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Kinematic reconstruction of the Caribbean region since the Early Jurassic

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ABSTRACT

The Caribbean oceanic crust was formed west of the North and South American continents, probably from Late Jurassic through Early Cretaceous time. Its subsequent evolution has resulted from a complex tectonic history governed by the interplay of the North American, South American and (Paleo-)Pacific plates. During its entire tectonic evolution, the Caribbean plate was largely surrounded by subduction and transform boundaries, and the oceanic crust has been overlain by the Caribbean Large Igneous Province (CLIP) since ~90 Ma. The consequent absence of passive margins and measurable marine magnetic anomalies hampers a quantitative integration into the global circuit of plate motions. Here, we present an updated, quantitatively described kinematic reconstruction of the Caribbean region back to 200 Ma, integrated into the global plate circuit, and implemented with GPlates free software. Our reconstruction includes description of the tectonic units in terms of Euler poles and finite rotation angles. Our analysis of Caribbean tectonic evolution incorporates an extensive literature review. To constrain the Caribbean plate motion between the American continents, we use a novel approach that takes structural geological observations rather than marine magnetic anomalies as prime input, and uses regionally extensive metamorphic and magmatic phenomena such as the Great Arc of the Caribbean, the CLIP and the Caribbean high-pressure belt as correlation markers. The resulting model restores the Caribbean plate back along the Cayman Trough and major strike-slip faults in Guatemala, offshore Nicaragua, offshore Belize and along the Northern Andes towards its position of origin, west of the North and South American continents in Early Cretaceous time. We provide the paleomagnetic reference frame for the Caribbean region by rotating the Global Apparent Polar Wander Path into coordinates of the Caribbean plate interior, Cuba, and the Chortis Block. We conclude that formation of the Caribbean plate, west of the North and South Americas, as a result of Panthalassa/Pacific spreading leads to a much simpler plate kinematic scenario than Proto-Caribbean/Atlantic spreading. Placing our reconstruction in the most recent mantle reference frames shows that the CLIP originated 2000–3000 km east of the modern Galápagos hotspot, and may not have been derived from the corresponding mantle plume. Finally, our reconstruction suggests that most if not all modern subduction zones surrounding the Caribbean plate initiated at transform faults, two of these (along the southern Mexican and NW South American margins) evolved diachronously as a result of migrating trench–trench–transform triple junctions.

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1. Introduction

Kinematic reconstruction of regional tectonic evolution comprises translation of qualitative geological data into a quantitative model, describing the relative motions of plates and regional tectonic units. Ideally, the reconstruction is quantified by sets of Euler poles and corresponding finite rotation angles. This has become normal practice in the global reconstruction of continents and continental fragments through geological time (e.g. Besse and Courtillot, 2002; Müller et al., 2008; Torsvik et al., 2008, 2012; Doubrovine et al., 2012; Seton et al., 2012), but regional examples are still few (e.g. van Hinsbergen et al., 2011, 2012, 2014). Generally, regional tectonic reconstructions compile relative motions through time, but when being linked to the global plate circuit of plate motions using a mantle reference frame, they become key input for the assessment of how lithospheric evolution is coupled to underlying mantle processes and mantle structure (Spakman and Hall, 2010).

In the present paper, we aim to develop this kind of kinematically quantified tectonic evolution model of the Caribbean region since ~200 Ma, tied to the North and South American plates. This model can be incorporated in a current or future global plate circuit of choice of either relative of absolute motion. There is currently no generally accepted global plate circuit, but through time the differences in relative plate motions among the models proposed have become gradually smaller (e.g., Gordon and Jurdy, 1986; Müller et al., 2008; Torsvik et al., 2008, 2012; Doubrovine et al., 2012; Seton et al., 2012). In this study, we incorporate our plate model into the South America–Africa and North America–Africa frame of the global plate circuit of Torsvik et al. (2012).

The Caribbean plate is a largely oceanic tectonic plate (3500 km E–W by 1000 km N–S), bounded by convergent margins in the east (Lesser Antilles subduction zone), west (Central American subduction zone), and along the northeastern margin of South America (South Caribbean Deformed Belt), and strike-slip-dominated boundaries in the north and southeast, accommodating a relative westward movement of the Americas with respect to the Caribbean plate (Wilson, 1966; Burke et al., 1978; Pindell and Barrett, 1990). Marine magnetic anomalies and fracture zones are usually key in reconstructing past plate motions, providing a direct indication of the position of passive margins back in time. However, only sparse evidence exists of marine magnetic anomalies within the Caribbean plate (Ghosh et al., 1984) and ages of the anomalies are unconstrained. In conjunction with the either strike-slip dominated or convergent nature of the plate boundaries and the consequent absence of passive margins, it has proved challenging to reconstruct the movement of the Caribbean plate relative to its surrounding plates back in time. In modern global plate reconstructions (e.g. Seton et al., 2012), the Caribbean region is included using the Euler poles provided in the seminal paper of Ross and Scotese (1988). Based on the reconstruction of Pindell and Barrett (1990), Ross and Scotese (1988) provided poles for a large number of tectonic elements in the Caribbean region, placed in context of the history of seafloor spreading in the Central and South Atlantic (based on poles of Klitgord and Schouten, 1986), magnetic anomalies of the Cayman Trough (using Rosencrantz and Sclater, 1986; Rosencrantz et al., 1988), and various geological studies predating 1988.

Since Ross and Scotese (1988), further progress in understanding Caribbean kinematics has been made owing to increase of the geological database, as well as major leaps forward in our understanding of the fundamental behavior and geological expression of subduction zone evolution. Incorporation of these new data and concepts into plate tectonic models of the Caribbean has been compiled in the form of qualitative tectonic reconstructions by e.g. Burke (1988), Pindell et al. (1988, 1998), Pindell and Barrett (1990), Meschede and Frisch (1998), Müller et al. (1999), Kerr et al. (2003), Pindell and Kennan (2009) and Kennan and Pindell (2009) and generally published in the form of time-sequences of paleogeographic maps. No kinematic parameters in terms of Euler pole rotations are available which renders these reconstructions less useful for quantitative approaches of linking driving mantle processes to tectonic evolution of the region. van Benthem et al. (2013) discussed the possible links between Caribbean tectonic evolution and remnants of more than 100 Myr of subduction, now detected in the upper and lower mantle. To step forward from this analysis and allow for a quantitative assessment of the coupling between Caribbean tectonic evolution, driving processes and mantle structure (as a memory of past processes), a kinematically quantified reconstruction of the region will be key for testing first order evolution hypotheses using geodynamic modeling of crust–mantle evolution.

To this end and for incorporation of 25 years of geological observations and research of the region since Ross and Scotese (1988), we aim to construct a kinematically quantified reconstruction of Caribbean region evolution since the Early Jurassic using the state-of-the-art of global plate reconstructions for casting the regional reconstruction in a global plate motion frame. We use the freely available software package *GPlates* (http://www.gplates.org; Boyden et al., 2011) as the versatile platform for converting geological data and interpretations into Euler poles and finite rotations.

The organization of our paper is as follows. We first provide a research philosophy and approach, and a description of the main geological features used for regional correlation. Then, we review the structural geology of the Caribbean region, as well as the geology of units that can serve for correlation across major fault zones key geological features of the Caribbean region to obtain kinematic constraints as basis for our reconstruction, starting with the Caribbean plate interior, and then in a clockwise journey across the region from Central America, over Cuba to the Lesser Antilles, and via the Leeward Antilles to the northern Andes. We provide this fairly extensive geological review for two reasons: firstly, to help demonstrate the relationship between Caribbean plate movements and geology, and thus to show where our reconstruction is based on, and secondly, because it has been many years since local works around the Caribbean have been synthesized into a coherent model. Next, we will provide a restoration of the Caribbean from the Present back in time and present selected time-slices from the continuous reconstruction. Finally, the resulting reconstruction will be used to evaluate the origin of the lithosphere of the Caribbean plate, the origin of the Caribbean Large Igneous Province, the initiation of subduction zones, and the role of absolute plate motions in the evolution of the Caribbean region.

2. Approach

In reconstructing past plate motions, oceanic magnetic anomalies are the most robust data source. In an extensional setting, crustal material has little opportunity to disappear from the rock record, and extensional geological records are most complete at the end of the tectonic event. Conversely, the geological record of convergent plate boundaries are the most incomplete at the end of the tectonic event, and in the most extreme case, crust may entirely subduct without leaving a rock record at all. In addition, compression normally emerges rocks which may disappear from the geological record belt by erosion. Reconstructed amounts of shortening will consequently be minimum estimates of the amount of convergence accommodated in the shortened zone and therefore, extensional records are preferred. To illustrate this, Europe–Africa motion can be determined by shortening records from the Alps, providing a minimum estimate of convergence. However, combining North and Central Atlantic ocean spreading reconstructions that quantify the movement of Europe relative to North America and the movement of North America relative to Africa, respectively, will provide a much more accurate estimate (e.g. Dercourt et al., 1986; Dewey et al., 1989; van Hinsbergen and Schmid, 2012; van Hinsbergen et al., 2014).

For the Caribbean plate, the only oceanic extensional records that can be used in a quantitative sense are the magnetic anomalies in the Cayman Trough, an oceanic pull-apart basin on the Caribbean–North American plate boundary, where the Swan Island and Oriente faults accommodate left-lateral transform motion (Fig. 1). The Cayman Trough provides kinematic data for the Caribbean plate with respect to the Cuban segment, formerly part of the Caribbean plate but welded to the North American plate since the Paleogene collision of the Caribbean plate with the Bahamas borderlands (Pindell and Barrett, 1990). Before opening of the Cayman Trough (i.e., prior to an estimated 49.4 Ma; Leroy et al., 2000), other data sources are needed to reconstruct relative plate motions, such as major strike-slip fault displacements, continental extensional or shortening estimates, paleomagnetism (indicating rotations and paleolatitudes of plates or plate fragments), plate boundary volcanism and metamorphic belts, and obducted ophiolites indicating former intra-oceanic subduction zones.

The boundary conditions for our Caribbean reconstruction are provided by the relative movements between the North American and African plates, and the South American and African plates using poles from Central and South Atlantic ocean reconstructions of Torsvik et al. (2012) using the timescale of Gradstein et al. (2004) (with the exception of the age of M0, where we adopt a 120.8 Ma age following He et al. (2008)). Fig. 2 shows the resulting relative motion of South America relative to North America. These movements constrain the opening evolution of the Proto-Caribbean Ocean (i.e. the westward continuation of the Central Atlantic Ocean) between the Americas during Jurassic break-up of Pangea (Dickinson and Coney, 1980; Pindell, 1985).

The starting point of our reconstruction is the present-day situation. For the last 49.4 Myr, the motion of the Caribbean plate relative to the Cuban segment is constrained by oceanic spreading in the Cayman Trough. The first step in reconstructing is therefore reversing the oceanic Cayman extension, recorded by magnetic anomalies (Leroy et al., 2000). Prior to opening of the Cayman Trough and before welding of the Cuban segment to North America, reconstruction requires less straightforward data and becomes more difficult and uncertain. Some types of data are considered to constrain motion more precise than others. We

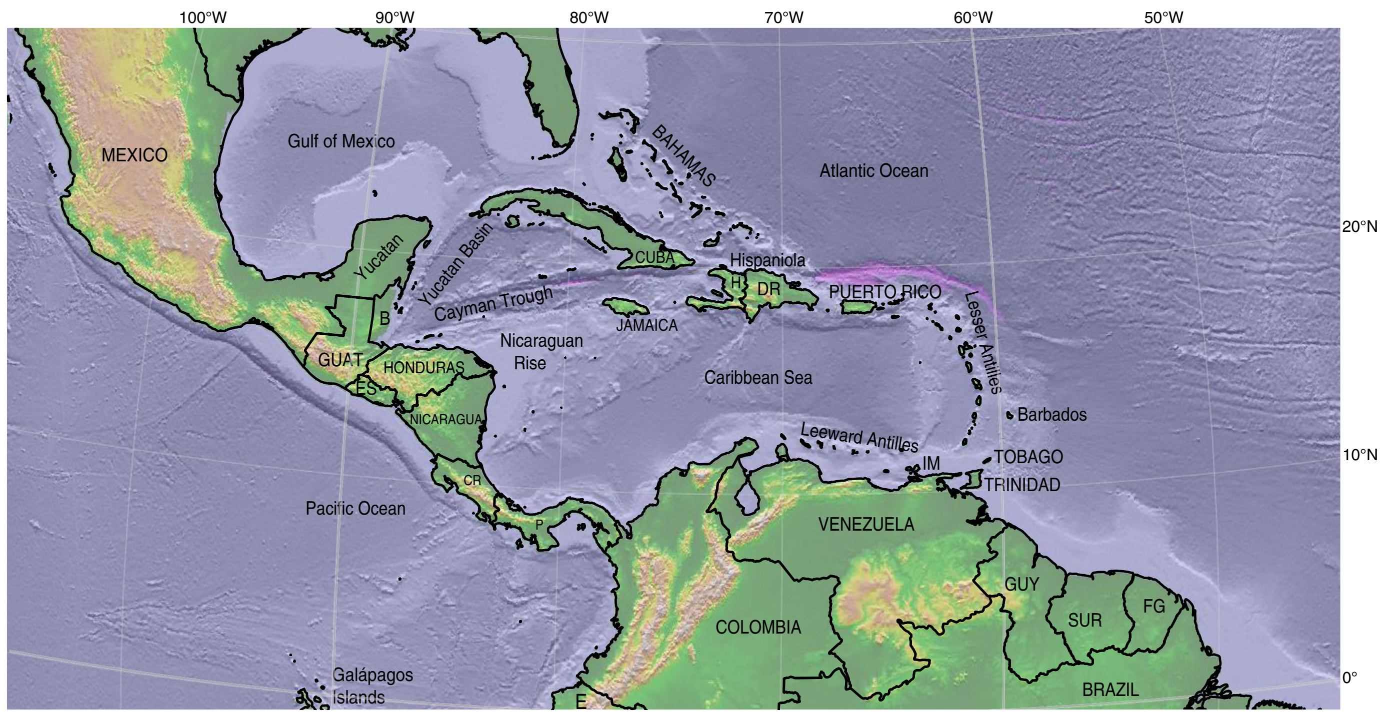

Fig. 1. Geographic map (3D Globe projection) of the Caribbean region. Countries: C, Cuba; H, Haïti; DR, Dominican Republic; FG, French Guyana; Sur, Suriname; Guy, Guyana; E, Ecuador; P, Panamá; ES, El Salvador; Guat, Guatemala; B, Belize. IM, Isla Margarita.

therefore composed an interpretation hierarchy of data types, arranged from higher to lower degree of certainty (Table 1). After interpretations from extensional records from the Atlantic and Cayman Trough we first consider interpretations from continental transform and strike-slip records. The precise amount of fault displacement may not be known in cases, but the presence of strike-slip faults provides hard constraints on the orientation and sense of relative motion. Thirdly, we consider continental extension. Even though continental extension is more difficult to reconstruct than oceanic spreading, it provides a maximum geological record at the end of deformation. Subsequently, continental collisional records provide constraints on the age and nature of collision. In collision zones, crustal material can (and frequently will) disappear by subduction, but remnants of the preceding situation can be preserved in the form of ophiolites, high-pressure (HP) metamorphic rocks, syn-kinematic sedimentary basins, or thin- or thick-skinned fold–thrust belts. We then consider interpretations based on volcanic data. Here, we particularly focus on ages of magmatism (e.g., oceanic island arc rocks), to infer the presence of a convergent plate boundary for that period. The geochemical literature of the Caribbean contains a large number of models invoking major geodynamic events such as subduction polarity reversals or slab break-off episodes, and complex subduction zone configurations (e.g., Meschede and Frisch, 1998; Kerr et al., 1999; Neill et al., 2011; Hastie et al., 2013). Such interpretations are not a priori followed here, but rather, our reconstruction may serve as an independent kinematic basis to evaluate the kinematic feasibility of such scenarios. The before-last category is the existence of geological and geophysically imaged features like magnetic boundaries, basement ages or comparable stratigraphic sections as potential correlation tools across tectonic boundaries. Finally, we use paleomagnetic data. Paleomagnetic data are normally high-quality quantitative constraints in kinematic reconstructions, but in the Caribbean area, only few data are available, and not always of the quantity and quality that passes modern quality criteria. Furthermore, paleomagnetic results from the northern and southern plate boundary zones often represent local strike-slip rotations, instead of large scale block rotations (MacDonald, 1880; Mann and Burke, 1984; MacDonald et al., 1996).

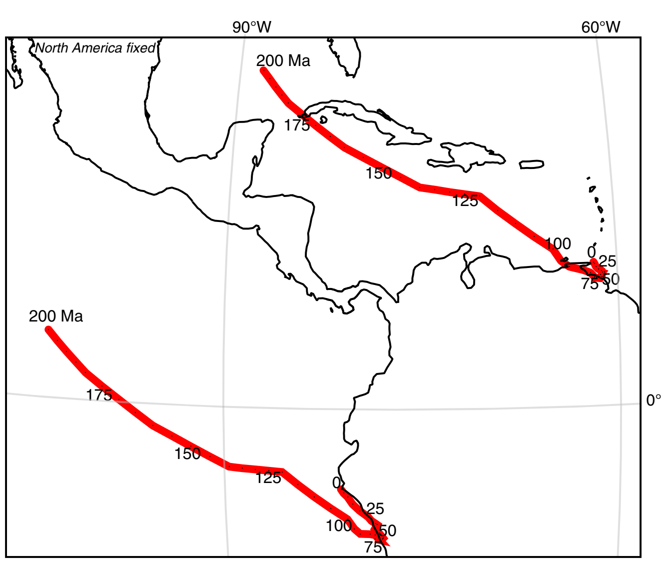

Fig. 2. 200–0 Ma motion paths for two locations on the South American continent, relative to a stable North America, based on the plate circuit of Torsvik et al. (2012). 3D Globe projection. The distance between two black dots represents 5 million years.

Our hierarchy is chosen such that the uncertainty of kinematic interpretations based on these increases with every next step. This ranking allowed us to choose between contradicting interpretations. We stress, however, that this hierarchy is used to choose between preferred interpretations of certain data sets: we have not discarded data. Our reconstruction is at every step tested against the basic concepts of plate tectonics (Cox and Hart, 1986). We took the following, conservative assumptions: (1) A tectonic terrane/block is rigid, unless there is evidence for the contrary. (2) A tectonic terrane can only move independently relative to the major plates if it is bordered by plate boundaries and conversely: if two blocks have different movements, there must be a plate boundary in between. (3) Triple junctions are stable, or, for a short period of time, unstable and transforming, or falling apart into two stable triple junctions. (4) Without evidence for a subduction

Table 1
The reconstruction of the Caribbean region is made using a hierarchy of data types, in decreasing order of certainty. Where interpretations based on different data types are mutually exclusive, this hierarchy is used to select the preferred interpretation.

Data type	Application	Reference
Oceanic extensional records	Atlantic; Cayman Trough	Torsvik et al. (2012); Leroy et al. (2000)
Transform and strike-slip records	Motagua; Cuba; Andes	Burkart (1983, 1994); Rosencrantz (1990), Cruz-Orosa et al. (2012a); Kerr et al. (1998), Trenkamp et al. (2002)
Continental extensional records	Nicaragua	Phipps Morgan et al. (2008)
Continental collisional records	Siuna–Chortis; Great Arc–Yucatan; Cuba–Bahamas; Great Arc–South American continent	Venable (1994), Rogers et al. (2007); Pindell and Kennan (2009), Ratschbacher et al. (2009), Martens et al. (2012), Solari et al. (2013); Meyerhoff and Hatten (1968), Knipper and Cabrera (1974), Pardo (1975), Iturralde-Vinent et al. (2008); e.g. Kennan and Pindell (2009)
Volcanism	Subduction related (Great Arc of Caribbean; Lesser Antilles Arc; Central American Arc)	E.g. Stanek et al. (2009); Briden et al. (1979); Denyer et al. (2006), Buchs et al. (2010)
	Oceanic plateau material (CLIP)	E.g. Sinton et al. (1998)
Matching basement types/magnetic boundaries, etc.	Chortis–Western Mexico	Rogers et al. (2007)
Paleomagnetism	Cuba; Chortis; Aruba and Bonaire; Tobago; Puerto Rico	Tait et al. (2009); Molina Garza et al. (2012); Stearns et al. (1982); Burmester et al. (1996); Reid et al. (1991)

polarity reversal and a straightforward kinematic evolution of such a reversal, the direction of subduction is assumed to remain the same. These assumptions ensure that a tectonic terrane will always be part of one of the large tectonic plates (the North or South American, Caribbean or Pacific plate), unless there is evidence for microplate motion and ensure that we always reconstruct the most simple plate kinematic scenario, within the restrictions provided by the geological data.

These assumptions provide a test for the geological and paleomagnetic interpretations and limit the amount of contradicting reconstruction scenarios. When geological interpretations were clearly inconsistent with the basic rules of plate tectonics, we refuted these interpretations in order to keep the reconstruction kinematically consistent. The final plate-tectonic model is supported by as many geological data and interpretations as possible, but the 'rules of plate tectonics' are considered to be superior to regional interpretations and scenarios based on individual tectonic terranes.

We used the freely available software package GPlates to make the reconstruction (http://www.gplates.org; Boyden et al., 2011). The surface of the Caribbean plate is divided into undeformable polygons, reconstructed using a reconstruction tree starting with North America (except for the NW Andes, which are reconstructed relative to South America). When reconstructing back in time, the polygons may overlap (indicating extension) or drift apart (indicating shortening). GPlates interpolates motion of polygons with constant rates between constrained situations, leading to a visualization and mathematical description of continuous plate motion. The rotation and shape files are given in the online appendix.

3. Previous work and correlation concepts

In previous studies of the Caribbean region, several ideas and concepts were developed that may form a useful basis for correlation in our reconstruction, and which will be used in the subsequent review of geological data. These concern Burke's (1988) 'Great Arc of the Caribbean', García-Casco et al.'s (2008a) 'Caribeana continental promontory' and the Caribbean Large Igneous Province (CLIP).

3.1. Great Arc of the Caribbean

Volcanic arc material is present along the perimeter of most of the present-day Caribbean plate (e.g. in Central America, on the Greater Antilles (Cuba, Jamaica, Hispaniola and Puerto Rico), on the Aves ridge and the Lesser Antilles and in the Northern Andes; Figs. 1, 3). Mann and Burke (1984) and Speed (1985) were the first to suggest that there may be a coherence between different arc segments, based on similarities in age and composition between arc material on the Greater Antilles, the Aves ridge–Lesser Antilles Arc system and the Netherlands Antilles, in contradiction with the then prevailing concept of independent island arcs developed in different times at different places. Burke (1988) introduced the term 'Great Arc of the Caribbean', being a volcanic arc that developed at the subduction plate boundary between the future Caribbean plate and the Proto-Caribbean Ocean (that was part of the North and South American plates and connected to the Central Atlantic Ocean) and that migrated towards the east relative to the Americas. As the Great Arc entered the Atlantic realm, the northern segment collided progressively from west to east with Yucatan and the Bahamas platform, leaving arc material behind in Guatemala and on the islands of Cuba, Hispaniola, Jamaica and Puerto Rico. The central segment, the present-day Aves ridge, remained in this concept an active subduction system while the southern segment collided with the northwestern margin of South America, leaving tectonic slivers of arc material behind in Ecuador and Colombia. Subsequent transform motion transported fragments of Great Arc material to Venezuela, the Leeward Antilles and Tobago (Burke, 1988; Fig. 3). Our reconstruction tests this hypothesis and, in addition to other sources of information, uses fragments of the Great Arc as kinematic markers of (particularly strike-slip) displacements.

3.2. Caribeana continental promontory

Arc material of the Great Arc of the Caribbean is found on the Greater Antilles to intrude in and overlie ophiolitic complexes, which in turn overlie accretionary wedges of deformed sediments, all related to Proto-Caribbean subduction and subsequent collision of the Caribbean plate with the Yucatan and Bahamas borderlands. Joyce (1983) and Wadge et al. (1984) found evidence for a fourth element in the subduction complexes, defined by García-Casco et al. (2008a) as the continental 'Caribeana' terrane. Caribeana is described as 'a conceptual paleogeographic domain characterized by Mesozoic sedimentary piles that occupied a portion of the Proto-Caribbean oceanic domain' (García-Casco et al., 2008a). The rocks of Caribeana are HP–LT metamorphic (up to eclogite facies) metasedimentary and -volcanic rocks found in thrust piles with a metamorphic grade decreasing within every next tectonic unit (e.g. in the Sierra de Escambray on Cuba). These rocks are found in, form west to east, the Cangre, Pinos, Escambray, Asunción (on Cuba) and Samaná (on Hispaniola) terranes (Figs. 3 and 4). Geophysical data and samples dredged from the ocean floor also provide evidence of metasedimentary complexes offshore eastern Yucatan and offshore eastern Hispaniola, northern Puerto Rico and the northeastern Virgin Islands (East Yucatan terrane and Puerto Rico Trench terrane) (García-Casco et al., 2008a). Although these rocks were previously interpreted to have been derived from the Bahamas borderlands, the accretionary prism of Cuba demonstrates that the Caribeana rocks of Cuba were underthrusted by oceanic sedimentary rocks between ~60 and ~45 Ma

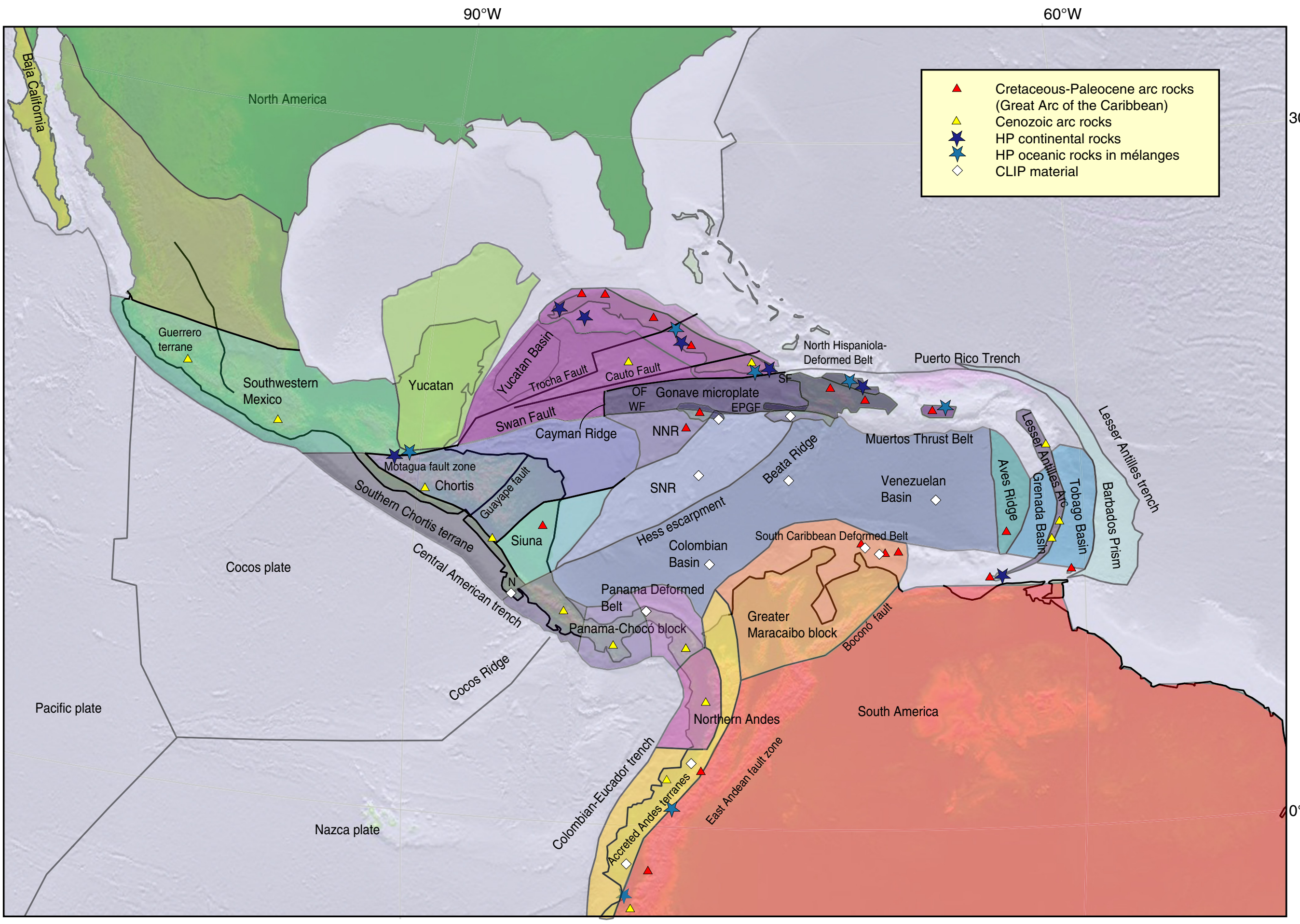

Fig. 3. Tectonic map of the Caribbean region, showing the tectonic blocks as described in Section 3 and locations of arc material (from Iturralde-Vinent et al., 2008; Krebs et al., 2008; Buchs et al., 2010), HP rocks (from García-Casco et al., 2006, 2008a; Iturralde-Vinent et al., 2008; Krebs et al., 2008) and CLIP material (from Geldmacher et al., 2003; Buchs et al., 2010; Lewis et al., 2011). Tectonic features: OF, Oriente Fault; WF, Walton Fault; SF, Septentrional Fault; EPGF, Enriquillo–Plantain Garden Fault; NNR, North Nicaraguan Rise; SNR, South Nicaraguan Rise; N, Nicoya Peninsula.

before collision of the Caribbean plate, overlying arc and accreted Caribeana fragments with North America (García-Casco et al., 2008a; Iturralde-Vinent et al., 2008; van Hinsbergen et al., 2009). Caribeana is thus envisaged as a NW–SE elongated submarine domain with (stretched) continental basement, likely positioned as a promontory of the southeastern Yucatan block. We will use the Caribeana HP belt as correlation marker in our reconstruction, in tandem with the Great Arc of the Caribbean.

3.3. Caribbean Large Igneous Province (CLIP)

The majority of the oceanic crust of the Caribbean plate is anomalously thick (15–20 km), as measured by seismic refraction studies (Burke et al., 1978). Exceptions are the Cayman Trough and the Yucatan, Grenada and Tobago Basins (Figs. 1 and 3). Deep sea drilling (DSDP and ODP) into this thick crust showed that the ocean floor consists of basaltic rocks, mainly tuffs and fine-grained intrusives, interbedded with roughly 80 Ma pelagic sediments (Donnelly et al., 1973). Based on these results, Donnelly et al. (1973) proposed that a large basalt-flooding event occurred in the Late Cretaceous. This flood basalt formed a major oceanic plateau, now known as the Caribbean Large Igneous Province (CLIP) (Burke et al., 1978; Burke, 1988; Saunders et al., 1996). The same material as found in drill cores from the Venezuelan and Colombian basins and the Beata Ridge has also been found on-land (e.g. on Hispaniola, Costa Rica, Panama, and in the Northern Andes, see Fig. 3). Radiometric dating of DSDP and ODP drill samples and of samples from Haiti, Curaçao and Western Colombia suggests that the massive basalt flooding happened between 91 and 88 Ma (Kerr et al., 1997; Sinton et al., 1997, 1998; Hauff et al., 2000; Révillon et al., 2000; Hoernle et al., 2002). This short-lived magmatic event has often been attributed to the plume-head stage of the Galápagos hotspot (e.g. Duncan and Hargraves, 1984; Hill, 1993). Hoernle et al. (2004), however, propose a 70 Myr (139–69) history for the CLIP, but the deviating ages are found in multiple different oceanic igneous structures, accumulated on the western boundary of the Caribbean plate during the subduction of the Farallon plate. An example is the magmatic rock suite of the Nicoya Peninsula of western Costa Rica, containing fragments of plateau basalt and hotspot volcanoes of ages ranging between 139 and 111 Ma (Hoernle et al., 2004). In our study, these and equivalent rocks of deviating age are not considered to be part of the Caribbean Large Igneous Province, but as accreted fragments of Panthalassa/Pacific crust. The total area of the CLIP is ~$6 * 10^5$ km^2, but since part of the plateau was thrusted onto the South American margin of Colombia and Ecuador, the plateau may originally have been more than twice this size (Burke, 1988). CLIP material is used in the reconstruction as a correlation marker for the interior of the Caribbean plate.

4. Review

The following section provides a review of the geological data of the Caribbean region that can be used for kinematic restoration. A summary of the temporal and spatial constraints on key geological markers is

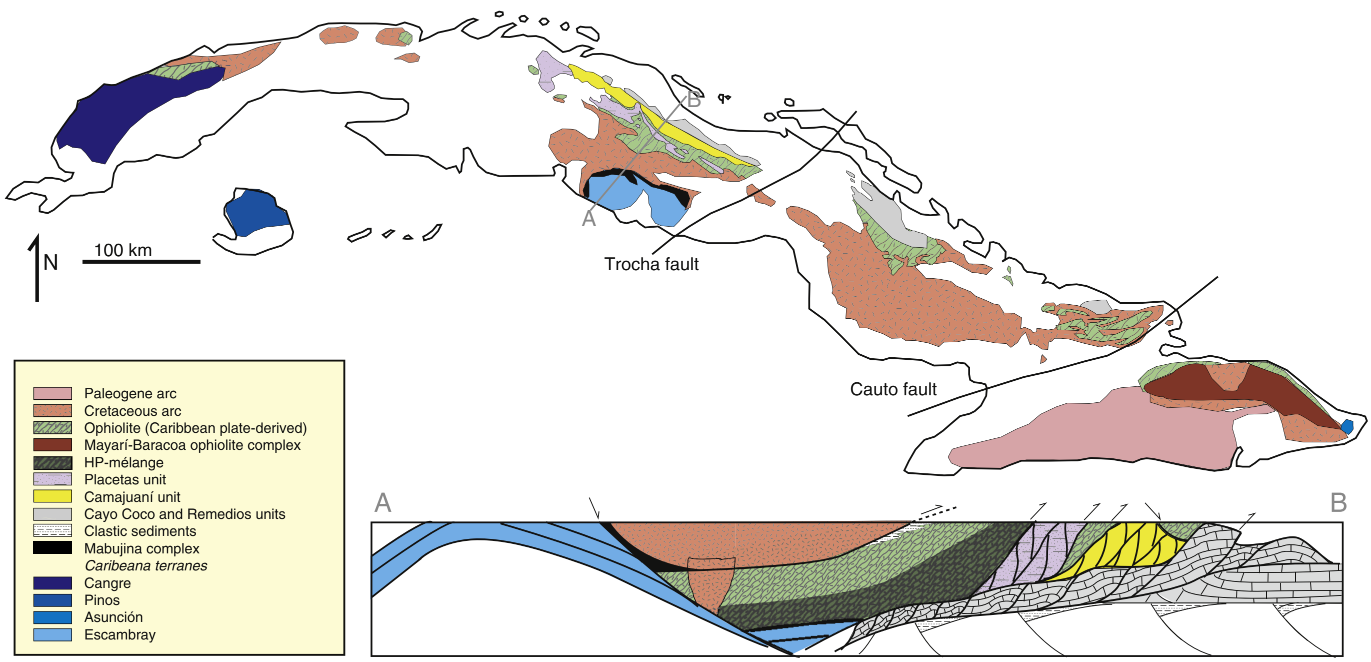

Fig. 4. Map and cross section of Cuba displaying the major geological units.
Based on García-Casco et al. (2001), Iturralde-Vinent et al. (2008), and Lázaro et al. (2013).

given in Fig. 5. The most important quantitative constraints on fault displacements provided in the literature, inferred by us in the review below, applied in the reconstruction for optimal fits, and predicted by the final model is given in Table 2. Geological rock units are described per area from old to young, and from bottom to top. The review starts with the interior of the Caribbean plate, and subsequently describes the plate boundaries clockwise, starting with Central America, the Motagua fault zone, the Cayman Trough and the Cuban segment in the north, followed by the eastern subduction system, the southern transform boundary and finalizing with the Northern Andes. In addition to the geological data, paleomagnetic data and seismic tomographic constraints on mantle structure (relevant for the subduction systems) is described.

4.1. Caribbean Sea floor

The Caribbean Sea floor contains several basins and highs with various basement types. From northwest to southeast, it contains the North and South Nicaraguan Rise, Colombian Basin, Beata Ridge, and the Venezuelan Basin (Fig. 3). Other basins and highs formed within the Caribbean plate — the Yucatan Basin, Cayman Trough, Aves Ridge, Grenada Basin, Lesser Antilles Arc and Tobago Basin — will be described later. The boundary between the Nicaraguan Rise and the Colombian Basin is a major fault scarp, the Hess escarpment, which has been interpreted as inactive since the Late Cretaceous, except for the southwestern part that has been active in recent times (Case et al., 1990; Bowland, 1993; Mauffret and Leroy, 1997). The Colombian Basin, Beata Ridge, and Venezuelan Basin are underlain by oceanic plateau crust, interpreted as part of the CLIP (Donnelly et al., 1973). Despite the presence of the CLIP, some magnetic anomalies were identified by Ghosh et al. (1984) on the ocean floor in the eastern basins, which they interpreted as typically spreading-related, although their age remains uncertain. These anomalies are currently NE–SW trending. The original orientation of the anomalies is not known, as a result of the unknown ages, and rotations of the Caribbean plate following spreading. Therefore, the probable orientation of the spreading ridge and, perpendicular to that, the direction of paleo-spreading cannot be determined accurately. The oceanic basement of La Désirade Island in the Lesser Antilles, formed in a back-arc supra-subduction setting, has a 143.74 ± 0.33 Ma U–Pb age (Mattinson et al., 2008; Neill et al., 2010), suggesting that ocean spreading continued at least until the latest Jurassic. There is no evidence for spreading within the Caribbean plate since the emplacement of the CLIP basalts (except for the Grenada Basin in the far east and the Yucatan Basin and Cayman Trough in the north), so ages of CLIP material in the Venezuelan and Colombian Basins provide a minimum age for sea-floor spreading. Since ~91–88 Ma, the Caribbean plate interior has not been increasing in size anymore and has only been reduced by subduction below South America.

The Nicaraguan Rise floors the Caribbean Sea between the Hess escarpment and the Cayman Trough and covers an area of ~$4 * 10^5$ km^2. The North Nicaraguan Rise consists of thinned continental crust with correlative outcrops in northern Central America and calc-alkaline Upper Cretaceous–Paleocene island arc rocks (Arden, 1975; Perfit and Heezen, 1978; Lewis and Draper, 1990; Lewis et al., 2011; Ott et al., 2013). The South Nicaraguan Rise is composed of thick oceanic plateau crust similar to the Venezuelan and Colombian Basins and is interpreted as CLIP material (Case et al., 1990; Mauffret and Leroy, 1997).

The role of the Nicaraguan Rise in tectonic reconstructions of the Caribbean region is controversial. Some reconstructions show 500–700 km of Eocene–Oligocene convergence between the Nicaraguan Rise and north Hispaniola (Sykes et al., 1982; Müller et al., 1999) based on paleomagnetic data suggesting as much as ~8° of latitudinal convergence between south and north Hispaniola (van Fossen and Channell, 1988). Pindell and Barrett (1990) and Pindell et al. (2012) suggested that the Nicaraguan Rise and Jamaica formed a contiguous island arc, below which several hundred kilometers of Proto-Caribbean lithosphere subducted in the Late Cretaceous. Lewis et al. (2011) postulated that an extension of the South Nicaraguan Rise was subducting below the North Nicaraguan Rise and caused Late Cretaceous magmatism in the North Nicaraguan Rise. In contrast, Mann et al. (2007) suggested strike-slip as the main process to bring the Nicaragua Rise and southern Hispaniola in place. Recently, van Benthem et al. (2013) found no evidence for Nicaraguan Rise subduction in seismic tomographic images of underlying mantle structure, confirming the model of Mann et al. (2007).

Recent GPS observations indicate that the Caribbean plate interior is not rigid, but internally deforming with 1–3 mm/yr. A two plate model best explains the data, but the plate boundary is unknown. Relative motions between the western and eastern Caribbean plate are

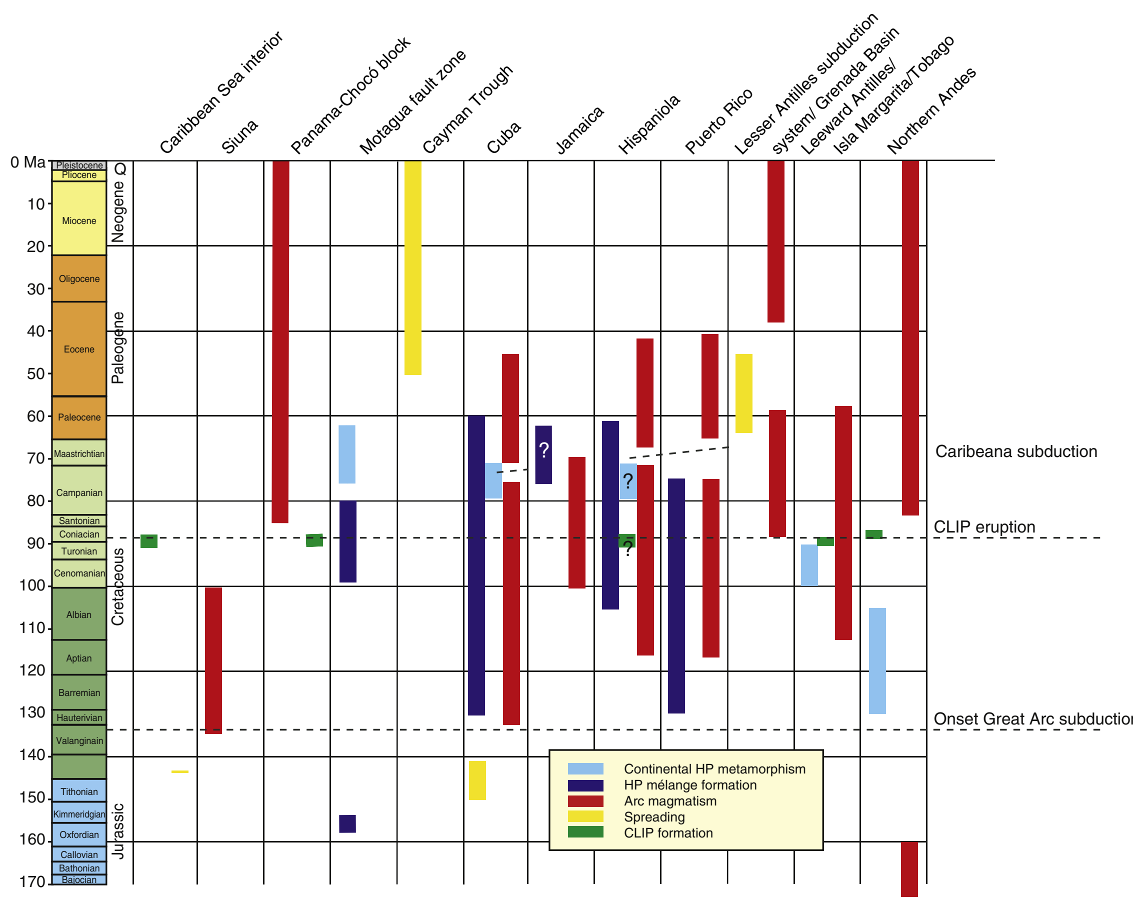

Fig. 5. Overview of temporal constraints on geological phenomena used for regional correlation per area. For references, see text.

suggested to be accommodated in the Nicaraguan Rise or Beata Ridge (Mattioli et al., 2014).

In summary, the Caribbean Sea can be divided into two parts: oceanic crust formed before emplacement of the CLIP (South Nicaraguan Rise, Colombian and Venezuelan Basins and Beata Ridge) and locally, crust formed by extension after emplacement of the CLIP (Yucatan Basin, Cayman Trough, Grenada and Tobago Basins). Paleospreading directions in modern orientations were NW–SE, and limited age constraints suggest that spreading occurred at least in late Jurassic time.

4.2. Central America

The western boundary of the Caribbean plate is the Central America Trench, accommodating eastward subduction of the Cocos plate and before the Miocene the Farallon plate (Barckhausen et al., 2008), below the Caribbean plate. The Central American land bridge can be divided into several tectonic blocks, from north to south including: the Chortis Block, the Southern Chortis terrane, the Siuna block and the Panama–Chocó block (Fig. 3).

The Chortis Block (southern Guatemala, Honduras and northern Nicaragua) exposes crystalline Paleozoic and older continental basement that was probably part of the North American continent prior to the Cenozoic. The block is bounded by the Motagua left-lateral strike-slip fault zone to the north and a geological basement transition interpreted as a former passive margin (Rogers et al., 2007) to the south and west. A major tectonic feature within the Chortis Block is the Guayape fault system, which is currently inactive (DeMets et al., 2007), but which acted as a strike-slip fault system in the Cretaceous–Miocene (Finch and Ritchie, 1991). A two-stage model was proposed by Finch and Ritchie (1991), with more than 50 km sinistral strike-slip displacement probably related to Cenozoic sinistral movement along the Motagua fault zone followed by a dextral phase of smaller displacement. The area south of the Guayape fault (named Eastern Chortis by Rogers et al., 2007), is interpreted as an extended continental margin of the Chortis Block, formed when it was still part of the North American plate during the Jurassic opening of the Proto-Caribbean ocean (Rogers et al., 2007). James (2006) interpreted the Guayape fault as an Upper Jurassic normal fault associated with rifting.

The area to the southwest of the Chortis Block is the Southern Chortis terrane (southwestern Guatemala, Honduras, Nicaragua and El Salvador). As inferred from geochemistry of Quaternary lavas (Carr et al., 2003), the Southern Chortis terrane probably has no (pre-)Paleozoic basement. One single known exposure of basement contains metavolcanic amphibolite of unknown age (Markey, 1995). The Southern Chortis terrane is interpreted as an accreted island arc (Rogers et al., 2007), similar in nature to parts of the Guerrero island arc composite terrane that fringes the continental basement of western Mexico to the northwest. The Guerrero terrane accreted to western Mexico in the Cretaceous (Tardy et al., 1994; Centeno-García et al., 2011). The age of accretion of the Southern Chortis terrane to the Chortis Block is uncertain. Because the Southern Chortis terrane is not directly relevant for reconstruction of the Caribbean plate, and the timing of accretion is unknown, we kept it fixed to the Chortis Block.

The sinistral movement along the Motagua fault system indicates an eastward movement of the Chortis Block with respect to the Yucatan block (Figs. 1 and 3). It was suggested that the Chortis Block was located along the southwestern margin of Mexico prior to its eastward displacement (e.g. Dengo, 1985; Ross and Scotese, 1988; Pindell and Barrett, 1990). Tectonic models (e.g. Ross and Scotese, 1988; Pindell and Barrett, 1990) incorporate this interpretation and suggest that the Chortis Block was transferred from the North American to the Caribbean plate by the formation of the Motagua fault. Rogers et al. (2007) summarized

Table 2
Quantitative constraints mentioned in the text on fault displacements, shortening, extension and rotation in the Caribbean region (1) provided in the literature; (2) inferred in this paper through correlation of geological units; (3) applied in our reconstruction along structures without direct geological constraints to yield an optimal fit and (4) predicted by the final model. All details of the reconstruction are given in the online Appendix in the form of rotation and polygon files for *GPlates* free software.

Structure	Displacement	Age range	Reference
Displacement estimates from literature			
Puerto Rico rotation	25°ccw rotation	11–4 Ma	Reid et al. (1991)
Maracaibo block versus South America	250 km right-lateral slip	11–0 Ma	Ross and Scotese (1988)
Chortis Block	insignificant rotation	17–0 Ma	Molina Garza et al. (2012)
Western Panama–South America motion	1000 km NE–SW convergence	35–0 Ma	Montes et al. (2012)
Tobaga, Aruba and Bonaire	90°cw rotation	65–0 Ma	Burmester et al. (1996); Stearns et al. (1982)
Panama Deformed Belt	250 km NE–SW convergence	35–0 Ma	Montes et al. (2012)
Cayman Trough	900 km	49.4–0 Ma	Leroy et al. (2000)
Displacement estimates based on geological correlation (see text)			
Nicaraguan Basin extension	100 km E–W extension	15–0 Ma	
Nicaraguan forearc	4°ccw rotation	15–0 Ma	
Nicaraguan forearc–Chortis motion	150 km right-lateral slip	15–0 Ma	
Chortis Block	32°ccw rotation	Largely 38–33 Ma	
Muertos Trench convergence	30 km N–S shortening	40–0 Ma	
La Trocha fault Cuba	20 km left-lateral slip	44–40 Ma	
Cauto fault Cuba	15 km left-lateral slip	40–38 Ma	
Septentrional Fault Hispaniola	500 km left-lateral slip	50–0 Ma	
Puerto Rico Trench	300 km left-lateral slip	40–0 Ma	
Cuba–Yucatan slip along Belize margin	900 km left-lateral slip	70–45 Ma	
Subduction erosion Cuba	140 km	70–45 Ma	
Tobago/Grenada basin extension	75 km E–W extension	55–40 Ma	
Mayarí–Baracoa ophiolite emplacement	≤135 km N–S convergence	75–70 Ma	
Mayarí–Baracoa back-arc extension	≤135 km N–S extension	80–75 Ma	
Chortis–Jamaica and North Nicaraguan Rise motion	550 km left-lateral slip	85–70 Ma	
Displacement estimates to optimize fits (see text)			
Siuna–Chortis–North Nicar. Rise and the South Nicar. Rise	300 km left-lateral slip	50–38 Ma	
Siuna–Chortis–North Nicar. Rise and the South Nicar. Rise	100 km right-lateral slip	38–32 Ma	
Hess Escarpment	150 km right-lateral slip	50–30 Ma	
Caribbean plate rotation vs North America	7°ccw rotation	50–0 Ma	
Cuban segment–Caribbean plate interior displacement	400 km left-lateral slip	70–50 Ma	
Caribbean plate rotation vs North America	17°ccw rotation	70–50 Ma	
Siuna–Chortis–North Nicar. Rise and the South Nicar. Rise	600 km left-lateral slip	70–50 Ma	
Nicaraguan Rise/offshore Chortis Block extension	100 km N–S extension	75–70 Ma	
Caribbean plate rotation vs North America	9°ccw rotation	100–70 Ma	
Caribbean–North America motion	900 km left-lateral motion	135–100 Ma	
Caribbean plate rotation vs North America	5°ccw rotation	135–100 Ma	
Model predictions			
Caribean–South America strike-slip	1000 km right-lateral slip	50–0 Ma	
SCDB subduction (eastern Venezuela)	250 km	50–0 Ma	
SCDB subduction (Colombia)	800 km	50–0 Ma	
Caribbean–South American motion; transform component	700 km right-lateral slip	70–50 Ma	
Caribbean–South American motion; subduction component	500 km convergence	70–50 Ma	
Caribbean–South American motion; transform component	1300 km right-lateral slip	100–70 Ma	
Caribbean–South American motion; subduction component	200 km convergence	100–70 Ma	
Pre-drift extension in Proto-Caribbean basin	300–400 km	200–170 Ma	

arguments for this scenario and suggested that the Chortis–Southern Chortis terrane boundary corresponds to the west Mexican continental basement–Guerrero terrane boundary. This correlation provides an estimate of displacement of the Chortis Block relative to North America of ~1000 km since formation of the Motagua fault. Because of the absence of evidence for convergence between southern Mexico and the Chortis Block during displacement and because the southern Mexican margin is oriented WNW–ESE, the motion of Chortis towards the E–W striking Motagua fault zone would require a counterclockwise rotation of the Chortis Block (Rogers et al., 2007). Paleomagnetic data from Miocene volcanics in western Honduras (17–14 Ma) show no significant rotation of these volcanics and constrain any rotation to pre-Middle Miocene (Molina Garza et al., 2012).

South of the Chortis composite terrane is the Siuna block, composed of volcanics, serpentinized peridotite and associated ultramafic cumulates, and carbonate-rich sediments. The Siuna block is interpreted as a Lower Cretaceous island arc, probably part of the Great Arc of the Caribbean, developed on ocean floor and accreted to (thrusted over) the Chortis Block in the Late Cretaceous (Venable, 1994; Rogers et al., 2007).

The southern part of Central America, between the Siuna block and the South American continent, is known as the Panama–Chocó block and contains the countries Panama and Costa Rica. Boundaries of this block are the Panama Deformed Belt in the northeast, the Central American trench in the west (where the Cocos plate, Cocos Ridge and Nazca plate are subducting), a diffuse thrust belt in the Cordillera Central of Costa Rica in the northwest, and the suture with the South American continent in the southeast (Buchs et al., 2010). The bulk of the Panama–Costa Rica land bridge is composed of upper Campanian–Neogene volcanic arc rocks (Denyer et al., 2006; Buchs et al., 2010; Montes et al., 2012) underlain by oceanic plateau crust, as suggested by geochemical characteristics of recent magmas (Feigenson et al., 2004; Gazel et al., 2009). $^{40}Ar/^{39}Ar$ ages of oceanic plateau fragments range from 139 to 71 Ma (Sinton et al., 1997; Hoernle et al., 2002, 2004). Middle Turonian–Santonian and Coniacian–Santonian ages of radiolarites intercalated with arc-derived volcanic material from the Nicoya Peninsula of Costa Rica indicate that the volcanic arc has been active since at least the Santonian (Bandini et al., 2008). Geochemical data suggest that at ~75 Ma, a protoarc formed (represented by dikes and lava flows of the Azuero Protoarc Group with a geochemistry similar to CLIP material). In the Maastrichtian, the arc system matured (Lissinna et al., 2006; Buchs et al., 2010; Wegner et al., 2011). Formation of a protoarc indicates subduction initiation, formation of a western Caribbean plate boundary and therefore separation of the Farallon and Caribbean plates (Pindell and Kennan, 2009). Pindell and Kennan (2009) proposed an age of 80–88 Ma for subduction initiation, considering that a slab needs several million years to reach depths where melt can be generated. The Pacific margin of Costa Rica and Panama is characterized by accreted fragments of Upper Triassic to Middle Miocene radiolarites and volcanic rocks from seamounts derived from the far interior of the Farallon/Cocos plate (Feigenson et al., 2004; Denyer et al., 2006; Baumgartner et al., 2008). Prior to collision with the South American continent, the Panama–Chocó block formed a straight volcanic arc, that started to segment into the western, central, eastern and Greater Panama blocks during Late Eocene–Early Oligocene times (~28–38 Ma, Farris et al., 2011; Montes et al., 2012). Segmentation and deformation was achieved by vertical axis rotation of the individual blocks and local folding and faulting (Rockwell et al., 2010; Montes et al., 2012). In our reconstruction, we follow the model of Montes et al. (2012) for the Panama–Chocó block.

Finally, the western margin of Central America, including the Southern Chortis terrane is presently moving as a forearc sliver relative to the Caribbean plate (Von Huene et al., 1980; Ranero et al., 2000). This forearc sliver contains the Nicaraguan Basin, where extension was recorded in basins filled by volcanics ranging in age from 25 Ma to Recent (Phipps Morgan et al., 2008; Molina Garza et al., 2012). Phipps Morgan et al. (2008) interpreted these volcanics as recording 100 km of E–W extension in the last 15 Myr. GPS studies show that the eastern part of the Chortis Block (eastern Honduras and Nicaragua) is currently moving with the same eastward velocity and direction relative to North America as the interior of the Caribbean plate. The westernmost part of the Chortis Block, on the other hand, is moving with a slightly lower eastward relative velocity, resulting in arc-normal extension (DeMets et al., 2007). Furthermore, DeMets (2001) shows that the forearc sliver is transported northwestward relative to the Caribbean plate. Slip directions of earthquakes on the boundary between the forearc sliver and the Chortis Block are deflected 10° clockwise from the plate convergence direction, indicating partitioning of oblique Cocos–Caribbean plate convergence (DeMets, 2001; LaFemina et al., 2009). LaFemina et al. (2009) suggested that the collision of the Cocos Ridge contributes to the northwestward motion of the forearc sliver. The buoyant, thickened crust of the Cocos Ridge acts as an indenter (Gardner et al., 2013) and arc-parallel forearc motion, as well as relative (north)eastward motion of the Panama–Chocó block represents tectonic escape (Kobayashi et al., 2014). Phipps Morgan et al. (2008) noticed that the Motagua fault zone at present does not crosscut the Central American forearc and does not continue towards the Central American trench. With plate rigidity, the south Mexican trench–Central American trench–Motagua transform triple junction should move eastward along the Motagua transform. Phipps Morgan et al. (2008) suggested that the extension in the Nicaraguan Basin and the northwestward motion of the forearc sliver are the result of the stationary position of this triple junction due to a strong, undeforming Cocos plate. The triple junction is unstable (see Fig. 2b, c of Phipps Morgan et al., 2008) and because the Cocos plate is not tearing, the eastward movement of the Caribbean plate relative to North America is not accommodated in the Cocos plate, but by internal deformation in the two overriding plates. This deformation is manifested in transpressional structures in the North American plate (Guzmán-Speziale, 2010) and extension and forearc motion in the Caribbean plate ("zipper" process, Authemayou et al., 2011).

In summary, the Central American land bridge consists of three types of crust: continental (Chortis terrane), volcanic arc upon regular oceanic crust (Siuna block, southern Chortis terrane) and volcanic arc upon oceanic plateau crust (Panama–Chocó block). Subduction initiation of the Farallon plate below the Caribbean plate in the Panama–Chocó Block occurred ~88–80 Ma. Caribbean–North American plate motion culminated in thrusting of the Caribbean plate over the southern Chortis margin in late Cretaceous time, and since the Cenozoic formation of the Motagua fault zone, the Chortis Block moved eastwards over ~1000 km towards its modern position.

4.3. Motagua fault zone

The northern boundary of the modern Caribbean plate is a left-lateral transform boundary that extends from western Guatemala to the Lesser Antilles subduction zone (Fig. 3). The western part of the plate boundary consists of a continental arcuate strike-slip fault system in central Guatemala and the oceanic Cayman Trough. This intracontinental fault system separates the North American Yucatan block (southern Mexico, Belize and northern Guatemala, also referred to as the Maya block) in the north from the continental Chortis Block in the south, which at present belongs to the Caribbean plate (see above).

The Motagua fault zone exposes a complex amalgamation relict ocean floor, metamorphic complexes, and volcanics, in roughlywest–east trending belts, separated by faults and shear zones (Ratschbacher et al., 2009; Fig. 6). The four main features in the fault zone are, from north to south, the Polochic fault, the Baja Verapaz shear zone (a 5–10 km wide greenschist facies mylonite zone), the Motagua fault and the Jocotán fault (Ortega-Obregón et al., 2008; Ratschbacher et al., 2009).

The basement of the Yucatan block is exposed in the Santa Rosa Group north of the Polochic fault and in the Rabinal complex between the Polochic fault and the Baja Verapaz shear zone. The Santa Rosa Group contains Paleozoic sediments and locally some felsic intrusions. The Rabinal complex contains Lower Paleozoic, low-grade volcanosedimentary rocks and granitoids (Ratschbacher et al., 2009; Solari et al., 2013). Metamorphism of the Rabinal Granite is dated at 70.1 ± 0.6 Ma (^{40}Ar–^{39}Ar white mica age, Solari et al., 2013). There is no obvious difference in Paleozoic lithostratigraphy and magmatism between the Santa Rosa Group and the Rabinal complex (Ortega-Gutiérrez et al., 2007; Ratschbacher et al., 2009).

South of the Motagua fault, two metamorphic complexes are defined that are interpreted as the basement of the Chortis Block: the Sanarate complex in the west and the Las Ovejas complex in the east (Ratschbacher et al., 2009). The Sanarate complex contains Jurassic metapelites and the Las Ovejas complex is characterized by Precambrian–Paleozoic volcano-sedimentary rocks that have undergone Cenozoic amphibolite facies metamorphism (Ratschbacher et al., 2009).

Between the Baja Verapaz shear zone and the Motagua fault to the south, the Chuacús complex is exposed, containing Paleozoic–Triassic, high-grade metamorphic volcano-sedimentary rocks and granitoids. The Chuacús complex has been interpreted as a part of the Yucatan block (Dengo, 1969; Donnelly et al., 1990; Ratschbacher et al., 2009), or as the separate Jacalteco terrane accreted to the Yucatan block by shearing along the Baja Verapaz shear zone (Ortega-Gutiérrez et al., 2007; Ortega-Obregón et al., 2008; Solari et al., 2011). (Ultra-)high pressure metamorphism in the Chuacús complex has been dated at ~76–62 Ma (U–Pb, Rb–Sr and ^{40}Ar–^{39}Ar cooling ages of white mica and amphibole; Ratschbacher et al., 2009; and references therein; Martens et al., 2012). The Baja Verapaz shear zone is a 5–10 km wide, south dipping shear zone, thrusting (with a minor sinistral strike-slip component) the Chuacús complex onto pre-Silurian low-grade metasedimentary rocks of the Yucatan block (Ratschbacher et al., 2009). Shearing in the Baja Verapaz shear zone is simultaneous with metamorphism in the Chuacús complex, dated by 74–66 Ma white micas from sheared shales (K–Ar dating, Ortega-Obregón et al., 2008) and ~70 Ma white micas in mylonite gneiss (Ratschbacher et al., 2009).

The metamorphism in the Rabinal Granite and the Chuacús complex, and shearing in the Baja Verapaz shear zone is interpreted to result from

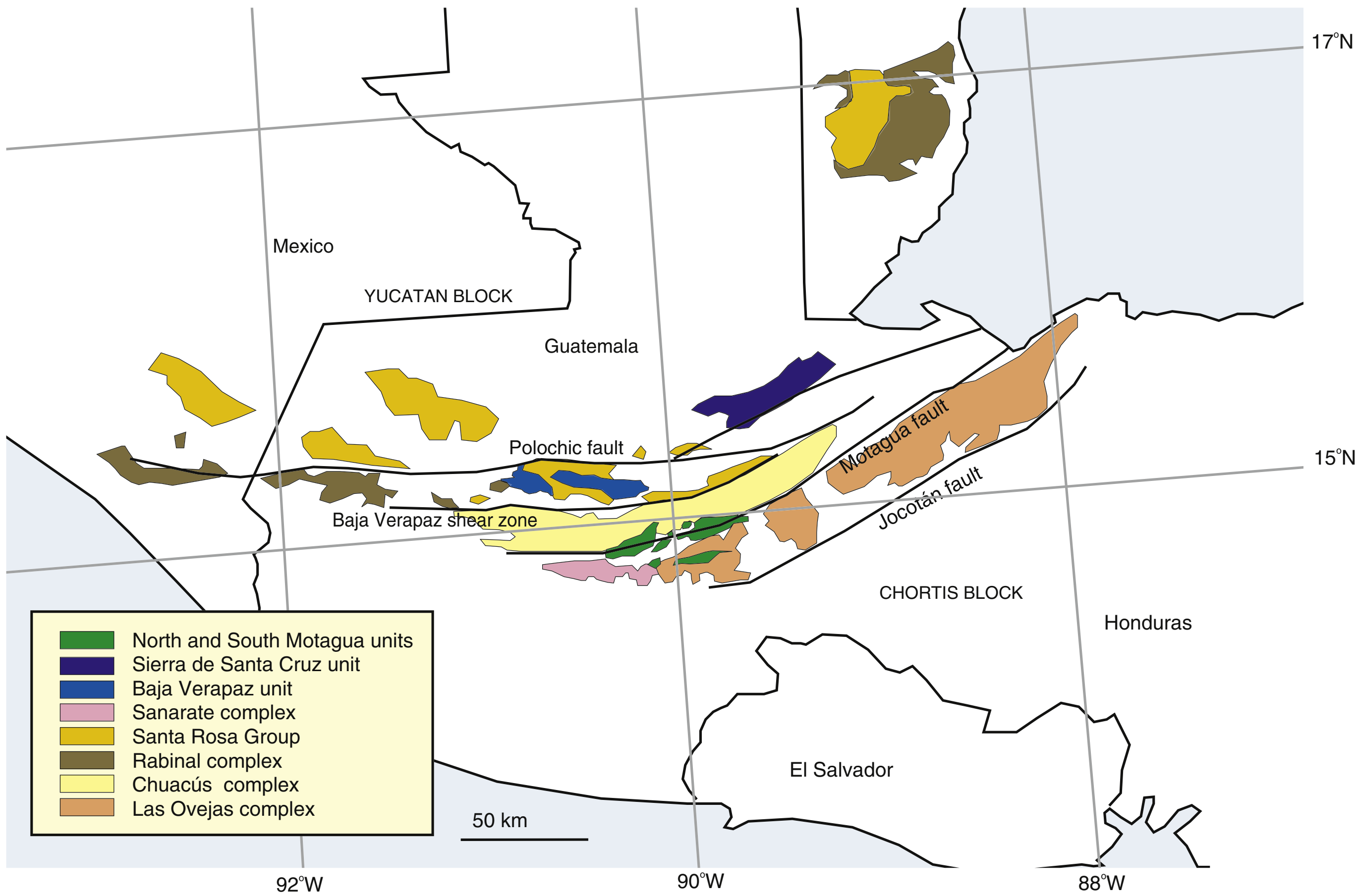

Fig. 6. Basement geology of the Motagua fault zone.
Based on Ratschbacher et al. (2009).

Campanian–Maastrichtian oblique collision of the Great Arc of the Caribbean thrusting over the southern margin of the Yucatan block. During collision, continental material of the Yucatan promontory entered the subduction zone and was subjected to HP metamorphism (Pindell and Kennan, 2009; Ratschbacher et al., 2009; Martens et al., 2012; Solari et al., 2013).

Overlying the Yucatan block are two northward-emplaced harzburgite-dominated ophiolitic nappes of the El Tambor complex: the Sierra de Santa Cruz unit in the east, north of the Polochic fault and the Baja Verapaz unit south of the Polochic fault (Giunta et al., 2002a; Solari et al., 2013). These units comprise serpentinized harzburgites, layered gabbros, dolerites and scarce basalts with island-arc affinity (Giunta et al., 2002a; Ratschbacher et al., 2009). The Baja Verapaz unit overlies the Chuacús complex and the Sierra de Santa Cruz unit overlies Maastrichtian–Danian turbidite fan successions of the Sepur Formation (Wilson, 1974; Beccaluva et al., 1995; Giunta et al., 2002a). This formation is interpreted to have formed in a foreland basin and is the youngest unit involved in deformation produced by thrusting of nappes. The arrest of convergence between the ophiolite units and the Yucatan peninsula is therefore constrained by the age of the Maastrichtian–Danian Sepur Formation (Martens et al., 2012; Solari et al., 2013).

Also part of the El Tambor complex are the North and South Motagua units, north and south of the Motagua fault. The North and South Motagua units contain HP/LT serpentinite-matrix mélanges (Harlow et al., 2004) and ophiolitic sheets of MORB-type geochemical affinity (Beccaluva et al., 1995; Giunta et al., 2002a). The North Motagua unit overlies the Chuacús complex of the Yucatan block and the South Motagua unit overlies the Chortis Block. These units are unconformably overlain by Eocene continental molasses of the Subinal Formation (Giunta et al., 2002a). ^{40}Ar/^{39}Ar cooling ages from blueschist blocks within the serpentinites are 77–65 Ma for the northern, and 124–113 Ma (Harlow et al., 2004) to 144–132 Ma (Brueckner et al., 2009) for the southern mélange. U–Pb zircon dating of jadeitites, phengite jadeitites and mica-albite rocks provided ages of ~98–80 Ma and ~154–158 Ma for the North and South Motagua unit mélanges, respectively. These ages are interpreted to represent the age of crystallization of the rocks during active subduction (Flores et al., 2013). The mélange units are either interpreted to have formed in two subduction–collision events (one in the Late Jurassic–Early Cretaceous and the other in the Late Cretaceous), followed by juxtaposition through eastward movement of the Chortis Block along the Motagua fault zone (e.g. Harlow et al., 2004), an interpretation we will follow in our restoration. Alternatively, these data have been interpreted to reflect a long-lasting single subduction event (e.g. Brueckner et al., 2009). According to the latter authors, exhumation of the southern mélange in the Early Cretaceous caused them to be unaffected by later events, whereas the northern mélange was reworked, or accreted below the southern, in a Late Cretaceous subduction event.

The youngest part of the history of the Motagua fault zone is the formation of a sinistral brittle strike-slip duplex with several restraining and releasing bends, displacing the Chortis Block relative to the Yucatan block (Giunta et al., 2002a). Burkart (1994) proposed that the Jocotán fault was active from 20 Ma to 10 Ma, the Polochic fault from 10 Ma to 3 Ma, and the Motagua fault from 3 Ma to present, the latter being the present-day plate boundary. Despite attempts to quantify the displacement along the faults (e.g. Burkart, 1983; Donnelly et al., 1990), there is no estimate of cumulative offset by direct correlation across the fault segments. We will therefore use the correlation of the western continental margin of Chortis to the western continental margin of Mexico as proposed by Rogers et al. (2007). The present displacement rate along the Motagua fault zone is ~21 mm/year (DeMets et al., 2007).

In summary, the plate boundary zone between the Yucatan and Chortis Block records events starting with HP metamorphism between 158 and 124 Ma in the South Motagua unit as a result of subduction of oceanic crust connected to the Chortis Block below southwestern Mexico. HP metamorphism in the North Motagua unit indicates subduction of oceanic crust with an opposite polarity underneath the Great Arc since the early Late Cretaceous until ~70 Ma, when an ophiolite and overlying arc (represented by the North Nicaraguan Rise and Jamaica (see below)) collided with southern Yucatan. This oblique collision resulted in a phase of sinistral shearing in the Baja Verapaz shear zone and burial and HP–LT metamorphism of the Yucatan passive margin (i.e., the Chuacús complex). Left-lateral strike-slip faulting in the Cenozoic moved the Chortis Block to the east and positioned the North and South Motagua units adjacent to each other. The short-lived character of the arc systems in the Motagua fault zone indicates that they are not part of the Great Arc of the Caribbean, which is located further south (Siuna block).

4.4. Cayman Trough

The Cayman Trough is an oceanic pull-apart basin that formed along the eastward extension of the Motagua fault zone. It is a rectangular depression on the transform plate boundary between the Caribbean plate and the Cuban segment, currently part of the North American plate, extending from the Belize margin to Jamaica. The trough is underlain by oceanic crust accreted along an ~110 km long N–S trending spreading center (CAYTROUGH, 1979). The northern and southern boundaries of the trough are the Oriente and Swan Island faults, respectively. Holcombe et al. (1973) were the first to provide evidence for seafloor spreading in the Cayman Trough and Macdonald and Holcombe (1978) identified magnetic anomalies. Although identification of the anomalies was problematic and based on models, Leroy et al. (2000) re-interpreted the anomalies to reflect A1 (0.8 Ma), A3a (6.3 Ma), A5 (11.0 Ma), A6 (19.7 Ma), A8 (26.2 Ma), A13 (33.7 Ma), A20 (42.8 Ma) and A22 (49.4 Ma) and suggested an age of 49.4 Ma for onset of opening of the trough. Early (?) Paleocene–Early Eocene rifting preceded formation of oceanic crust (Mann and Burke, 1990). We follow the interpretations of Leroy et al. (2000) and assume constant, symmetric spreading between the anomalies. The Cayman Trough is frequently assumed to reflect Caribbean–North American plate motion, but a note of caution is relevant here: in the first few Myr of opening of the Cayman Trough, the Cuban segment still moved with respect to the North American plate (until collision with the Bahamas platform, ~45 Ma, see below). Furthermore, the eastern half of the Cayman Trough is part of the Gonave block, which slightly moves relative to the Caribbean plate along the Walton–Enriquillo–Plantain Garden fault system (see Section 4.7). It is thus more accurate to state that the Cayman Trough records Cuban segment–Gonave block motion.

4.5. Cuban segment

The Cuban segment consists of the Yucatan basin and the island of Cuba, which exposes a lower North America/Proto-Caribbean-derived fold–thrust belt overlain by a Caribbean plate-derived ophiolite and volcanic arc sequence. The Yucatan basin is an oceanic basin bounded by the Cayman Trough to the south, the Yucatan peninsula to the west and Cuba to the north (Rosencrantz, 1990). It can be divided into two segments; a western deep basin and the eastern two-thirds including the topographically heterogeneous domains of the Cayman rise. The western basin is a NNE–SSW striking rectangular deep underlain by oceanic crust that formed during the Paleocene to Middle Eocene and is interpreted as a large, currently inactive pull-apart basin that formed as a result of left-lateral transform motion of the Cuban segment along the Belize margin (Rosencrantz, 1990). The Cayman rise contains volcanic arc material, probably resting upon oceanic crust of pre-Cenozoic age (Rosencrantz, 1990). Cuba is a fold–thrust belt, composed of rock sections of pre-Jurassic to Eocene age that developed as a result of convergence between the North American and Caribbean plates in Mesozoic and Cenozoic times (Iturralde-Vinent, 1988, 1994, 1996; Iturralde-Vinent et al., 2008).

To the west, the Cuban fold–thrust belt rests upon Neoproterozoic basement of the Yucatan borderlands overlain by Paleogene clastic successions, and to the north and east on the Bahamas carbonate platform (Fig. 4), which is underlain by ~200 Ma Central Atlantic Magmatic Province (CAMP)-related volcanic rocks (Somin and Millán, 1981; Renne et al., 1989; Iturralde-Vinent, 1994, 1998, 2006). Thrust slices tectonically imbricating the North American borderlands and the Proto-Caribbean basin underlie, or are incorporated in, Cretaceous–Paleocene serpentenite-matrix-hosted subduction mélanges (Iturralde-Vinent, 1994, 1998; Kerr et al., 1999; García-Casco et al., 2002, 2006; Iturralde-Vinent et al., 2008; van Hinsbergen et al., 2009). These thrust slices comprise four tectonostratigraphic units, from north to south including the Cayo Coco, Remedios, Camajuaní and Placetas units (Ducloz and Vuagnat, 1962; Khudoley, 1967; Meyerhoff and Hatten, 1968, 1974; Hatten et al., 1988). The Cayo Coco and Remedios units consist of evaporates and carbonate rocks. The Camajuaní and Placetas units contain calcareous and siliceous rocks that are interpreted as deposits from the basin between the Bahamas carbonate platform and Caribeana (Ducloz and Vuagnat, 1962; Meyerhoff and Hatten, 1968, 1974; Díaz Otero et al., 1997; Pszczolkowski and Myczynski, 2003). Collision of the Caribbean plate with the Bahamas and Yucatan borderlands started in the latest Paleocene to ~45 Ma (Meyerhoff and Hatten, 1968; Knipper and Cabrera, 1974; Pardo, 1975; Bralower and Iturralde-Vinent, 1997; Gordon et al., 1997), as shown by biostratigraphy of foreland basin deposits and was finalized by the early Late Eocene as shown by shallow marine deposits unconformably overlying the thrusted sequences (~40–35 Ma; Iturralde-Vinent et al., 2008). Folding in the Bahamas foreland continued until the Miocene (Massaferro et al., 2002). Overthrusting the thrust slices mentioned above is an ophiolite-arc terrane, outcropping south of the continental margin units (Hatten et al., 1958; Pardo, 1975; Draper and Barros, 1994; Iturralde-Vinent, 1997; Fig. 4). It contains, from north to south, an ophiolite belt, a volcanic–sedimentary complex, a plutonic complex, and the up to amphibolite-facies metamorphic Mabujina complex. The age of formation of the oceanic lithosphere of the ophiolites, containing harzburgites and other elements of the PENROSE ophiolite suite, is constrained to Late Jurassic by Tithonian (145–152 Ma) radiolarites (Fonseca et al., 1985; Iturralde-Vinent and Marí-Morales, 1988; Iturralde-Vinent, 1994, 1996; Llanes et al., 1998; Lewis et al., 2006). Ages of the igneous rocks range from 160 to 50 Ma of which only the Late Jurassic to Early Cretaceous dates are considered to represent formation of oceanic lithosphere (see review by Iturralde-Vinent et al., 1996). The ophiolitic complexes are geochemically fingerprinted to supra-subduction environments (García-Casco et al., 2006). Island Arc Tholeiite (IAT) to MORB signatures are found in the basaltic rocks from the Cajálbana ophiolite (western Cuba), metamorphosed in an Early Cretaceous (130 Ma) volcanic arc environment (García-Casco et al., 2001). Boninitic rocks of unknown age (Fonseca et al., 1989; Kerr et al., 1999) and calc-alkaline signatures are found in the northern ophiolite belt towards the paleo-forearc (Andó et al., 1996; García-Casco et al., 2001).

The northern part of the ophiolite is separated from the underlying thin-skinned fold-thrust belt by a 1.5 km thick serpentinite-matrix mélange with HP metamorphic (eclogite, amphibolite, greenschist), plutonic, ultramafic, as well as non-metamorphic sedimentary rocks (MacGillavry, 1937; Bush and Sherbacova, 1986; Somin and Millán, 1981; Millán, 1996; García-Casco et al., 2006; van Hinsbergen et al., 2009). K–Ar and U–Pb zircon ages from samples in HP–LT rocks range from ~130 to 60 Ma, suggesting that subduction started not later than in the Early Cretaceous (Somin and Millán, 1981; Somin et al., 1992; Iturralde-Vinent et al., 1996; García-Casco et al., 2006; Lázaro et al., 2009; Rojas-Agramonte et al., 2010). Aptian amphibolite blocks from eastern Cuba record anomalously high geothermal conditions (~700 °C and 14 to 15 kbar) and counterclockwise PTt-paths, suggesting that these blocks originated from very young subducted oceanic lithosphere (García-Casco et al., 2008b; Blanco-Quintero et al., 2010). Therefore, García-Casco et al. (2008b) and Blanco-Quintero et al. (2010) suggested the presence of a trench–trench–ridge triple junction, situated north of Eastern Cuba around 120 Ma. On top of the northern ophiolite is a >3 km thick volcanic–sedimentary and plutonic complex that is interpreted to belong to the Great Arc of the Caribbean (Burke, 1988; Cruz-Orosa et al., 2012b). U/Pb ages date arc magmatism to have lasted from at least 133 to 80 Ma (see review by Stanek et al., 2009; Rojas-Agramonte et al., 2011). Arc magmatism terminated in the late Campanian in west-central Cuba, but continued in east Cuba into the Paleogene (García-Casco et al., 2001, 2008a, 2008b). The volcanic arc of Cuba is located only ~20 km to the south of the suture zone and the late Cretaceous forearc, typically 166 ± 60 km wide (Gill, 1981), is largely missing. van Hinsbergen et al. (2009) concluded that this is the consequence of subduction erosion of the forearc after the Late Cretaceous to Paleocene arrest of the arc exposed on Cuba, and before the final collision of the ophiolite with the Bahamas platform in the Middle Eocene. This tectonic erosion of the Cretaceous forearc may explain the absence of post-Campanian arc-related rocks on Cuba, as this arc may have shifted southwards into the Yucatan Basin during subduction erosion.

In eastern Cuba (Fig. 4), Cretaceous volcanic arc rocks assigned to the Great Arc (the Purial metavolcanics) have been metamorphosed at blueschist facies (Boiteau et al., 1972; Somin and Millán, 1981) and are overlain by the Upper Cretaceous (Santonian) Mayarí–Baracoa ophiolite, which is associated with an underlying metamorphic sole (Güira de Jauco Amphibolite Complex; Lázaro et al., 2013). The sole formed sometime in de Late Cretaceous dated by K/Ar ages of 72 ± 3 and 58 ± 4 Ma (Somin and Millán, 1981). Peak pressures recorded in the sole are 8.5–8.7 kbar (ca. 30 km depth; Lázaro et al., 2013). Non-metamorphic arc rocks from this area contain geochemical evidence for two distinct slab components (Marchesi et al., 2007).

Lázaro et al. (2013) interpreted this sequence as evidence that subduction in eastern Cuba jumped, probably locally, from below the Great Arc into the overriding plate, towards a Campanian back-arc spreading ridge, to allow for the formation of the metamorphic sole below the Mayarí–Baracoa ophiolite and subduction-related metamorphism of the Purial complex. Subduction initiation here occurred in the late Campanian (modeled as 75 Ma) and buried the original arc below back-arc ocean floor.

In several places on Cuba, underlying the ophiolites and volcanic arc complexes are metasedimentary and meta-volcanic rocks probably derived from continental crust, exposed in the Cangre, Isla de Iuventud, Sierra de Escambray and Asunción complexes (García-Casco et al., 2001, 2006, 2008a). The Escambray complex in central Cuba contains three thrust slices of metacarbonates and quartz-mica schists with tectonic slivers of metagabbro, greenschist and serpentinite (Somin and Millán, 1981; Millán and Somin, 1985; Stanek et al., 2006). The upper thrust slice is eclogite-facies metamorphic, the middle reaches blueschist metamorphism and the lower thrust slice is lowest-grade at greenschist metamorphic. This downward decrease in metamorphic grade is interpreted as the result of stacking of HP-metamorphic thrust slices in a subduction channel during an episode of continental subduction (Millán, 1997; Stanek et al., 2006; García-Casco et al., 2008a). The metamorphic peak occurred during the latest Campanian (Schneider et al., 2004; García-Casco et al., 2006, 2008a; Stanek et al., 2006). After 70 Ma, exhumation started and HP metamorphic rocks reached the surface at about 45 Ma (Kantchev, 1978). The Escambray complex is imbricated north and south by the HT arc-related Mabujina-complex (Cruz-Orosa et al., 2012b) and its final exhumation is thought to have occurred in a metamorphic core-complex (Pindell et al., 2005; García-Casco et al., 2008a). These HP-metamorphic, continent-derived metasediments are interpreted to belong to Caribeana, which underthrusted the ophiolites ~75–70 Ma ago — simultaneously with the Chuacús complex in Guatemala, see Section 4.3, well before the final suturing of the ophiolite with North America in the Eocene (García-Casco et al., 2008a).

Finally, Cuba and the Yucatan Basin are cut by large-scale strike-slip faults, the two largest being the La Trocha and Cauto faults (Fig. 4). Being poorly exposed, the amount and nature of displacement on these faults remain poorly constrained. Rosencrantz (1990) suggested that the La Trocha fault acted as a sinistral strike-slip fault with an offset of less than 50 km. Uppermost Cretaceous sediments are cut by the La Trocha fault and syntectonic sedimentation (>1200 m of growth strata), indicating a component of normal fault motion, occurring between the Paleocene and Eocene (Cruz-Orosa et al., 2012a). A model has been proposed (Mann and Burke, 1990; Pindell and Barrett, 1990; Mann et al., 1995; Mann, 1997; Pindell et al., 2005), showing a gradual change in motion of the Caribbean plate, with Cuba at its leading edge, with respect to the North American plate during collision with the Bahamas borderlands: the Caribbean plate escaped towards a free face in the Atlantic ocean. During this process, the Caribbean–North American relative plate motion changed from NNE to E, and the plate boundary jumped stepwise from the Belize margin (west-Yucatan deep) and the subduction zone north of Cuba, to the Cauto fault and subsequently the La Trocha fault, to eventually the Oriente fault north of the Cayman Trough, leaving the Cuban segment welded to the American plate. These regional changes are recorded in the nature and structure of the La Trocha fault. The La Trocha fault evolved from a left-lateral strike-slip fault during shortening in the Cuban orogen to a post-welding normal fault (Cruz-Orosa et al., 2012a).

Paleomagnetic data from a small section of Cretaceous volcanic and sedimentary rocks in central Cuba show a counterclockwise rotation with respect to the North American plate of 43 ± 16° since the mid-Cretaceous (Renne et al., 1991) to ~70–80° between 120 and 90–45 Ma (Tait et al., 2009). These latter authors suggested a local, strike-slip related origin for these rotations.

In summary, the Cuban segment plays a crucial role in the geological history of the Caribbean region. The Cuban orogen records tectonic events starting with the onset of formation of oceanic crust in the Proto-Caribbean Ocean since ~160 Ma, followed by the establishment of a subduction zone below the oceanic Caribbean lithosphere at or before ~130 Ma. Subduction of the Proto-Caribbean oceanic crust led to the formation of the Great Arc of the Caribbean in the Early Cretaceous. In the Campanian, the Caribeana promontory reached the subduction zone. This first led to a short-lived jump of subduction to a previously opened back-arc basin and led to burial and blueschist metamorphism of Great Arc rocks below the Mayarí–Baracoa ophiolite in eastern Cuba. Ongoing Caribbean–North America convergence subsequently resulted in HP metamorphism in Caribeana-derived units with an ~70 Ma peak. Ongoing SSW-ward subduction led to a subduction transform fault along the margin of Belize, in which a releasing bend formed the western Yucatan pull-apart basin. This basin opened until the collision of the Cuban ophiolite with North America and the Bahamas platform ~45 Ma ago. Following Caribeana subduction and accretion, and prior to Eocene collision, the Placetas and Camajuani complexes were accreted from subducting Proto-Caribbean lithosphere or extended North American continental crust, followed by underthrusting of the Cayo Coco and Remedios complexes. After arrest of the Cuban arc in the late Cretaceous and before Eocene collision, the bulk (~100–200 km) of the Cretaceous forearc was removed from the Cuban ophiolite by subduction erosion. Collision was associated with a change in relative motion of Caribbean plate versus North America from NNE to E, stepwise accommodated along the increasingly more easterly trending Cauto and La Trocha faults, transferring the plate boundary towards the Cayman Trough, leaving the Cuban segment welded to the North American plate.

4.6. Jamaica

The island of Jamaica is located between the Yucatan Basin in the north and the Caribbean Sea in the south (Figs. 1, 3). Jamaica's basement consists of Lower Cretaceous to Paleocene volcanic and plutonic arc-related rocks with a similar composition as those found on the North Nicaraguan rise, and metamorphic rocks (Abbott et al., 1999; Mitchell, 2006; Hastie et al., 2010; Lewis et al., 2011). The metamorphic rocks are mainly schists (Westphalia Schist and Mt. Hibernia Schist) and serpentinites (Abbott et al., 1996, 1999, 2003). Trace element geochemistry of the Westphalia Schist is consistent with an island arc setting (West et al., 2014). The blueschist facies Mt. Hibernia Schist (cropping out east of the Blue Mountain Fault, on the eastern side of the island) has CLIP-related geochemistry and has been exhumed in Maastrichtian times, suggesting Late Cretaceous (~75 Ma) subduction of CLIP material. The HP rocks are overlain by upper Maastrichtian sediments (McFarlane, 1974; Robinson, 1994; Mitchell, 2006; West et al., 2014).

Jamaica formed as a restraining bend between the Wagwater and Blue Mountain transpressional faults. These NW-striking faults are interpreted as transpressional stepovers in the left-lateral strike-slip system of the Walton fault, an eastward propagation of the Swan Island fault to the east of the Cayman spreading center, and the Plantain Garden fault of Hispaniola. Uplift of the island started in the Late Miocene (Mann et al., 2007). Jamaica can thus be regarded as an exposed portion of the North Nicaraguan rise.

Tectonic reconstructions (e.g. Ross and Scotese, 1988; Pindell and Kennan, 2009) locate the island of Jamaica adjacent to the southern margin of Yucatan between ~70 and ~50 Ma after restoring the opening of the Cayman Trough, matching the blueschists on Jamaica with the blueschists of the Rabinal and the Chuacús Complexes and forming a link between the Great Arc rocks of Siuna and Cuba. Prior to ~70 Ma, the island was part of the active Great Arc.

4.7. Hispaniola

The island of Hispaniola is divided into the countries Haiti in the west and the Dominican Republic in the east. It is located on the transform plate boundary between the North American and Caribbean plates. Hispaniola can be divided into two tectonic domains, separated by the Enriquillo–Plantain Garden fault: the northern two-thirds and the southern one-third.

The northern two-thirds of the island contain metasedimentary rocks, interpreted as accreted continental material from the Bahamas carbonate platform overlain by an ophiolitic complex and associated arc intrusives and overlying Lower Cretaceous to Middle Eocene volcanic arc rocks. Between the accreted Bahamas platform rocks and the ophiolite is a serpentinite-matrix mélange with HP rocks (blueschists and eclogites) (Escuder-Viruete et al., 2006, 2013; Krebs et al., 2008, 2011). The ophiolitic rocks (dated at 136.4 $\pm$ 0.32 Ma, U–Pb age of Escuder-Viruete et al., 2011c) and serpentinite-matrix mélanges can be found on the northernmost part of the island, north of the Septentrional fault in e.g. the Río San Juan and Puerto Plata complexes (Draper and Lewis, 1991; Draper and Nagle, 1991; Escuder-Viruete et al., 2013). These complexes are ~50 km apart and are essentially the same but displaced by a strike-slip fault (Krebs et al., 2008). Blocks in the mélange recorded peak metamorphism at different times, ranging from 103.6 $\pm$ 2.7 Ma dated from an eclogite, 80.3 $\pm$ 1.1 Ma from an omphacite blueschist and 62.1 $\pm$ 1.4 Ma for a jadeite blueschist (Lu–Hf and Rb–Sr ages; Krebs et al., 2008). Krebs et al. (2008) interpreted the subduction zone forming these HP blocks as active between ~130 and ~55 Ma. The mélange contains blocks originating from both the upper (Caribbean) and downgoing (Proto-Caribbean Ocean) plate. This suggests that subduction erosion took place at the base of the Caribbean plate in the Late Cretaceous (Escuder-Viruete et al., 2011c). On the Samaná Peninsula, east of the Río de San Juan complex, the HP metamorphic Samaná terrane is exposed (Escuder-Viruete et al., 2011a, 2011b). This is a more coherent unit interpreted as part of Caribeana (García-Casco et al., 2008a).

In the Cordillera Central, south of the Septentrional fault, and overlying the mélange, ophiolitic and Cretaceous–Eocene metamorphic, plutonic and island-arc volcanic rocks (Tireo formation), as well as covering non-metamorphic rocks are found (Kesler et al., 1991; Escuder-Viruete et al., 2006, 2013). Escuder-Viruete et al. (2006) showed that subduction below the oceanic lithosphere from which the ophiolites were derived started shortly before 116 Ma, indicated by a volcanic complex with three units: boninites and LREE depleted IAT volcanics at the bottom, an intermediate unit dated 116 Ma, and normal IAT volcanics at the top. The non-metamorphic rocks are interpreted to have deposited in a forearc basin (Escuder-Viruete et al., 2013). Upper Campanian–Maastrichtian non-volcanic sediments are locally overlying the Tireo formation, and volcanics from this period are missing (Lewis et al., 1991; García-Casco et al., 2008a). This hiatus was interpreted by García-Casco et al. (2008a) to reflect a period of inactivity of the Great Arc during underthrusting of Caribeana (Samaná terrane), but it may also be the result of younger strike-slip motions, transporting bodies of arc material eastwards, leading to gaps and duplications (K. Burke, 2013, personal communication).

Volcanic arc rocks and associated sediments are regionally angularly unconformably covered by Upper Eocene to Recent sediments that post-date island arc magmatism (Mann et al., 1991; Dolan et al., 1998; Krebs et al., 2008). Collision of the volcanic complex of Hispaniola with the Bahamas borderlands is dated at Middle–Late Eocene (Vila et al., 1987; Cribb et al., 1989; De Zoeten and Mann, 1991) and is documented by this angular unconformity separating folded Upper Paleocene–Lower Eocene sediments from an Upper Eocene basal conglomerate (De Zoeten and Mann, 1991).

The southern one-third of the island, on the other hand, exposes flood basalts, interpreted as CLIP material (Maurrasse et al., 1979; Sen et al., 1988; Mann et al., 1991).

Eocene–Recent North American–Caribbean transpressional plate motion is accommodated by plate boundary-perpendicular motion in the offshore North Hispaniola Deformed Belt, north of the island and strike-slip motion, localized along seismically active faults: the Septentrional fault in the north (i.e. an eastward extension of the Oriente fault), and the Enriquillo–Plantain Garden and Los Pozos faults in the south (Mann et al., 1984, 1991; Pindell and Barrett, 1990; Hayes et al., 2010; Prentice et al., 2010). The western part of Hispaniola is part of the Gonave block, bounded by the Cayman spreading center in the west, the Oriente–Septentrional fault system in the north, the Walton–Enriquillo–Plantain Garden fault system in the south and NS-oriented thrust faults in Hispaniola in the east (Fig. 3). The western part of Hispaniola is moving faster to the east with respect to North America than the eastern part of Hispaniola and is uplifting as a result of this differential motion (Mann et al., 1995; DeMets and Wiggins-Grandison, 2007).

In summary, Hispaniola contains a segment of the Great Arc of the Caribbean, active from before 116 Ma until collision with the North American continental margin in the Middle–Late Eocene (with a hiatus in the upper Campanian–Maastrichtian). The interruption in volcanic activity is attributed to the collision of Hispaniola with the Samaná terrane of Caribeana, simultaneously, or slightly later than the underthrusting of the Escambray terrane in Cuba. Eocene–Recent strike-slip faulting moved Hispaniola eastward with respect to the Cuban segment, whereby left-lateral strike-slip was partitioned over two fault systems: the Oriente–Septentrional fault system in the north and the Walton–Enriquillo–Plantain Garden fault system in the south. The latter fault system juxtaposed oceanic Caribbean crust overlain by CLIP lavas against Great Arc of the Caribbean-equivalent rocks of central Hispaniola.

4.8. Puerto Rico

Puerto Rico lies east of Hispaniola and exposes Cretaceous–Eocene island arc rocks associated with southwest dipping subduction, underlain by a serpentinite-matrix mélange (Bermeja complex), containing blocks of serpentinized peridotite, altered basalt, amphibolite and chert (Larue, 1991; Schellekens, 1991, and references therein; Jolly et al., 2008; Laó-Dávila et al., 2012). K–Ar whole rock ages of the amphibolite blocks range between ~130 and 75 Ma (Bandini et al., 2011; and references therein). Radiolarian chert date the 'Mariquita Chert Formation' of the Bermeja complex at lower Middle Jurassic to lower Upper Cretaceous (upper Bajocian–lower Callovian to upper lower Albian–lower Middle Cenomanian; Bandini et al., 2011), reflecting the minimum age of the oceanic crust that subducted below the arc of Puerto Rico. The serpentinites have been emplaced by thrusting in two events: in Maastrichtian–Paleocene and Late Eocene–Early Oligocene times (Laó-Dávila et al., 2012).

The record of arc magmatism contains an Upper Cretaceous–Danian hiatus (~75–65 Ma) (Jolly et al., 1998; García-Casco et al., 2008a), roughly equivalent to the hiatus on Hispaniola. García-Casco et al. (2008a) suspected that this may also be an expression of the collision of the Caribbean plate with Caribeana, although HP–LT metamorphic rocks that underwent late Cretaceous metamorphism have only been found offshore Puerto Rico (Puerto Rico Trench terrane; García-Casco et al., 2008a), and not on the island itself. Final arrest of arc magmatism was in the Eocene (Jolly et al., 1998). The volcanic arc rocks are overlain by Eocene–Pliocene sediments, including sandstones, shales and limestones, deposited in both shallow and deep water (Larue, 1991). An angular unconformity between strongly folded, deep-marine middle Eocene rocks and gently folded middle–upper Oligocene shallow-marine and nearshore rocks indicate a tectonic event in the Upper Eocene, simultaneously with cessation of the volcanic arc (Dolan et al., 1991; Mann et al., 2005). This unconformity is interpreted to date the collision of Puerto Rico with the Bahamas (Dolan et al., 1991). Paleomagnetic data from this Oligocene and younger series showed that the island underwent an ~25° counterclockwise rotation between upper Miocene (~11 Ma) and Pliocene (~4 Ma) times (Reid et al., 1991), interpreted by Mann et al. (2002) as the result of oblique collision with the Bahamas borderlands, and GPS data confirm the absence of active rotation (Jansma et al., 2000), which we adopt in our reconstruction.

The Puerto Rico trench, north of the island, lies along the eastward extension of the North Hispaniola Deformed Belt and curves southward towards the East, to connect with the Lesser Antilles trench. A south-dipping slab can be traced to a depth of 240 km underneath Puerto Rico (van Benthem et al., 2013). The area is a zone of transition between transform motion along the northern plate boundary and convergent motion along the eastern plate boundary of the Caribbean plate, resulting in a curved slab that can be traced to below eastern Hispaniola (van Benthem et al., 2013). Focal mechanisms record almost pure strike-slip motion, indicating very strong obliquity of subduction (Molnar and Sykes, 1969; McCann and Sykes, 1984; Dillon et al., 1996; Dolan et al., 1998). This highly oblique convergence may explain the absence of active arc magmatism above the Puerto Rico slab; dehydration melting and magmatism already occurred earlier beneath the Lesser Antilles Arc, before the slab arrived underneath Puerto Rico farther to the west (Calais et al., 1992; van Benthem et al., 2013). Relative westward motion of the slab edge below Hispaniola results in surface deformation (van Benthem et al., 2014).

South of the island of Puerto Rico, the Muertos Thrust Belt is located and a north-dipping seismic zone can be traced to 100 km depth (van Benthem et al., 2013), associated with underthrusting of the Caribbean oceanic interior in the Muertos Trough (Byrne et al., 1985; Dillon et al., 1996; Dolan et al., 1998). The Muertos Thrust Belt is interpreted as a back thrust response to the oblique Bahamas collision and subduction in the Puerto Rico trench (Mann et al., 2002). In the footwall of the thrust belt, beneath the basal detachment, Lower Miocene and older sedimentary rocks are found. Younger sediments are incorporated into the thrust belt itself, which is active today (Ten Brink et al., 2009). Thrusting is thought to have started in the Late Miocene (Mann et al., 2002). The relative plate velocity between Puerto Rico and the Caribbean plate is very small (~1 mm/yr, GPS measurements, Jansma et al., 2000). Northwestern Puerto Rico is tectonically separated from the eastern part of Hispaniola by the N–S Mona rift, opening with ~5 mm/yr (Jansma et al., 2000; Hippolyte et al., 2005).

In summary, the island of Puerto Rico exposes rocks ascribed to the Great Arc of the Caribbean, which formed until Late Eocene collision of the Caribbean plate with the Bahamas platform. Interruption of arc volcanism indicates that Puerto Rico may have collided with a part of Caribeana. Since collision with the North American plate, highly oblique convergence between the North American and Caribbean plates has led to underthrusting of North American lithosphere in the Puerto Rico trench and Caribbean lithosphere in the Muertos Trough below Puerto Rico.

4.9. Eastern Caribbean subduction system: Lesser Antilles Arc

The eastern part of the Caribbean plate consist of a series of N–S trending bathymetric zones including, from west to east, the Aves ridge, the Grenada Basin, the Lesser Antilles Arc, the Tobago Basin and the Barbados Accretionary Prism (Figs. 1 and 3). The Lesser Antilles Arc is the active volcanic arc associated with westward subduction of Atlantic oceanic lithosphere below the Caribbean plate. Tomographic images indicate at least 1100 km of subduction (van der Hilst, 1990; van Benthem et al., 2013). Ages of arc volcanism range from 38 Ma to present (K–Ar, Briden et al., 1979). The Barbados prism is an accretionary prism that formed along the Lesser Antilles subduction zone between the Early Eocene and the present (Speed and Larue, 1982). The prism is exposed on the island of Barbados. The southern half of the prism is extraordinarily large due to favorable conditions for accretionary buildup: sediment that constitute the southern half of the prism were supplied by the Orinoco River that runs through Colombia and Venezuela and flows out into the Atlantic Ocean close to Trinidad and Tobago. Since formation of the Orinoco river delta, a large amount of sediment is scraped off the downgoing Atlantic oceanic lithosphere and accreted to the Caribbean plate (Speed and Larue, 1982).

The Aves ridge is a remnant island arc, interpreted to have been part of the Great Arc of the Caribbean (Burke, 1988; Bird et al., 1993). The ridge contains granodiorite, diabase, porphyritic basalt and metabasalt (Fox and Heezen, 1975; Neill et al., 2011) and has a similar seismic velocity structure and crustal thickness as the Lesser Antilles Arc (Christeson et al., 2008). The arc was active in at least the Late Cretaceous and Paleocene (Fox and Heezen, 1975; Pinet et al., 1985; Bouysse, 1988) indicated by ages of dredged and drilled volcanic rocks ranging from 88 to 59 Ma (Neill et al., 2011). No CLIP material has been reported from the Aves ridge.

The Grenada basin is an oceanic basin of which the origin is debated. Different models for basin opening have been proposed, as outlined by Bird et al. (1993), Bird et al. (1999) and Aitken et al. (2011). These models vary in spreading direction (east–west (e.g. Tomblin, 1975; Bird et al., 1999), north–south (e.g. Pindell and Barrett, 1990) or north-east–southwest (e.g. Bouysse, 1988)), but all consider the Grenada Basin to be a backarc basin. In the models of Bouysse (1988) and Pindell and Barrett (1990), the Grenada Basin formed as a result of right-lateral shear between the Great Arc and the South American continent. Aitken et al. (2011) proposed a different model of evolution of the Lesser Antilles subduction system based on seismic, well and onshore geological data. The basement of the Grenada and Tobago basins, on both sides of the active Lesser Antilles Arc, is considered to be the extended oceanic forearc of the Aves Ridge, widened (in E–W direction) in the Paleocene–Middle Eocene. Due to roll-back of the Proto-Caribbean downgoing slab, volcanism ceased in the Aves ridge and migrated to the Lesser Antilles Arc, dividing the former forearc into two basins, the Grenada and Tobago Basins. We adopt this hypothesis in our reconstruction.

4.10. Trinidad, Tobago, Isla Margarita and the Leeward Antilles

The Leeward Antilles (Aruba, Curaçao, Bonaire, Las Aves, Los Roques, La Orchilla, La Blanquilla, Los Hermanos and Los Testigos), Isla Margarita (West Indies) and Trinidad and Tobago are located in the Caribbean Sea off the coast of Venezuela, near the Caribbean–South American plate boundary (Fig. 1).

The island of Trinidad exposes a continental basement, Cretaceous mafic volcanic rocks and Mesozoic and younger sediments and is cut by right-lateral strike-slip fault systems accommodating Caribbean–South American motion. The sediments bear evidence for Jurassic extension — probably related to the opening of the Atlantic ocean — and Miocene compression and dextral wrenching, leading to low-grade metamorphism on the northern part of the island (Saunders, 1974; Frey et al., 1988; Donovan, 1994; Pindell and Kennan, 2009; Neill et al., 2014). The volcanics document the opening of the Proto-Caribbean Ocean at 135.0 ± 7.3 Ma (crystallization age of a mafic volcanic rock, Neill et al., 2014). Trinidad is considered to be part of the South American continent, with Neogene deformation being related to right-lateral dextral wrenching between South America and the Caribbean plate (Pindell and Kennan, 2009).

The island of Tobago, on the other hand, contains three different types of oceanic island arc rocks, exposed in roughly north–south belts; the North Coast Schist, the Tobago Volcanic Group and the Tobago Pluton. The southern part of the island is covered by Pliocene to recent sediments (Snoke et al., 2001a; Neill et al., 2012). The Volcanic Group and Pluton are of Albian age (~110 to 103 Ma based on faunal and $^{40}Ar–^{39}Ar$ hornblende dating; Sharp and Snoke, 1988; Snoke et al., 1990; Snoke and Noble, 2001) and were formed by partial melting of MORB-crust (Neill et al., 2013). The North Coast Schist contains meta-igneous and metasedimentary greenschist and amphibolite facies metamorphic rocks. Greenschist rocks of the Palatuvier Formation have a U–Pb age of 128.7 ± 0.2 Ma (Neill et al., 2012). Greenschist facies metamorphism is interpreted to be regional, and related to an episode between ~129 and 110 Ma of dextral shearing, whereas amphibolite facies metamorphism is interpreted to relate to contact metamorphism associated with intrusion of the Tobago pluton (Snoke et al., 2001b; Neill et al., 2012). Magmatism and metamorphism on Tobago is considered to be part of the Great Arc of the Caribbean (Burke, 1988).

Isla Margarita exposes Paleozoic–Mesozoic metamorphic rocks of both continental (metapelites, marbles, gneisses) and oceanic (metabasalts, carbonaceous schists) character (Maresch, 1975; Maresch et al., 2009). The island contains two eclogite facies metamorphic complexes, structurally overlain by greenschist facies rocks. The age of metamorphism is constrained by the age of the protolith of the youngest HP unit (the Guayacán gneiss, crystallization at ~116–106 Ma, U–Pb age) and the intrusion of the El Salado Metagranite (island arc affinity, ~86 Ma), which intrudes the HP units and has not been subjected to HP metamorphism itself. Peak metamorphism probably occurred between 100 and 90 Ma (Maresch et al., 2009). The HP rocks are interpreted to have formed in a subduction zone at a depth of 50 km. Juxtaposition of the greenschist units happened upon exhumation, after 80 Ma (Maresch et al., 2009). Maresch et al. (2009) interpreted the Isla Margarita HP rocks to be derived from the NW South American continental margin upon Cretaceous collision with the Caribbean plate, after which they were translated to their current position by right-lateral wrenching between the Caribbean and South American plates.

Aruba, Curaçao, La Blanquilla and Gran Roque of the Los Roques contain magmatic rocks that are geochemically very similar to the island arc rocks from the southern Aves Ridge (Neill et al., 2011). Granodiorites and tonalites of La Blanquilla are dated at 75.5 ± 0.9 Ma and 58.7 ± 0.5 Ma respectively, the Aruba batholith at 88.6 ± 0.5 Ma (U–Pb ages of van der Lelij et al., 2010; Wright and Wyld, 2011) and dikes of diorite on Curaçao at 86.2 ± 1.1 Ma (Wright and Wyld, 2011). Bonaire contains island arc rocks with ages of 94.6 ± 1.4 Ma, 98.2 ± 0.6 Ma and ~112 Ma (Thompson et al., 2004; Wright and Wyld, 2011). These rocks are interpreted to have formed above a depleted mantle (Wright and Wyld, 2011). No CLIP material has been found on La Blanquilla and Bonaire (Neill et al., 2011). On Aruba, Curaçao and Gran Roque however, gabbros and dolerites of CLIP affinity are present (Giunta et al., 2002b). CLIP material of the Curaçao Lava Formation has been dated by Sinton et al. (1998) at 88.6 ± 0.5 Ma ($^{40}Ar/^{39}Ar$), CLIP material of the mafic Aruba Lava Formation at ~90 Ma (White et al., 1999). The Leeward Antilles area has been subjected to shortening by inversion of normal faults and right-lateral strike-slip faulting related to collision with the South American continent and subsequent segmentation in extensional basins by a phase of oblique normal faulting (Escalona and Mann, 2011).

Paleomagnetic data from arc volcanics suggest a clockwise rotation of Tobago, Aruba and Bonaire of 90° since the Late Cretaceous with respect to a stable South America (Stearns et al., 1982; Burmester et al., 1996).

In summary, the Leeward Antilles and Tobago are interpreted as fragments of the Great Arc of the Caribbean on the former leading edge of the Caribbean plate, which has been in oblique collision with the northern margin of South America for most of the Cenozoic (Snoke et al., 2001b; Escalona and Mann, 2011; Neill et al., 2012). The present day position of the Leeward Antilles (500–600 km west of the southern tip of the Aves Ridge) may be the result of Cenozoic dextral shearing of the Caribbean plate along the South American margin, slowing down the motion of the Leeward Antilles. Tobago, on the other hand, is currently located ~350 km east of the southern tip of the Aves Ridge and became further separated upon opening of the Grenada and Tobago basins in the former forearc of the Aves ridge. Isla Margarita is interpreted as a fragment of the northwestern South American margin that was metamorphosed when the Great Arc of the Caribbean collided and the South American continental margin underthrusted ~100–90 Ma ago. This fragment was accreted to the Caribbean plate, intruded by arc magmas and displaced eastwards along the South American margin towards its present position (Maresch et al., 2009).

4.11. Northern Andes

The Northern Andes (northwesternmost Peru, western Ecuador, western Colombia and the northwestern corner of Venezuela) is a complex tectonic region of intense deformation. It is bounded by the Colombian–Ecuador trench in the west, the Panama–Chocó block in the northwest, the South Caribbean Deformed Belt in the north and a major fault system in the east, including the East Andean fault zone and the Boconó fault (Pennington, 1981; Kellogg and Vega, 1995; Egbue and Kellogg, 2010; Fig. 3). The Northern Andes and the Panama–Chocó block together accommodate the strain associated with the triple junction between the Nazca, South American and Caribbean plates (Cortés and Angelier, 2005).

The Northern Andes can be divided into two basement provinces: a continental eastern province and a western province composed of thrusted fragments of continental and oceanic crust, including island arc volcanics and associated sedimentary rocks and slivers of oceanic plateau basalts (Kerr et al., 2003; Cortés and Angelier, 2005; Kennan and Pindell, 2009; Fig. 7). The eastern province contains Neoproterozoic gneisses and schists, and unmetamorphosed to low-grade metamorphic Paleozoic sediments (Restrepo-Pace, 1992; Restrepo-Pace et al., 1997) intruded by plutons ranging in age from ~241 to 80 Ma and overlain by a thin Cretaceous sedimentary cover (Litherland et al., 1994; Noble et al., 1997; González, 2001; Villagómez et al., 2008, 2011; Montes et al., 2010). The continental basement is interpreted as the former southern passive margin of the Proto-Caribbean Ocean, conjugate to the southeastern Chortis margin. The plutons are interpreted as part of a broad volcanic arc, related to eastward dipping subduction below the South American continent.

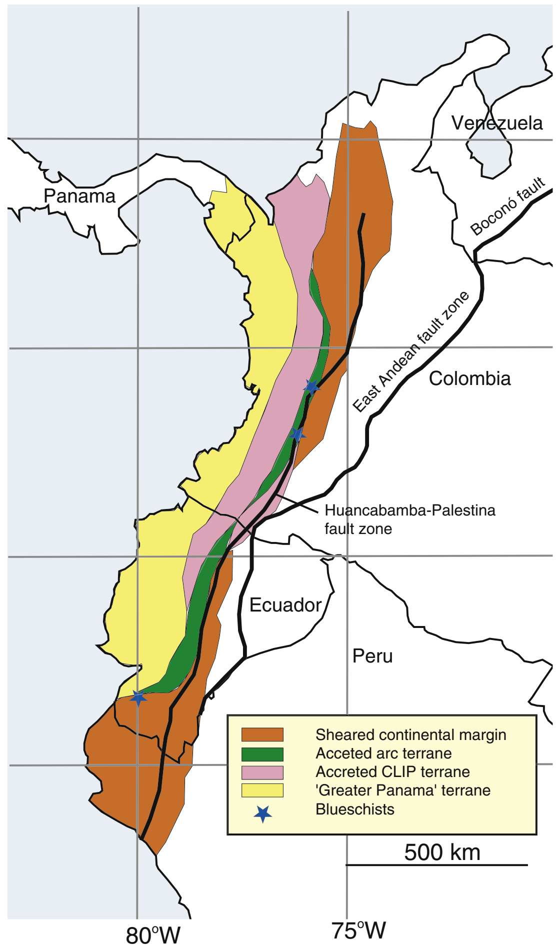

Fig. 7. Tectonic terrane map of the Northern Andes. Based on Kennan and Pindell (2009).

The western province can be subdivided into three belts. The easternmost belt of the western province exposes high to intermediate pressure metamorphic rocks (blueschists, amphibolites and eclogites), Paleozoic and Mesozoic schists and amphibolites and Lower Cretaceous metatuffs, pillow basalts and volcanoclastic sediments (Kennan and Pindell, 2009). Geochemical data from the volcanics suggest a theoleiitic to andesitic island-arc or backarc origin with some influence of underlying continental crust, and magmatism took place between ~180 and 145 Ma (Nivia et al., 2006; Villagómez et al., 2011). All volcanic rocks of this belt are classified by Kennan and Pindell (2009) as sheared and accreted fragments on the Trans-American Arc and its successor, the Great Arc of the Caribbean. K–Ar, Lu–Hf and $^{40}Ar/^{39}Ar$ ages of the metamorphic rocks of 133–126, 120–110 and 68–61 Ma and whole rock K–Ar ages of mafic HP/LT rocks of 125 $\pm$ 15, 110 $\pm$ 10 and 120 $\pm$ 5 reflect peak metamorphism and onset of cooling, and suggest an exhumation event starting in the Early Cretaceous (McCourt et al., 1984; Maya-Sánchez, 2001; Maya-Sánchez and Vásquez-Arroyave, 2001; John et al., 2010; Bustamante et al., 2011). The trace element signatures of the 133–126 Ma rocks provide evidence for subducted seamounts and MORB-crust (John et al., 2010). The 120–110 Ma rocks have a continental origin and are interpreted to result from subduction of the continental margin of South America during collision with, and underthrusting below the Great Arc (Kennan and Pindell, 2009), following subduction of oceanic South American plate. The 67–61 Ma rocks have been interpreted by Bustamante et al. (2011) as part of a tectonic mélange, related to a younger accretion event.

The central belt comprises slivers of oceanic plateau basalts and associated ultramafic rocks and sediments. The plateau basalts are dated between ~100 and 87 Ma and geochemistry indicates an intra-oceanic plateau origin interpreted as related to the CLIP (Vallejo et al., 2006; Villagómez et al., 2011). The basalts are thrusted over, as well as imbricated against the western margin of the South American continent (Kerr et al., 1997). This suggests that the South American continent was originally in a downgoing plate position relative to the Caribbean plate after the formation of the CLIP, followed by a subduction polarity reversal and underthrusting of the Caribbean plate below South America, as seen today. In the northwestern part of the northern Andes, intrusive rocks of 54–59 Ma (U–Pb zircon ages, Bayona et al., 2012) are found. This period of magmatism is thought to be related to oblique and shallow subduction of the edge of the CLIP, following arc–continent collision (Cardona et al., 2011; Bayona et al., 2012). The westernmost belt of the Northern Andes contains Santonian–Campanian boninites, tholeiites and calc-alkaline basalts of an intra-oceanic island arc origin, underlain by oceanic plateau basalts, and covered by Paleogene and younger forearc basin sediments (Kennan and Pindell, 2009; Borrero et al., 2012). Kennan and Pindell (2009) propose that these rocks were formed above a northeast dipping subduction zone at the trailing edge of the Caribbean plate, as a result of subduction of the Farallon plate. The Santonian–Campanian age of the boninites indicate an age shortly postdating the onset of subduction (e.g. Stern et al., 2012). This westernmost terrane is interpreted as an extension of the Panama–Chocó block (or 'Greater Panama') that subducted beneath the South American continent until the collision of the present-day Panama–Chocó arc with Colombia in the Late Miocene–Pliocene (Duque-Caro, 1990; Coates et al., 2004; Kennan and Pindell, 2009). Kennan and Pindell (2009) suggested that the buoyant upper crustal parts of the lithosphere were scraped off the downgoing plate and accreted to the South American overriding plate, while their original lower crustal and mantle underpinnings subducted. The present-day Panama–Chocó block behaves as a rigid indenter and is thought to be the main reason for the Andean compressional phase in Colombia (Taboada et al., 2000; Trenkamp et al., 2002; Vargas and Mann, 2013). Ongoing Panama–Chocó–South America convergence is estimated by GPS measurements at 25 mm/yr (Trenkamp et al., 2002). Cenozoic subduction of the Caribbean plate below the South American continent and the Panama–Chocó block led to the formation of the Panama Deformed Belt and the South Caribbean Deformed Belt (Kennan and Pindell, 2009; van Benthem et al., 2013). Subduction in the South Caribbean Deformed Belt initiated diachronously from west to east in Middle-Eocene to Late Neogene times. Before subduction initiation, in Late Cretaceous–Miocene times, the belt acted as a back thrust associated with subduction of the South American plate below the Great Arc (Kroehler et al., 2011).

Within the eastern province, the Huancabamba–Palestina fault zone is exposed, which can be traced more or less continuously from northern Peru to northern Colombia. This anastomosing fault zone records brittle, dextral strike-slip motion. The southern part of the fault zone has been active throughout the Late Cretaceous and Paleogene, the northern part is still active today (Kennan and Pindell, 2009). The amount of Cretaceous to Eocene offset of the fault zone has been estimated at ~300 km (Kennan and Pindell, 2009). The boundary between the eastern and central belt in the western province is the Romeral suture, which recorded significant dextral strike-slip fault movement since the Cretaceous (Kerr et al., 1998; Trenkamp et al., 2002; Kennan and Pindell, 2009). There is no cumulative estimate of dextral strike-slip displacement in the Northern Andes from the Cretaceous to present day. GPS data shows that the Maracaibo block of northern Colombia and northwestern Venezuela (the fault bounded crustal feature defined by the Maracaibo, Oca and Santa Marta fault zones, Mann and Burke, 1984) is currently moving to the northeast with respect to a stable South America with 6 mm/yr. This movement is interpreted as northeastward tectonic escape following the collision between the South American continent and the Panama–Chocó block (Mann and Burke, 1984; Freymueller et al., 1993; Kellogg and Vega, 1995;

Trenkamp et al., 2002). Note that the “Greater Maracaibo” block defined in Fig. 3 also includes areas west of the Santa Marta fault and north of the Oca fault.

In summary, the geology and geometry of the Northern Andes is the result of a series of events starting with eastward subduction of the Farallon plate below the west coast of South America in pre-late Jurassic time, when South America was still part of Pangea. During the break-up of Pangea, the Northern Andes became a passive margin bordering the South American continent from the Proto-Caribbean Ocean. Since the Early Cretaceous onset of subduction below the Caribbean plate and the associated volcanism in the Great Arc of the Caribbean, the Caribbean plate moved with an easterly direction relative to South America, consuming Proto-Caribbean ocean floor. Ages of continental metamorphic HP/LT rocks date the onset of collision of the Caribbean plate and Great Arc with the passive margin of South America in the Early Cretaceous. After collision, westward subduction ceased and the leading edge of the Caribbean plate moved dextrally alongside the South American continent towards the northeast, accommodated along major strike-slip faults. Ongoing convergence between the Caribbean plate and the South American margin with its accreted terranes led to the formation of an eastward dipping subduction zone in the Maastrichtian, accretion of plateau basalts and resulting progressive westward back-stepping of the subduction zone (Millward et al., 1984; Kerr et al., 1997; Vallejo et al., 2006). Since the Eocene, Panama arc crust has been indenting Colombia, resulting in the accretion of the upper crustal parts of the extension of the Panama–Chocó block, collision of the present day Panama–Chocó block in the Late Miocene–Pliocene (Kennan and Pindell, 2009) and escape tectonics in Colombia and Venezuela.

5. Reconstruction

We now integrate the constraints described above and summarized in Fig. 5 and Table 2, and the first-order kinematic interpretations per sub-region into a kinematic restoration of the Caribbean plate relative to the Americas, starting in the present and presented backward in time. Per time interval, the motions of plates or plate fragments are described forwards, as this is easier to comprehend. All figures are shown in a North America-fixed frame just as all motions (except for the NW South American terranes) are described relative to a fixed North America. The hierarchy of interpretations indicated in Table 1 is retained here in the order in which the constraints for the reconstruction are used. *GPlates* rotation and shape files are given in the online appendix.

5.1. 50 Ma–present

5.1.1. Caribbean–North American plate boundary evolution: the Motagua–Cayman–Oriente fault system and the Cuban segment

The youngest part of the reconstruction, between 50 Ma and the present, is primarily based on oceanic extension recorded in the Cayman Trough that currently forms the Caribbean–North American plate boundary (Leroy et al., 2000). The Cayman Trough recorded ~900 km of sinistral motion between the Caribbean plate and the Cuban segment since the estimated 49.4 Ma (Leroy et al., 2000). As a consequence, reversing the oceanic spreading in the Cayman Trough displaces the Caribbean plate ~900 km to the west with respect to the Cuban segment, which since the early Late Eocene was firmly welded to the North American Plate.

To the west, the Cayman Trough connects to the Motagua fault zone, which since the incorporation of the Chortis Block to the Caribbean plate must have accommodated Caribbean–North American (Yucatan) plate motion (Burkart, 1994; DeMets et al., 2007). Following the interpretation of Rogers et al. (2007), we reconstructed the Chortis Block to move along the southern margin of Mexico in the Early Eocene. Due to the difference in orientation between the Motagua fault zone (~E–W trending) and the margin of southern Mexico (NW–SE trending), this restoration requires a counterclockwise rotation of Chortis relative to North America. We reconstructed a 32° counterclockwise rotation between ~50 Ma and the present. The bulk of this rotation is reconstructed between ~38 and 33 Ma, when Chortis passed the bend in the margin of southwestern Mexico, corresponding with a phase of the largest rotation (~1.7°/Myr). After 17 Ma, the rotation is negligible, consistent with paleomagnetic constraints from Molina Garza et al. (2012) that could not demonstrate significant rotation of volcanics of the Chortis Block in western Honduras since this time.

The island of Hispaniola is dissected by multiple left-lateral strike-slip faults, accommodating the Caribbean–North American plate motion. The southern part of Hispaniola contains CLIP material and no faults are documented separating it from the Caribbean plate interior. Therefore, we kept the southern ‘CLIP’ part of Hispaniola fixed to the Caribbean plate interior. The cumulative displacement of the strike-slip faults and the North Hispaniola Deformed Belt must be equal to the amount of spreading in the Cayman trench, but how this displacement is partitioned is not directly geologically constrained. We therefore align the tectonostratigraphy of Hispaniola, with Caribeana-related terranes and Lower Cretaceous to Middle Eocene arc material, with their equivalents of southeastern Cuba (following suggestions in Rojas-Agramonte et al., 2005, 2008). As a result, we partition the Caribbean–North American motion with ~550 km displacement on the North Hispaniole Deformed Belt, accommodating relative motion between southern Cuba and northern Hispaniola and aligning the tectonostratigraphies of Hispaniola and southeastern Cuba at ~50 Ma, and the remaining ~350 km on the Enriquillo–Plantain Garden/Los Pozos fault, separating northern (Great Arc material) and southern (CLIP material) Hispaniola, similar to tectonic reconstructions of the area of Calais and Mercier de Lépinay (1995) and Rojas-Agramonte et al. (2005, 2008).

The island of Puerto Rico contains Cretaceous–Eocene volcanics, interpreted as Great Arc rocks (Jolly et al., 1998). We restored ~300 km displacement between Puerto Rico and Hispaniola, accommodated by strike-slip motion in the Puerto Rico Trench (Molnar and Sykes, 1969; McCann and Sykes, 1984), to locate Puerto Rico adjacent to eastern Hispaniola at 50 Ma, following Ross and Scotese (1988). We reconstructed ~30 km Neogene convergence between the Caribbean plate and Puerto Rico, representing underthrusting of the Caribbean plate in the Muertos Trough (Byrne et al., 1985; Dillon et al., 1996; Dolan et al., 1998) and restore a 25° counterclockwise rotation between 11 and 4 Ma following Reid et al. (1991).

The island of Jamaica and the North Nicaraguan Rise, containing Great Arc rocks (Abbott et al., 1999; Lewis et al., 2011), are currently not aligned with the arc rocks of the other Great Antilles. Because Jamaica and the North Nicaraguan Rise are located south of the main strike-slip boundary between the Caribbean plate and the Cuban segment, we restored them as part of the Caribbean plate since 50 Ma, which results in restoring Jamaica to southern Yucatan at 50 Ma. Cooling ages of the Chuacús complex (~70 Ma) coincide with the age of cessation of arc magmatism on Jamaica. This is consistent with collision of the Great Arc with southern Yucatan resulting in the end of subduction and exhumation of HP rocks, currently exposed in Guatemala.

Collision of the Caribbean plate and leading Cuban segment with the Bahamas platform started in the latest Paleocene to ~45 Ma and was finalized by early Late Eocene (Bralower and Iturralde-Vinent, 1997; Iturralde-Vinent et al., 2008). The collision was associated with a gradual change in Caribbean plate motion from NNE-wards along the Belize margin transform to eastwards along the Cayman Trough, accommodated along progressively more easterly oriented left-lateral strike-slip faults (e.g. Mann et al., 1995). We reconstructed the final collision of western Cuba at 45 Ma and reconstructed an arbitrary but small 20 km of displacement along the La Trocha fault between 44 and 40 Ma and 15 km of displacement along the Cauto fault between 40 and 38 Ma, accommodating relative motion between western and central Cuba and central and eastern Cuba, respectively (based on

Leroy et al., 2000; Cruz-Orosa et al., 2012a). This way, the ages of collision are younging to the southeast and the relative motions between western, central and eastern Cuba lead to a gradual but quite rapid change from NNE to E-ward relative Caribbean–North American motion.

Finally, in latest Cretaceous–Eocene time, the Cuban segment moved relative to North America. As a result, translating the Cayman Trough kinematics in terms of Caribbean–North America plate motion before final welding of the Cuban segment with North America in the early Late Eocene, requires restoration of the Cuba–North America motion. This motion is reconstructed by aligning the Caribeana rocks of Cuba with those from Guatemala around 70 Ma, i.e. 900 km SSW-wards relative to today (see Section 5.2). The direction of the motion of the Cuban segment is parallel to the Belize margin and opened the Yucatan pull-apart basin. The 50 Ma position of the Caribbean plate shown in Fig. 8a is an interpolated position between these reconstruction steps, assuming a continuous motion rate. The Cuban segment (and as a result also the Motagua–Cayman–Oriente fault system and the Caribbean plate interior) is reconstructed ~200 km SSW relative to North America at 50 Ma (Fig. 8a).

5.1.2. Caribbean–Farallon/Cocos–Nazca plate boundary evolution: the Central America Trench

The western Caribbean plate boundary is characterized by subduction of the Farallon, and subsequently, Cocos and Nazca plates in the Central America Trench. The continuous Upper Cretaceous–Present record of arc volcanism in the Panama–Chocó block indicates ongoing active subduction throughout the 50–0 Ma period. The eastward motion of the Chortis Block along the margin of southwestern Mexico results in a migrating trench–trench–transform triple junction at the northwestern tip of Chortis. After passing of the Chortis Block and the consequent migration of the triple junction, the former transform boundary along the southern margin of Mexico evolved into a subduction zone, subducting Farallon/Cocos plate below Mexico. Following Phipps Morgan et al. (2008), the triple junction has migrated since the Middle Miocene even though the nature of the plate boundaries did not change. This implies that the eastward motion of the Chortis Block must have been accommodated by intra-plate deformation. We reconstructed the post-Middle Miocene history of the Chortis Block by restoring 100 km E–W extension in the Nicaraguan Basin since 15 Ma (Phipps Morgan et al., 2008), 4° counterclockwise rotation of the forearc with respect to Chortis and 150 km right-lateral strike-slip motion between the Chortis Block and its western forearc (DeMets, 2001), following the scenario proposed by Phipps Morgan et al. (2008).

5.1.3. Eastern Caribbean plate boundary evolution: the Lesser Antilles subduction system

The eastern plate boundary of the Caribbean plate is characterized by subduction of the Atlantic North and South American plates below the Caribbean in the Lesser Antilles subduction system. Accumulation of the Barbados Accretionary Prism since the Early Eocene indicates ongoing convergence during the 50–0 Ma period. Arc volcanics in the Lesser Antilles Arc and Aves Ridge are dated ranging from 38 to 0 Ma (Briden et al., 1979) and from 88 to 59 Ma (Neill et al., 2011) respectively, suggesting that the transition of the active arc from the Aves Ridge to the Lesser Antilles Arc took place between 59 and 38 Ma. We followed the model of subduction system evolution of Aitken et al. (2011), considering the Grenada Basin to be the forearc of the Aves Ridge that widened by E–W extension in the Paleocene–Middle Eocene, after which active volcanism ceased in the Aves Ridge and migrated to the Lesser Antilles Arc, separating the former forearc into the Grenada and Tobago Basins. We kept the island of Tobago fixed to the eastern edge of the Tobago Basin; restoring E–W Paleogene extension in the eastern Caribbean forearc brings Tobago closer to the southern tip of the Aves Ridge (see also Section 5.2.5).

5.1.4. Caribbean–South American plate boundary evolution

In the 50–0 Ma period, South America is slightly moving to the northwest relative to North America (according to the North America–Africa–South America plate circuit of Torsvik et al. (2012), see also Pindell and Kennan (2009) and references therein). Combining this with the reconstructed motion of the Caribbean plate relative to North America (northeastward between 50 and 45 Ma and changing to eastward after 45 Ma) results in mainly right-lateral strike-slip motion between the Caribbean plate and South America for the period 50–45 Ma. Since between 45 and 40 Ma, accommodation of N–S convergence along the northern Caribbean plate boundary ceased, ongoing N–S convergence between North and South America became accommodated along the Caribbean–South America plate boundary. The South American–Caribbean plate boundary was thus subjected to highly oblique convergence in this period, accommodated by oblique subduction of the Caribbean plate below South America forming the South Caribbean Deformed Belt and the Maracaibo subduction zone, accompanied by right-lateral strike-slip faulting. The west to east younging transition in the South Caribbean Deformed Belt from back thrusting to subduction (Kroehler et al., 2011) follows the position of the Great Arc and marks the relative eastward Caribbean–South America motion. We reconstructed a total of ~250 km (below eastern Venezuela) to 850 km (below Colombia) of Caribbean plate subduction alongside ~1000 km strike-slip motion between 50 and 0 Ma.

Together with the Caribbean plate interior, an extension of the Panama–Chocó block subducted below South America, accreting arc material to the continent, until collision of the modern Panama–Chocó block with the Northern Andes in the Late Miocene–Pliocene (Duque-Caro, 1990; Coates et al., 2004; Kennan and Pindell, 2009). As a result of tectonic escape, following the collision, the Maracaibo block of northern Colombia and northwestern Venezuela has been moving to the northeast (Freymueller et al., 1993; Kellogg and Vega, 1995; Trenkamp et al., 2002). Because kinematic estimates of Ross and Scotese (1988) remain pertinent for the NW South American region in the light of data published since then, we followed their reconstruction and restored ~250 km of northeastward Maracaibo block movement relative to South America between 11 and 0 Ma. After collision, and due to ongoing convergence, the Panama Deformed Belt formed, north of the Panama–Chocó block. We reconstructed ~1000 km of NW–SE convergence between the western Panama block and the South American continent and ~250 km of NE–SW shortening across the Panama Deformed Belt, following reconstruction of the western, central and eastern Panama blocks by Montes et al. (2012).

The Late Cretaceous–Paleocene Great Arc rocks of the Leeward Antilles and the 100–90 Ma HP rocks of Isla Margarita are moving with a motion direction similar to the Caribbean plate but with a lower velocity. This results in a 50 Ma position of the islands near the southern edge of the Aves Ridge, north of central Colombia. This is an intermediate position on their journey along the northern margin of South America between the location of formation and the modern location. Dextral shearing with the South American margin results in extension between the Leeward Antilles and Isla Margarita between 50 and 0 Ma and a more westward final position of the Leeward Antilles.

5.1.5. 50 Ma reconstruction

The 50 Ma reconstruction, following from the geological and kinematic constraints listed in Section 4, and the reconstruction choices outlined above, contains some kinematic inconsistencies, which we propose to solve as follows. Firstly, rotating the Chortis Block during its westward restoration along the Mexican margin, and reconstructing the Caribbean plate southward prior to 38 Ma following the constraints on the Cuba–North America collision creates a gap between the Chortis Block and the interior of the Caribbean plate at 50 Ma, corresponding to a location around the Nicaraguan Rise. This reconstructed gap between polygons would require post-50 Ma shortening of 150–200 km between the Caribbean plate interior and the Chortis Block, which is inconsistent with seismic tomographic constraints (van Benthem et al., 2013) and for which there is no geological evidence. An alternative restoration that remains consistent with the available facts requires an ~7° counterclockwise rotation of the Caribbean plate (including the Cuban segment) at 50 Ma, around a Caribbean–North American Euler pole that is chosen such that 70–45 Ma pure strike-slip along the Belize margin transform is preserved (inset Fig. 8a). This rotation results in a slight oblique convergence between Cuba and the Bahamas platform and takes away the necessity to infer major post-50 Ma intra-Caribbean shortening. It still requires decoupling of the Caribbean plate interior from the Chortis Block, however, along a dextral strike-slip fault. We consider the Hess escarpment the most suitable candidate for such a strike-slip fault and predict ~150 km of dextral strike-slip motion on the Hess escarpment, or a structure parallel to that within ~100–200 km, between 50 and 30 Ma, during the main phase of rotation of the Chortis Block (inset Fig. 8a). However, more strike-slip motion is required between the Chortis Block and the Caribbean plate interior, and to avoid gaps and overlaps as effectively as possible, we predict

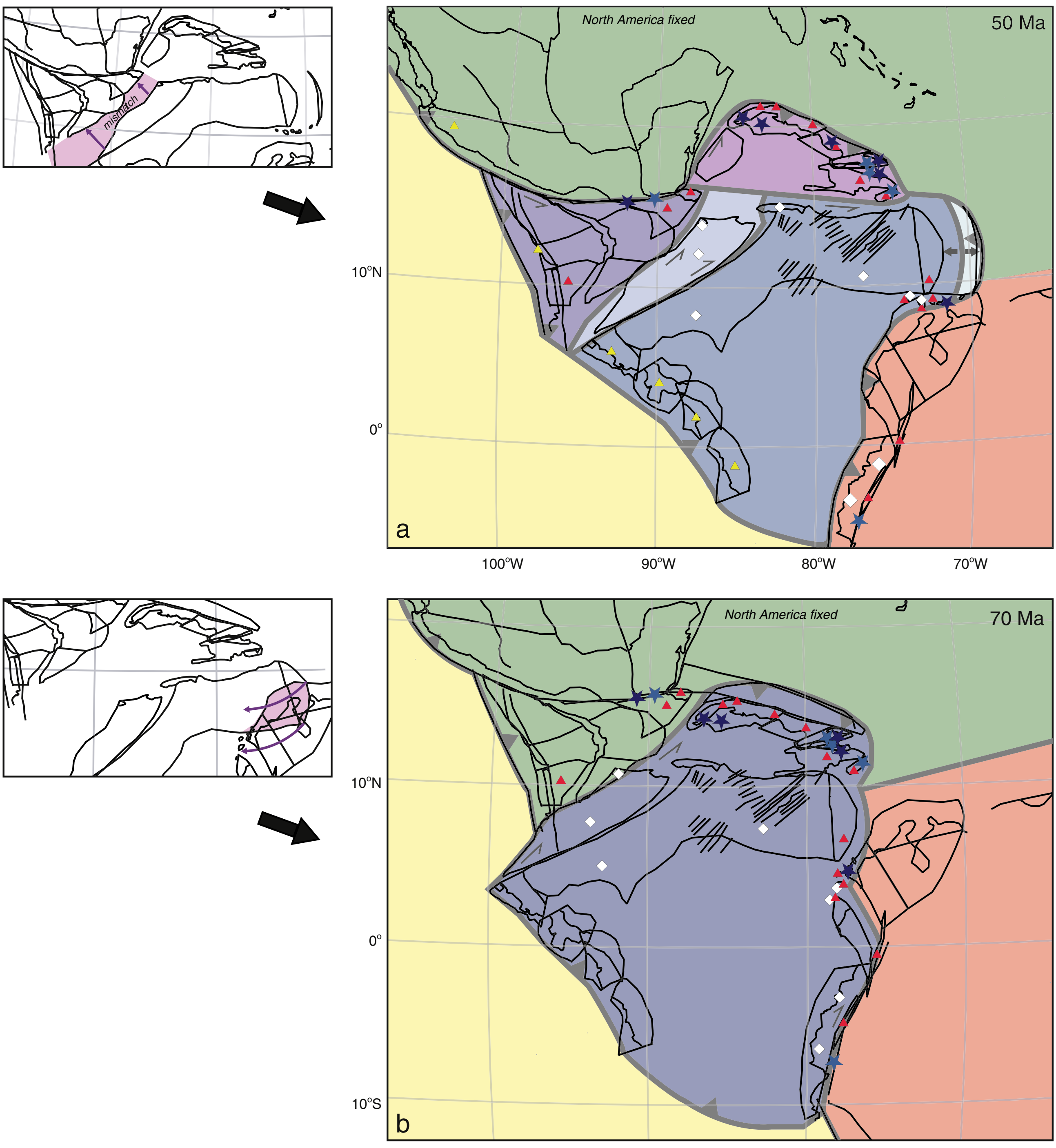

Fig. 8. Snapshots of the reconstruction at selected time slices. Tectonic blocks in the same color-area move, at the time of the reconstruction snapshot, with the same velocity and direction. There is no internal deformation within a color-area. The line fragments in the Caribbean plate interior represent the magnetic anomalies identified by Ghosh et al. (1984). Insets in panels a and b indicate overlaps that were avoided by strike-slip motions in the Caribbean plate interior, and by invoking rotations of the Caribbean plate as a whole (see text for further explanation). (a) 50 Ma; (b) 70 Ma; (c) 100 Ma; (d) 135 Ma; (e) 170 Ma; (f) 200 Ma.

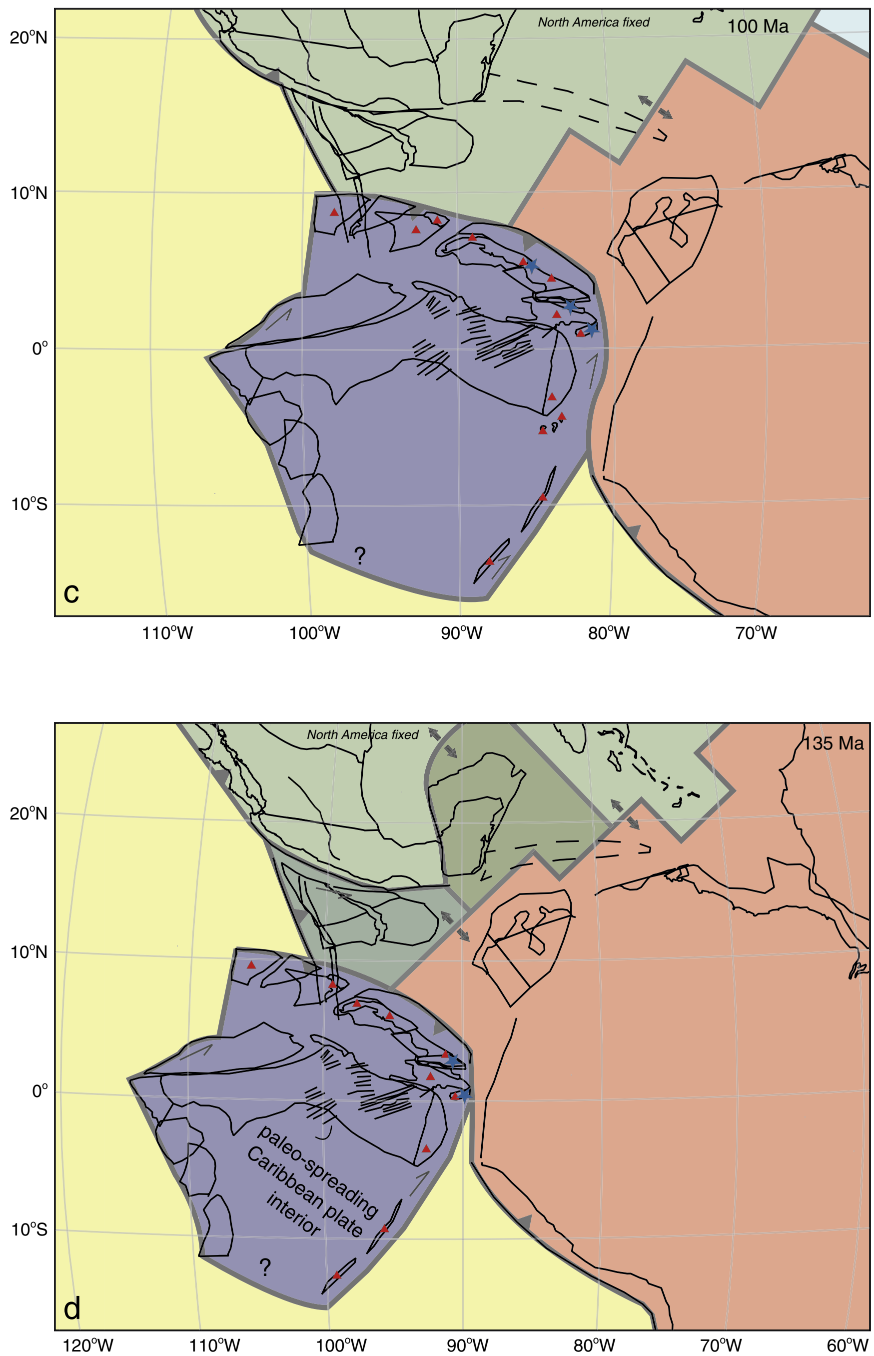

Fig. 8 (*continued*).

the remaining strike-slip motion to be accommodated between the Siuna–Chortis–North Nicaraguan Rise and the South Nicaraguan Rise (300 km of left-lateral motion between 50 and 38 Ma and 100 km of right-lateral motion between 38 and 32 Ma). Present-day internal deformation of the Caribbean plate is supported by GPS data, suggesting separated western and eastern Caribbean plates (Mattioli et al., 2014). The resulting 50 Ma reconstruction is shown in Fig. 8a.

5.2. 70–50 Ma

5.2.1. Caribbean–North American plate boundary evolution: transform motion and subduction in the Cuban subduction zone

The main constraint for the 70–50 Ma period is the Paleocene–Middle Eocene opening of the western Yucatan pull-apart basin (Rosencrantz, 1990), indicating transform motion parallel to the margin of Belize and convergence between the Great Arc at the leading edge of the Caribbean plate and the Bahamas platform as recorded in the geology of the Greater Antilles. The convergence of the Great Arc of Cuba, Hispaniola, and Puerto Rico with the Bahamas platform was accommodated by subduction of the Proto-Caribbean Basin and its northern North American passive margin underneath the Caribbean plate. Evidence for subduction is provided by Cretaceous–Eocene arc volcanism on e.g. Hispaniola and in the Yucatan basin (Larue, 1991; Schellekens, 1991; Abbott et al., 1999; García-Casco et al., 2001, 2008a; Escuder-Viruete et al., 2006, 2013; Mitchell, 2006; Stanek et al., 2009). HP rocks in these serpentinite-matrix mélanges range in age from 130 to 60 Ma in the Greater Antilles (Somin and Millán, 1981; Somin et al., 1992; Iturralde-Vinent et al., 1996; García-Casco et al., 2006; Krebs et al., 2008; Bandini et al., 2011); rocks

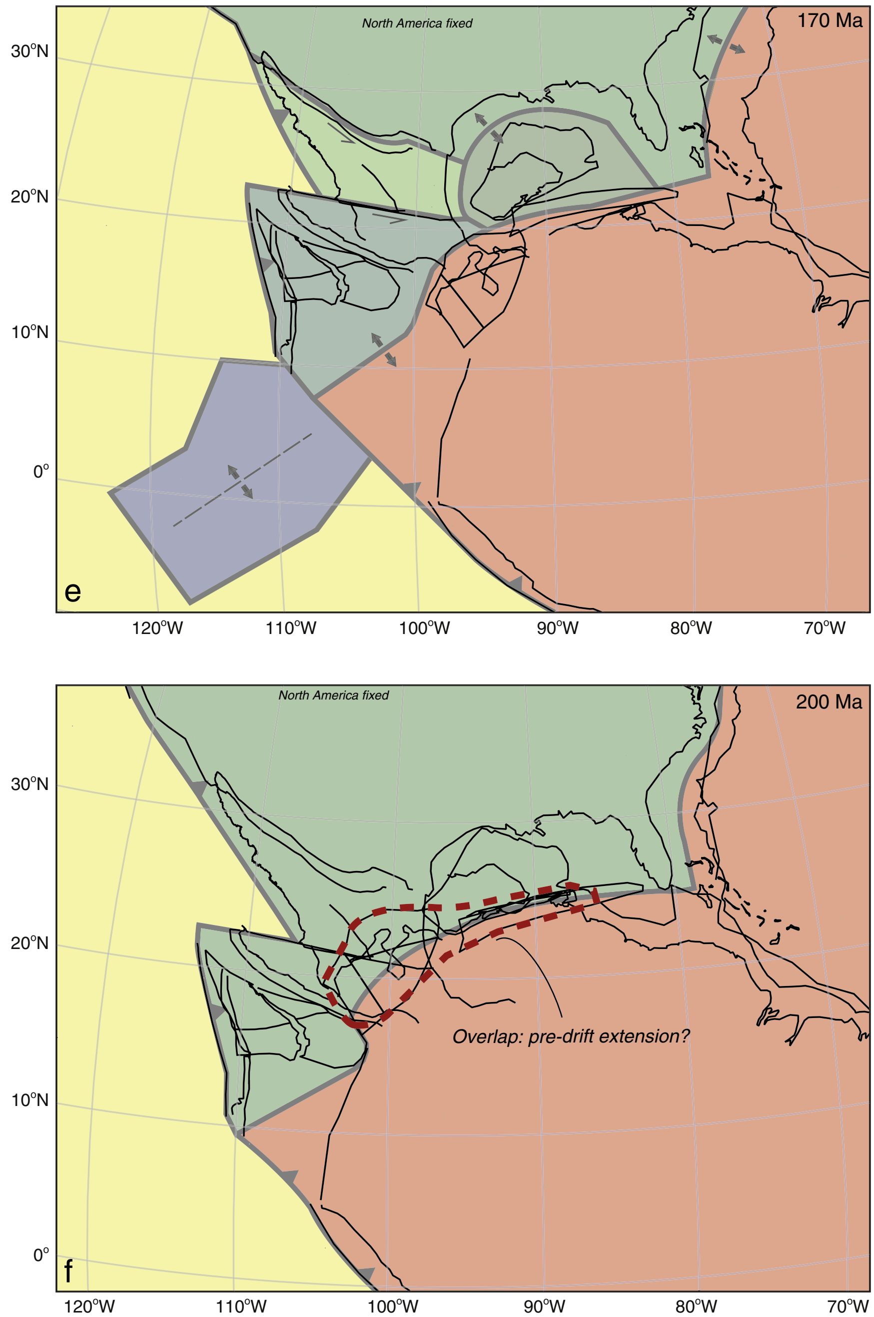

Fig. 8 (*continued*).

that were incorporated in the serpentinite mélanges after the Cretaceous are mainly non-metamorphic thin-skinned accreted sedimentary units with Paleocene and early to middle Eocene foreland basin units offscraped from the downgoing slab and incorporated as deformed slivers in the serpentinite mélange (Iturralde-Vinent et al., 2008; van Hinsbergen et al., 2009). The NNE–SSW trending, 450 km long western Yucatan pull-apart basin (Rosencrantz, 1990) is reconstructed as a transform plate boundary, most likely developing above a slab edge (STEP-fault sensu Govers and Wortel, 2005, see van Benthem et al., 2013); this provides a minimum estimate for the amount of Greater Caribbean SSW-ward subduction since 70 Ma.

As described in Section 5.1.1, we restored the island of Jamaica and the North Nicaraguan Rise as a part of the Caribbean plate interior after 50 Ma, and restored these units adjacent to southern Yucatan and its high-pressure metamorphic southern margin, the Chuacús complex. Cooling ages of the Chuacús complex (~76–62 Ma; Ratschbacher et al., 2009 and references therein; Martens et al., 2012) and latest Campanian–Paleocene ages arc volcanism in Jamaica (McFarlane, 1974; Robinson, 1994; Mitchell, 2006) indicate the timing of collision and overthrusting of the Great Arc over the southern margin of Yucatan. Before the inception of opening of the Cayman trench we have therefore kept the island of Jamaica and the North Nicaraguan Rise fixed relative to Yucatan between 70and 50 Ma, as a part of the North American plate, similar to the reconstructions of Ross and Scotese (1988) and Pindell and Kennan (2009). Restoring the Yucatan basin then aligns the Cuban and Hispaniola parts of the Great Arc with those of Jamaica and the North Nicaraguan Rise at 70 Ma. In addition, the Caribeana complexes of Cuba and Hispaniola align with their Chuacús counterpart in Guatemala. The Caribeana metamorphic terranes recorded Campanian peak metamorphism and post-70 Ma exhumation, indicating that underthrusting and accretion of Caribeana units below the leading edge of the Caribbean plate was finalized by the latest Cretaceous (Kantchev, 1978; García-Casco et al., 2006, 2008a; Stanek et al., 2006). This is supported by the late Campanian–Danian interruption of arc volcanism in Cuba, Hispaniola and Puerto Rico (Lewis et al., 1991; Jolly et al., 1998; García-Casco et al., 2001, 2008a). As a result, the leading Great Arc (represented by Jamaica, the North Nicaraguan Rise, Cuba, Hispaniola and Puerto Rico) is reconstructed ~900 km SSW-ward relative to its modern position versus North America at 70 Ma. Latest Cretaceous–Paleocene subduction erosion of the Caribbean plate resulted in the almost complete disappearance of the forearc of the Great Arc (van Hinsbergen et al., 2009); we have restored a conceptual 160 km wide forearc at 70 Ma, i.e., the average trench–arc distance in modern subduction zones (Gill, 1981).

After restoring the Chortis Block to the west and rotating it between 50 and 0 Ma, the boundary between the Southern Chortis terrane and the Chortis Block coincides with the boundary between Guerrero and stable western Mexico (Rogers et al., 2007). There is no indication for movement of the Chortis Block relative to North America prior to the opening of the Cayman Trough, hence we kept it fixed relative to North America between 70 and 50 Ma. With the Chortis Block and Great Arc fragments of Jamaica and the North Nicaraguan Rise being part of the North American plate, subduction of Proto-Caribbean lithosphere below the Great Arc of the Greater Antilles implies that the plate boundary configuration between 70 and 50 Ma was different compared to the 50–0 Ma period. Between 70 and 50 Ma, the Caribbean–North American plate boundary was partly a SSW–NNE transform boundary, accommodating pure left-lateral motion (with a releasing bend creating the Yucatan basin) between Belize and western Cuba, and farther to the SW between the Chortis Block and the Caribbean plate interior (see also Section 5.2.5). To the NE, the plate boundary was a WNW–ESE striking, SSW dipping subduction zone (Fig. 8a and b).

5.2.2. Caribbean–Farallon plate boundary evolution: the Central America Trench

The continuous upper Campanian–Neogene geological record of arc volcanism in the Panama–Chocó block (Denyer et al., 2006; Buchs et al., 2010) indicates ongoing active eastward subduction of the Farallon plate below the Caribbean plate throughout the 70–50 Ma period and is reconstructed as such.

5.2.3. Eastern Caribbean plate boundary evolution: the Lesser Antilles subduction system

The record of arc volcanism in the Aves Ridge (88–59 Ma; Neill et al., 2011) indicates ongoing subduction of proto-Caribbean lithosphere of the North and South American plates below the Caribbean plate in the Lesser Antilles subduction system, which is reconstructed this way.

5.2.4. Caribbean–South American plate boundary evolution

The South American continent underwent a very minor (<100 km) eastward motion with respect to North America in the 70–50 Ma period (Pindell and Kennan, 2009; Torsvik et al., 2012). The NNE-ward motion of the Caribbean plate inferred from the Yucatan basin predicts transpressional right-lateral motion between the Caribbean plate and the NNE striking margin of the South American continent. This prediction is consistent with Upper Cretaceous and Paleogene dextral strike-slip recorded from the Northern Andes (Kennan and Pindell, 2009) and the Leeward Antilles area (Escalona and Mann, 2011). We reconstructed ~700 km of dextral strike-slip between 70 and 50 Ma, alongside ~500 km of convergence between the Caribbean plate interior and the South American continent, accommodated by subduction of the Caribbean plate below South America for which evidence is provided by accreted Santonian–Paleogene 'Greater Panama' thrust slices accreted in the Northern Andes. Based on paleomagnetic studies (Stearns et al., 1982; Burmester et al., 1996), we reconstructed a 90° rotation of the Netherlands Antilles and Tobago area between 50 and 65 Ma, lining up the arc material between the Aves Rige and arc material of the Northern Andes at 65 Ma. We followed Aitken et al. (2011) for opening of the Tobago and Grenada Basins, and fixed the island of Tobago to the eastern edge of the Tobago Basin.

5.2.5. 70 Ma reconstruction

The 70 Ma reconstruction, following from the geological and kinematic constraints listed in Section 4, and the reconstruction choices outlined above, contains some kinematic inconsistencies, which we discuss in this section. Our reconstruction demonstrates that a NNE-ward motion of the Caribbean plate between 70 and 50 Ma relative to the America's is consistent with the geological records of all plate boundaries. The eastern margin of the Caribbean plate, formed by the Aves ridge and its eastern forearc, must hence be restored sufficiently far to the west so as to move along the NW South American margin prior to ~50 Ma, and to avoid overlap between the Caribbean plate and South America. Restoring the eastward Caribbean plate motion associated with the opening of the Cayman Trough, however, does not bring the Aves ridge sufficiently far west: subsequent reconstruction of the Yucatan margin would result in an overlap of ~300 km of the Aves Ridge and the Northern Andes and a gap of ~600 km between the Caribbean plate interior and the Siuna and Chortis blocks around 70 Ma. This would suggest that the eastern part of the Caribbean plate interior underwent ~300 km of post-70 Ma extension, and the northwestern part ~600 km of shortening, for which there is no evidence. We propose to solve this inconsistency by inferring ~400 km of sinistral motion between the Caribbean plate and the Chortis Block in the west, and the Cuban segment in the east prior to opening of the Cayman Trough (inset Fig. 8b). We predict the sinistral motion to be accommodated by pre-drift extension of the margins of the Cayman Trough (Mann and Burke, 1990) and by predecessors of the Oriente–Septentrional and Walton–Enriquillo–Plantain Garden fault systems south of the Cuban segment, which, as a result, is modeled as already a microplate 20 Myr prior to Cayman Trough opening. This restoration also requires the existence of a strike-slip system south of the Chortis Block, for which there is no evidence in the onshore geology. We inferred ~150 km of dextral strike-slip motion along the Hess escarpment between 50 and 30 Ma, but inferring significant transform motion along the Hess escarpment for the 70–50 Ma period would result in a major overlap between the South Nicaraguan Rise, and the Yucatan Basin and Cuba, and a gap south of the Siuna block at 70 Ma. We therefore accommodate the transform motion in this period between the South Nicaraguan Rise and the Siuna–Chortis–North Nicaraguan Rise, considering the South Nicaraguan Rise to be part of the Caribbean plate interior. Reconstructing transform motion along all plate boundaries mentioned above (in absence of evidence for major shortening or extension) requires restoring 17° counterclockwise rotation of the Caribbean plate with respect to North America (inset Fig. 8b). The resulting 70 Ma reconstruction is shown in Fig. 8b.

5.3. 100–70 Ma

5.3.1. Caribbean–North American plate boundary evolution: Proto-Caribbean Ocean subduction in the Greater Antilles subduction zone, east Cuban subduction jump, and minor transform motion

In the 100–70 Ma period, the Caribbean–North American plate boundary was characterized by subduction of Proto-Caribbean oceanic lithosphere below the Caribbean plate, as indicated by Great Arc volcanism and HP metamorphism recorded in blocks in serpentinite-matrix mélanges (see Section 5.2.1). Prior to collision with Yucatan, Jamaica and the North Nicaraguan Rise were all part of the active Great Arc and we restored these segments as fixed to the Caribbean plate interior before 70 Ma, aligned with the Cuban part of the Great Arc. The Siuna block contains Lower Cretaceous arc magmatic rocks, indicating that it was probably also part of the Great Arc before its Late Cretaceous collision (modeled at 85 Ma) with the Chortis Block (Venable, 1994). Prior

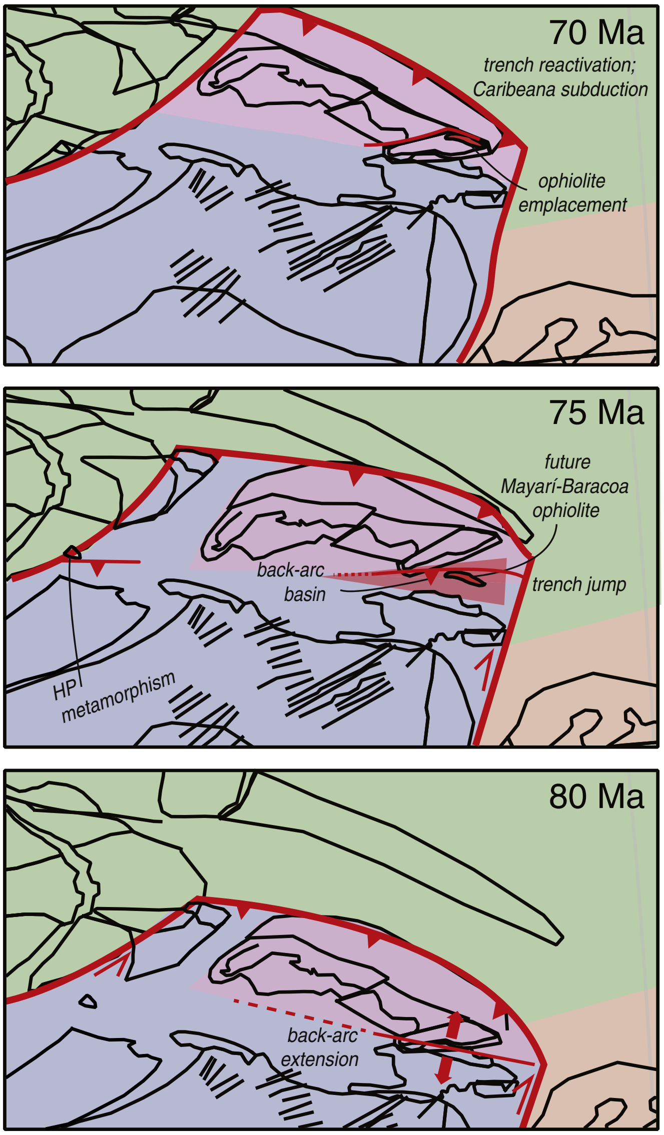

Fig. 9. Reconstruction of formation and emplacement of the Mayarí–Baracoa back-arc ophiolite over the Puriol arc fragment of Eastern Cuba, as recently postulated by Lázaro et al. (2013).

to the collision in the Late Cretaceous, we therefore aligned the Siuna block as the westward continuation of the Great Arc and reconstructed it fixed to the Caribbean plate interior. Between the 85 Ma collision of the Siuna ophiolite terrane with the Chortis Block, and collision of the Jamaica arc with the margin of southern Yucatan around 70 Ma, we reconstructed ~550 km of transform motion between the eastern Chortis Block to the west and Jamaica and the North Nicaraguan Rise to the east, parallel to the strike of the Belize margin.

To account for the emplacement of the Mayarí–Baracoa ophiolite in East Cuba with its Campanian metamorphic sole, and the consequent blueschist facies metamorphism recorded in east Cuban Great Arc rocks, we reconstructed the opening of a narrow back-arc basin between 80 and 75 Ma, following Lázaro et al. (2013). We linked the short-lived jump of the subduction zone from below the Great Arc into the back-arc to a conceptual 75 Ma Great-Arc–Caribeana collision, and accommodated the 75–70 Ma Caribbean–North America convergence in the former back-arc. We modeled the 80–75 Ma opening by a small counterclockwise rotation of the Cuban Great Arc relative to the Caribbean plate interior and a maximum, arbitrary but small extension of ~135 km between Eastern Cuba and the Caribbean plate. Subsequent inversion of this back-arc is modeled as a clockwise rotation of equal magnitude, and 135 km maximum convergence, sufficient to bury the Great Arc rocks to blueschist facies conditions (Fig. 9).

Easternmost Jamaica (blueschist metamorphic CLIP material) collides with the Siuna terrane at 76 Ma and consequently, ~100 km shortening is reconstructed between the South Nicaraguan Rise and easternmost Jamaica to account for subduction of CLIP material and blueschist metamorphism of the Mt. Hibernia Schist (West et al., 2014; see Fig. 9). After 72 Ma, easternmost Jamaica is kept fixed to the North Nicaraguan Rise.

In conclusion, the Caribbean–North American plate boundary in the 100–70 Ma period is a WNW–ESE subduction zone, consuming Proto-Caribbean oceanic lithosphere. Because the Chortis Block is located closer to the subduction zone than Yucatan–Caribeana, collision with Chortis preceded collision with Yucatan. This led to a transform fault subperpendicular to the general trend of the subduction zone, accommodating strike-slip motion between the Chortis Block and Jamaica–North Nicaraguan Rise between 85 and 70 Ma.

5.3.2. Caribbean–Farallon plate boundary evolution: origin of the Central America Trench

The oldest dated arc magmatic rocks in the Panama–Chocó block are of Campanian age (Denyer et al., 2006; Buchs et al., 2010). However, middle Turonian–Santonian and Coniacian–Santonian ages of radiolarites intercalated with arc-derived material on the Nicoya peninsula (Bandini et al., 2008) and Santonian–Campanian boninites in the accreted Greater Panama terranes in the Northern Andes (Kennan and Pindell, 2009) suggest that the volcanic arc has been active since at least the Santonian (~85 Ma). Boninites are closely related to subduction initiation (e.g., Stern et al., 2012), and we modeled the Central American subduction zone to form only after 85 Ma. Prior to subduction initiation, there was no convergent plate boundary between the Caribbean plate and the Farallon plate, nor is there evidence for an ocean spreading center of Cretaceous age in Central America; the Caribbean plate was hence either part of the Farallon plate prior to 85 Ma, or moved relative to the Farallon plate along a transform plate boundary.

5.3.3. Eastern Caribbean plate boundary evolution: the Lesser Antilles subduction system

The oldest reported arc magmas from the Aves Ridge (88 Ma; Neill et al., 2011), and island arc rocks on Aruba, Curaçao, La Blanquilla and Gran Roque (~58–89 Ma; van der Lelij et al., 2010; Wright and Wyld, 2011), which are geochemically very similar to those from the Aves Ridge, indicate that the Lesser Antilles subduction system has been active since at least the Coniacian. The Aves ridge is hence part of the Great Arc of the Caribbean, and the leading edge of the Caribbean plate was somewhat irregular in shape and orientation: the Aves ridge strikes almost orthogonal to the strike of the Greater Antilles segment of the Great Arc. To have simultaneous arc volcanism in both segments, subduction must have been highly oblique below the eastern Caribbean margins between ~90 and ~60 Ma.

5.3.4. Caribbean–South American plate boundary evolution

In the 100–75 Ma period, South America moved to the southeast relative to North America, which must have been accommodated by the last stages of spreading in the Proto-Caribbean Ocean. At ~75 Ma, this motion reversed, associated with an ~5 Myr period without significant motion between the two continents (Pindell and Kennan, 2009; Torsvik et al., 2012). Major Late Cretaceous dextral faulting and shearing was reconstructed in the Northern Andes (Kerr et al., 1998; Trenkamp et al., 2002; Kennan and Pindell, 2009). In addition, convergent motion resulted in the emplacement and imbrication of slivers of ~87 Ma CLIP magmas over the South American margin in Colombia, and associated deposition of ultramafic-derived sediments in the central belt of the Northern Andes (Kerr et al., 1997; Vallejo et al., 2006; Kennan and

Pindell, 2009). This requires restoration of transpression between the Caribbean plate and South America, but quantitative kinematic data that constrain the amount of convergence and translation are not available. We reconstructed a Caribbean–South America dextral strike-slip component of ~1300 km between 100 and 70 Ma, and a convergence component of ~200 km.

HP rocks on Isla Margarita represent deep underthrusting of the margin of the South American continent. Peak metamorphism was interpreted to occur around 100–90 Ma (Maresch et al., 2009), which we adopt as the timing of final collision between the overriding Caribbean plate with the downgoing South American continent (see also Section 5.4). Our restoration places Isla Margarita near the coast of present northern Peru during peak metamorphism, east of the southern edge of the Aves Ridge. Greenschist-facies metamorphism on Isla Margarita prevailed during exhumation, as a result of shearing during the island's journey towards its present position, north of Venezuela (Maresch et al., 2009).

Following our reconstruction philosophy, we interpret the arc rocks in the Northern Andes as part of the Great Arc of the Caribbean and test the hypothesis that the Great Arc was one, continuous island arc. Plate kinematically, this is the most simple scenario (and therefore preferred), but we acknowledge another interpretation; the Northern Andes Arc may have been a separate subduction zone with a reversed polarity.

5.3.5. 100 Ma reconstruction

The 100 Ma position of the northern, leading edge of the Caribbean plate is an interpolated position between the 135 Ma (see Section 5.4) and the 70 Ma positions, assuming a continuous motion rate. The Caribbean plate is reconstructed another ~900 km SSW of its 70 Ma position relative to North America, and we restored an ~9° counterclockwise rotation between 100 and 70 Ma to reconstruct transpressional motion with South America. The resulting 100 Ma reconstruction is shown in Fig. 8c. At 75 Ma, the North Nicaraguan Rise overlaps with the eastern part of the Chortis Block. Without moving the Chortis Block further west prior to 50 Ma, it is not possible to avoid this overlap. Our reconstruction therefore predicts ~100 km of extension within or between the offshore portions of the Chortis Block and the North Nicaraguan Rise between 75 and 70 Ma.

Between 91 and 88 Ma, the bulk of the plateau basalts of the CLIP was been formed. Shortly after formation, the CLIP may have been twice its present size, illustrated by the size of the blue area in Fig. 8b, minus the Cuban segment and the Aves Ridge. The center of the original CLIP would have been ~1000 km south of southern Hispaniola.

5.4. 135–100 Ma

5.4.1. Caribbean–North/South American plate boundary evolution: origin of the Great Arc of the Caribbean

K–Ar ages from samples of the HP metamorphic blocks in the Cuban serpentinite mélange range from ~130 to 60 Ma (Somin and Millán, 1981; Somin et al., 1992) and the oldest arc magmatic rocks have been dated at 133 Ma (Rojas-Agramonte et al., 2011), indicating that subduction of the Proto-Caribbean Ocean started in at least the Hauterivian. Pindell et al. (2012) modeled an age of 135 Ma for subduction initiation. We have adopted a 135 Ma age for initiation of subduction below the Great Arc of the Caribbean, although we note that this is a minimum age, and subduction may have started earlier. Prior to the onset of subduction and the formation of the Great Arc of the Caribbean, there was no convergent boundary between the Caribbean(/Farallon) plate and the North and South American parts of the young Proto-Caribbean Ocean.

In the Northern Andes, sheared and accreted Lower Cretaceous island arc rocks and continental HP rocks of 120–110 Ma, found in the same tectonic belt, record the collision of the Great Arc with the South American continent (Nivia et al., 2006; Kennan and Pindell, 2009). This provides constraints on the position of the leading edge of the Caribbean plate in the late Early Cretaceous. In the Late Jurassic–earliest Cretaceous, prior to collision, the Northern Andes formed a passive margin between the South American continent and the Proto-Caribbean Ocean (see Section 5.5) and the accreted arc rocks represent the leading edge of the Caribbean plate. We therefore fixed the arc terranes of the Northern Andes to the Caribbean plate prior to the ~100 Ma arc–continent collision. In the 135–100 Ma period, SE-ward motion of the South American continent relative to North America (Pindell and Kennan, 2009; Torsvik et al., 2012) must have been accommodated by ~1100 km of Proto-Caribbean Ocean spreading. The 135 Ma position of the Caribbean plate, at the time of subduction initiation, is constrained by the relative positions of the North and South American continents. The Caribbean plate is reconstructed to the west of the American continents, aligning the Great Arc more or less with the western margins of the North and South American continents, but with an opposite subduction polarity (i.e. westward below the Great Arc, and eastward below the Americas). This alignment was acquired by reconstructing ~900 km of eastward motion and a 5° counterclockwise rotation of the Caribbean plate between 135 and 100 Ma.

5.4.2. Chortis–Southern Mexico: evolution of the Motagua mélanges

HP rocks within the serpentinite mélange of the South Motagua unit (ophiolitic El Tambor complex) are dated at 124–113 Ma and 144–132 Ma ($^{40}Ar/^{39}Ar$ cooling ages; Harlow et al., 2004; Brueckner et al., 2009) and 158–154 Ma (U/Pb zircon crystallization ages; Flores et al., 2013). This suggests Late Jurassic–earliest Cretaceous subduction of the Chortis Block below the margin of southern Mexico until collision and exhumation of the subducted continental margin between 144 and 113 Ma. We reconstructed this by giving the Chortis Block a motion similar to the motion of South America between 156 and 120 Ma, resulting in transpressional motion between the Chortis Block and southern Mexico and oblique collision around 120 Ma. This essentially means that we reconstructed a North–south America plate boundary jump from north of Chortis to its south in the Early Cretaceous.

5.4.3. 135 Ma reconstruction

The resulting 135 Ma reconstruction is shown in Fig. 8d. Around 130 Ma, after full development of the eastern Caribbean subduction system and prior to collision of the Great Arc with parts of the North and South American plates, and subduction of the Caribbean plate interior below the South American continent, the Caribbean plate lithosphere and the Great Arc of the Caribbean are at their maximum sizes.

5.5. 200–135 Ma

The 143.74 ± 0.33 Ma U–Pb age of ocean floor from La Désirade Island in the Lesser Antilles (Mattinson et al., 2008), and Late Jurassic ages of sediments related to the Cuban ophiolites (Iturralde-Vinent and Marí-Morales, 1988; Iturralde-Vinent, 1994, 1996; Llanes et al., 1998) show that the oceanic crust of the Caribbean plate formed during this time. Spreading was probably perpendicular to the orientation of the magnetic anomalies identified below the CLIP by Ghosh et al. (1984). Given the uncertainties in our reconstruction and in the imaged anomalies of Ghosh et al. (1984), the spreading responsible for these anomalies may have been subparallel to the NW–SE direction of spreading in the Proto-Caribbean Ocean and the early stages of opening of the Central Atlantic Ocean (Labails et al., 2010): the difference in reconstructed orientation of the anomalies and of Proto-Caribbean spreading is up to ~25°. We will discuss the likelihood of a Proto-Caribbean or Pacific origin of the Caribbean plate lithosphere in the discussion section, by assessing the plate boundary configurations and their plate kinematic feasibility.

Because kinematic estimates of Ross and Scotese (1988) remain pertinent for southwestern Mexico, we followed their reconstruction and reconstructed southwestern Mexico westward prior to 143 Ma by

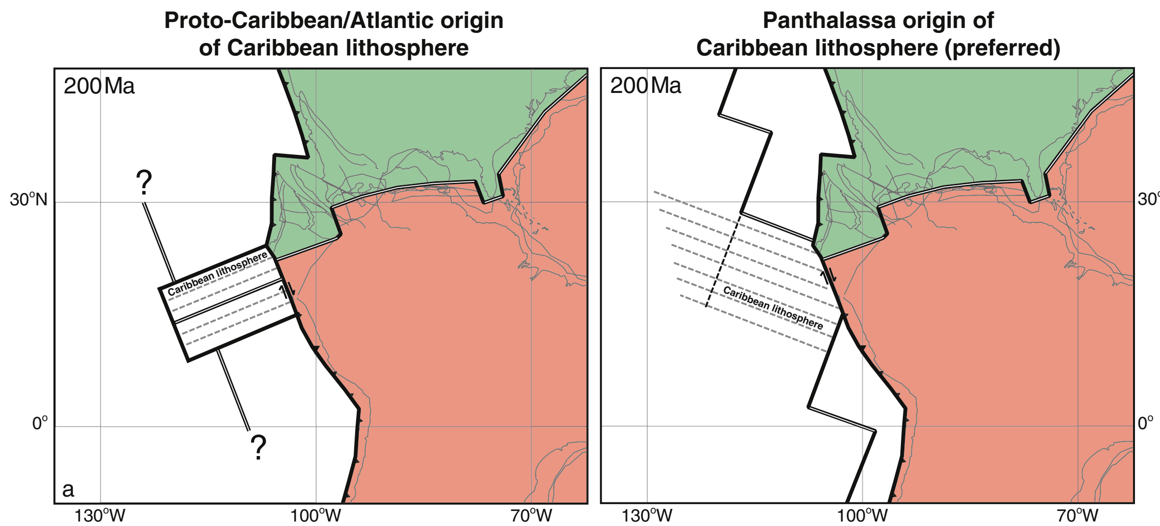

Fig. 10. Conceptual plate boundary configurations at 200 Ma.

motion in the Trans-Mexican volcanic belt. The reconstructed NW-ward motion of northern Mexico relative to North America along the Mojave Megashear is based on Müller et al. (2008). For opening of the Gulf of Mexico, we follow the reconstruction of Pindell and Kennan (2009), reconstructing Yucatan NW-ward prior to 130 Ma. The resulting 170 Ma reconstruction is shown in Fig. 8e.

Finally, a 200 Ma reconstruction is derived with a total of 300–400 km of overlap between the Northwestern Andes and Mexico. This overlap results from treating these margins and blocks as rigid, and would suggest 300–400 km of extension being accommodated by pre-drift extension associated with the opening of the Proto-Caribbean Ocean. Such a number is within the range of extension accommodated in modern passive margins (Torsvik et al., 2008). The 200 Ma reconstruction is shown in Fig. 8f.

6. Discussion

6.1. Constraints on Jurassic plate boundaries

Between ~200 and ~135 Ma, during the break-up of Pangea, the Caribbean plate did not exist yet as a separate plate. Africa and South America were rifting and drifting away from North America, opening the Central Atlantic and Proto-Caribbean oceans. Subduction already occurred below the Americas around and before 170 Ma (Litherland et al., 1994; Noble et al., 1997; González, 2001; Villagómez et al., 2008; Kennan and Pindell, 2009; Boekhout et al., 2012), as well as in offshore intra-oceanic subduction zones (van der Meer et al., 2012; Buchs et al., 2013; Sigloch and Mihalnyuk, 2013) but little is known about exact directions of plate motion within the Panthalassa Ocean, because pre-Cretaceous plate reconstructions of this region are conceptual and not constrained by any magnetic anomaly data (Seton et al., 2012; van der Meer et al., 2012; Sigloch and Mihalnyuk, 2013). The absence of arc magmatism older than 133 Ma suggests that there was no subduction zone between the future Caribbean lithosphere and the Proto-Caribbean Ocean yet before that time, although we note again that this is the minimum age of subduction initiation.

We use our reconstruction to assess whether the Caribbean lithosphere before inception of Great Arc subduction may have formed (west of the North and South American continents) due to North–south America spreading (and can be considered 'Atlantic' or 'Proto-Caribbean' in origin), or whether a Panthalassa origin is more likely. We note here that both scenarios would qualify as 'Pacific origin' in the old 'in-situ' (James, 2006) versus 'Pacific' (Pindell et al., 2006) origin debate. As described in Sections 5.4.3 and 5.5, the orientation of the magnetic anomalies identified by Ghosh et al. (1984) strike within 25° from the orientation perpendicular to spreading in the Proto-Caribbean Ocean, and the ages of formation of Caribbean oceanic crust are quite similar to spreading directions and ages that would have formed due to the Proto-Caribbean (Atlantic) spreading. It may thus be possible that the Proto-Caribbean Ocean and the Caribbean lithosphere were connected, or perhaps separated by a transform plate boundary that became inverted to become the Great Arc subduction zone. Such a transform was conceptually termed the 'inter-American transform' by Pindell et al. (2012) and the question thus is: was the inter-American transform within Proto-Caribbean lithosphere, or between Proto-Caribbean and Panthalassa lithosphere?

Due to the similar strike of the 'inter-American transform' and the North and South American subduction zones, a Proto-Caribbean origin of Caribbean lithosphere creates some difficulties for the plate boundary configuration between 200 and 135 Ma. Fig. 10a shows a conceptual, and quite complex configuration, whereby Panthalassa lithosphere is subducting below the North and South American continents, while maintaining absence of convergence between Panthalassa and Proto-Caribbean lithospheres. Alternatively, as shown in Fig. 10b, a much simpler plate boundary scenario may be invoked, also consistent with the reconstructed spreading directions of the Caribbean plate interior, with a Panthalassa origin for the Caribbean plate. This scenario is very similar to the modern Pacific–Juan de Fuca–North American–San Andreas system. This scenario may also explain the reconstructed age and paleo-spreading direction of the Caribbean lithosphere, the presence of an intra-American transform and subduction zones below the North and South American subduction zones. We stress that the scenarios of Fig. 10 are conceptual, and are thought-exercises to test viability of an intra-American or Panthalassa origin of the Caribbean lithosphere; nevertheless, we note that a Panthalassa origin requires the simplest scenario.

6.2. Caribbean plate origin debate

In both plate kinematic scenarios shown in Fig. 10, the position of origin of the Caribbean plate is the same: west of the North and South American plates. However, this position has been subject of debate (e.g. James, 2006; Pindell et al., 2006). Here, we emphasize the two main plate kinematic arguments against the 'in-situ' model of James (2006, 2009, where the Caribbean plate originates between the North and South American continents as a result of Proto-Caribbean spreading), and illustrate why we are convinced that this model is in violation

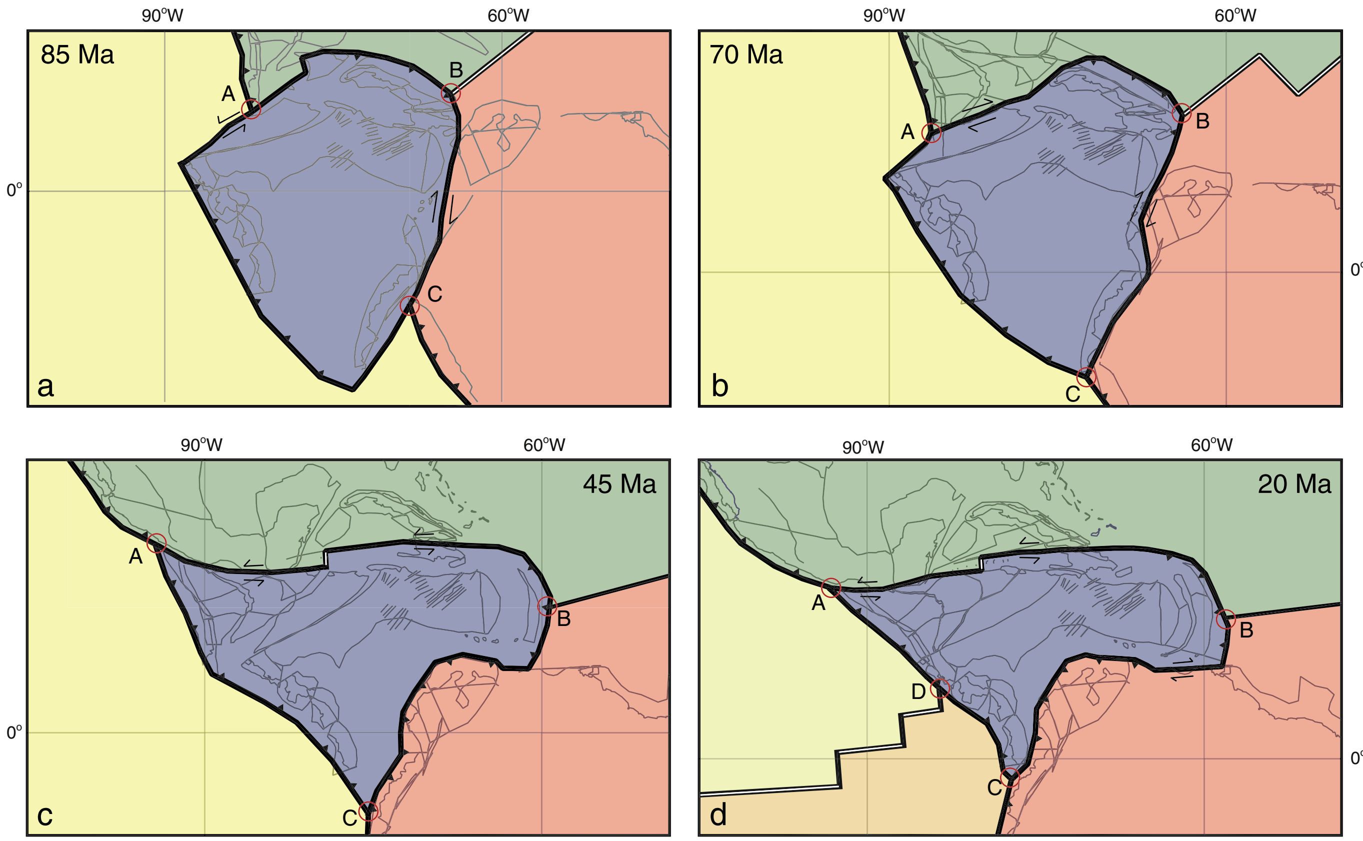

Fig. 11. Plate boundary configurations at 85, 70, 45 and 20 Ma.

with geology and basic plate kinematics. For many other arguments, see e.g. Pindell and Barrett (1990); Pindell et al. (2006, 2012).

For the last 50 Myr, the motion of the Caribbean plate relative to North America is constrained in numerous ways. We, as most authors, put faith in the oceanic spreading history of the Cayman Trough, even though the magnetic anomalies of the Trough are poor (Leroy et al., 2000). However, in conjunction with assessment of the 1000 km long west to east migrating foreland basins of northern South America (Pindell et al., 1998; Pindell and Kennan, 2007; Escalona and Mann, 2011), as well as the close match between the Cayman Trough's length and orientation and the Eocene–Recent motion history of North America relative to the hotspot reference frame to which the Caribbean Plate is fixed (Pindell and Kennan, 2009, see their Fig. 25), we are confident that the Cayman Trough does in fact record most (but not all; Sykes et al., 1982; Mann and Burke, 1990) of the Eocene–Recent displacement between the Caribbean and North American plates, which occurred in an E–W (~080°) direction relative to a stable North America.

Furthermore, the record of ongoing subduction below the Aves Ridge and Lesser Antilles Arc since at least the Late Cretaceous (Fox and Heezen, 1975; Pinet et al., 1985; Bouysse, 1988; Neill et al., 2011) and tomographic evidence of at least 1100 km of subduction (van der Hilst, 1990; van Benthem et al., 2013) indicate that there has been significant relative motion between the Caribbean plate and the Atlantic/Proto-Caribbean Ocean, whereas the east coasts of North and South America are passive margins (i.e. no motion between the Americas and the Atlantic Ocean). This difference is direct evidence that the Caribbean plate is a separate plate that has been moving eastwards relative to North and South America during the Cenozoic.

Another significantly different Caribbean model is the 'plateau-collision' model of Duncan and Hargraves (1984), Burke (1988), Kerr et al. (2003) and Hastie and Kerr (2010). This model includes a subduction polarity reversal in the Great Arc subduction zone after collision of the CLIP with the Great Arc in the Late Cretaceous. Following our reconstruction philosophy, we do not include such a polarity reversal, although this model has no kinematic problems and we acknowledge the possibility of this scenario.

Table 3
Global Apparent Polar Wander Path of Torsvik et al. (2012) rotated into the coordinates of the Caribbean plate interior, the Chortis Block, and the Cuban segment. Codes behind the regions refer to the codes in the GPlates reconstruction given in the online Appendix.

Age	A95	Chortis frame (2023)		Cuban frame (2038)		Caribbean plate interior (2007)	
		PoleLat	PoleLon	PoleLat	PoleLon	PoleLat	PoleLon
0	1.9	−88.5	353.9	−88.5	353.9	−88.5	353.9
10	1.8	−86.7	336.0	−86.4	342.2	−86.6	337.0
20	2.6	−82.3	345.4	−83.7	343.2	−84.3	337.1
30	2.6	−76.4	348.5	−82.1	338.7	−84.3	314.6
40	2.9	−65.1	355.7	−80.1	337.2	−83.2	306.0
50	2.8	−55.8	5.2	−72.2	353.3	−77.3	347.3
60	2.1	−53.5	11.6	−74.7	351.0	−73.9	2.5
70	2.5	−53.7	14.1	−79.1	334.0	−72.5	5.2
80	2.9	−54.8	12.6	−76.6	321.5	−71.0	354.7
90	2.5	−56.4	9.4	−71.7	321.4	−66.3	350.5
100	3.3	−58.1	6.4	−65.4	327.7	−59.7	352.2
110	3.3	−56.1	17.2	−64.4	345.7	−57.4	6.3
120	2.6	−54.1	15.3	−59.7	348.4	−52.6	7.6
130	2.8	−48.4	14.5	−56.7	349.4	−49.6	8.0
140	6.0	−42.7	17.6				
150	6.4	−43.5	8.2				
160	5.1	−35.9	4.3				
170	4.6	−32.3	3.2				
180	3.4	−38.4	357.5				
190	2.9	−41.6	348.9				
200	2.8	−39.5	340.0				

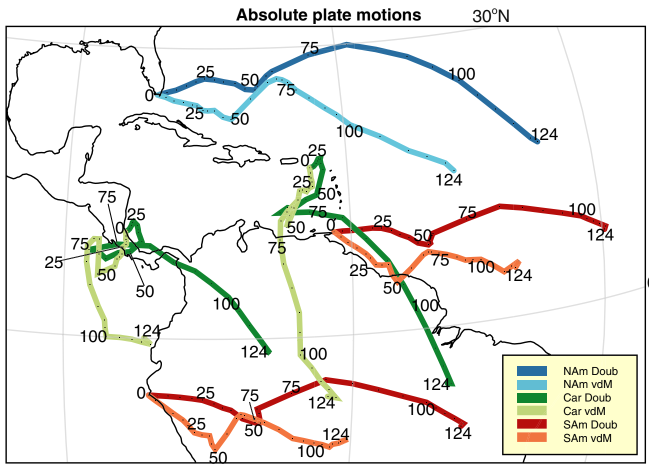

Fig. 12. Absolute plate motions of the Caribbean (Car), South American (SAm) and North American (NAm) plates. Doub, global moving hotspot reference frame of Doubrovine et al. (2012); vdM, slab-fitted mantle reference frame of van der Meer et al. (2010).

6.3. Cretaceous and younger plate boundary configurations and triple junctions

Since the onset of subduction in the Panama–Costa Rica Arc, the Caribbean plate was a separate tectonic plate, bounded on all sides by plate boundaries. Prior to Central American subduction initiation, the nature of the triple junctions at the ends of the boundary between the Farallon plate and future Caribbean plate remain unknown. From late Cretaceous times until the Present, the Caribbean plate has three triple junctions: the North American–Caribbean–Farallon (A), North American–South American–Caribbean (B) and Caribbean–South American–Farallon (C) triple junctions (Fig. 11a). Since Miocene fission of the Farallon plate into the Cocos and Nazca plates (Barckhausen et al., 2008), there was a fourth: the Cocos–Caribbean–Nazca triple junction (D), which is a stable ridge–trench–trench triple junction subducting Cocos and Nazca lithosphere in the Central American trench (Fig. 11d). We here assess the feasibility of our reconstruction by testing whether the triple junctions are consistent with the basic rules of plate tectonics.

Between ~88 and 50 Ma, the northern triple junction (North-American–Caribbean–Farallon, A) is a stable trench–transform–transform triple junction south of the Siuna block (Fig. 11a). Chortis–Siuna is part of the North American plate. In this triple junction, the Farallon plate subducts below the North American plate and the Caribbean plate has a transform motion relative to Farallon and North America. At ~50 Ma, the Chortis Block transferred to the Caribbean plate and as a consequence, the triple junction jumped towards the western end of the Motagua fault zone and transformed into a transform–trench–trench triple junction (Fig. 11c) where Farallon subducted below the Caribbean and North American plates, and the Caribbean–North America motion is a pure transform. Subduction below Mexico initiated along the former transform upon ESE-ward motion of the triple junction relative to North America; initiation of subduction below southern Mexico must hence have been strongly diachronous, younging ESE-ward.

Between 135 Ma and the age of cessation of Proto-Caribbean spreading inferred from reconstructions of the Atlantic Ocean (~75 Ma), the eastern triple junction (North American–South American–Caribbean, B) is a stable ridge–trench–trench triple junction (Fig. 11a), at which the spreading center of the Proto-Caribbean/Central Atlantic Ocean subducted. After ~75 Ma, the Lesser Antilles subduction system continued to subduct the North and South American plates, but the relative motion between these two is very minor. The plate circuit of Torsvik et al. (2012) suggests that the North–south American plate contact was mildly transpressional.

The southwestern triple junction (Caribbean–South American–Farallon, C) has undergone a similar evolution as the northwestern triple junction A. Since ~88 Ma, it has been a stable trench–transform–transform triple junction (Fig. 11a), where the Farallon plate subducted below South America and the Caribbean plate had a transform motion relative to Farallon and South America. Because the relative motion between the Caribbean and South American plates was accommodated along several faults and shear zones in the Northern Andes, the plate boundary is diffuse and the location of the triple junction is not precisely constrained; it may have changed positions several times during the history of dextral motion in the Northern Andes.

Around 75 Ma, the Panama–Costa Rica Arc arrived at the triple junction, transforming it into a trench–trench–transform triple junction (Fig. 11b), where the Farallon plate subducted below the Caribbean and South American plates. The Northern Andes evolved from a passive margin that collided with the Great Arc, to a transform boundary, to an overriding plate, whereby the Caribbean plate subducted below South America (Fig. 11c). Comparable to the transition from eastward Panthalassa subduction below Gondwana to westward subduction below Panthalassa/Caribbean lithosphere, the subduction switch seems to have been facilitated and in time separated by a stage of transform motion. Similar to the Mexican margin, renewed eastward subduction below South America was likely diachronously initiated along that transform, behind a NE-ward migrating trench–trench–transform triple junction.

6.4. Global Apparent Polar Wander Path in Caribbean coordinates

Our reconstruction may provide a reference for future paleomagnetic research in the Caribbean region. We have rotated the Global Apparent Polar Wander Path of Torsvik et al. (2012) into coordinates of three most prominent regions, using the Euler poles given in the online appendix: the Caribbean plate interior, the Chortis Block, and the Cuban segment (Table 3). These poles may be used to paleomagnetically test the predictions of our reconstruction, as well as serve as reference for future paleomagnetic analyses in the Caribbean region.

Table 4
Comparison of tectonic reconstructions on the basis of amounts of subduction. Maracaibo; subduction of the Caribbean plate below the South American continent.Modified from van Benthem et al. (2013).

Subduction zones	Tomography	Plate reconstructions			
	van Benthem et al. (2013)	Meschede and Frisch (1998)	Müller et al. (1999)	Pindell and Kennan (2009)	This study
Lesser Antilles	1100	1100	1100	1100	1100
Maracaibo	~1000	0–3000	1000	800	750–850
Muertos	<200	?	200 ± 94	0–150	30
Nicaraguan Rise	<200	~200	800	~300	0
Great Arc	1200–2100	1000	x	2000–3000	2200–2800

6.5. Caribbean motion in an absolute reference frame

Around 140 Ma, the absolute motion of the North American plate changed from NNW to W (using the slab-fitted absolute reference frame from van der Meer et al. (2010) based on the True-Polar Wander corrected paleomagnetic reference frame of Steinberger and Torsvik (2008), see online Appendix). With the Farallon/Panthalassa plate(s) being anchored in the mantle by slabs below or offshore the Americas, such a change in absolute plate motion may have driven the initial E–W convergence between the Proto-Caribbean lithosphere — with passive margins attached to the American continents — and the Caribbean lithosphere — connected (probably through a transform fault) to Panthalassa lithosphere (Fig. 10). This would be an interesting case study for analyzing absolute plate motion forcing of subduction initiation. Subduction probably initiated along a transform fault (the intra-American transform; Pindell et al., 2012). The transformation of this inter-American trench-to-trench transform resulted in a continuous subduction zone from northern North America to southern South America, but the direction of subduction in the central part, subducting the Proto-Caribbean Ocean instead of Panthalassa lithosphere, was eastwards instead of westwards. The former transform–transform–ridge triple junction between the inter-American trench and the Proto-Caribbean spreading center transformed into a trench–trench–ridge triple junction (Figs. 10 and 11). Evidence for this triple junction has been found in Aptian amphibolite blocks from eastern Cuba (García-Casco et al., 2008b; Blanco-Quintero et al., 2010). After subduction initiation, the future Caribbean plate was overriding the continuously spreading Proto-Caribbean Ocean and its spreading center.

The western Caribbean subduction system formed at 88–80 Ma (Pindell and Kennan, 2009). This implies that, if the Caribbean plate originated as a fragment of the Farallon plate, the Caribbean LIP (91–88 Ma) would actually be a Farallon LIP at the time of formation (Pindell and Kennan, 2009). In our reconstruction, the center of the present day CLIP was around 90 Ma located ~2000 km east of the site of the present day Galápagos hotspot in the slab-fitted mantle reference frame of van der Meer et al. (2010), and ~3000 km in the global moving hotspot reference frame of Doubrovine et al. (2012). If the CLIP represents the arrival of the Galápagos plume head below the lithosphere, this plume must have undergone significant motion relative to the mantle in a direction opposite to that of the overriding plate motion (i.e. against the mantle wind). We therefore consider it unlikely that the CLIP formed as a result of the plume-head stage of the Galápagos hotspot, but may source from a different plume event.

The North and South American plates moved WNW-ward between 100 and 70 Ma and SW-ward between ~70 and 50 Ma and resumed their WNW-ward motion at ~50 Ma (Fig. 12). The Caribbean plate had a constant N (van der Meer et al., 2010) to NW-ward (Doubrovine et al., 2012) motion between 100 and 70 Ma. Around 70 Ma, the motion stagnated and, in particular for the last 40 Myr, the Caribbean plate has been more or less stable relative to the mantle, particularly in longitude. It is remarkable that the relatively small Caribbean plate may be one of the most stable plates in an absolute reference frame. The oppositely dipping subduction zones in the west and east of the Caribbean plate likely function as an anchor keeping the subduction zones, and therefore the Caribbean plate, relatively stable with respect to the mantle.

Since 50 Ma, the Caribbean plate is at first order bounded by its present plate boundaries: subduction of the adjacent oceanic lithosphere in the Lesser Antilles subduction system and the Central American Trench and transform motion between the Caribbean plate and the North and South American plates. We note that collision of the leading edge of the Caribbean plate with Caribeana and Yucatan did not have a marked influence on relative plate motions, whereas the collision with the Bahamas platform coincided with a major change of direction of motion, onset of spreading in the Cayman Trough and the end of subduction. In particular in the slab-fitted mantle reference frame of van der Meer et al. (2010), the absolute motion of the Caribbean plate underwent a major change at 50 Ma (Fig. 12). We therefore suggest that, as factors influencing relative plate motion, the interaction between the Caribbean plate and the North and South American plates may be inferior to absolute plate motion and mantle anchors.

6.6. Comparing tectonic reconstructions

By comparing the amount of subduction in different subduction zones predicted by a reconstruction, different reconstructions can be quantitatively compared and tested against dimensions of relics of subduction images by seismic tomography recently estimated by van Benthem et al. (2013). Table 4 lists three of the tectonic reconstructions mentioned in the introduction. Our reconstruction predicts quite similar subduction budgets as Pindell and Kennan's (2009), except for the Nicaraguan Rise where we modeled no subduction, consistent with the absence of evidence of any slab that may be attributed to a Nicaraguan Rise subduction zone (van Benthem et al., 2013). The differences with Meschede and Frisch (1998) and Müller et al. (1999) are considerably larger. For the former, this difference is mainly the much larger amount of subduction below the Great Arc of the Caribbean in our model, and for the latter, our predicted amount of Muertos, Nicaraguan Rise and Great Arc subduction strongly deviates. Our modeled amounts of subduction are consistent with the estimates derived from seismic tomography (van Benthem et al., 2013), although we reconstructed a somewhat larger amount of subduction below the Great Arc. The tomographic estimates are, however, based on poorly constrained slab shrinking factors associated with slabs penetrating the lower mantle.

7. Conclusions

This paper reviews the tectonic history of the Caribbean region and presents a quantitative, kinematically consistent reconstruction back to 200 Ma, relative to a stable North America. The poles describing the relative motion of the tectonic elements of the Caribbean plate are given in the online appendix. The reconstruction can easily be adapted to newly obtained data and may therefore form the basis of further quantitative geodynamic and kinematic research concerning the Caribbean region.

Based on our reconstruction, we conclude the following.

(1) The Caribbean plate formed west of the North and South American continents. The direction and age of spreading in the Caribbean plate was similar to the direction of spreading in the Proto-Caribbean Ocean, but a Panthalassa origin of the Caribbean lithosphere can be explained by a much simpler plate kinematic scenario.
(2) The Northern Andes and Mexico have accommodated 300–400 km of Early Jurassic pre-drift extension. This number is within the range of pre-drift extension accommodated in modern passive margins.
(3) During its formation, the center of the Caribbean Large Igneous Province was located 2000–3000 km east of the present-day position of the Galápagos hotspot. It is therefore not likely that the CLIP formed as the result of the Galápagos plume-head stage, but may for a separate plume event instead.
(4) Subduction initiation below the Great Arc of the Caribbean around or before 135 Ma may be the result of westward absolute plate motion of the Americas over their western subduction zones acting as anchors in the mantle; such an event would initiate E–W convergence between the future Caribbean lithosphere and the Proto-Caribbean lithosphere between Americas and have initiated subduction along a transform fault separating these.
(5) The major Caribbean plate motion changes from NE-ward to E-ward relative to North America around 50 Ma are widely ascribed to collision of the Caribbean plate with the Bahamas.

However, we note that this change coincides with a southwestward to westward absolute plate motion change of the Americas, and may be dominated by far-field rather than local stress changes.

(6) All subduction zones surrounding the Caribbean plate, including the Great Arc subduction zone, the Lesser Antilles Trench, the Central American trench, the South Mexican and Colombian, as well as the Maracaibo subduction zone, appear to have initiated along former transform faults. When a migrating triple junction changes a transform plate boundary into a subduction zone, subduction initiates highly diachronously. Regional subduction polarity changes are facilitated by, and intervened by phases of transform motion between plates.

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Appendix A. Supplementary data

Supplementary material 1. readme.
Supplementary material 2. shape file.
Supplementary material 3. rotation file.

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