacPhee 1999; Ross & Scotese 1988; Meschede & Frisch 1998; Kerr et al. 1999; Mann 1999; Pindell & Kennan 2001).

7.4 Origin of the Yucatán Basin

Many models consider Late Cretaceous to early Paleocene back-arc or intra-arc spreading causing the formation of the Yucatán Basin (e.g. Hall & Yeung 1982, Pindell & Dewey 1982, Pindell 1994), but the reliability of these models remains ambiguous:

- At least parts of the Yucatán Basin may be older than Late Cretaceous / Paleocene (Rosencrantz 1990; chapter 3.3.4);
- Some models (e.g. Hutson et al. 1998; Meschede & Frisch 1998) would have produced magnetic anomalies at approximately right angles to those of Hall & Yeung (1982), and
- a spreading axis is not identified.

The assumed extent of the PVA (Fig. 37) would have isolated the Yucatán Basin from its original affiliation to the Caribbean Plate. Hence, no or only minor back-arc / intra-arc spreading is necessary to explain the origin of basin. The Albian reconstruction in Fig. 32 more or less aligns the tentatively identified series of magnetic anomalies of the Yucatán Basin with those of the Venezuelan Basin which would ascribe the origin of oceanic crust of the Yucatán Basin to pre-Oxfordian rifting along the Caribbean spreading system (chapter 6.4). At least from geometric inference, this interpretation seems feasible, although it does not rule out any Late Cretaceous to early Paleocene back-arc or intra-arc stretching.

7.5 Onset of the northern Caribbean transform margin

As outlined above, relative northward motion of the Caribbean Plate was impeded from the late Campanian – Maastrichtian onward which resulted in the commencement of north-dipping subduction and the emergence of the PVA. Due to a continuous clockwise rotation of North America – Caribbean relative motion, the PVA probably faced increasingly oblique subduction in its northwestern portion during the Paleocene to Middle Eocene (Fig. 35, 37, 40, 42).

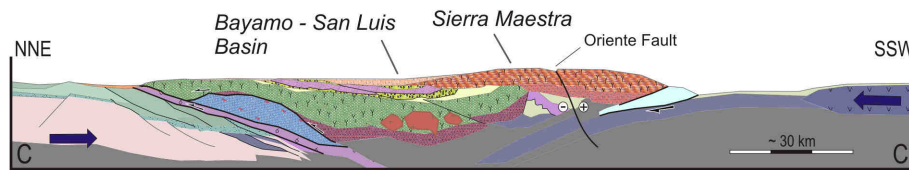
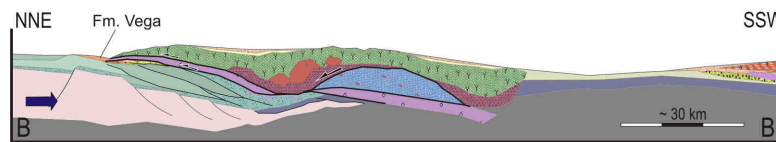
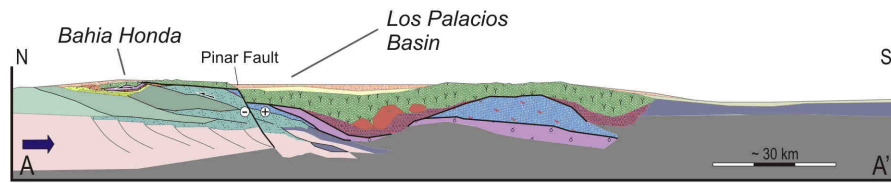
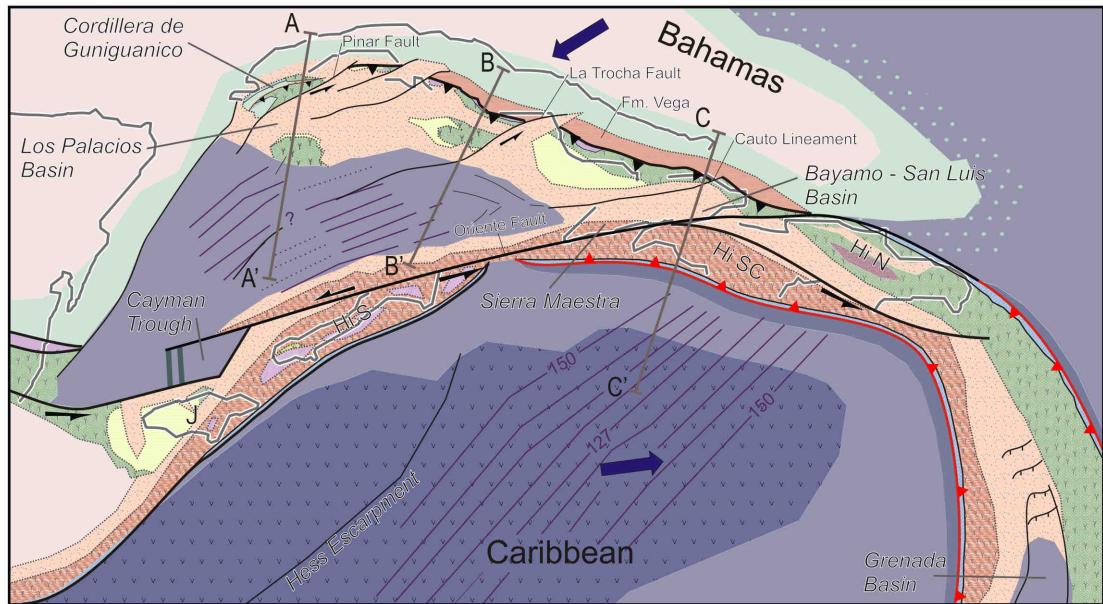
The change in direction of convergence also affected the final episodes of collision between the GCA and the Bahamas margin. From Early Eocene time, transtensional basins (Cuenca Los Palacios, Cuenca Las Vegas, Cuenca de la Broa, Cuenca Central, Cuenca Cauto; Fig.9) started to accumulated several thousand meters of sedimentary infill (Puscharovskiy et al. 1989; Saura et al. 2008). The associated major northeast trending sinistral strike-slip faults (Pinar, Matanzas, La Trocha, Cauto Faults) may have started to crosscut the compressive structures in the Cuba orogenic belt even earlier during the Paleogene (Fig. 37).

Fig. 40: (Next page) Eocene tectonic reconstruction of the northwestern Caribbean.

Hi N = northern Hispaniola; J = Jamaica; Hi S = southern peninsula of Hispaniola;

Hi SC = south-central Hispaniola.

Eocene, 40Ma



- | | | |
|--|--|--|
| carbonate turbidites, turbiditic sandstone (K ₂ m) | megaturbidite (K ₂ m) | debris flow deposits e.g. Fm. Vega |
| hemipelagic turbidites, volcanoclastic debris (K ₂ cp) | sedimentary "cover" of CVA (K ₂ cp-m) | olistostromes (E _{1,2}) |
| North American Paleomargin deposits (J ₃ - K ₂) | Cretaceous volcanic arc suite (CVA) (K ₁ - K ₂) | Cenozoic basin deposits |
| riftbasin deposits (J ₁ -J ₃) | granitoids | Paleogene volcanic arc suite (E _{1,2}) |
| continental basement | amphibolites earliest products of CVA (J ₃ - K ₁) | Nipe-Cristal / Moa-Baracoa oceanic rocks |
| thickened Caribbean oceanic crust | subduction-accretion complex (protoliths J ₁ - J ₃) | olistostromes SE-Cuba (K ₂ m) |
| oceanic crust | ophiolite obduction (protoliths: ? J - K ₁) | |
- magnetic anomalies relative plate motion

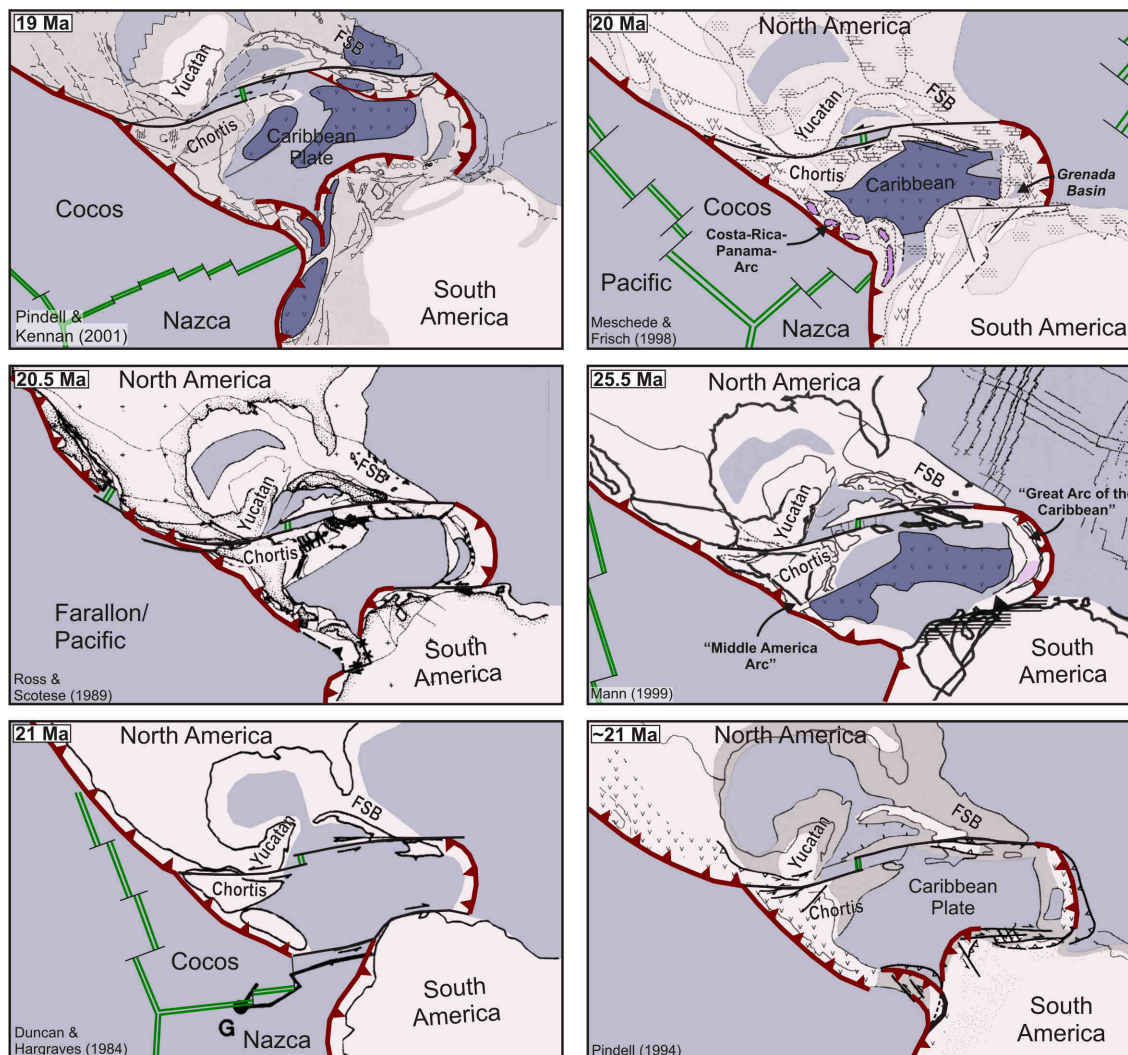


Fig. 41: Comparison of Miocene reconstructions of the Caribbean region (compiled from Duncan & Hargraves 1984; Pindell 1994; Ross & Scotese 1988; Meschede & Frisch 1998; Mann 1999; Pindell & Kennan 2001).

Simultaneous Middle Eocene cessation of PVA activity was probably caused by the arrival of the CLIP at the subduction zone (Fig. 40). Deformation of the rocks of the Sierra Maestra, which are folded with increasing intensity from south to north, might be attributed to the collision of the CLIP. This collision stopped any northward component of relative movement of the Caribbean Plate. As a result, North America – Caribbean relative motion began to be taken up mostly at the site of the sinistral Oriente Fault and the Cayman Trough. Spreading at the Mid-Cayman Rise probably began during the late Early Eocene (Rosencrantz et al. 1988; Leroy et al. 2000; Fig. 8) and coupled the Oriente Transform Fault with the Swan Island Transform and the Motagua–Pólochic Fault System. The Caribbean Plate including the Cortís Block and the Nicaragua Rise effectively started relative eastward movement towards the free space in the eastern Atlantic.

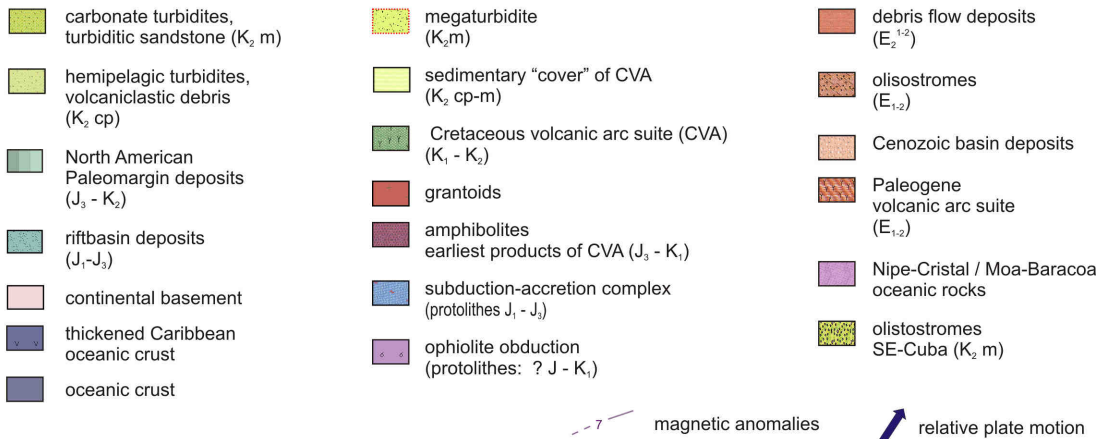
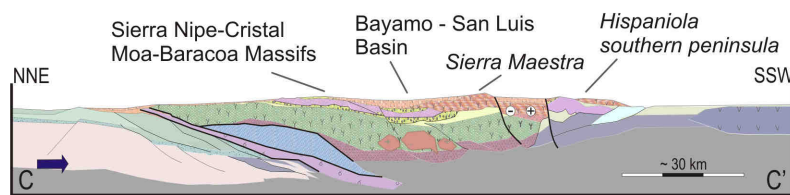
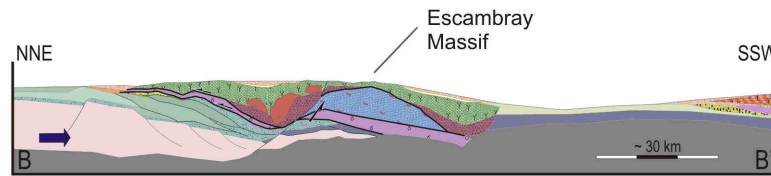
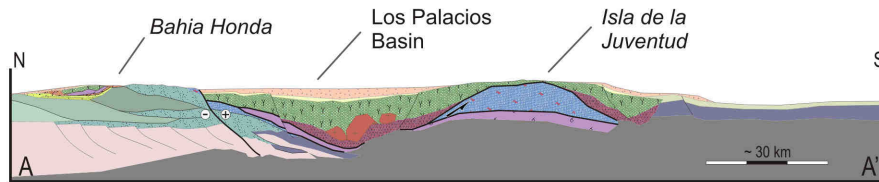
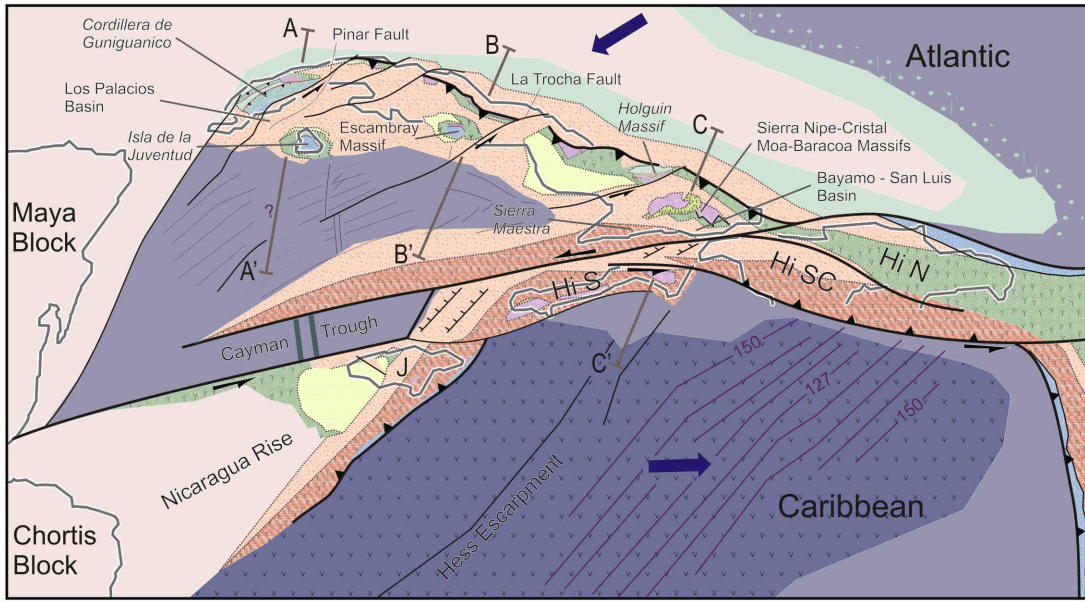
As evident on Hispaniola, the present day Oriente Transform Fault splits eastward into a multibranch fault system. Restoring the south-central part of the island including

the Presqu' ile du Nord-Ouest, the Montagne Noir, the Chaine de Matheux, and the Sierra de Neiba to a Middle Eocene position off the coast of eastern Cuba (Fig. 37, 40) accounts for an approximate Cenozoic displacement of 200 to 300 km between the Cordillera Central and south-central Hispaniola. With regard to an estimated amount of 800 to 1000 km of sinistral offset along the Oriente Fault, most of its eastern prolongation must be accommodated between the south-central part and the southern peninsula of Hispaniola. Restoring an approximate amount of 800 to 1000 km of relative eastward movement places the southern peninsula of Hispaniola to a Middle Eocene position south of the Cayman Rise (Fig. 37, 40).

According to this reconstruction, the Cayman Ridge and the Sierra Maestra only preserve the northernmost fringes of the PVA which was almost completely dismembered by Cenozoic strike-slip since the Middle Eocene reconfiguration of the northern Caribbean Plate boundary.

Fig. 42: (Next page) Miocene tectonic reconstruction of the northwestern Caribbean.
Hi N = northern Hispaniola; J = Jamaica; Hi S = southern peninsula of Hispaniola;
Hi SC = south-central Hispaniola.

Miocene, 20Ma



8 Conclusions

Various plate tectonic models of the Caribbean have been proposed over the last decades. With regard to the northwestern Caribbean, geodynamic reconstructions are particularly controversial in terms of the number of involved arcs, subduction polarity, and timing of collision.

A careful review of the local geological data of Cuba, its relations to Hispaniola and Jamaica as well as a comparison of prevalent plate tectonic models, allow for the proposal of refined Late Cretaceous to Miocene tectonic reconstructions (Fig. 32, 34, 35, 37, 40, 42). Inferences from these reconstructions comprise some essential elements for any evolutionary synthesis of the Caribbean region:

- The Mabujina amphibolites on Cuba and the Duarte Complex on Hispaniola point to a primitive initial GCA-activity as early as 130 Ma. During the Aptian – Albian, the GCA was fully developed in its juvenile stage.
- Initiation of the GCA along the northern fringe of the Chortís Block appears to be the most adequate concept for a northwestern Caribbean reconstruction. It provides a link between (1) middle Late Cretaceous collision processes along the Matagua suture zone, (2) the Turonian termination of GCA-activity on Jamaica, (3) the late Campanian onset of collision in the Cuba orogenic belt, and (4) the general inference of southwest-dipping polarity of the arc.
- Continuous southwest-dipping polarity of the GCA, at least from the Aptian-Albian until the late Campanian onset of collision with the Bahamas margin, can be inferred from (1) its Late Cretaceous approach towards the North American margin, (2) the magnitude of top to the north directed tectonic transport in the Cuba orogenic belt after the late Campanian onset of collision, and (3) the internal structures of the metamorphic fore-arc assemblages and their evolution on the north side of the arc.
- The collision of the GCA with the Bahamas margin hampered relative northward motion of the Caribbean Plate from the late Campanian onward. Continued northward push finally resulted in north-dipping subduction and the emergence of a Paleocene to Middle Eocene volcanic arc (PVA) which spanned the northwestern Caribbean along the southern margin of the Yucatán Basin while the Chortís Block and the Nicaragua Rise were still in a paleoposition to the south of the Maya Block. The eastern bend of the PVA is probably represented by the Aves Ridge.
- The Maastrichtian commencement of north-dipping subduction was accompanied by superposition of oceanic crust and large-scale north-directed gravity sliding on the upper plate, as documented by ophiolitic slide-masses and Maastrichtian olistostromes in Jamaica, eastern Cuba, and the southern peninsula of Hispaniola.

- North-dipping subduction and the emergence of the Paleogene volcanic arc isolated the Yucatán Basin from the Caribbean Plate. Instead of a proposed Late Cretaceous – Paleocene origin by back-arc or intra-arc spreading, the oceanic crust of the Yucatán Basin probably has its origin in the Jurassic Caribbean spreading system.
- Sinistral strike-slip and associated basin formation in the Cuba orogenic belt point to continuous clockwise rotation of North America – Caribbean relative motion during the Paleocene. Relative northward motion of the Caribbean stopped after the Eocene arrival of thickened oceanic crust at the north-dipping subduction zone of the PVA.
- After the late Early Eocene commencement of spreading at the Mid-Cayman Rise, North America – Caribbean relative motion was taken up along the sinistral Oriente Fault and its eastern and western prolongations. This transform margin dissects and dismembered the former extend of the PVA with estimated amounts of 800 to 1000 km offset along the Oriente Fault since the Middle Eocene. If south-central Hispaniola is restored to a Middle Eocene position to the south of eastern Cuba, most of the total Oriente offset must be accommodated at the northern bounding-faults of the southern peninsula of Hispaniola.

Whether labelled as "intra-American" or "Pacific", almost all of the views imply significant amounts of eastward Caribbean migration relative to the Americas. The controversies essentially pinpoint in different hypotheses about the Cretaceous development of the "Great Caribbean Arc" and the role of the Costa Rica – Panamá – Arc.

Presuming a "northern Chortís" initiation of the GCA and considering the size and shape of the Caribbean Plate when subducted crust is extracted from beneath the PVA and South America, an Early Cretaceous situation is inferred which represents a syntheses between the concepts of Meschede & Frisch (1998) and Pindell & Kennan (2001), each advocates of the apparently incompatible "intra-American" and "Pacific" concepts respectively.

Concepts of a pre-Campanian east-dipping polarity of the GCA can be ruled out from the current state of knowledge about the collision history in Cuba. A Campanian polarity reversal due to collision of the "Caribbean Large Igneous Province" is objected by (1) evidence for Paleocene to Middle Eocene north-dipping subduction of Caribbean oceanic crust, and (2) evidence for an actual collision of the CLIP no earlier than Eocene, when north-dipping subduction was choked and the northern Caribbean transform margin was established.

Hypotheses about a pre-Aptian east-dipping polarity of the GCA are equivocal. They can neither be ruled out nor clearly reconfirmed by geological evidence from the northwestern Caribbean. After all, from regional analysis and reconstruction a pre-Aptian east-dipping polarity of the GCA appears to be dispensable.

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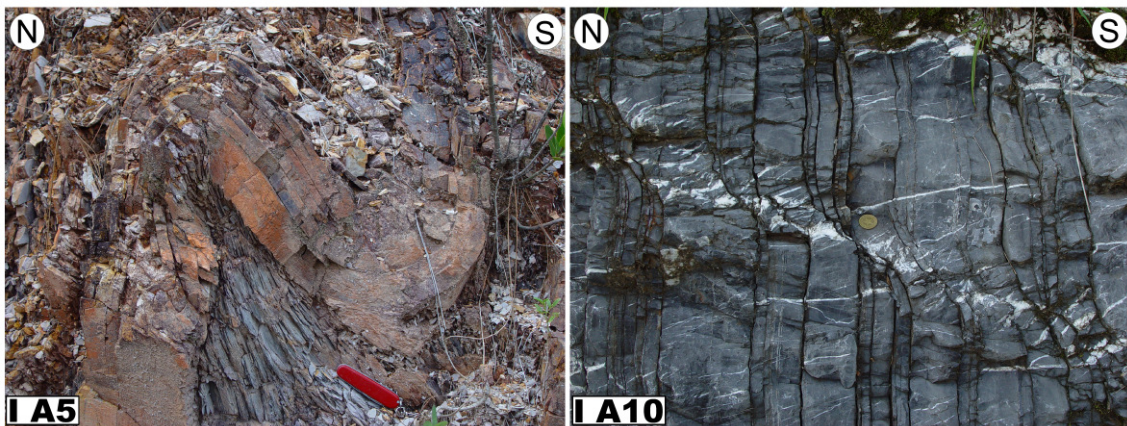
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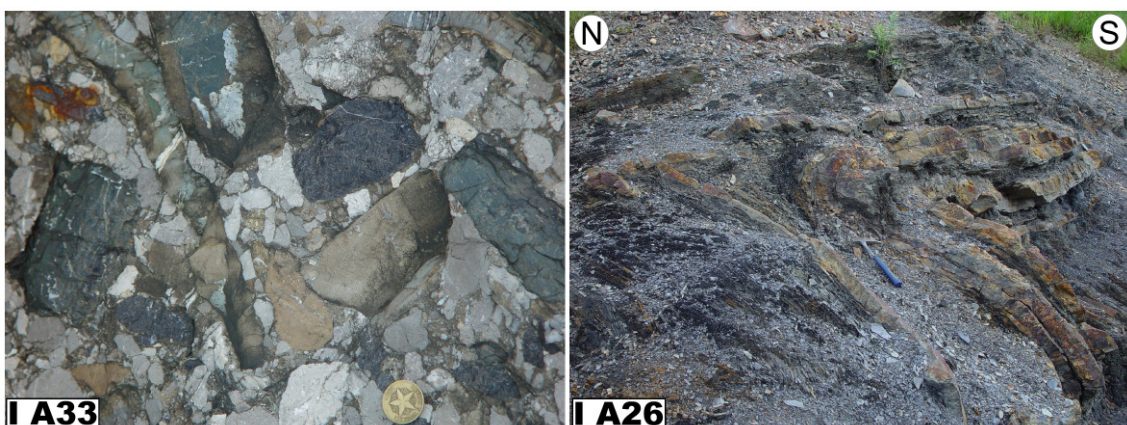
Appendix

Photo documentations of exemplary outcrops and lithotypes from Cuba

I. Riftbasin and paleomargin deposits



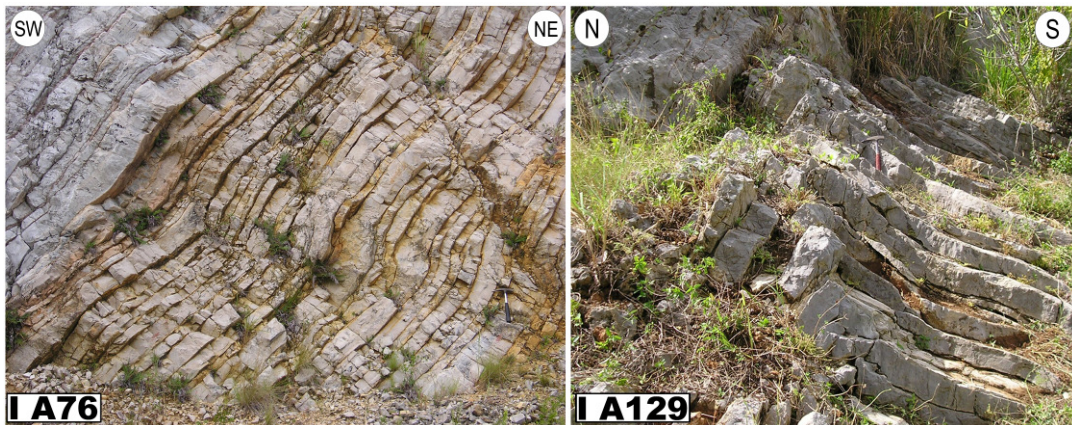
I A5: Tightly folded beds of the San Cayetano Formation (J_1 - J_{3ox}), Cordillera Guaniguanico, Southern Rosario belt; 22°41'40"N, 83°44'08"W. **I A10:** Upright and kinked dark lutite beds of the Artemisa Formation (J_3 - K_1), Cordillera Guaniguanico, Southern Rosario belt; 22°44'10"N, 83°35'10"W.



I A33: Basal breccia of the Cacarajícara Formation, Cretaceous – Tertiary boundary megaturbidite, Cordillera Guaniguanico, Northern Rosario belt; 22°52'10"N, 83°02'38"W. **I A26:** Tightly folded beds of the San Cayetano Formation (J_1 - J_{3ox}), Cordillera Guaniguanico, Sierra de los Organos; 22°39'07"N, 83°21'59"W.



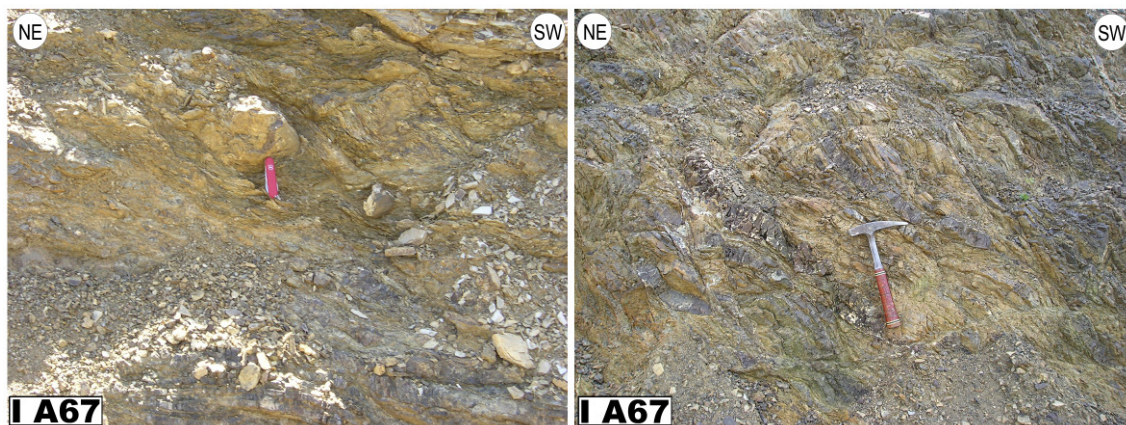
I A86: Gently folded Cretaceous platform deposits of the Remedios belt, central Cuba, “El Ygre” quarry; 22°18'02"N, 79°14'49"W. **I A131:** Cretaceous platform deposits of the Remedios belt with subhorizontal shear zones and drag folds, central Cuba, “Sierrezuela” quarry; 22°22'06"N, 79°26'34"W.



I A76: Folded fine grained limestones of the Trocha Formation (J_3k-t), Camajuani belt, central Cuba, “Los Asores” quarry; 22°22'58"N, 79°35'13"W. **I A129:** Late Jurassic fine grained limestones with kink folds, Camajuani belt, central Cuba; 22°23'08"N; 79°33'51"W.

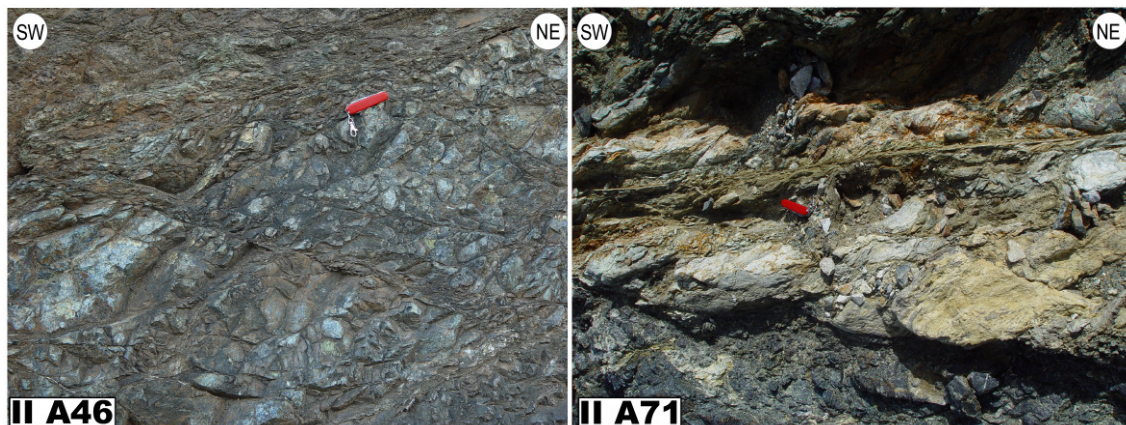


I A65: Phacoid duplexes of the Carmita Formation (deep water limestones, $K_{1-2al-cm}$) exposed in a tectonic window within the thrust sheets of the northern ophiolites, Placetas belt, central Cuba, 22°27'25"N, 79°50'49"W.



I A67: Intensely sheared and disrupted cherts, mudstones, and limestone beds of the Carmita Formation ($K_{1-2a-l-cm}$), central Cuba, directly below the thrust sheets of the northern ophiolites (c.f. Fig. 22); $22^{\circ}27'44''N$, $79^{\circ}46'39''W$.

II. Northern Ophiolite Belt



II A46: Sheared serpentinites of the Northern Ophiolite Belt, La Habana province; $22^{\circ}54'52''N$, $81^{\circ}50'58''W$. **II A71:** Serpentinite mélangé of the Northern Ophiolite Belt, central Cuba; $22^{\circ}23'00''N$, $79^{\circ}50'22''W$.

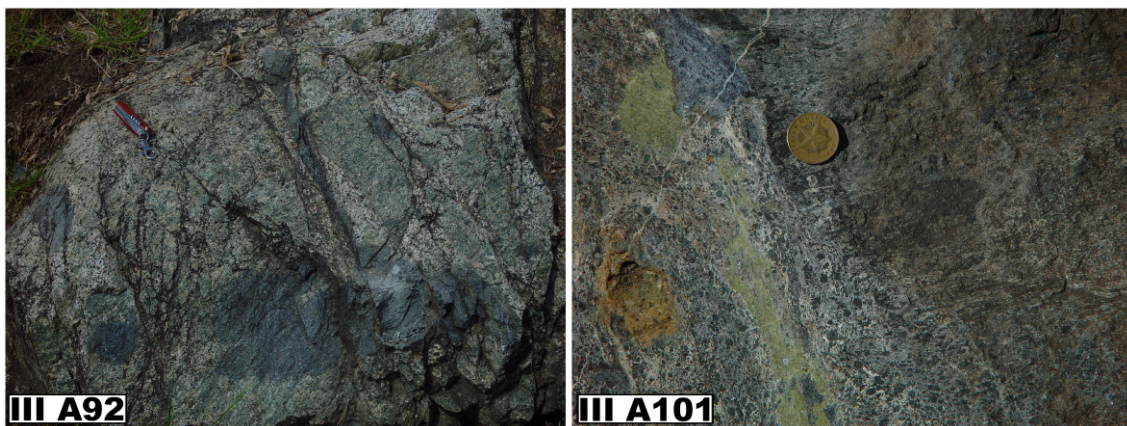
III. Cretaceous volcanic arc



III A29: Folded and sheared cherts and volcanics of the Encrucijada Formation ($K_{1-2a-al}$), Bahía Honda unit; $22^{\circ}52'29''N$, $83^{\circ}09'34''W$.



III A99: Faulted ash layers of the Tasajera Formation (K2cn-cp), central Cuba; 22°19'11"N, 79°57'57"W.



III A92: Granite with amphibolitic and mica dominated xenoliths, central Cuba; 22°09'33"N, 80°00'37"W. **III A101:** Amphibolite of the Mabujina unit; 22°05'40"N, 79°58'39"W.

IV. Foreland basin deposits



IV A31: Caotic and sheared deposit of the Manacas Formation (E_{1-2}), Cordillera Guaniguanico, Northern Rosario belt; 22°50'43"N, 83°03'09"W. **IV A75:** Conglomerate of the Vega Formation (E_2^{1-2}); Clastic components dominated by Cretaceous limestones and cherts from the Remedios and Placetas zones (paleomargin deposits), central Cuba; 22°21'12"N, 79°34'16"W.

V. Piggyback basin deposits



V A27



V A146

V A27: Interbedded conglomerates and sandstones of the Capdevilla Formation (E_2^1), western Cuba, Los Palacios Basin; 22°38'32"N, 83°21'06"W. **V A146:** Sedimentary breccia of the Haticos Formation (E_1), resting unconformably on serpentinite body of the northern ophiolites; Clastic components dominated by serpentinites, gabbros, andesites, and cherts; eastern Cuba, southern fringe of Holguin Massif; 20°51'24"N, 75° 56'33"W.

VI. Nipe-Cristal / Moa Baracoa ophiolite massifs



VI A171a



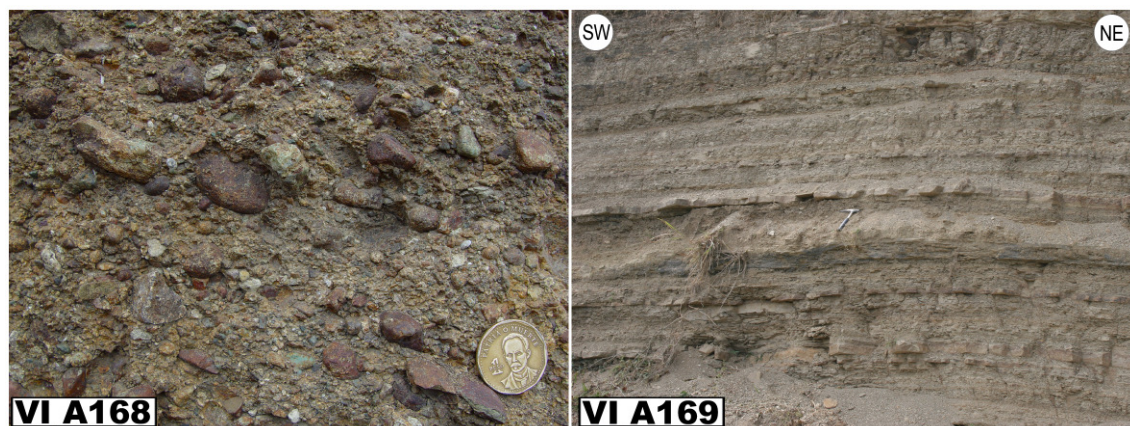
VI A171b



VI A171c



VI A171: Moderately deformed serpentinite body resting on an undeformed conglomerate of the Picota Formation (K_2m); The contact between the conglomerate and the serpentinites is slightly sheared; Clastic components of the Picota Formation are dominated by lithotypes from the ophiolite suite, eastern Cuba, near Mayarí Arriba; $20^{\circ}24'56''N$, $75^{\circ}34'54''W$.

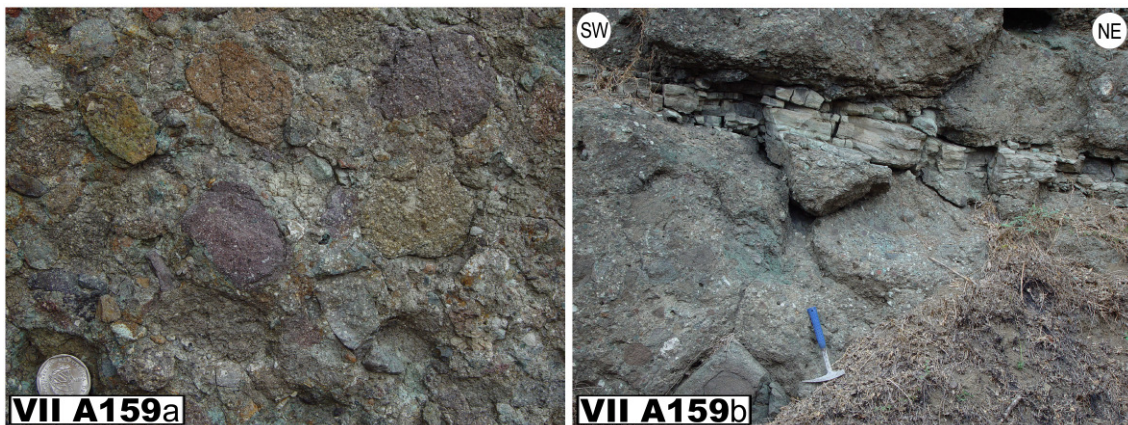


VI A168: Conglomerate of the lower Mícará Formation (K_2m-E_1) in which the Picota Formation (A171) occurs as lenticular intercalations; eastern Cuba, near Mayarí Arriba; $20^{\circ}23'18''N$, $75^{\circ}33'07''W$. **VI A169:** Sandstones and shales of the upper Mícará Formation, eastern Cuba, near Mayarí Arriba; $20^{\circ}25'29''N$; $75^{\circ}33'32''W$.

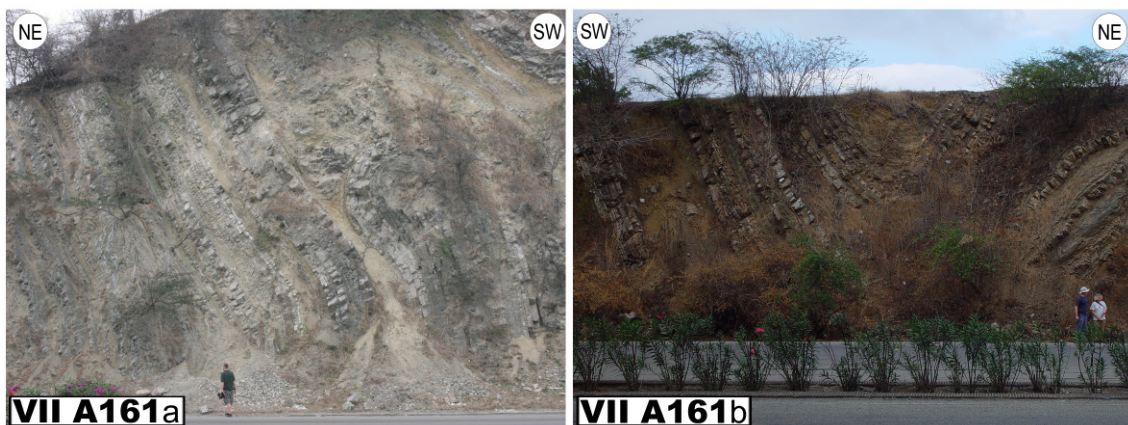


VI A202: Limestone beds and basal contact of a limestone breccia (debris flow deposit) of the Gran Tierra Formation (E_1) overlaying the Mícaro Formation in the Mayarí Arriba area, eastern Cuba; $20^{\circ}25'17''N$, $75^{\circ}34'20''W$.

VII. Paleoge volcanic arc



VII A159: Volcanic agglomerate and tuff layer of the El Cobre Group (E_{1-2}), eastern Cuba, Sierra Maestra; $20^{\circ}08'26''N$, $75^{\circ}59'49''W$.



VII A161: Folded tuff and limestones of the El Cobre Group, eastern Cuba, Sierra Maestra, north of Santiago de Cuba; $20^{\circ}04'13''N$, $75^{\circ}48'07''W$.

VIII. Back-arc deposits of the Paleogene volcanic arc



VIII A154



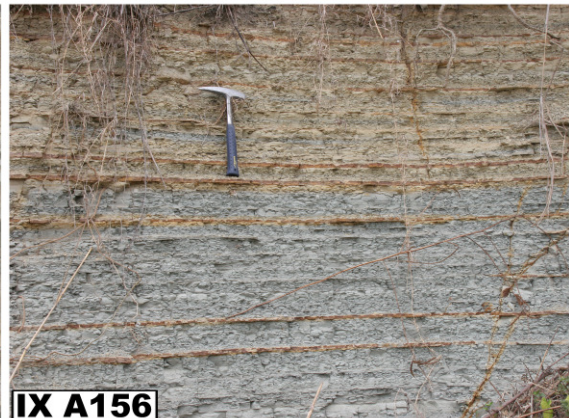
VIII A143

VIII A154: Tuff bed within a micritic limestone succession of the Miranda Formation (E_{1-2}), eastern Cuba, Bayamo – San Luis Basin; 20°18'10"N, 75°55'19"W. **VIII A143:** Weathered tuff bed within laminated deep-water limestones of the Vigia Formation (E_{1-2}), eastern Cuba, southern fringe of Holguin Massif; 20°50'04"N; 76°07'21"W.

IX. Cover deposits of the Paleogene volcanic arc



IX A157



IX A156

IX A157: Limestone beds of the Puerto Boniato Formation (E_2^2), eastern Cuba; northern edge of Sierra Maestra; 20°09'22"N, 75°59'10"W. **IX A156:** Interbedded shales and sandstone layers of the San Luis Formation (E_2^2), eastern Cuba; Bayamo - San Luis Basin; 20°14'06"N, 75°57'51"W.



IX A163



IX A187

IX A163: Conglomerate of the Camarones Formation (E_2^2), eastern Cuba, Bayamo – San Luis Basin; 20°09'32"N, 75°42'20"W. **IX A187:** Conglomerate of the San Luis Formation (E_2^2); Clastic components dominated by volcanic rocks (El Cobre Grp) and limestones (?Puerto Boniato Fm), eastern Cuba, Bayamo – San Luis Basin; 20°11'34"N, 75°53'57"W.

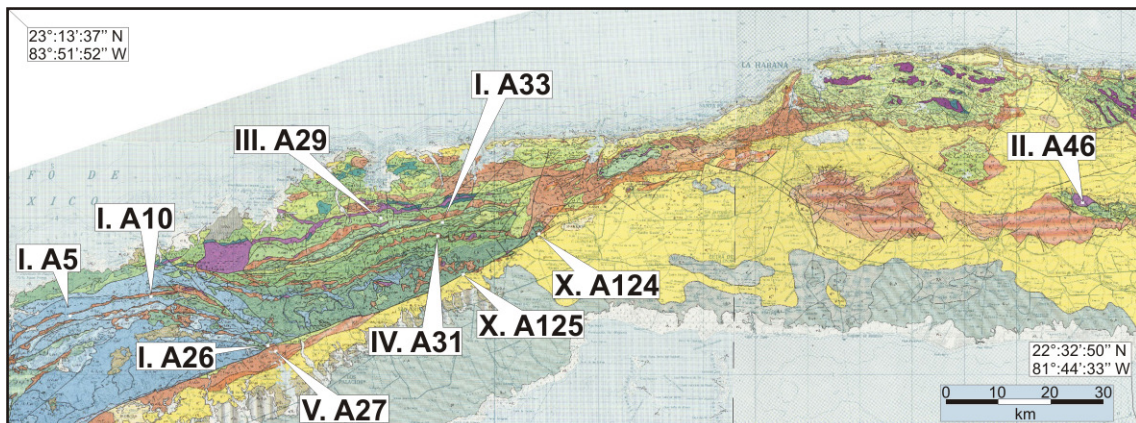
X. Pinar Fault Zone



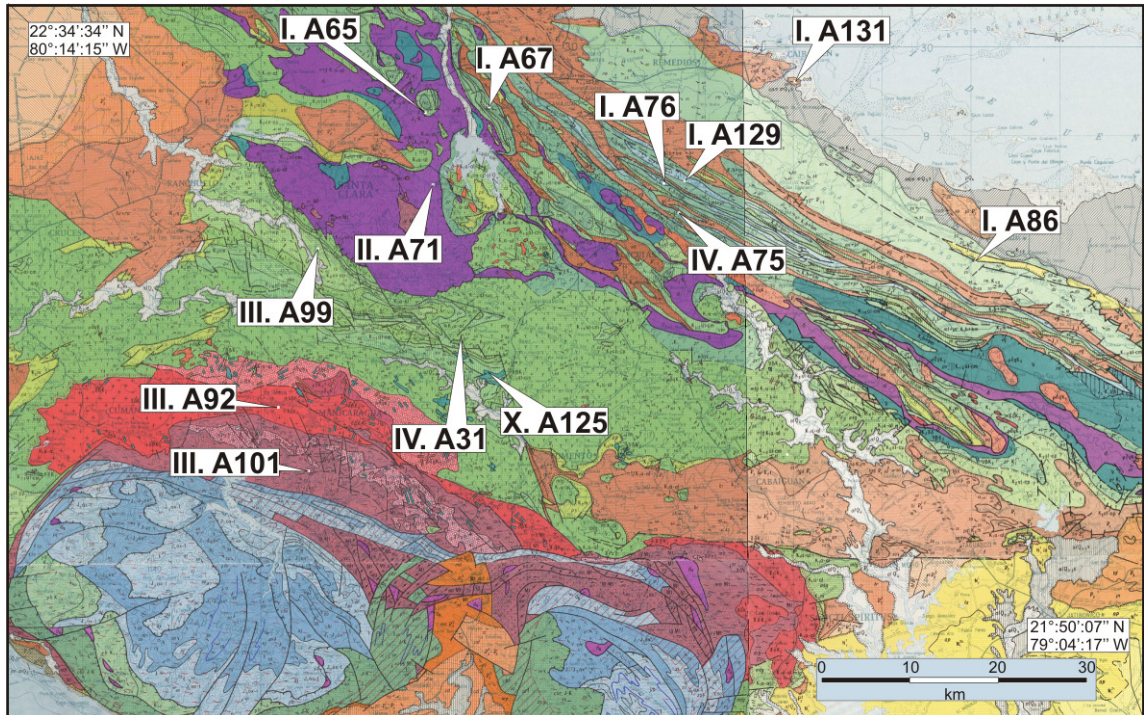
X A124: Tectonite of the Pinar Fault Zone, western Cuba, with shear fabric and rotated serpentinite lense indicating sinistral sense of shear, direction of view is perpendicular to the ground; 22°51'06"N, 82°51'50"W. **X A125:** Slickenside within limestone of the Arroyo Cangre Formation (J₂ – J_{3ox}) at the Pinar Fault Zone. Down dipping lineation indicates left-lateral normal faulting; 22°46'47"N, 83°00'13"W.

Location maps

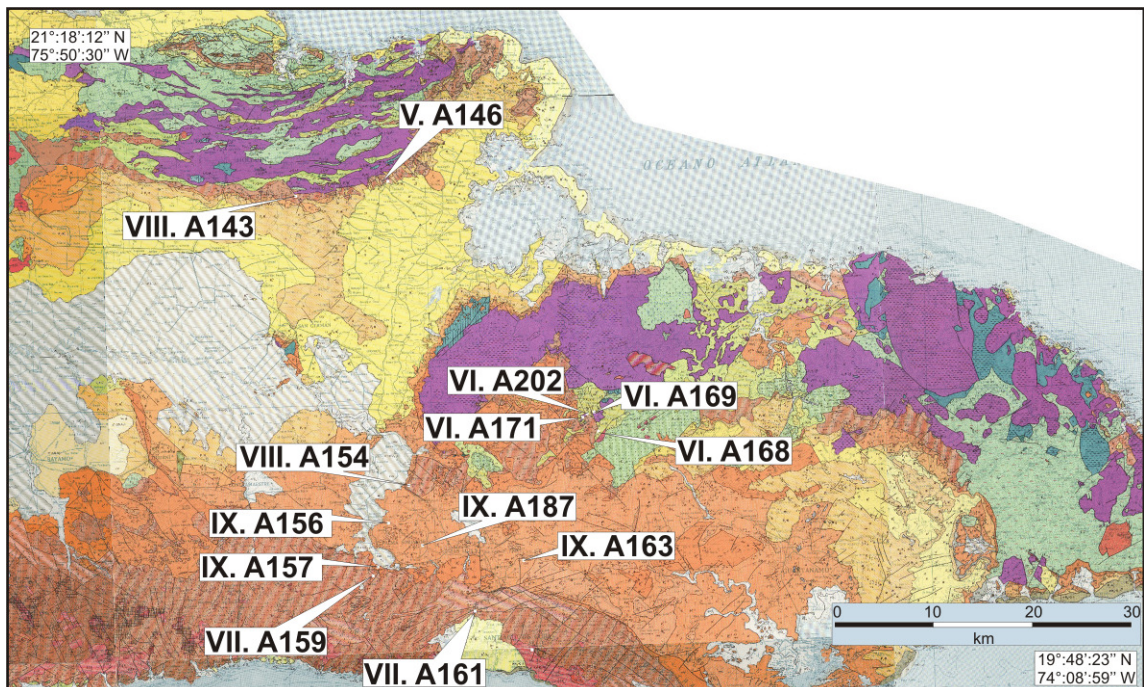
Western Cuba



Central Cuba



Eastern Cuba



Erklärung

Hiermit erkläre ich, dass diese Arbeit bisher von mir weder an der Mathematisch-Naturwissenschaftlichen Fakultät der Ernst-Moritz-Arndt-Universität Greifswald noch einer anderen wissenschaftlichen Einrichtung zum Zwecke der Promotion eingereicht wurde.

Ferner erkläre ich, daß ich diese Arbeit selbständig verfasst und keine anderen als die darin angegebenen Hilfsmittel benutzt habe.

Greifswald, den 12.02.2009

LEBENS LAUF

Max Sommer

Anschrift	Glockenkreuz 12, 53894 Mechernich-Weyer
Geburtsdatum	03.02.1975
Geburtsort	Mannheim
Familienstand	verheiratet, 3 Kinder

Schulbildung

08/1981 - 06/1985	Fritz-Gansberg-Schule, Grundschule Wiesbaden
08/1985 - 06/1986	Leibnitz Gymnasium, Wiesbaden
08/1986 - 06/1991	Städtisches Ruhr-Gymnasium, Witten
08/1991 - 06/1992	St. Francis High-School, Traverse City, Michigan, USA
08/1992 - 06/1994	Städtisches Ruhr-Gymnasium, Witten Abschluß: Hochschulreife (Note 2,1)

10/1994 - 10/1995	Wehrdienst
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Hochschulstudium

10/1995 - 01/2002	Studium der Geologie und Paläontologie an der Ernst-Moritz-Arndt-Universität Greifswald Diplomarbeit: „Strukturgeologische Untersuchungen zur Krustenentwicklung im zentralen Schwarzburger Antiklinorium (Thüringisches Schiefergebirge / Saxothuringikum)“ (Note 1,1) Abschluß: Diplom-Geologe (Note 1,2)
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Berufstätigkeit / praktische Tätigkeiten

05/2000 - 07/2000	Praktikum bei Umweltplan GmbH, Stralsund
09/2000 - 10/2000	Praktikum bei DURTEC GmbH, Neubrandenburg
seit 05/2002	Promotionsstudent am Institut für Geographie und Geologie der Universität Greifswald
seit 09/2005	Geologe beim Geoinformationsdienst der Bundeswehr

Greifswald, den 12.02.2009