Morphostructure at the junction between the Beata ridge and the Greater Antilles island arc (offshore Hispaniola southern slope)

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ABSTRACT

Oblique convergence between the Caribbean plate's interior and the inactive Greater Antilles island arc has resulted in the collision and impingement of the thickened crust of the Beata ridge into southern Hispaniola Island. Deformation resulting from this convergence changes from a low-angle southward-verging thrust south of eastern Hispaniola, to collision and uplift in south-central Hispaniola, and to left-lateral transpression along the Southern peninsula of Haiti in western Hispaniola. Using new swath bathymetry and a dense seismic reflection grid, we mapped the morphological, structural and sedimentological elements of offshore southern Hispaniola. We have identified four morphotectonic provinces: the Dominican sub-basin, the Muertos margin, the Beata ridge and the Haiti sub-basin. The lower slope of the Muertos margin is occupied by the active Muertos thrust belt, which includes several active out-of-sequence thrust faults that, were they to rupture along their entire length, could generate large-magnitude earthquakes. The interaction of the thrust belt with the Beata ridge yields a huge recess and the imbricate system disappears. The upper slope of the Muertos margin shows thick slope deposits where the extensional tectonics and slumping processes predominate. The northern Beata ridge consists of an asymmetrically uplifted and faulted block of oceanic crust. Our results suggest that the shallower structure and morphology of the northern Beata ridge can be mainly explained by a mechanism of extensional unloading from the Upper Cretaceous onward that is still active residually along the summit of the ridge. The tectonic models for the northern Beata ridge involving active reverse strike–slip faults and transpression caused by the oblique convergence between the Beata ridge and the island arc are not supported by the structural interpretation. The eastern Bahoruco slope an old normal fault that acts as a passive tear fault accommodating the sharp along-strike transition from low-angle thrusting to collision and uplifting.

Keywords: Caribbean plate; Hispaniola block; Beata ridge; Muertos thrust belt; Island arc; Collisional tectonics

1. Introduction

The deadly 2010 Haiti earthquake has renewed the interest of the scientific community in the study of active processes in the northern Caribbean (e.g., Calais et al., 2010; Hornbach et al., 2010; McHugh et al., 2011; Mercier de Lépinay et al., 2011; Prentice et al., 2010; ten Brink et al., 2011). However, although the first order tectonic framework of the NE Caribbean is well understood, there are many under water regions where the knowledge of the tectonics remains limited. For example, the southern margin of Hispaniola Island and the northern Beata ridge were previously studied using mostly widely-spaced and low-resolution seismic reflection profiles (e.g., Byrne et al., 1985; Cooey, 1982; Driscroll and Diebold, 1998; Jany, 1989; Ladd et al., 1977, 1981; Mauffret and Leroy, 1999; Mauffret et al., 2001; Mercier de Lépinay et al., 1988). This sparse coverage complicated the understanding of the along-strike variations of tectonic structures in the region. Furthermore, the paucity of seismic activity along structures such as the Beata Ridge makes it difficult to identify possible active faulting.

This work presents new high-resolution, systematic swath bathymetry and seismic reflection data and combines them with previous swath bathymetry and re-processed single-and multi-channel reflection profiles to provide a combined interpretation of the morphotectonic features of offshore southern Hispaniola. We focus on the identification and characterization of recent tectonic features in the region and provide well-defined targets to carry out future studies for seismic and tsunamigenic hazard assessment. We use the newly-integrated data set to review previous hypotheses for the evolution of the Beata ridge and its interaction with the island arc and to discuss the tectonic implications of the ongoing deformation in offshore southern Hispaniola.

2. Tectonic setting

Based on GPS-derived velocities and geological data, the Caribbean plate is moving relative to the North American plate at a rate of 20.0 ± 0.4 mm/y towards 074° ± 1° (DeMets et al., 2010) (Fig. 1). This eastward motion has been taking place during most of the Cenozoic controlled by left-lateral transcurrent systems (Mann et al., 2002). As a consequence of this movement, the eastern Greater Antilles (i.e., Hispaniola and Puerto Rico) have evolved to an inactive intra-oceanic island arc consisting of a 250 km-wide band of deformation, in which transtension and transpression as well as microplate and block tectonics take place (e.g., Byrne et al., 1985; Chaytor and ten Brink, 2010; Dillon et al., 1996; Dolan et al., 1998; Jansma and Mattioli, 2005; Larue and Ryan, 1998; Mann and Burke, 1984; Mann et al., 1995; Masson and Scanlon, 1991; ten Brink and López-Venegas, 2012). The eastward motion of the Hispaniola block is being impeded relative to the motion of the Caribbean plate's interior due to the collision with the Bahamas banks (Mann et al., 2002). This collision has resulted in the development of strain partitioning in the forearc and in the inner island arc (Calais et al., 2002) and convergence in the retroarc (Fig. 1; Granja Bruña et al., 2009; Mercier de Lépinay et al., 1988). The convergence in the retroarc has yielded the 650 km-long Muertos margin showing two main structural and morphological features: the Muertos thrust belt and the Muertos trough (Matthews and Holcombe, 1976). The convergence in the retroarc is mainly driven by the oblique subduction/collision between the Caribbean and North American plates along the Hispaniola and Puerto Rico trenches (ten Brink et al., 2009). As a result of the steep bathymetric gradients, gravitational forces are also likely to be involved in the compressive stresses in the retroarc (i.e., slope stresses or gravitational spreading), but in this case they may not be primary driving forces (ten Brink et al., 2009). The

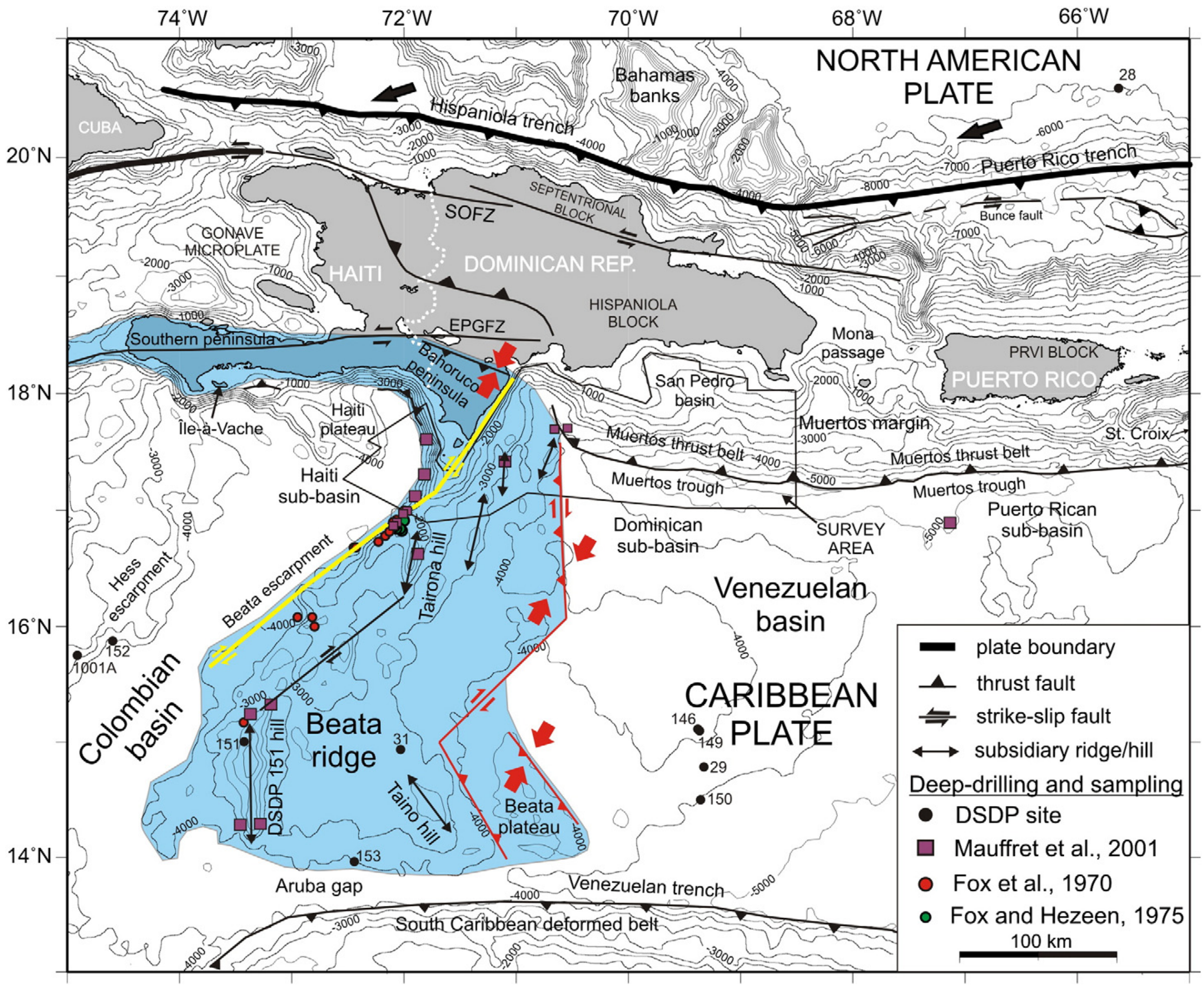


Fig. 1. Contoured bathymetry map of the central Caribbean showing a summarized tectonic setting (modified from Case and Holcombe, 1980; Driscroll and Diebold, 1998; Mauffret and Leroy, 1999). Isobaths based on satellite-derived bathymetry gridded at 1 arc-minute intervals (Smith and Sandwell, 1997) using free software Generic Mapping Tools (GMT; Wessel and Smith, 1998). The inset shows the study area. Blue shading shows the approximate extension of the Beata ridge (Mauffret et al., 2001). Thick black arrows show the relative convergence direction between the North American and Caribbean plates. Thick red arrows show the 030° oriented compression yielded by the oblique convergence between the Caribbean plate's interior and the island arc (Mauffret and Leroy, 1999; Mercier de Lépinay et al., 1988; van den Berghe, 1983). Red lines mark active reverse and strike–slip boundaries along the eastern flank of the Beata ridge (Driscroll and Diebold, 1998; Mauffret and Leroy, 1999). The yellow line marks a suggested major active strike–slip fault along the Beata escarpment and eastern Bahoruco slope (Mauffret and Leroy, 1999). EPGFZ = Enriquillo–Plantain Garden Fault Zone. SOFZ = Septentrional–Oriente Fault Zone. PRVI BLOCK = Puerto Rico–Virgin Islands Block.

oblique convergence between the North American and the Caribbean plates that started during the Jurassic and the Muertos margin could have originated in Eocene times (e.g., Heubeck and Mann, 1991; Leroy et al., 2000; Pindell and Kennan, 2009). Model slip rates of the current convergence in the western Muertos margin based in GPS-derived velocities estimates 5.3 mm/y (Benford et al., 2012). Studies on the deformational features in the central Muertos margin (Granja Bruña et al., 2009), together with sandbox kinematic (ten Brink et al., 2009), gravity (Granja Bruña et al., 2010) and wide-angle seismic modeling (Llanes Estrada et al., 2012), suggest the absence of a true subduction of the Caribbean plate's interior beneath the island arc along the Muertos Trough. The Muertos thrust belt is more likely a retro wedge forming part of a bi-vergent crustal wedge system surrounding the eastern Greater Antilles island arc during unidirectional subduction (ten Brink et al., 2009).

The Beata ridge is a thickened aseismic block of pre-Cenozoic oceanic crust that forms a NNE–SSW trending bathymetric high located in the interior of the Caribbean Large Igneous Province (e.g., Edgar et al., 1971; Ewing et al., 1960; Fox et al., 1970; Mauffret and Leroy, 1997). Seismic data showed that the crust under the ridge is at least 20 km thick, while the crust in the Venezuelan basin is 10–15 km thick, and the adjoining north-eastern Colombian basin is 5 km thick (Diebold et al., 1999; Ewing et al., 1960; Leroy, 1995). Such variations are interpreted as the differential thickening of a normal oceanic crust during the emplacement of the Upper Cretaceous plume-related magmatism and the subsequent tectonic evolution (Driscroll and Diebold, 1998; Mauffret and Leroy, 1999). Geological and petrological studies based on samples and deep drilling suggest that the Beata ridge is composed mainly of intrusive rocks (e.g., gabbros and dolerites) resulting in an imbricate sill/dike complex (Fig. 1; Edgar et al., 1973; Fox and Heezen, 1975; Fox et al., 1970; Mauffret et al., 2001; Révillon et al., 2000). The thickened crust block is essentially submerged except for its northern boundary where it forms the basement of the Bahoruco and Southern peninsulas along southern Hispaniola (Maurasse et al., 1979). The Beata ridge has undergone at least three major phases of intraplate deformation:

- A poorly documented E–W compression phase occurred in the Middle Cretaceous coincident with the emplacement of the uppermost and most voluminous volcanic extrusive elements of the igneous province (Diebold, 2009).
- A phase of E–W extensional unloading yielded by underplating and causing the uplifting of the Beata ridge as a result of lithospheric flexure in the Upper Cretaceous (Driscroll and Diebold, 1998; Mauffret and Leroy, 1999; Mauffret et al., 2001).
- A recent stage of transpressive tectonics (Miocene-Present) related to the convergence between the Caribbean plate's interior and Hispaniola (Mauffret and Leroy, 1999; Mercier de Lépinay et al., 1988).

The Beata ridge separates the abyssal seafloors of the Venezuelan and Colombian basins, except in the Aruba gap located at the southern end of the ridge (Fig. 1). We use the sub-basin nomenclature of Mauffret and Leroy (1997) to refer to smaller regions of the extensive Venezuelan and Colombian basins close to neighboring countries (i.e., Dominican sub-basin, Haiti sub-basin, Puerto Rican sub-basin). The underlying oceanic crust of the Dominican sub-basin is unusually thick (≤15 km) and consists of an upper volcanic sequence dominated by sub-horizontal reflections traceable for tens of kilometers and a lower volcanic sequence more homogeneous and generally devoid of reflecting horizons (Diebold, 2009; Edgar et al., 1971). This sub-basin section contains sedimentary seismic horizons from the Upper Cretaceous to recent times (Fig. 2) whose age and petrology were established from the correlation with the DSDP 146/149 sites (e.g., Diebold et al., 1981; Edgar et al., 1971, 1973; Kroehler et al., 2011; Matthews and Holcombe, 1985). Knowledge of the Haiti sub-basin is somewhat more limited than for the Dominican sub-basin since seismic surveys and DSDP drilling has been less extensive (Case et al., 1990). This basin is underlain by a relatively thin 5 km-thick oceanic crust and filled by a thick sedimentary cover mainly consisting of horizontal turbidite layers (Ewing et al., 1960; Mauffret and Leroy, 1997).

As a result of the oblique component of convergence between the island arc and the Caribbean plate's interior, the northern Beata ridge impinges on south-central Hispaniola (Heubeck and Mann, 1991; Jany, 1989; Ladd et al., 1981; Mercier de Lépinay et al., 1988; Pubellier et al., 2000; Vila et al., 1990) (see thick red arrows in Fig. 1). This impingement has altered the pre-existing E–W trending, intra-oceanic island arc structure of southern Hispaniola. The main consequences of the impingement are the termination of the Muertos thrust belt and the uplift and accretion of a part of the oceanic igneous crust of the Beata ridge to form the Bahoruco and Southern peninsulas in southern Hispaniola (Maurasse et al., 1979; Sen et al., 1988; van den Berghe, 1983).

3. Data

The study region was mapped during the CARIBENORTE cruise aboard the Spanish R/V Hespérides using the hull-mounted multibeam Kongsberg Simrad EM 120 system (Carbó-Gorosabel et al., 2010). The survey covered an area of ≈40,000 km^2, between water depths of −500 m and −5550 m (Fig. 3 and Appendix A1). We completed the study area using data acquired during previous cruises (GEOPRICO-DO cruise, Carbó-Gorosabel et al., 2005; TSUCAR cruise, Andrews et al., 2010). Locally the overlap between different cruises yielded multibeam data artifacts (see ma in Fig. 3 and Appendix A2).

Approximately 215 km of single-channel seismic reflection profiles were also collected in the western end of the Muertos margin and the eastern Bahoruco slope during the CARIBENORTE cruise (Carbó-Gorosabel et al., 2010) (see green lines in Fig. 3). In addition to new seismic data, we have re-processed and re-interpreted single- and multi-channel seismic reflection profiles provided by different institutions and data bases (Fig. 3): The Academic Seismic Portal managed

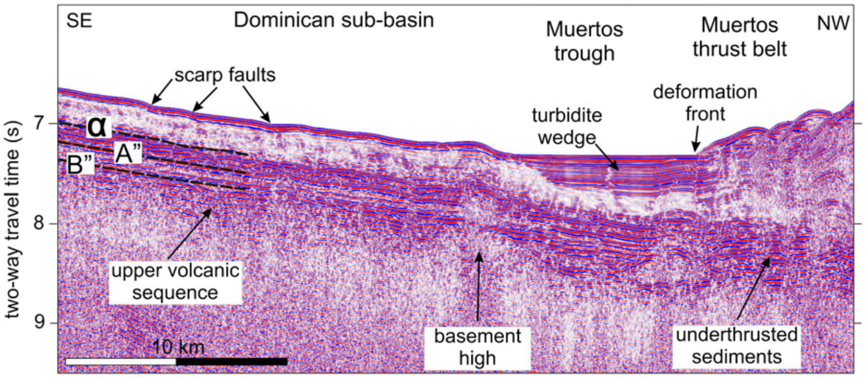


Fig. 2. Representative multi-channel seismic reflection profile (MCS) showing the seismic units of the Dominican sub-basin. Vertical exaggeration (V.E.) is times 6 on the seafloor. See location in Fig. 3.

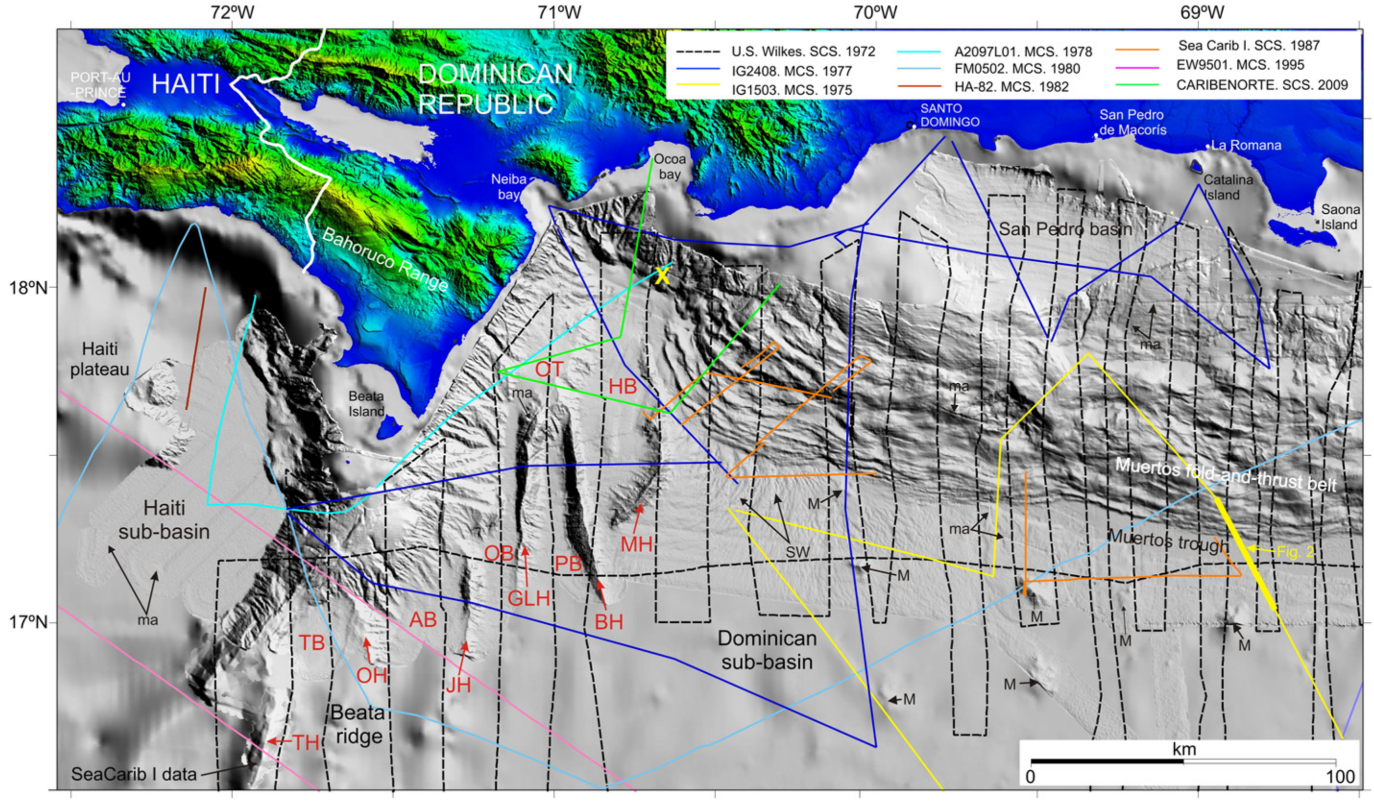


Fig. 3. Digital elevation model (DEM) illuminated from the NE showing the location of the seismic-reflection profile data set. The offshore model is derived from 150 m-gridded swath bathymetry data (rough aspect regions) and from GEBCO data gridded at 1 arc-minute intervals (smooth aspect regions). The onshore digital elevation model comes from SRTM90 data collected by the National Aeronautics and Space Administration (NASA) illuminated from the NE. Old seabeam bathymetry data from the Sea Carib I cruise (1987) at the Tairona hill (TH) was provided by Institut Français de Recherche pour l'Exploitation de la Mer (IFREMER). Different names for the offshore tectonic and physiographic features have been established in agreement with the Autoridad Nacional de Asuntos Marítimos (ANAMAR) of the Dominican Republic. The names previously used by Mauffret et al. (2001) now change from "Dyoris hill" to "Goleta Leonor hill", "North Beata hill" to "Michelen hill" and "Tairona ridge" is now the "Tairona hill". ma = multibeam artifacts related to the overlapping among different swath data cruises. X = Location of a geomagnetic anomaly yielded by an underthrusted hill of the Beata ridge (Mauffret and Leroy, 1999). M = Igneous mound. SCS = Single-Channel Seismics. MCS = Multi-Channel Seismics. Seismic cruises are color-coded together with acquisition year. SW = Sediment waves. TH = Tairona hill. TB = Taína basin. OH = Ojeda hill. AB = Anamar basin. JH = Juancho hill. OB = Orion basin. OT = Orion terrace. GLH = Goleta Leonor hill. PB = Paraiso basin. BH = Barahona hill. HB = Hespérides basin. MH = Michelen hill.

by the University of Texas-Institute for Geophysics (IG2408, IG1503 and FM0502 cruises; Shipley et al., 2005), the National Archive of Marine Seismic Surveys managed by the U. S. Geological Survey (HA-82 cruise; courtesy of WesternGeco), the Marine Geoscience Data Management System managed by Lamont-Doherty Earth Observatory (EW9501 cruise), the Woods Hole Oceanographic Institution (A2097L01 cruise), the National Geophysical Data Center managed by the National Oceanic and Atmospheric Administration (US Wilkes 1972 cruise) and the Institut Français de Recherche pour l'Exploitation de la Mer (Sea Carib I cruise). Several multi-channel stacked profiles from IG1503 and IG2408 cruises have been partially interpreted by Ladd et al. (1977, 1981). A detailed re-processing by means of the post-stack Memory Stolt f–k migration method has improved the old seismic profiles (Appendix A3; Stolt, 1978).

4. Description and interpretation of the main morphotectonic provinces

Based on along- and across-strike morphological, structural and sedimentological criteria derived from the combined interpretation of the bathymetry and seismic data, we can distinguish between four morphotectonic provinces (see inset map in Fig. 4). These are in the Caribbean plate's interior the Dominican sub-basin (DSB), the Beata ridge (BR) and the Haiti sub-basin (HSB) provinces, and in the southern island arc slope the Muertos margin (MM) province.

4.1. The Dominican sub-basin province

The Dominican sub-basin is located in the north-western corner of the Venezuelan basin and is bounded by the base of the Muertos margin (i.e., deformation front) and by the easternmost hills of the Beata ridge (see DSB in the inset map in Fig. 4). This sub-basin contains the incoming sedimentary section that is being actively accreted along the deformation front of the Muertos thrust belt (Fig. 2). The sedimentary section consists of sediments from the Coniacian (ages determined by correlation with the DSDP 146/149 sites; e.g., Kroehler et al., 2011) to recent times and is underlain by the upper volcanic sequence of the Caribbean Large Igneous Province (Figs. 5 and 6; Mauffret and Leroy, 1997). The basement surface is marked by a highly-reflective and smooth seismic horizon (i.e., B″ horizon; Fig. 2). The B″ seismic horizon was correlated with sills of Upper Cretaceous tholeitic basalts intruding pelagic carbonate sedimentary rocks of the Coniacian. Two seismic horizons are identifiable within the sedimentary section of the basin: the A″ horizon (Middle Eocene siliceous limestone and chert) and the α horizon (Lower Miocene). The shallower structure and the seafloor morphology of the Dominican sub-basin are mainly characterized by features yielded by recent sedimentary and tectonic processes, but locally there are also inherited features such as basement highs and mounds on the seafloor (e.g., Donnelly, 1973; Driscoll and Diebold, 1999; Jany, 1989; Masson and Scanlon, 1991; Matthews and Holcombe, 1985; Mauffret et al., 2001).

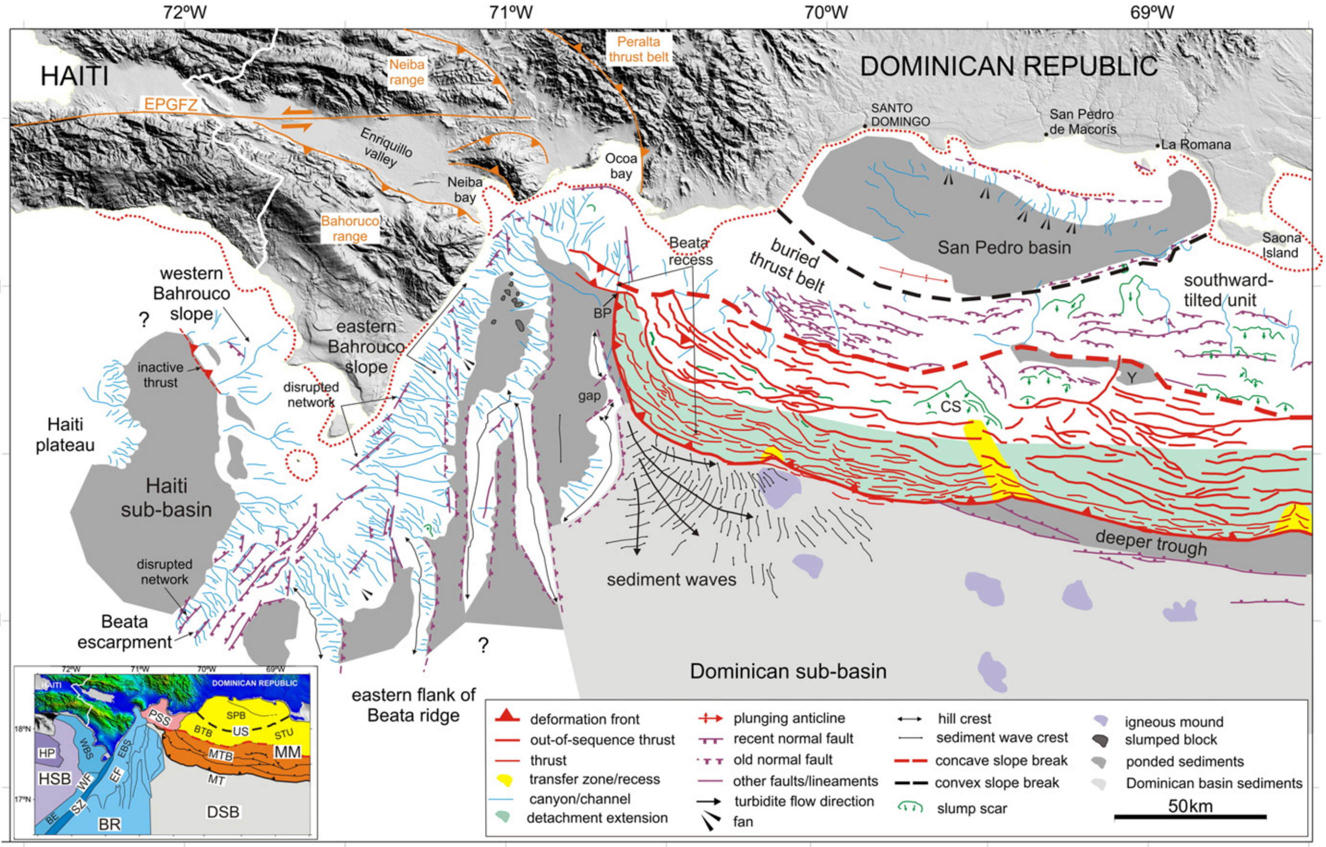


Fig. 4. Regional morphotectonic interpretation in map-view. Green area shows the reconstruction of the detachment surface beneath the Muertos margin. Orange lines mark major onshore structures. EPGFZ = Enriquillo–Plantain Garden Fault Zone. CS = Complutense slump. Y = big perched basin. BP = Branch point. The inset map shows the division into four morphotectonic provinces (DSB = Dominican sub-basin; BR = Beata ridge; HSB = Haiti sub-basin; MM = Muertos margin) and different zones within the provinces (HP = Haiti plateau; WF = Western flank of Beata ridge; BE = Beata escarpment; WBS = Western Bahoruco slope; SZ = Beata summit zone; EF = Eastern flank of Beata ridge; EBS = Eastern Bahoruco slope; PSS = Punta Salinas slope; MT = Muertos trough; MTB = Muertos thrust belt; BTB = Buried thrust belt; US = Upper slope; STU = Southward-tilted unit; SPB = San Pedro basin). Thick red dashed line marks the concave slope break. Thick black dashed line marks the convex slope break.

4.1.1. Basement highs and seafloor mounds

The basement highs in the Dominican sub-basin are characterized by vertical zones, devoid of coherent reflections, rising from the acoustic basement and disturbing the smooth and continuous horizons of the sedimentary cover (Fig. 5). The upper volcanic sequence frequently thickens towards the basement highs, which led Diebold et al. (1999) to suggest that they could be feeder conduits for the main upper volcanic sequence during the late phase of volcanism in the Upper Cretaceous. The upper sedimentary sequences are folded and are thinner than in the surroundings of the basement highs and show different thicknesses at both sides (Fig. 5). These observations agree with the hypothesis that the basement highs predate the large sediment accumulation and act as bathymetric barriers interfering with sediment dispersal in the Dominican sub-basin. Some of the basement highs penetrate the seafloor yielding mounds of variable size (see M in Fig. 3 and Appendix A4). These mounds rise 200–500 m above the surrounding seafloor and are blanketed by sediments (Fig. 5 and Appendix A1).

4.1.2. Sediment waves

Approximately 4000 km^2 of the seafloor in the north-western corner of the Dominican sub-basin is characterized by deep-water sediment waves (Figs. 3 and 4). The size of the sediment waves increases up-section as well as towards the NW, with wavelengths reaching 2–3 km and amplitudes of up to 100 m (Fig. 6). The sediment waves progressively fade as one moves away from the northwestern corner. At depth, these waves onlap the α horizon (Lower Miocene) towards the south and the east. The internal structure of the sediment waves consists of sub-parallel and closely-spaced reflectors defining series of anticlines and synclines affecting only to the upper sedimentary sequences (Fig. 6). The entire post-α horizon wedge of sediments thickens progressively north-westwards indicating a Miocene and younger sedimentary source from the northwest. This directional thickening contrasts with the geometry of the pre-α horizon sediments.

The existence of sediment waves suggests deposition and remobilization under high-energy bottom current conditions (Wynn and Stow, 2002). The shape of the sandwaves is asymmetric in cross-section indicating transport direction towards the steeper flank of the wave (see inset in Fig. 6). In map view, the sediment waves have a crescent shape and their orientation rotates with the change in the orientation of the deformation front (Fig. 4). Both the map view and the cross-sectional view of the sand waves suggest that they were formed by bottom currents sweeping down from the Neiba bay to the south and east towards the deepest part of the Muertos trough.

4.1.3. The Muertos trough

The Muertos trough is an elongated bathymetric depression formed along the base of the steep insular slope of the Muertos margin (Fig. 3 and Appendix A1). The bathymetric expression and the depth of the trough have significant along-strike variations controlled by active sedimentary and tectonic process (Figs. 4 and 7a–h). In the east of the study region, there is a deeper sector filled by a turbidite wedge resulting from the recent activity of dip–slip normal faults trending sub-parallel to the deformation front (see deeper trough in Fig. 4). These faults are identifiable up to 30 km southward of the deformation front and were interpreted as a combination of the deformational response to tensile

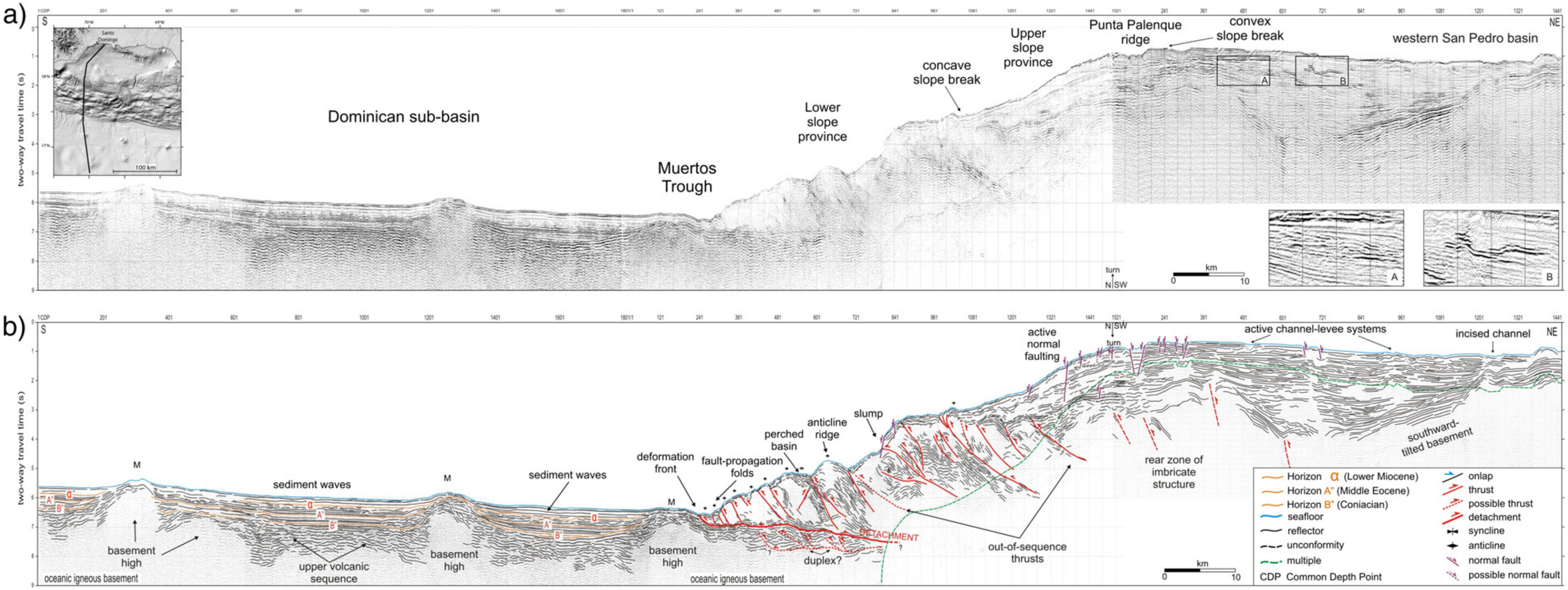


Fig. 5. a) Post-stack migrated multi-channel seismic profile (MCS). V.E. is 6.6× on the seafloor. CDP = Common Depth Point. The insets show enlarged sectors: A = low-amplitude and continuous reflectors. B = high-amplitude and discontinuous reflectors. b) Line drawing interpretation. M = seafloor mound.

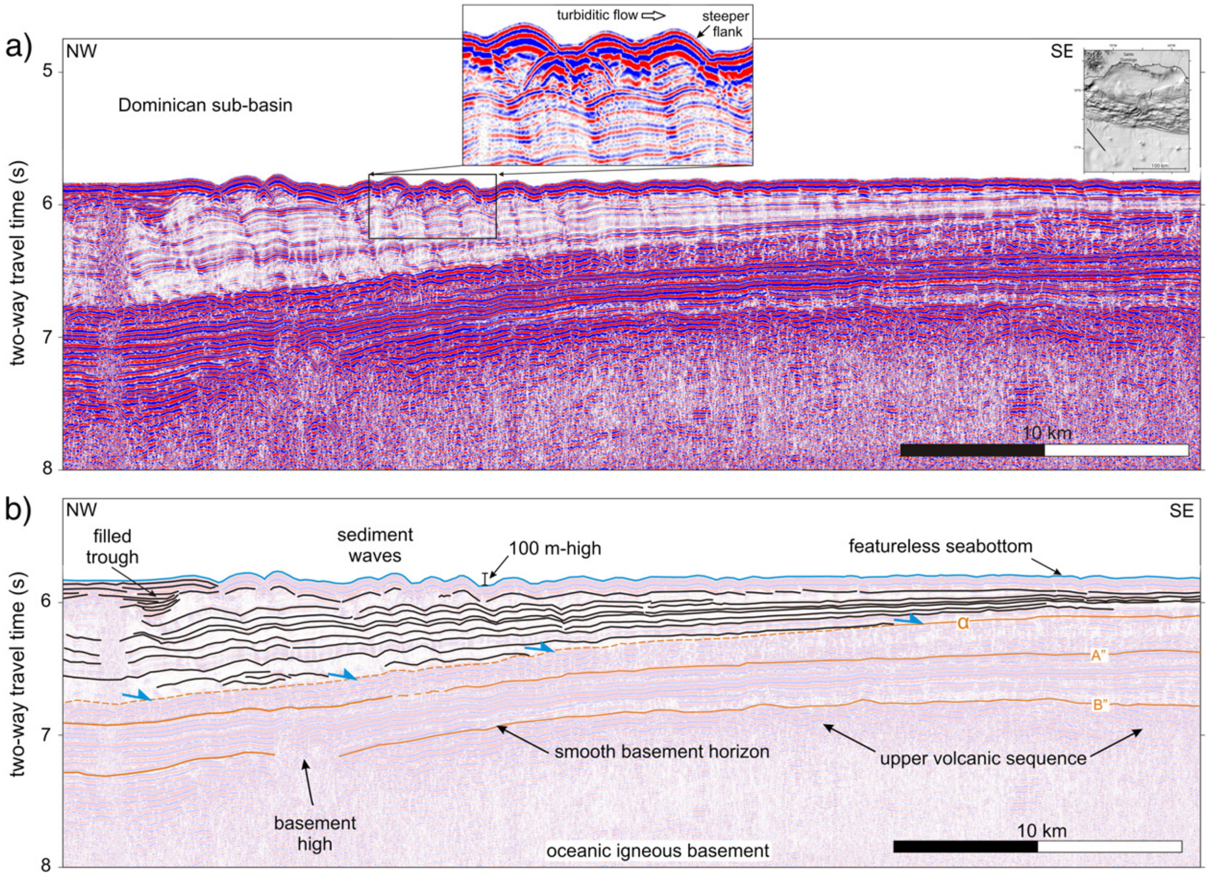


Fig. 6. a) Stacked MCS profile. V. E. is 6× on the seafloor. The inset shows an enlarged sector to image the asymmetry in the anticlines suggesting an active eastward turbiditic flow. b) Line drawing interpretation. See Fig. 5b for legend.

flexural stresses resulting from the bending of the Caribbean plate's interior and to the differential along-strike overburden of the Muertos thrust belt (Granja Bruña et al., 2009). Westward from this deeper sector up to near the Michelen hill, the Muertos trough becomes progressively shallower and narrower losing its distinct morphological expression (Fig. 3). Here the trough is filled by sediment waves lying obliquely to the deformation front (Figs. 5 and 7d). Westward, between the deformation front and the Michelen hill, there is a NNW–SSE trending 5 km-wide valley filled by a small wedge of ponded turbidites (Figs. 7e and 8). Farther north-westward the Michelen hill disappears and the Muertos trough becomes significantly wider and filled by a thick sequence of sub-horizontal turbidites (Figs. 7f–g and 9). The westward extension of the Muertos trough along the Neiba bay and onshore Enriquillo valley (Fig. 4) has been suggested by different authors (e.g., Biju-Duval et al., 1982; Dolan et al., 1991; Jany, 1989; Mann et al., 1991; Mercier de Lépinay et al., 1988), but that geometry is not well observed because of the large current offset between the onshore and offshore structures.

4.2. The Muertos margin province

The Muertos margin shows a good structural analogy with other well-studied offshore active thrust belts and accretionary prisms around the world (e.g.; North Panama Deformed Belt, Silver et al., 1995; Barbados, Brown and Westbrook, 1987; Makran, Smith et al., 2012). The western Muertos margin (MM) is divided into distinct zones along and across the slope (see inset map of Fig. 4): the active Muertos thrust belt (MTB), the upper slope (US) and the Punta Salinas slope (PSS). The Muertos thrust belt (MTB) and the upper slope (US) zones are separated by an along-strike concave slope break and form an insular slope ≈120 km-wide (see red dashed line in Fig. 4). The upper slope (US) shows three distinct zones (Fig. 4): the San Pedro basin (SPB), the buried thrust belt (BTB) and the southward-tilted unit (STU). The western end of the Muertos margin is occupied by the Punta Salinas slope (PSS) showing a steeper and narrower insular slope only ≈30 km-wide.

4.2.1. The Muertos thrust belt

The active Muertos thrust belt (or Muertos fold-and-thrust belt) lies along the lower insular slope between the deformation front and the concave slope break (Fig. 7a–e). This thrust belt shows a stepped slope resulting from long-lived active folding and thrusting in a thin-skin tectonic style over seismic horizons that extend northward from the Dominican sub-basin (Fig. 5 and Appendix A4; Granja Bruña et al., 2009). The northward steeply-dipping reflectors suggest highly deformed materials forming a fold-and-thrust belt with a prevailing south-verging imbricate structure over a detachment located within the Dominican sub-basin sedimentary section. This detachment extends beneath the deformed wedge to distances up to 25 km northwards from the deformation front in the eastern study region and decreases progressively westward until it disappears in Punta Salinas slope (see green area in Fig. 4). Using a sound velocity average of 2.5 km/s for the sedimentary materials of the thrust belt derived from wide-angle modeling (Llanes Estrada et al., 2012), we estimate the dip of the detachment to be 8°–10° between longitude 68.5°W and the Beata recess (Figs. 5 and 7a–d). In the Beata recess the detachment steepens up to 16° and the northward extension is reduced to <15 km (Figs. 7e–g, 8). Here the thrust belt lies over discontinuous northward-dipping reflections that may indicate the top of the oceanic basement rocks forming part of the easternmost Beata ridge.

The surface expression of the Muertos thrust belt is a highly-stepped slope characterized by an alternation of asymmetric elongated troughs

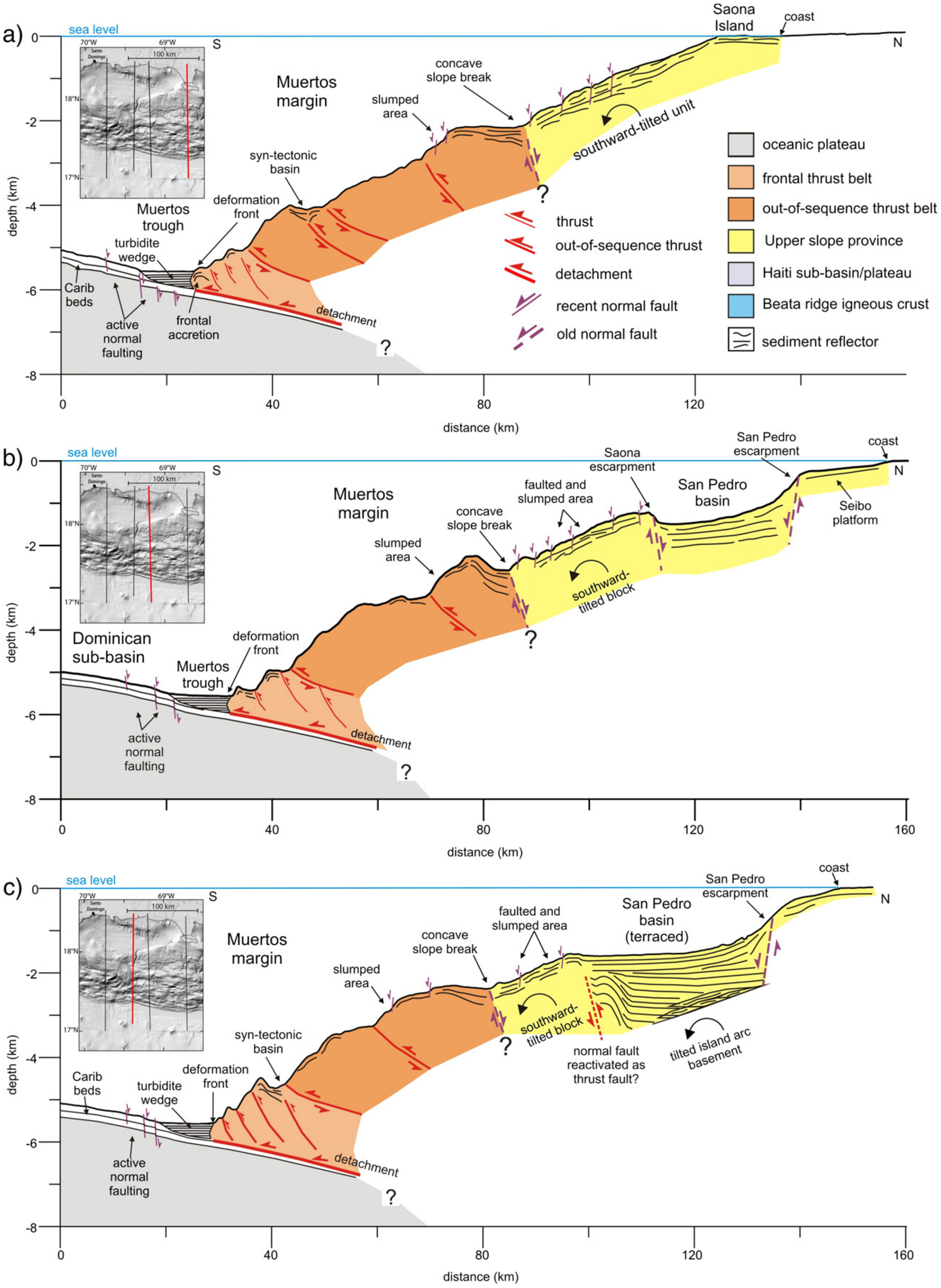


Fig. 7. Schematic cross-sections a) through j) deduced from this study. Bathymetric profiles were extracted from multibeam data gridded at 150 m intervals and completed at the ends with GEBCO data gridded at 1 arc-minute intervals. All cross-sections use the same scale to compare the along- and across-strike structure.

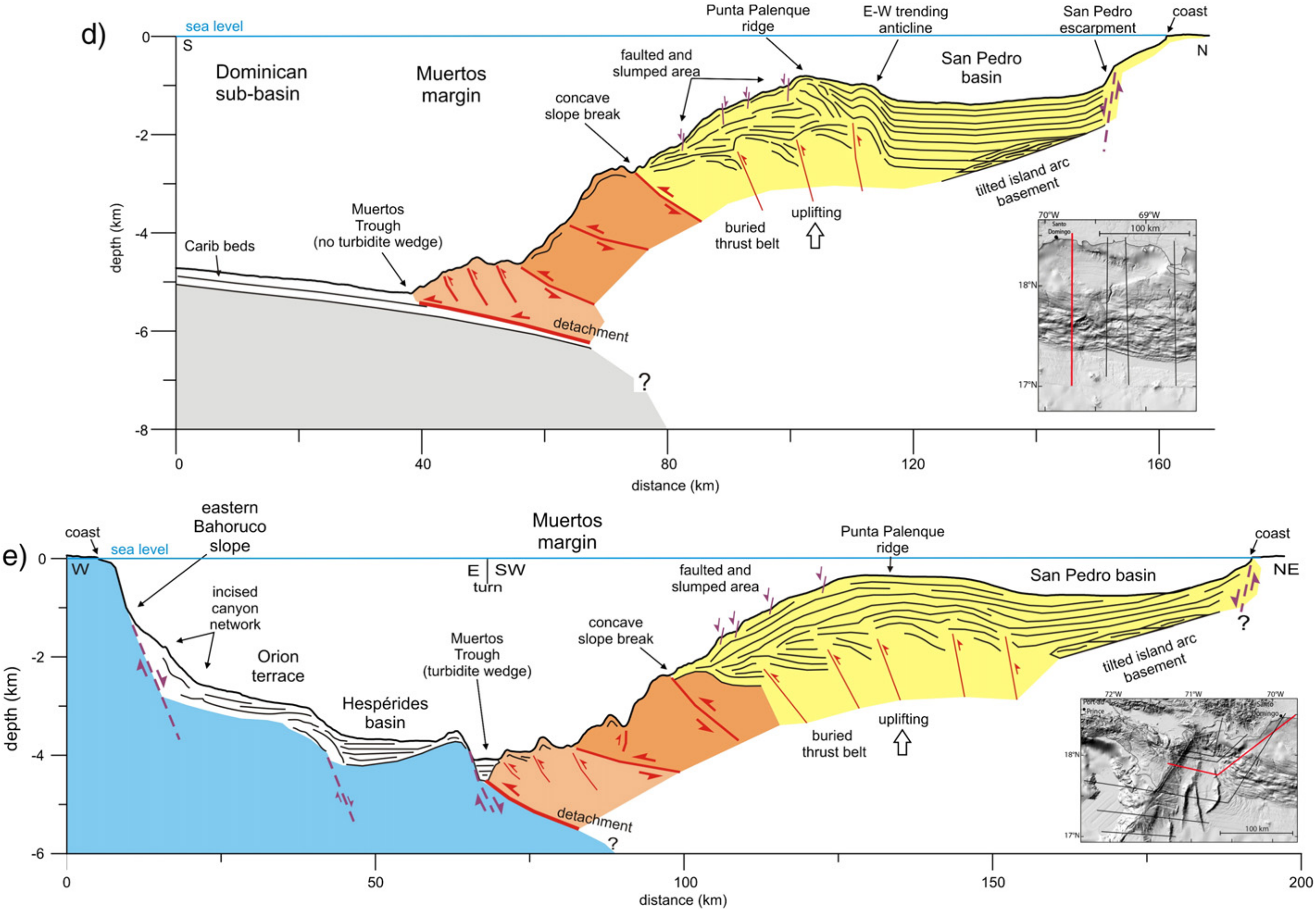


Fig. 7 (*continued*).

and anticline ridges oriented sub-parallel to the deformation front (Fig. 3), which are underlain by northward-dipping thrust faults (i.e., fault-propagation folds; Fig. 8). The individual anticline ridges (i.e., thrust slices) are characterized by curved or sinuous geometry in map view with low lateral continuity and can be rarely traced for more than 30 km along strike (Fig. 4). Anticline ridges show along strike plunging and are replaced by other anticline ridges yielding transfer or accommodation zones. In transfer zones there is interference from lateral- and oblique-thrust ramps (Calassou et al., 1993), resulting in an anostomosing architecture inherent to compressive thin-skinned imbricate belts (McClay, 1992). The size and spacing of the anticline ridges increase from the deformation front to the upslope, resulting in wider bathymetric troughs and perched basins upslope. These elongated basins show tilted sedimentary sequences formed because of the activity of adjacent thrust slices (Fig. 8). Upslope perched basins are connected with local canyon networks trapping the channelized turbiditic supplies (see Y in Fig. 4).

The sinuous morphology in map view of the deformation front is a result of tectonic erosion caused by the thrusting of basement highs and seamounts (c.f., Granja Bruña et al., 2009; Ranero and von Huene, 2000; Scholl et al., 1980; von Huene and Lallemand, 1990). The effects of the tectonic erosion propagate landward disrupting the east–west trending fabric of the imbricate thrust belt (see yellow areas in Fig. 4). Generally the thrusting of basement highs yields a local over-steepening of the taper slope and occasionally triggers slope failures (see the CS in Fig. 4 and Complutense slump in Appendices A1 and A2).

The western end of the Muertos margin shows an oroclinal bending of the thrust slices and a westward narrowing of the imbricate thrust belt to form the Beata recess (Fig. 4; Marshak, 2004). In the recess the imbricate structure undergoes a significant reorganization along strike forming wider anticline ridges and troughs (Figs. 7e, h and 8). This fact may be related to the significant steepening of the detachment and possibly changes in the shortening rate (Calassou et al., 1993).

Upslope within the active thrust belt there are several steep along-strike slope breaks forming bathymetric escarpments (Fig. 7a, b and see OOS in Appendix A2). These slope breaks are the surface expression of out-of-sequence, southward-verging, low-angle dipping faults that are thrusting over downslope imbricate systems (Figs. 5 and 7a–e). The well preserved morphology of the escarpments along strike and the frequent slope failures due to the ongoing over-steepening suggest that these structures are active. Some of these active out-of-sequence thrust faults can be traced for more than 100 km. This fact implies that if they rupture along their entire length, they could generate earthquakes of large magnitude (c.f., Bakun et al., 2012; ten Brink et al., 2011). The greater development of the imbricate system in the western Muertos margin may be associated to a higher shortening rate between the Caribbean plate's interior and the inland arc in the southern Hispaniola (c.f., McClay, 1992). This observation is in agreement with the higher convergence proposed from slip rate modeling for the western Muertos margin (Benford et al., 2012; Manaker et al., 2008). The out-of-sequence thrusts experience less oroclinal bending than the frontal thrust belt, and are therefore oriented more obliquely to the deformation front. This fact results in the progressive disappearance of the

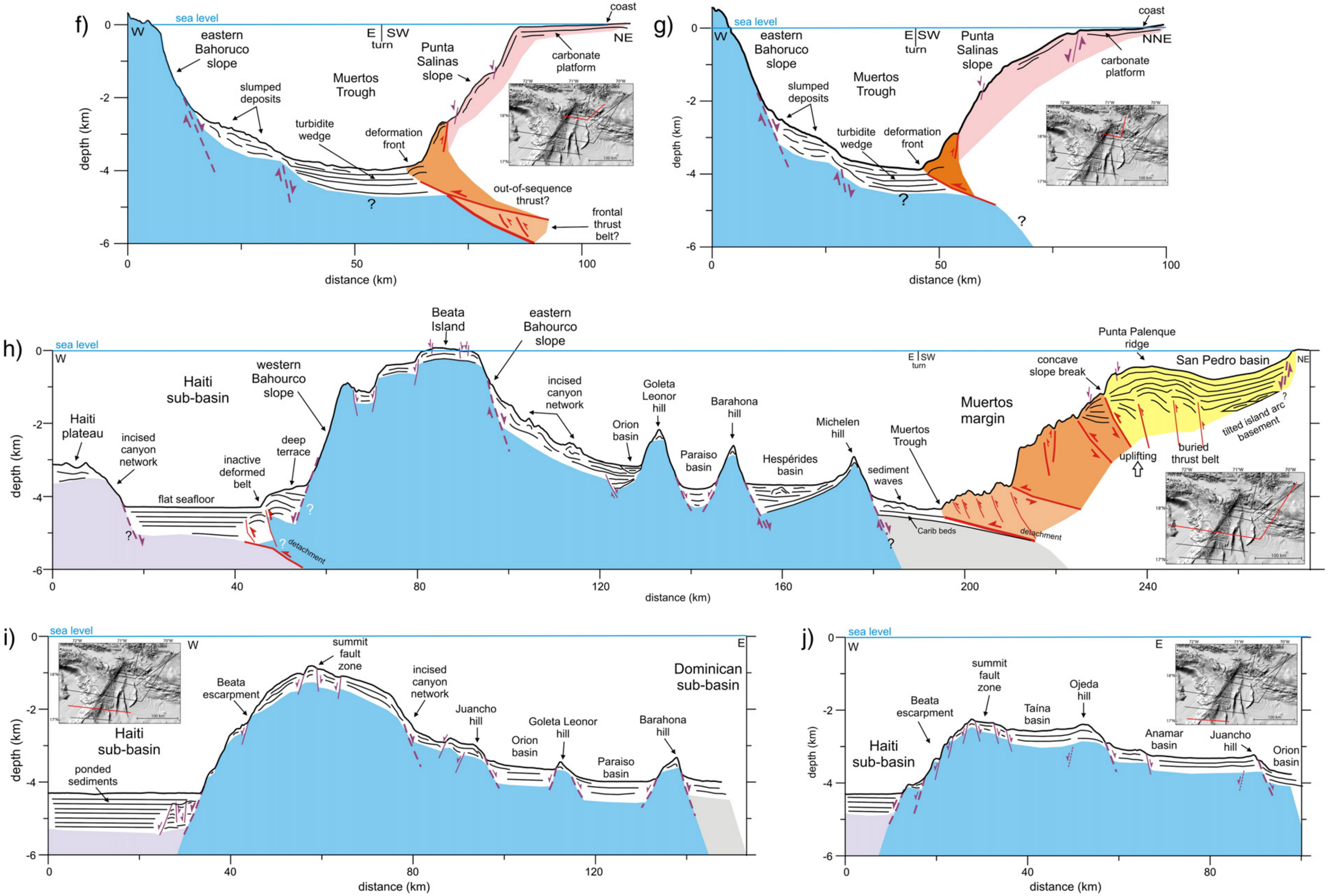


Fig. 7 (*continued*).

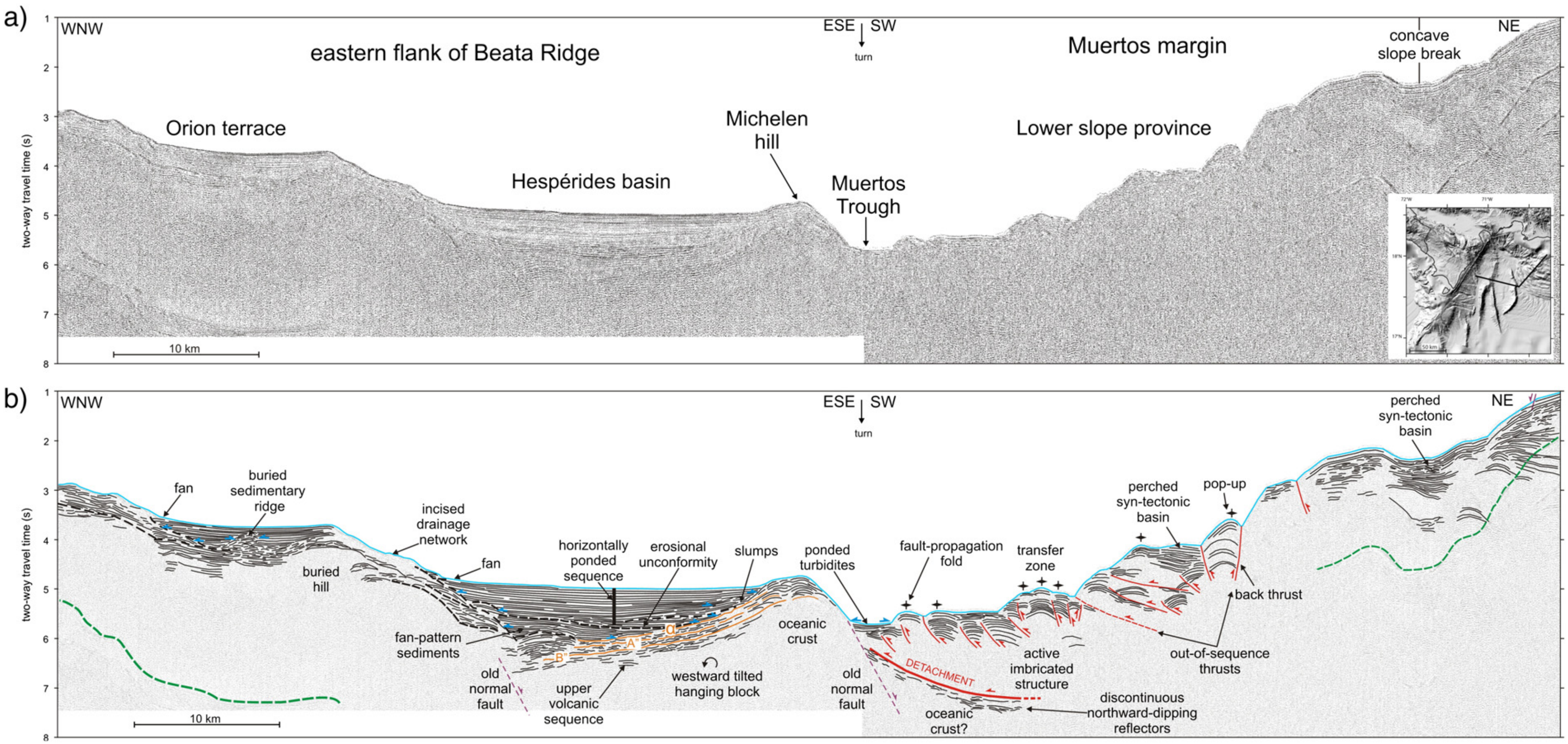


Fig. 8. a) Single-channel seismic profile (SCS). V. E. is 3.5× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend.

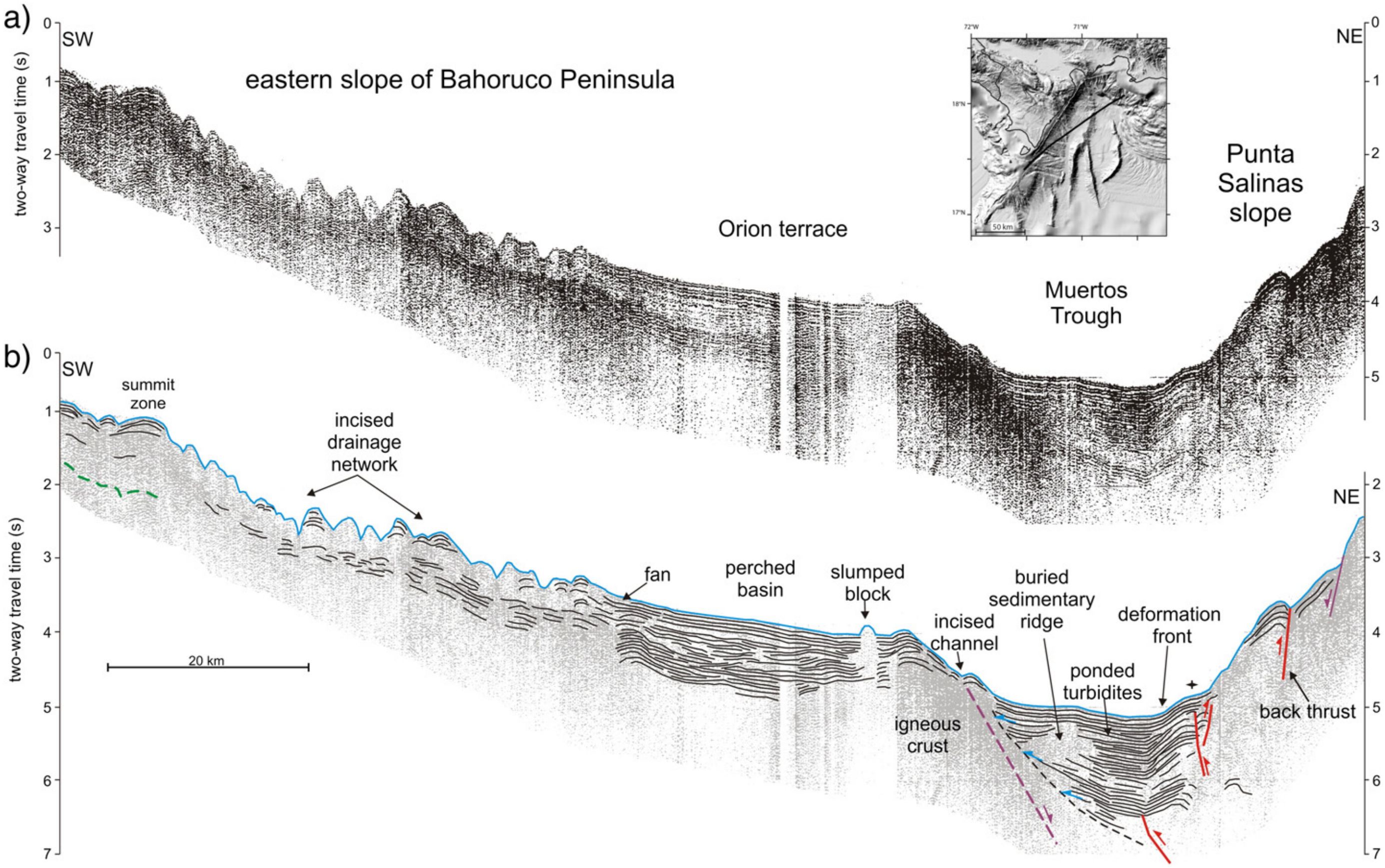


Fig. 9. a) SCS profile. V. E. is 9.5× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend.

downslope imbricate systems beneath the out-of-sequence thrust belts and the formation a branch point (see BP in Fig. 4).

4.2.2. The upper slope-San Pedro basin

The upper slope of the Muertos margin includes the offshore region lying between the concave slope break and the southern coast of the Dominican Republic (see US in the inset map of Fig. 4). This province has extended and thick slope deposits, burying the rear zone of the Muertos thrust belt, and southward-tilted island arc blocks (Fig. 7a–e). The San Pedro basin is the main tectonic and sedimentological feature observable in the Upper slope province (Fig. 10). This basin was classically interpreted as a forearc basin formed in a subduction setting along the Muertos Trough (e.g., Biju-Duval et al., 1982; Ladd et al., 1981), but it is located in the retroarc region and formed in a backthrust setting (Granja Bruña et al., 2010; ten Brink et al., 2009).

The San Pedro basin is an E–W trending bathymetric depression located at the top of the insular slope with maximum depths of ≈1600 m (Fig. 10a and Appendix A1). The depression is mainly bordered by structural highs and collects most of the southward drainage of eastern Dominican Republic, from Saona Island to Punta Palenque, resulting in a significant accumulation of sediments (a thickness of at least 3.4 s of two-way travel time (TWT); Fig. 11). The seafloor of the basin is smooth except near the boundaries where there is an incised drainage network forming channel–levee systems and fans (Figs. 10b and 12). These depositional systems yield seismic facies consisting of planar reflectors of low amplitude and high continuity (see inset A in Fig. 5a), alternating with high-amplitude reflectors that are discontinuous and of irregular morphology (see inset B in Fig. 5a) (c.f., Seely, 1979).

The San Pedro basin is flanked in the north by the WNW–ESE trending San Pedro escarpment formed between the edge of the insular shelf and the seafloor of the San Pedro basin and reaching a maximum height of ≈1200 m (Fig. 10c). The San Pedro escarpment is the exposed scarp of a main old southward-dipping normal fault (Fig. 12). Satellite-derived altimetry and seismic data allowed us to infer that the escarpment becomes progressively smaller along strike (Figs. 7a–e and 10c). The San Pedro basin is bounded in the south by an outer high that forms a convex slope break separating the ponded depression from the insular slope (see the inset map in Fig. 4). This outer high acts as a topographic barrier but shows a variable along-strike structure that controls the sediment infilling of the San Pedro basin (Fig. 7b–e). In the west the outer structural high is formed by the NW–SE-trending Punta Palenque ridge and in the east by the NE–SW-trending Saona escarpment. Between Punta Palenque ridge and the Saona escarpment there is a 17 km-long gap where the outer high is lower and is overcome by the basin sediments and forms a terrace (Figs. 7c and 10b). In this terraced area there is a highly-incised and sinuous channel connecting the San Pedro basin with the perched basins developed in the Muertos thrust belt (see ROA-IEO Channel in Fig. 10a). This terraced area separates the Upper slope province into western and eastern regions, with significant differences in the structural configuration and sediment infill in the San Pedro basin.

4.2.2.1. The western region. Basin stratigraphy preserves the record of its deformation over time. The shallow sediments are layered horizontally and onlap landward on deeper slightly tilted sediment wedges. These wedges overlie a tilted sedimentary layer, presumably covering island arc basement rocks (Fig. 11). The southward tilting of the planar surface therefore occurred prior to the major sediment accumulation. The sediment wedges may have been deposited in a pro-delta environment on a tilted surface, whereas the sub-horizontal layers were deposited in deeper water farther from the source, perhaps as turbidite sequences. From onshore deep wells in the southern coast of the Dominican Republic, this tilted surface was correlated with Upper Cretaceous island arc rocks (metavolcanics and tonalities; Rodgers, 1991; White, 1993).

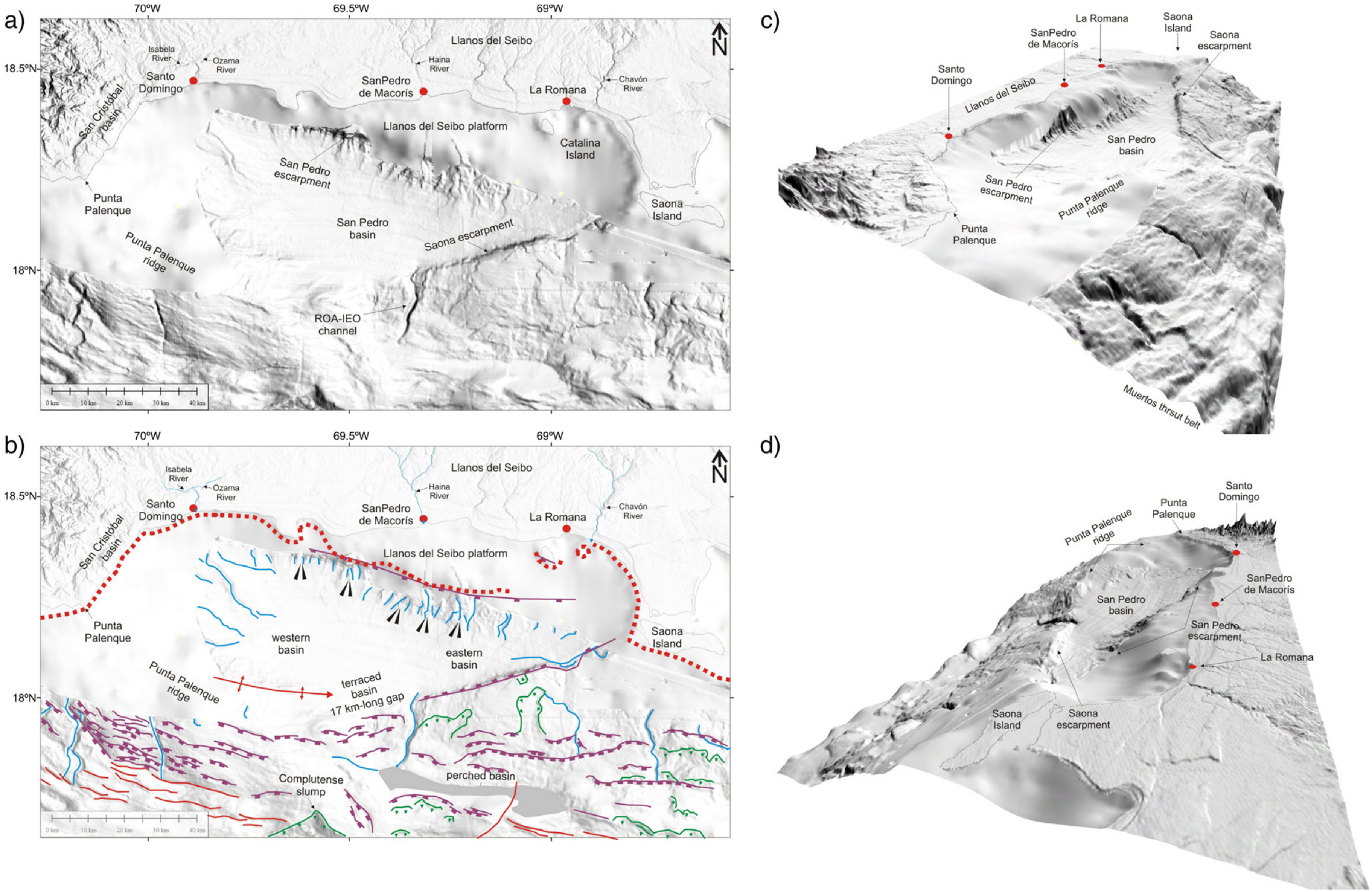


Fig. 10. Enlarged sector of the Upper slope province. a) DEM illuminated from NE as in Fig. 3. b) Morphotectonic interpretation in map-view. See legend in Fig. 4. c) Gray-shaded 3-D bathymetric image view from the SW. V.E. is 6×. d) Gray-shaded 3-D bathymetric image view from the NE. V.E. is 6×.

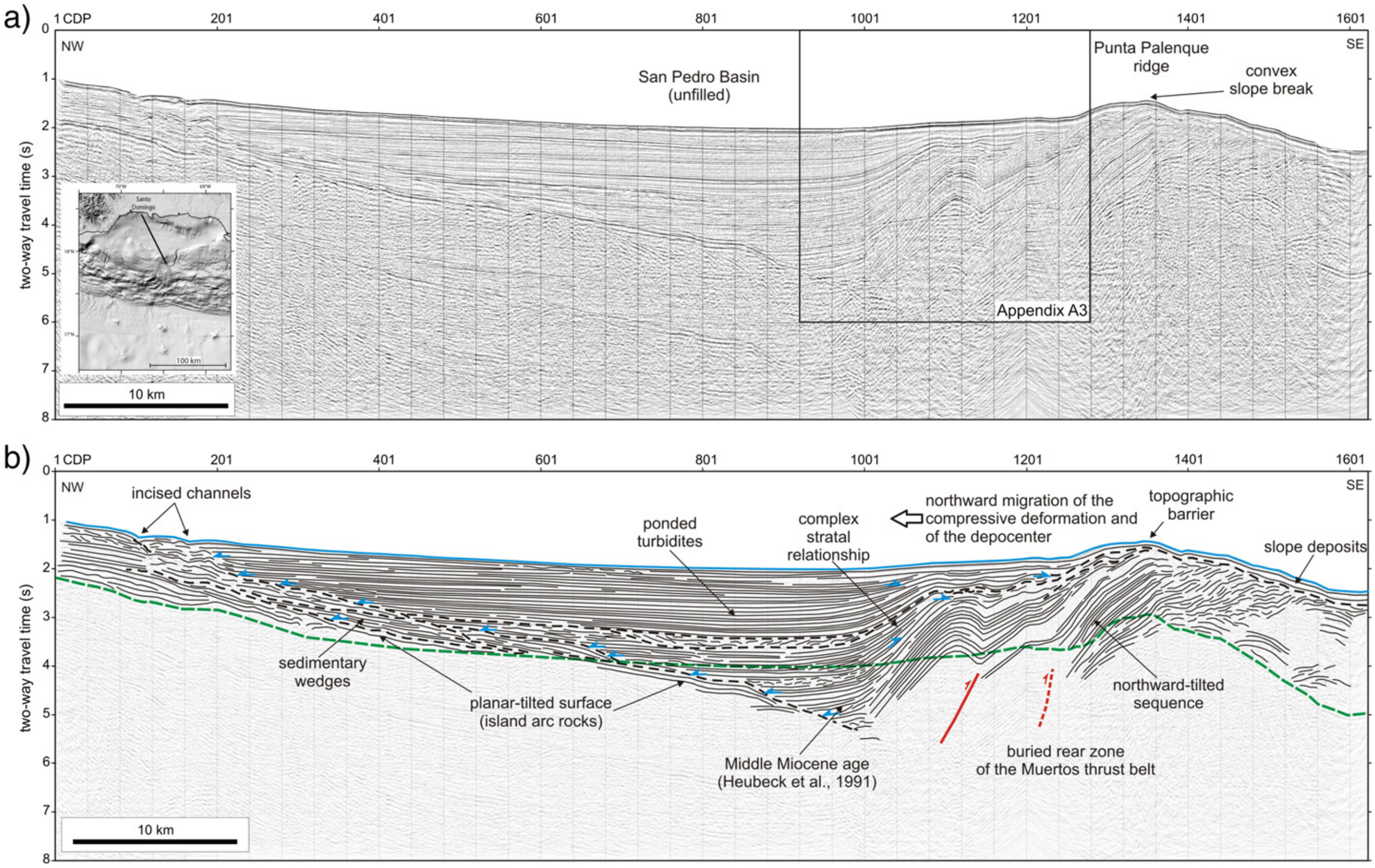


Fig. 11. a) Migrated MCS profile. V. E. is 4× on the seafloor. Inset shows the location of Appendix A3. b) Line drawing interpretation. See Fig. 5b for legend.

Several authors have tentatively correlated the seismic stratigraphy of the San Pedro basin with the onshore geology observed nearby on the Dominican Republic, though the results show very different ages (e.g., Biju-Duval et al., 1982; Heubeck et al., 1991; Ladd et al., 1981). Heubeck et al. (1991) reviewed the previous works, and correlated the Igenio Caei Formation near the onshore San Cristobal basin (Fig. 10a) with the lower sedimentary sequences of San Pedro basin, suggesting that the older sediments of the basin are of Middle Miocene age.

Seaward the basin is underlain by the rear zone of the Muertos thrust belt (Figs. 5 and 11). The rear zone of the thrust belt is formed by an imbricate structure composed of southward-verging blind thrusts. Some of these thrusts may still be active and form fault-propagation folds that deform the seafloor (see the E–W-trending anticline in Fig. 7d). The sub-horizontal sedimentary section of the basin is progressively tilted northward and folded, documenting the growth of the thrust belt and the northward migration of the compressive deformation. The landward and vertical growth of the thrust belt has resulted in the uplifting of the Punta Palenque ridge and the northward migration of the depocenter (Fig. 11). The sediment beds of the basin show a complex stratal relationship with the northward-tilted sedimentary sequences underlying the Punta Palenque ridge. This relationship results from the variable balance between sediment supply and growth of the thrust belt driven by the convergence rate (Fig. 11). The seaward slope of the Punta Palenque ridge shows a seafloor characterized by frequent sub-parallel scarps of tens of meters of height (Fig. 10a,b). These scarps are controlled by active bending-moment normal faulting arranged in echelon. These faults are the deformational response to the ongoing growth of the thrust belt and the consequent over-steepening of the slope. This over-steepening causes a local shallow extensional regime that contributes to the thrust belt can balance its critical taper angle (e.g., Dahlen, 1990).

4.2.2.2. The eastern region. The San Pedro basin is bounded in the south by the Saona escarpment (Fig. 10a). This north-facing escarpment rises to a maximum height of 500 m over the seafloor of the basin and progressively diminishes along the strike before fading away in the vicinity of Saona Island, where there is a flat insular shelf (Figs. 7a–b and 10d). Towards the south-west the escarpment is progressively buried by the sediments of the San Pedro basin in the terraced region. This scarp is the exposed scarp of a main old normal fault that separates the ponded horizontal sediment fill of the San Pedro basin from a southward-tilted and faulted, thick sedimentary sequence covering the insular slope (see southward-tilted unit in Fig. 12). The southward-tilted unit could correspond with old basin fills that were tilted similar to the southward-tilted surface underlying the western basin fill (Fig. 11). These tilted blocks would be controlled by northward-dipping normal faults, such as the one associated with the Saona escarpment, that were active before the Middle Miocene (Fig. 7a–c; Heubeck et al., 1991). Gravity flows, slumping and sliding are frequent processes in the southward-tilted unit, as well as incised channels, yielding a stepped bathymetry (Fig. 10b and Appendix A2).

4.2.3. The Punta Salinas slope

The Punta Salinas slope is a southwest-facing steep slope located in the western end of the Muertos margin (see PSS in the inset map of Fig. 4). This slope is incised by a dense canyon network that suggests an intense transport of sediment to the Muertos trough. Sediment transport along the Punta Salinas slope and the Neiva valley has resulted in a large amount of sediments in the western Muertos trough, forming a thick turbidite wedge (>1.3 sTWT; Fig. 9). More sediments transported along the trough axis and deposited in the Dominican sub-basin form the thick sequence of sediment waves described earlier (Fig. 6).

At the base of the slope there is a WNW–ESE trending narrow compressive belt deforming the Muertos trough turbidites (Fig. 9).

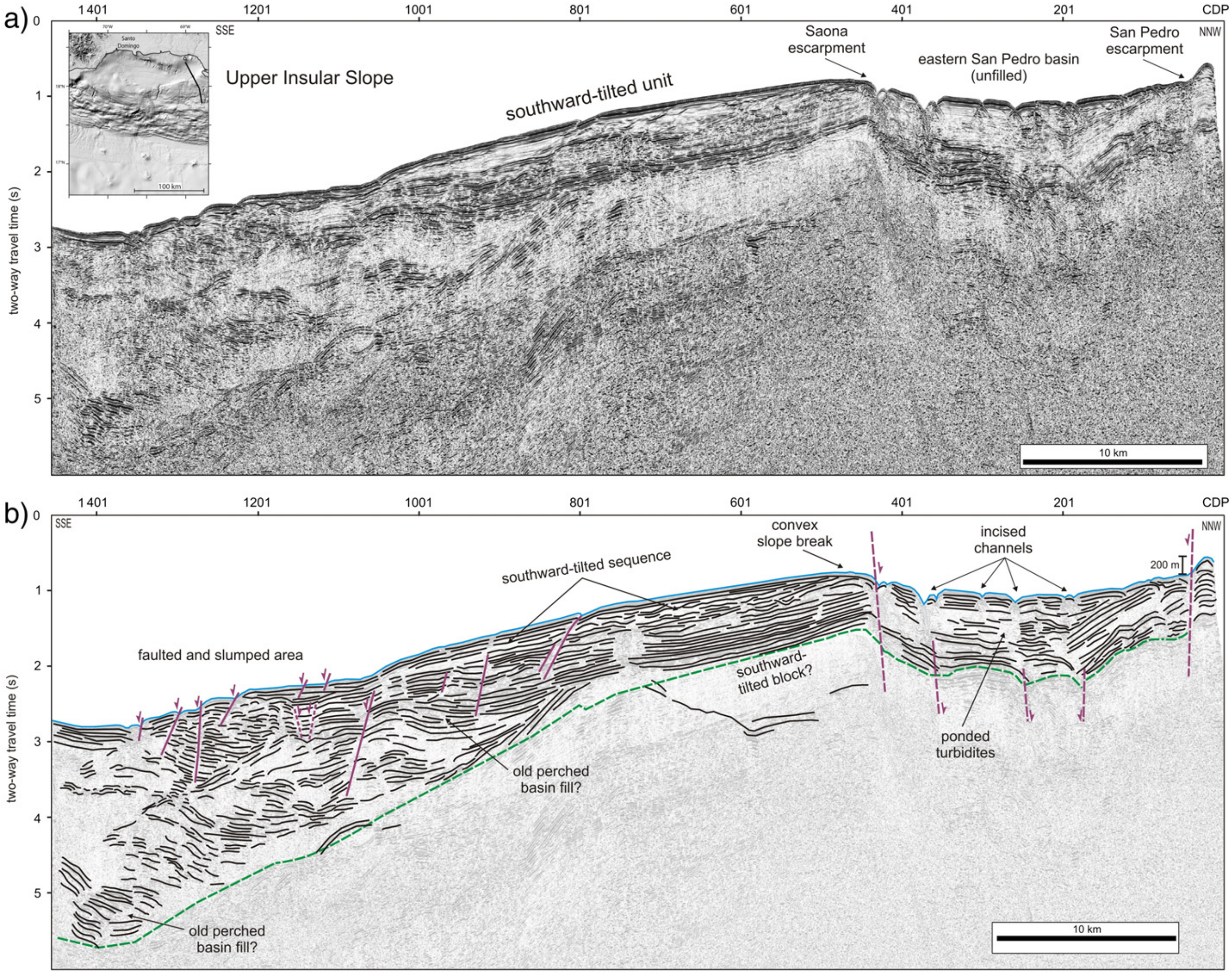


Fig. 12. a) Migrated MCS profile. V. E. is 6× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend.

The deformed belt does not show a well-developed imbricate structure and is bounded landward by a back thrust that forms an along-strike step in the insular slope. The Muertos trough turbidites are being actively folded suggesting active compressive stresses between the island arc and the Beata ridge. Additional evidence for compressive tectonics at the base of the Punta Salinas slope is a geomagnetic anomaly possibly caused by an underthrusted hill of the Beata ridge (see X in Fig. 3; Mauffret and Leroy, 1999).

4.3. The Beata ridge province

Most of the huge Beata ridge is submerged, but in the northern region it emerges locally to form the Beata and Alto Velo islands, and finally connects with the southward prolongation of the Bahoruco peninsula (Fig. 1). The submerged ridge has a roughly fan shape in map view and a length of 500 km from the Bahoruco peninsula to the Aruba gap. At latitude 14°N the Beata ridge loses its surface expression beneath the sediments of the Aruba gap. The northern zone of the Beata ridge is 130 km-wide and broadens towards the south reaching >400 km in width. In the southeastern region is the Beata plateau, a marginal unit of the Beata ridge, where active transpressional deformation was documented (Mauffret and Leroy, 1997). The depth along the summit of the ridge increases towards the south showing a succession of troughs and subsidiary ridges or hills (see Tairona, the DSDP 151 and Taíno hills in Fig. 1; Mauffret and Leroy, 1999).

The offshore northern Beata ridge shows a highly asymmetrical E–W cross-section (Fig. 1) interpreted classically as a complex and faulted huge horst (e.g., Driscroll and Diebold, 1998; Ewing et al., 1967; Fox and Heezen, 1975; Mauffret and Leroy, 1999; Roemer, 1973) where the eastern flank is significantly wider and less steep than the western flank (Fig. 7h, j). For a better analysis of the morphostructure, we have divided the Beata ridge into three distinct zones (see the inset map in Fig. 4): the eastern flank (EF), the summit zone (SZ) and the western flank (WF).

4.3.1. The eastern flank

The eastern flank of the Beata ridge consists of a series of N–S trending subsidiary hills which obliquely spread away from the NNE–SSW trending summit of the main ridge (see EF in the inset map of Fig. 4). To avoid confusion with the nomenclature between the Beata ridge and its subsidiary ridges, we use the term "hill" in reference to the subsidiary ridges. These hills form a series of steps that are deeper eastward and blur the boundary between the Beata ridge and the Dominican sub-basin (Fig. 3 and Appendix A4). Sediments filling the bathymetric troughs among the hills form perched basins and terraces (Figs. 13 and 14). The hills have steep sides reaching slopes of 20° and rise as much as 1600 m over the surrounding seafloor (Appendices A1

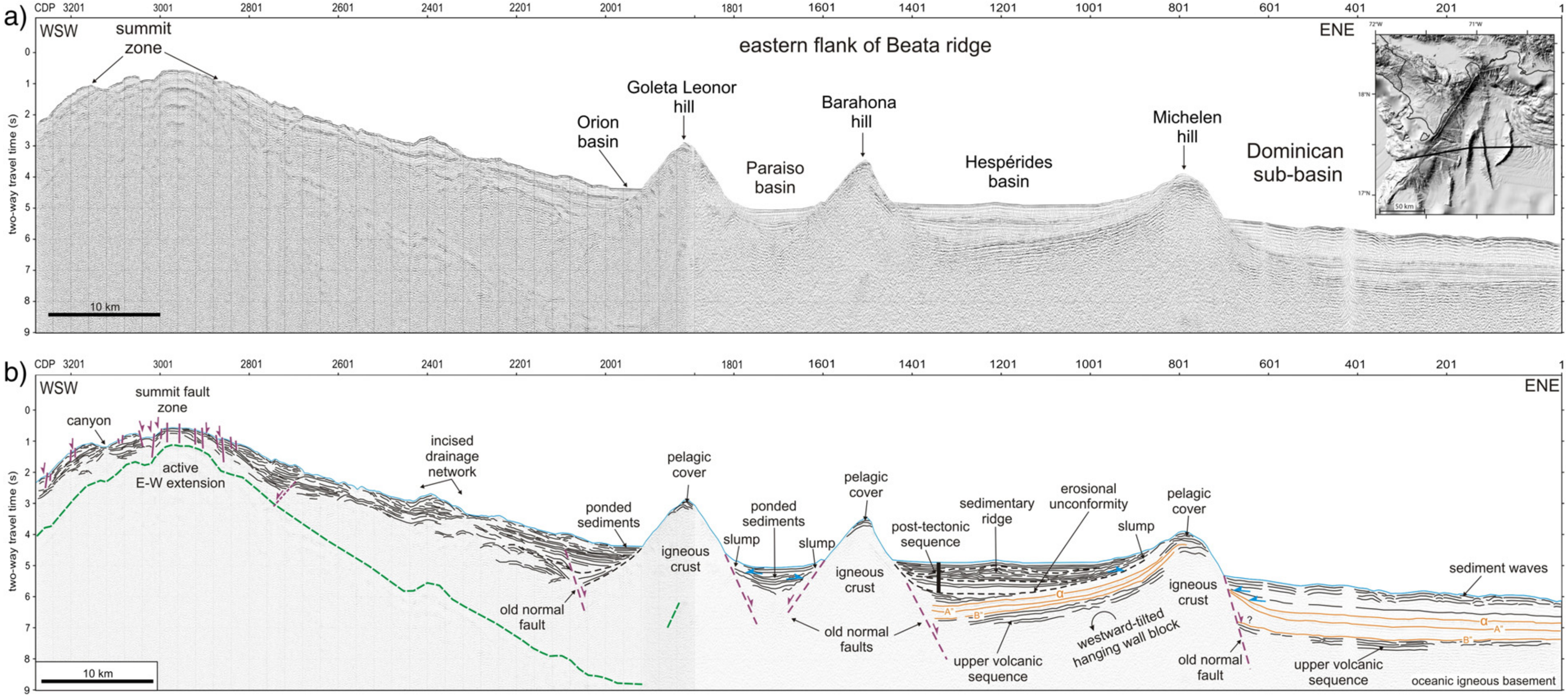


Fig. 13. a) Migrated MCS profile. V. E. is 3.7× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend.

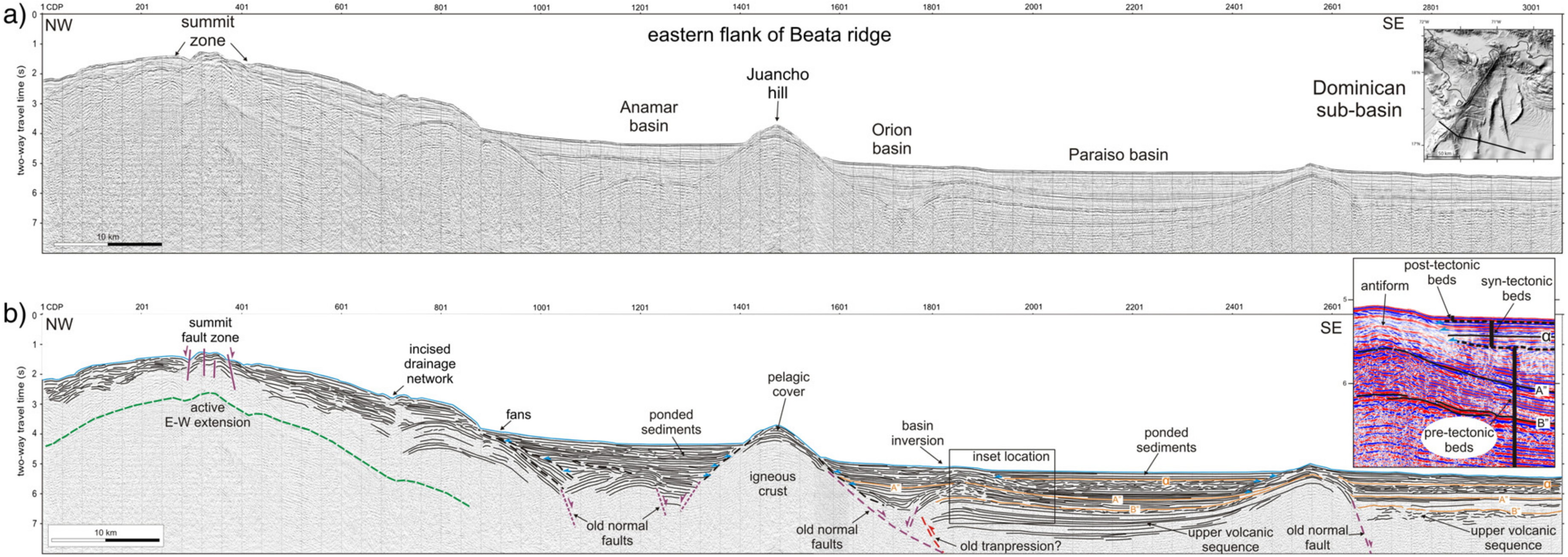


Fig. 14. a) Migrated MCS profile. V. E. is 3.7× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend. The inset shows the enlarged sector to image the inversion of the basin.

and A2). The hills reach lengths of 70 km (e.g., Goleta Leonor and Barahona hills) but plunge along strike and fade progressively beneath the sediments (Fig. 3). The highest hills have a largely triangular shape in cross-section view and are covered at the top by a thin blanket of sediment (Fig. 13). Sampling and seismic refraction experiments confirmed that the hills belong to the Upper Cretaceous oceanic crust underlying the Beata ridge (e.g., Edgar et al., 1971; Ewing et al., 1967; Fox et al., 1970; Mauffret et al., 2001; Révillon et al., 2000). Most hills have a linear geometry in map view, interpreted as being igneous constructions controlled by faulting (Diebold, 2009; Mauffret and Leroy, 1999).

The size of the perched basins and terraces (i.e., grabens and half-grabens) is variable depending on the accommodation space between the adjoining hills and their location in relation to the turbiditic supply (Fig. 4). There is a dense and incised canyon network in the eastern Bahoruco slope (see EBS in the inset map of Fig. 4) suggesting that the onshore supplies are significant sediment sources for the perched basins located in the northern zone (i.e., Orion and Hespérides basins; Fig. 4). Locally the basins can contain accumulations of blocks derived from slope failures in the eastern steep slope off Bahoruco peninsula (see slumped blocks in the Orion terrace; Figs. 4 and 9). The hills act as topographic barriers trapping and re-directing the turbiditic supply (e.g., Goleta Leonor hill). Perched basins located southwards collect only the sediment supplies from the local canyon network developed in the eastern flank and along the summit of the Beata ridge (e.g., Taína and Anamar basins; Figs. 4 and 7j). In this case, the sediment supplies are restricted to materials eroded from the summit of the Beata ridge and also minor local pelagic contributions, but can contain significant thicknesses of sediments (e.g., 1.6 sTWT in the Anamar basin; Fig. 14).

4.3.1.1. The Paraiso basin. The Paraiso basin occupies a N–S-elongated depression between the Barahona and Goleta Leonor hills (see PB in Fig. 3). The northern zone of the basin is narrow and isolated from the channelized supply by the striking Goleata Leonor hill (Fig. 13), but southwards the hill disappears beneath the sediments allowing the channelized sediment supplies to reach the basin (Figs. 7h, i and 14). The southern basin contains a thick sedimentary sequence from the Upper Cretaceous to the Present floored by the upper volcanic sequence (Fig. 14). Stratigraphic analysis of the seismic horizons permits two phases of deformation to be inferred. A first phase suggested by the westward thickening of the upper volcanic sequence and the A″–B″ interval as a result of a westward tilting of the oceanic crust underlying the basin. This phase of tilting finished soon after the deposition of the A″ horizon and suggests the existence of an eastward-dipping normal fault in the eastern side of the basin during this time. This tilting led to the coeval uplifting and formation of the hill located eastward since the seismic horizons of the upper volcanic sequence and the A″–B″ seismic interval continue towards the top of the hill. Before the deposition of the α horizon, a second phase of compressive deformation started yielding a local inversion in the western zone of the basin. This last phase has formed an antiform that seems to have been active until Pleistocene times (see inset in Fig. 14). Recent sediment beds cover the antiform and keep a constant thickness suggesting that this structure is not active. This antiform could be the expression of a deep reverse fault formed during the phase of transpressional deformation that started in the Lower Miocene. The two deformation phases experienced by the Paraiso basin have yielded a seismic architecture very different from that of the Dominican sub-basin.

4.3.1.2. The Michelen hill and the Hespérides basin. The Michelen hill shows a particular arcuate shape in map view, convex to the east, and an asymmetric cross-section with a steep eastern side and a gentle western side (see MH in Fig. 3). The eastern side is an exposed scarp of an old normal fault while the western side is the gently westward-tilted top surface of an oceanic igneous block (Figs. 8 and 13). This configuration results in a half-graben structure filled by substantial sediment accumulation and forms the Hespérides basin (see HB in Fig. 3). The Michelen hill plunges along strike but there is a structural 10 km-long gap where the hill is partially buried by the sediments of the Hespérides basin to form a terrace (see gap in Fig. 4). The northern zone of the hill lies parallel to the deformation front of the Muertos thrust belt, but there is a 5 km-wide trough filled by a wedge of horizontal turbidites between them (Fig. 8).

The Hespérides basin is a partly confined and elongated depression that traps an important part of the sediment supply from the Bahoruco peninsula resulting in a sedimentary section reaching a thickness of 2 sTWT (Figs. 8 and 13). The Hespérides basin contains the α–B″ interval overlying the upper volcanic sequence, and is both buried by wedges of fan-pattern sediments and a thick sequence of horizontally-ponded sediments. Stratigraphic analysis suggests an old phase of westward tilting and coeval uplifting of the Michelen hill. Uplifting is inferred from the presence of the basal sequences rising eastward that cover the summit of the Michelen hill. The substantial sediment accumulation of horizontally-ponded sediments filling the Hespérides basin suggests that the deformation in this basin ceased some time after the α horizon (Lower Miocene).

4.3.2. The summit zone

The summit of the Beata ridge mainly consist of a 10 km-wide band of stepped seafloor showing a succession of bathymetric scarps and lineaments oriented NNE–SSW (see SZ in the map inset of Fig. 4). These features are the surface expression of numerous normal faults that suggest active NW–SE oriented extension (see summit fault zone in Figs. 7i, 13 and 14). Faults offset the seafloor forming west- and east-facing bathymetric scarps, which extend tens of km along strike and reach heights of 250 m (Fig. 4). Coring in the summit region recovered Middle Eocene neritic chalk and Lower Oligocene-Pleistocene deep water carbonate oozes (Fig. 1; Fox and Heezen, 1975). Faulting activity has resulted in a variable depth along the summit region with a succession of minor elongated hills and troughs (Appendix A5). This fault zone continues north-eastward to the east of Alto Velo and Beata islands forming minor NNE–SSW trending steps disrupting the canyon network. However, there is no evidence for this fault zone continuing northeastward along the eastern Bahoruco slope. The eastern Bahoruco slope is a steep 2500–3000 m-high escarpment that may be the exposed scarp of an old southeast-facing normal fault (Fig. 7e–h). This fault would be similar to the faults bounding the subsidiary hills of the eastern flank of the Beata ridge formed during the more extensional unloading process in the Upper Cretaceous. At the base of the eastern Bahoruco slope there are minor steps disrupting the canyon network possibly related to buried N–S-oriented subsidiary hills (Fig. 4). The Beata fault zone seems to continue south-westward parallel to the Beata escarpment but satellite-derived altimetry and seismic data do not offer enough detail (see Fig. 10 in Driscoll and Diebold, 1999).

4.3.3. The western flank

The western flank of the Beata ridge is a huge westward-facing steep slope that rises from the flat abyssal plain of the adjoining Haiti sub-basin at a depth of 4300 m (see WF in the inset map of Fig. 4). This flank shows two distinct regions: the NW-facing Beata escarpment and the SW-facing western Bahoruco slope. The Beata escarpment shows a dense and incised canyon network flowing towards the Haiti sub-basin (Fig. 4). The slope varies between 8.5° and 11.5°, with locally higher slopes of 20° related to highly incised canyons (Appendix A2). There are several narrow slope steps disrupting and re-driving the drainage network that could be a result of minor normal faulting driven by gravitational forces because of the steep slope (Fig. 15 and Appendix A5). The cross-section view of this escarpment suggests that it is the exposed scarp of a major NW-dipping normal fault where the footwall block would be the Beata ridge and the hanging wall block the adjoining Haiti sub-basin (Fig. 7h–i). Sampling and diving along the Beata escarpment recovered gabbros and dolerites and only occasional pillow basalts (Fox and Heezen, 1975; Fox et al., 1970; Mauffret et al., 2001;

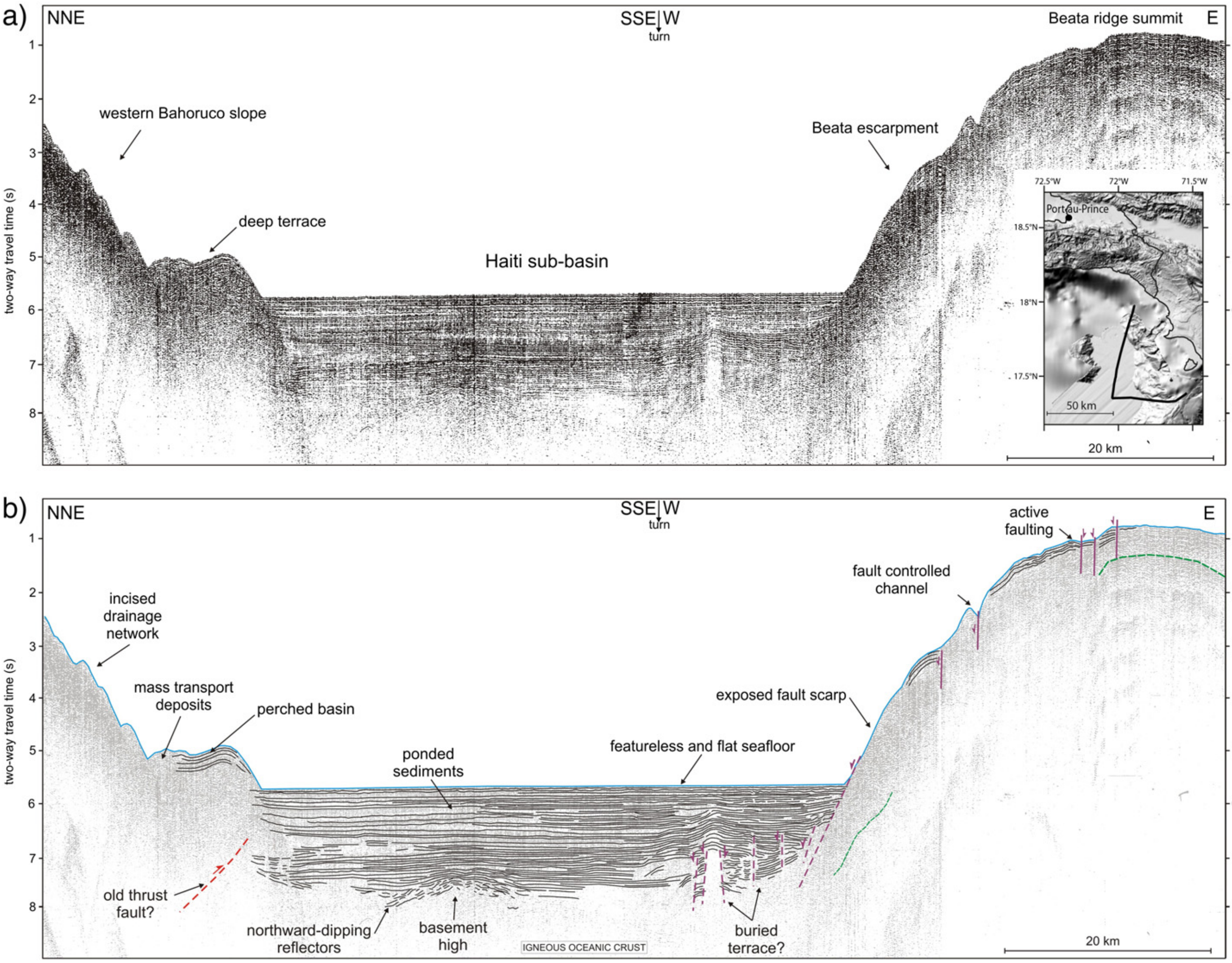


Fig. 15. a) SCS profile. V. E. is 9.5× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend.

Révillon et al., 2000) (Fig. 1). At the base of the western Bahoruco slope, at a depth of ≈3500 m, there is a deep terrace bounded seaward by a steep scarp (Fig. 7h and Appendix A1). This terrace is covered by products of erosional processes and supplies from the upward drainage network connected with the onshore rivers of the Bahoruco peninsula (Fig. 15). Samples in this deep terrace and along a canyon axis recovered Paleocene-Eocene shallow water carbonates and doleritic and grabbroic pebbles (Fig. 11; Mauffret et al., 2001).

4.4. The Haiti sub-basin province

The Haiti sub-basin has an almost flat and featureless seafloor with a quasi-constant average depth of 4300 m (Figs. 3 and 7h–i). The sub-basin fill consists of a 2–2.7 sTWT-thick sequence of continuous highly-reflective reflectors corresponding with ponded sediments (Figs. 15 and 16). The sedimentary beds are only disturbed by occasional growth faults in the center of the basin (see inset in Fig. 15) and are locally interlayered by chaotic reflections associated with mass transport deposits in the bounds of the basin (see MTD in Fig. 15). There is a northward thickening of the sedimentary cover controlled by a progressive northward sinking of the smooth basement surface (see B″ Horizon? in Fig. 16). Basal sediment horizons onlap the basement seismic surface southward and also show a gentle northward dipping. Upsection horizons lie progressively more horizontal suggesting that the northward-tilted geometry of the basement surface would be prior to the sedimentation of the recent sedimentary sequences. The seismic basement horizon could resemble the B″ horizon, and the underlying dipping reflectors could correspond with the upper volcanic sequence identified in the Dominican sub-basin. However, the oceanic crust underlying the Haiti sub-basin was not affected by the thickening processes experienced by the crust underlying the Dominican sub-basin crust and therefore the resulting crustal structure should be different (Driscoll and Diebold, 1999; Mauffret and Leroy, 1999). The stratigraphic control in the Haiti sub-basin was established by drilling far southward in the vicinity of the Hess escarpment, where probable equivalents of horizons A″ and B″ were identified in seismic lines across DSDP site 152 (Fig. 1; Bowland, 1993; Holcombe et al., 1990; Moore and Falhquist, 1976; Stoffa et al., 1981). However, we cannot identify the A″–B″ interval in the Haiti sub-basin in agreement with Mauffret and Leroy (1997, 1999).

At the northern edge of the sub-basin there is a wide synform which ends sharply northward in a region of chaotic reflections (Fig. 16). The synform fades upsection and the recent layers onlap horizontally the region of chaotic reflections. The existence of the synform could be a result of the compaction of the basin sediments and/or a result of the southward migration of compressive deformation from base of island arc slope. The last hypothesis suggests that this chaotic region could be interpreted as a buried wedge of highly-deformed sediments. The Haiti sub-basin sediments were deformed and folded by means of south-verging thrust faults. The buried wedge lies above isolated and

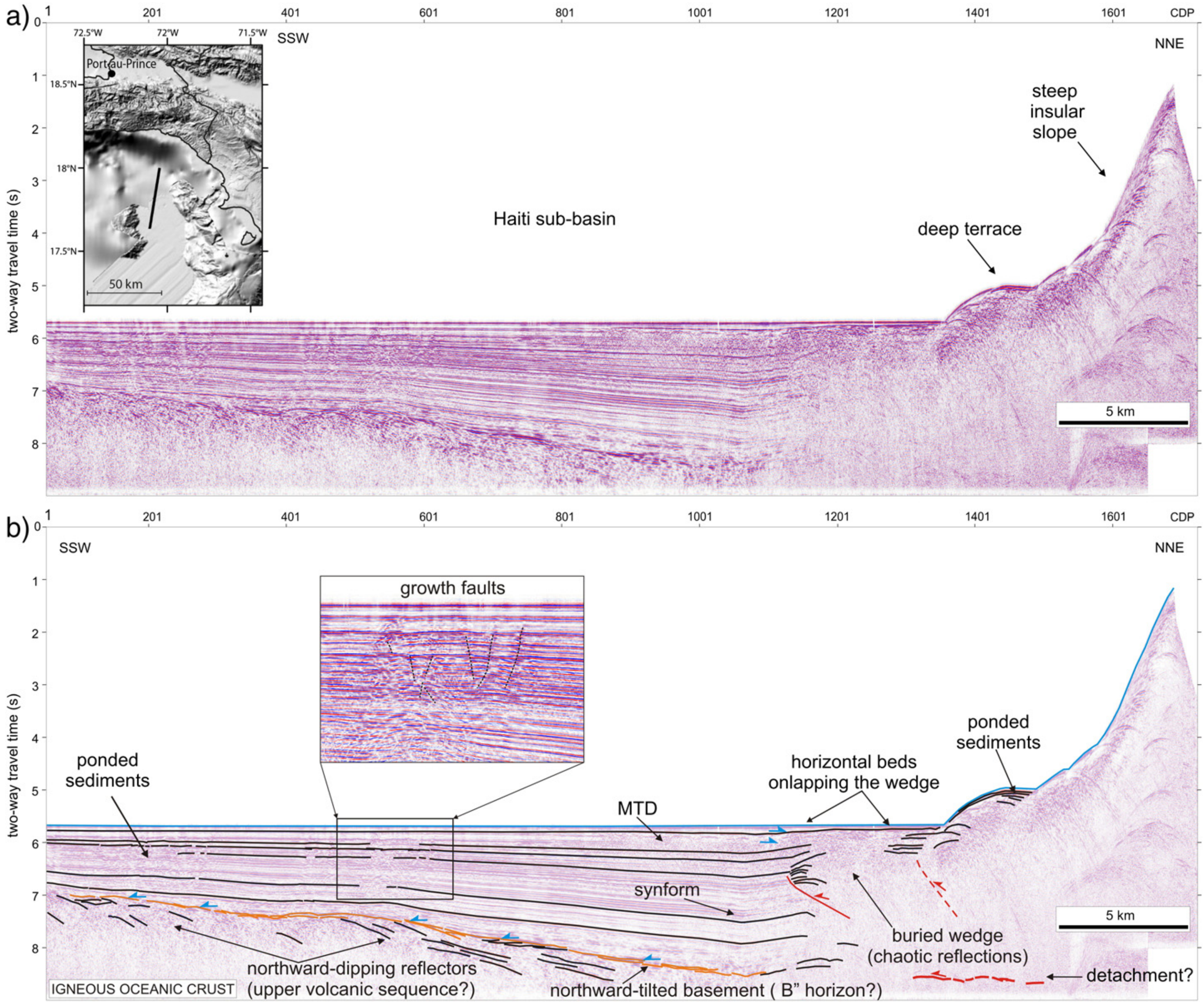


Fig. 16. a) MCS profile. V. E. is 2.8× on the seafloor. b) Line drawing interpretation. See Fig. 5b for legend. The inset shows an enlarged sector to image growth faults in the basin.

discontinuous sub-horizontal reflectors suggesting a possible detachment surface but seismic data do not offer enough resolution to establish whether this is the case (Fig. 16). This wedge could be the buried part of a narrow deformed belt that emerges and forms a deep terrace along the base of the western Bahoruco slope (Figs. 7h and 15). Recent horizontal sediments of the Haiti sub-basin are onlapping and burying the deformed wedge, suggesting that this compressive system is inactive (Fig. 16).

The Haiti plateau is poorly documented because it has not been studied previously (Fig. 4). The plateau rises 800 m over the seafloor of the Haiti sub-basin and has a steep eastern side highly-incised by a dense canyon network. On the basis of this morphology it was suggested that the plateau could consist of a NW–SE-trending thick block possibly related to the Caribbean Large igneous province (Mauffret and Leroy, 1999).

5. Discussion

5.1. Deformation and morphology in the northern Beata ridge

The morphology of the Beata ridge has been interpreted as a result of the overlapping of different deformation mechanisms which have occurred since its formation in the Upper Cretaceous. However, there is controversy regarding the importance of the latest phase of Neogene transpressive deformation experienced by the ridge (see red line in Fig. 1). Mauffret and Leroy (1997, 1999) suggested that an important part of the morphology of the Beata ridge has been yielded from the Lower Miocene by active NE–SW oriented compression driven by the collision between the Beata ridge and the island arc. This hypothesis is based on active reverse strike–slip structures observed in the eastern flank of the Beata ridge, yielded by a strong transpressional tectonic regime that reactivated the normal faults of the Upper Cretaceous volcanic plateau. This transpressional regime implies the existence of a Colombian plate (i.e., Colombian basin + Beata ridge) thrusting obliquely over a Venezuelan plate (i.e., Venezuelan basin) along the eastern flank of the Beata ridge. Driscoll and Diebold (1998) and Diebold (2009) proposed an alternative hypothesis where by the morphology of the Beata ridge is the result of an extensional unloading and consequent flexural rebound of the lithosphere formed by the Upper Cretaceous volcanic plateau within a single plate. Minor and local fault reactivation in the Neogene along the eastern flank of the Beata ridge is associated with accommodation zones that record the collision between the Beata ridge and the island arc.

5.1.1. Deformation along the eastern flank of the Beata ridge

The boundary between the Beata ridge and the Dominican sub-basin is located along the easternmost hills of the Beata ridge, but it remains buried beneath the sediments. The easternmost hills are controlled by pre-existing normal faults forming deep detachments between the Beata ridge (>20 km-thick crust) and the Venezuelan basin (<15 km-thick crust) (Driscoll and Diebold, 1999; Mauffret and Leroy, 1999). Such detachments have accommodated the extensional unloading and consequent flexural rebound of the thick crust of the Beata ridge from the Upper Cretaceous to soon after the Lower Miocene, after which they were partially reactivated as transpressional structures (Diebold, 2009; Mauffret and Leroy, 1999). The active transpressional tectonics is driven by the collision of the Beata ridge and the island arc resulting in the reactivation of pre-existing normal features along the eastern flank of the Beata ridge. Active transpressional structures have been well documented at the south-eastern end of the Beata ridge (see Beata plateau and Taíno hill in Fig. 1; Mauffret and Leroy, 1997) and in the eastern flank approximately 90 km south of Michelen hill (see Fig. 10 in Diebold, 2009). Based on these observations, Mauffret and Leroy (1999) suggested that the uplift of the easternmost hills of the northern Beata ridge (i.e., Michelen hill) is a result of transpressional tectonics after the Lower Miocene and that uplifting is still active. They therefore inferred that faults bounding the hills are active reverse strike–slip structures connected along strike by transversal strike–slip faults and lateral ramps, defining a step-over transpressive deformation front between the Beata ridge and the Venezuelan basin (Fig. 1). On the basis of seismic profiles, they interpreted the northward increase in the elevation of the hills as a result of greater amounts of transpressional stress in the proximity of the collision region. They estimated that the transpression would have caused at least 1 km of uplift in the Michelen hill and at least 0.6 km in Barahona hill.

Our observations in the northern Beata ridge only show local evidence of a late uplift event in the eastern flank, and we see no evidence for active reverse structures in recent times. This late uplift event was recorded locally in the western Paraiso basin not long before the deposition of the α horizon (Lower Miocene) and finished in Pliocene times (see the inset in Fig. 14). The estimates of the uplift experienced by the hills carried out by Mauffret and Leroy (1999) without the benefit of swath bathymetry are only feasible at the location of their seismic profiles. The N–S plunging morphology of the subsidiary hills means that the height of the hills is variable along strike and does not increase northward. The latter observation invalidates the hypothesis that the hills have an increasing elevation caused by the increasing transpressional stresses near the collision area. In the study region therefore the faults bounding the hills seem to be inherited features and they are not currently accommodating transpressional tectonics caused by the interaction between the Caribbean plate's interior and the island arc. These hills can be better explained by a mechanism of extensional unloading developed from the ridge formation in the Upper Cretaceous (i.e., underplating) which is still residually active in the summit region (i.e., Beata summit fault zone). Southward of the study region the morphology and topography of the hills seems to be also controlled by local fault reactivation along the eastern flank of the Beata ridge (Driscroll and Diebold, 1998; Mauffret and Leroy, 1999)

5.1.2. Origin and morphology of the Michelen hill

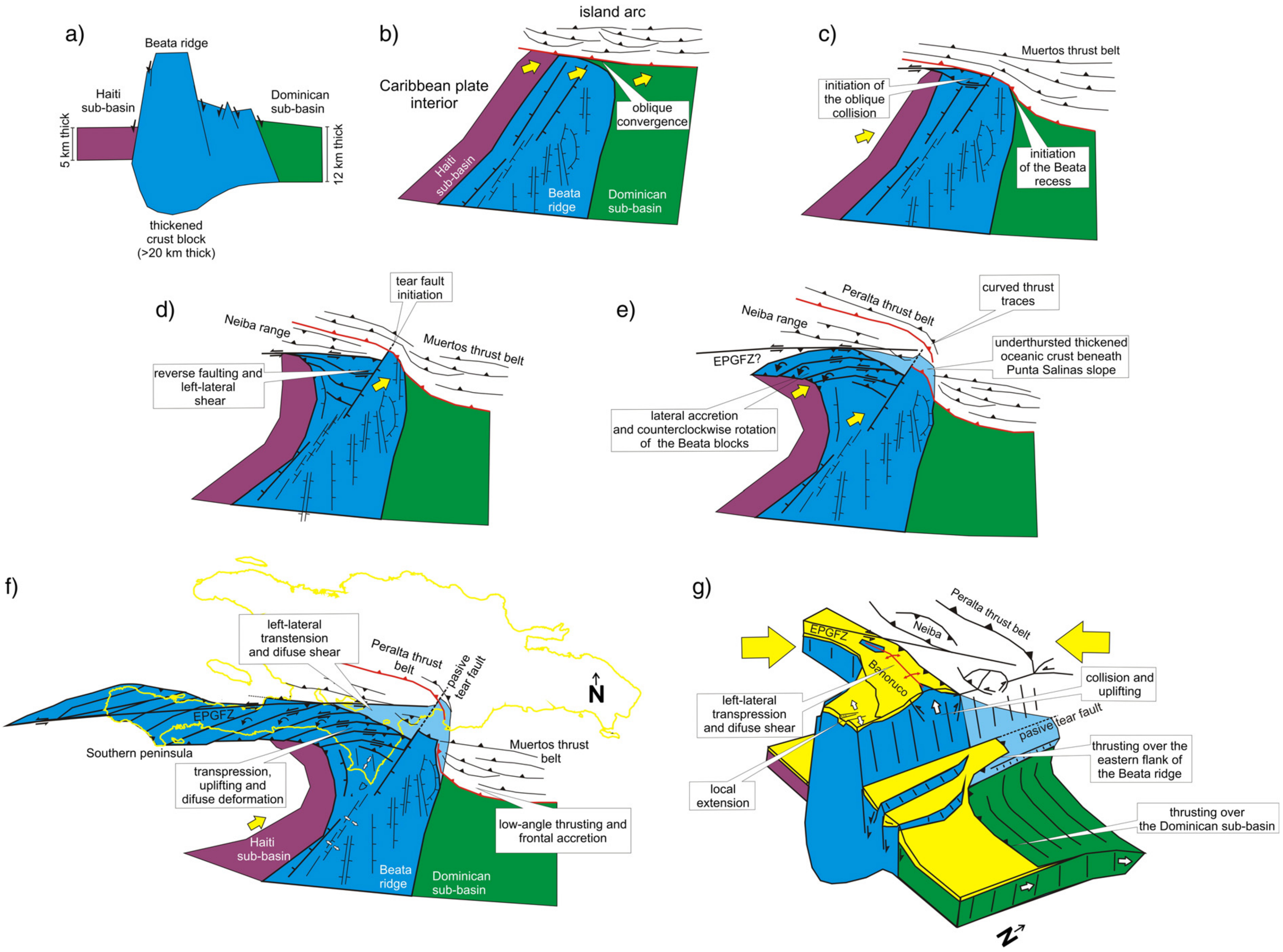
The particular arcuate shape in map view of the Michelen hill lying obliquely to the other linear N–S trending hills was interpreted as a result of the active collision of the Beata ridge and the Muertos thrust belt (Jany, 1989). This collision would have caused the Michelen hill to bend oroclinally and become parallel to the deformation front. In the same way the thrust belt would also experiment a significant oroclinal bend and form the Beata recess. Our observations on the basis of new data suggest that such an interpretation has obvious inconsistencies. This hypothetical ongoing collision should also be recorded in the faults bounding the Michelen hill and in the recent sedimentary record of the Hespérides basin. However, the last recorded deformational phase (i.e., extension and uplift) finished soon after the Lower Miocene and the normal faults bounding the Michelen hill seem to have been long inactive (Figs. 8 and 13). Between the Michelen hill and the deformation front of the Muertos thrust belt there is a 5 km-wide trough filled by a an undeformed turbidite wedge. The compressive deformation suggested by frontal anticline ridges did not reach the Michelen hill (Fig. 8). Therefore, there is no current tectonic interaction between the Michelen hill and the Muertos thrust belt that could explain the arcuate shape of the Michelen hill. We interpret this arcuate shape as a combination of a curved inherited igneous morphology modified by a later stage of uplifting and westward tilting that finished soon after the Lower Miocene.

5.1.3. The Beata escarpment and the eastern Bahoruco slope

Mauffret and Leroy (1999) suggested that the W-facing Beata escarpment could be part of a recent major right-lateral strike–slip fault. The NNE–SSW trending strike–slip fault would continue to the NE along the Beata summit and along the E-facing eastern Bahoruco slope (see yellow line in Fig. 1). This major fault and the reverse strike–slip faults observed along the eastern flank would result from the reactivation of the pre-existing normal faults due to the collision of the Beata ridge and the island arc from the Lower Miocene. Driscroll and Diebold (1998) interpreted the Beata escarpment as a major NW-dipping normal fault that offsets the igneous crust of the Beata ridge (i.e., the footwall block) 5 sTWT (≈3750 m) from the igneous crust of the Haiti sub-basin (i.e., the hanging wall block). The eastern Bahoruco would be a major SE-dipping normal fault that offsets ≈2500 m the Bahoruco peninsula (i.e., the footwall block) and the adjoining eastern flank of the Beata ridge (i.e., the hanging wall block). Thus the Beata escarpment and the eastern Bahoruco slope would be opposed-verging old normal faults formed as the Beata ridge rose as a result of the extensional unloading and consequent flexural rebound of the lithosphere during and soon after the emplacement of the Upper Cretaceous volcanic plateau.

Our interpretations agree with the hypothesis postulated by Driscroll and Diebold (1998). We consider that the Beata escarpment and the eastern Bahoruco slope are major normal faults formed early in the history of the Caribbean igneous province. We have not observed any evidence of active strike–slip structures along the Beata escarpment, the Beata summit fault zone or the eastern Bahoruco slope. In the Beata escarpment and in the eastern Bahoruco slope we only observe minor shallow faults disrupting locally the canyon network (see disrupted network in Fig. 4). The NNE–SSW oriented Beata summit fault zone is parallel to the Beata escarpment and to the Bahoruco slope and shows evidence of active tectonics. However the active tectonics is concentrated in a 10 km-wide band of deformation and characterized by discontinuous shallow normal faults. This deformation seems to respond to NW–SE oriented extension probably governed by a residual extensional unloading process started in the Upper Cretaceous

Fig. 17. Conceptual model based on this study and on Mercier de Lépinay et al. (1988), Pubellier et al. (2000) and Leroy et al. (2000). a) Schematic E–W cross-section of the oceanic crust of the Beata ridge and the adjoining basins. b) Stage previous to collision. Caribbean plate's interior moves towards the NE relative to the island arc. Yellow arrows show the convergence vector. c) Initiation of the oblique collision and the recess in the Muertos thrust belt in the Lower Miocene (7 Ma; Leroy et al., 2000). d) Initiation of the NE–SW trending passive tear fault along the eastern Bahoruco slope and the E–W trending left-lateral transpression in the Bahoruco peninsula. e) Vertical and lateral accretion of oceanic crust blocks from the thicker region of the Beata ridge westward of the passive tear fault. Disconnection between the onshore (i.e., Peralta thrust belt) and offshore thrust belts (i.e., Muertos thrust belt). EPGFZ = Enriquillo–Plantain Garden fault zone. f) Present configuration of westward detached slices of the Beata ridge forming the oceanic basement of the Southern peninsula. Yellow line shows the present coast of Hispaniola. g) Southern Hispaniola viewed from the SE showing the along-strike change in the tectonic regime yielded by the oblique collision. Yellow zones show the Cenozoic cover. Yellow arrows show the regional convergence direction. White arrows show local tectonic compression.


a)
Beata ridge
Haiti
sub-basin
Dominican
sub-basin
5 km thick
12 km thick
thickened
crust block
(>20 km thick)
b)
island arc
Caribbean plate
interior
Haiti
sub-basin
Beata
ridge
Dominican
sub-basin
oblique
convergence
c)
Muertos thrust belt
initiation of
the oblique
collision
initiation
of the Beata
recess
d)
tear fault
initiation
Neiba range
Muertos thrust belt
reverse faulting and
left-lateral
shear
e)
curved thrust
traces
Peralta thrust belt
Neiba range
underthursted thickened
oceanic crust beneath
Punta Salinas slope
EPGFZ?
lateral accretion
and counterclockwise rotation
of the Beata blocks
f)
left-lateral
transtension
and difuse shear
Peralta thrust
belt
pasive
tear fault
N
EPGFZ
Southern peninsula
Muertos thrust
belt
transpression,
uplifting and
difuse deformation
low-angle thrusting and
frontal accretion
Haiti
sub-basin
Beata
ridge
Dominican
sub-basin
g)
Peralta thrust belt
Neiba
EPGFZ
Banoruco
collision and
uplifting
left-lateral
transpression
and difuse shear
pasive tear fault
thrusting over the
eastern flank of
the Beata ridge
local
extension
thrusting over
the Dominican sub-basin
N

(Driscroll and Diebold, 1998). This fact implies that these structures are not currently accommodating any transpressional deformation driven by the collision of the Beata ridge and the island arc.

5.2. Tectonic implications of the oblique collision

The convergence between the Beata ridge and the island arc existed at least from the Lower Miocene and it is still active. This fact is supported from different approaches:

- Fault striation studies in the Bahoruco peninsula show active horizontal compression oriented 030° (see red arrows in Fig. 1; Mercier de Lépinay et al., 1988; van den Berghe, 1983; Vila et al., 1990).
- GPS-derived velocities show that the Bahoruco peninsula (i.e., an emerged part of the Beata ridge covered by Neogene sediments) is moving 5–9 mm/y faster to the NE than the Hispaniola interior block (Benford et al., 2012; Calais et al., 2010).
- Results from systematic geological mapping support a progressive turn of the Peralta thrust belt as a response to the accommodation of the impingement of the Beata ridge in southern Hispaniola (See Peralta thrust belt in Fig. 4; Hernáiz-Huerta et al., 2007)
- Seismicity at depths shallower than 30 km in the south-central Hispaniola block shows oblique reverse focal mechanisms that suggest NE–SW compression (Byrne et al., 1985; Granja Bruña et al., 2010).
- Offshore observations suggest a southward thrusting of the Hispaniola block over the eastern flank of the Beata ridge (Mauffret and Leroy, 1999; this study in the base of Punta Salinas slope).

As a result of the asymmetrical E–W cross-section and consequent variable thickness of the northern Beata ridge (Fig. 17a), the consequences of the oblique convergence with the island arc vary significantly from E to W. In addition, the deformational response is very different between the Beata ridge, formed by a thickened and faulted crystalline oceanic crust (i.e., “rigid block”), and the island arc, formed by accreted terranes derived from an intra-oceanic island arc (Pindell and Kennan, 2009). Therefore, the offshore northern Beata ridge mainly shows old inactive structures without recent deformation associated to the current collision.

Based on the results of this study and from the evolutionary models for the northern Caribbean boundary zone (e.g., Leroy et al., 2000; Pubellier et al., 2000), we propose a conceptual model to explain the implications of the oblique collision of the Beata ridge and the island arc (Fig. 17b–g). The Muertos thrust belt grows southwards over the Caribbean plate's interior at least from the Eocene onward and in the Upper Miocene starts to collide obliquely with the thickened crust block of the Beata ridge (Fig. 17c). This oblique collision leads to a significant steepening of the detachment, the thrust belt is oroclinally bent and the huge Beata recess is formed (Fig. 17b–f). ten Brink et al. (2009) successfully reproduced the impingement of thickened crust (i.e., the Beata ridge) into a retroarc thrust belt (i.e., Muertos thrust belt) using sand-box kinematic modeling. The results showed that when the thrust belt reaches the thickened crust block, the pre-existing thrust traces bend oroclinally, and the new thrusts initiate with a distinct curve towards the point of impingement (see Fig. 9c in ten Brink et al., 2009). At the Punta Salinas slope, the Beata ridge is becoming progressively thicker and the accommodation of the oblique convergence takes place by means of narrow compressive deformed belt, though without a well-developed imbricate system. This compressive belt suggests that there are active compressive stresses between the island arc and the Beata ridge and that part of the eastern flank is being underthrusted beneath the Punta Salinas slope (Fig. 17e). Westward the height of the Beata ridge rises rapidly in the eastern Bahoruco slope and the southward-verging thrusting evolves sharply to collision and uplifting in the Bahoruco peninsula (Fig. 17c–f).

The eastern Bahoruco slope is the exposed scarp of an old fault that accommodates the sharp transition from thrusting at the base of Punta Salinas slope to collision and uplifting in the Bahoruco peninsula and adjacent areas. This is a pre-existing structure formed in the Upper Cretaceous and probably reactivated during the extensional unloading experienced by the Beata ridge, but it shows no evidence of recent deformation related to the collision process. This sharp evolution from thrusting to collision would imply the existence of a NE–SW-oriented tear fault between the underthrusted eastern flank of the Beata ridge and the basement of the uplifted areas of south-central Hispaniola located westward (i.e., basement of the Neiba range and the Peralta thrust belt) (Fig. 17f). This tear fault would be the northward continuation of the eastern Bahoruco slope and could be related to the shallower seismicity in the south-central Hispaniola block. Here an oblique reverse focal mechanism suggests NE–SW compression (Granja Bruña et al., 2010). During the oblique collision with the island arc, the thicker area of the Beata ridge experiences an E–W oriented left-lateral shear forming series of horizontal, counterclockwise-rotating stacked slices of oceanic crust (Mercier de Lepinay et al., 1988). This mechanism of horizontal accretion of uplifted oceanic blocks would form the basement of the Bahourco and Southern peninsulas, and could also be implicated in the initiation of the left-lateral Enriquillo–Plantain Garden fault zone. Westwards of the Beata escarpment and western Bahoruco slope, the thickness of the oceanic crust underlying the Haiti sub-basin decreases sharply to 5 km. An additional part of the oblique convergence would be accommodated by means of a southward-verging thrust of the western Bahoruco slope over the Haiti sub-basin. This thrusting has formed a deep terrace at the base of the western Bahoruco slope but it is currently inactive in the study region. Farther towards the W in the offshore southern slope of the Southern peninsula of Haiti this southward thrusting could be active (Bien-Aime Momplaisir, 1986).

In sum, the oblique convergence of the Beata ridge and the island arc from the Upper Miocene onward has caused a sharp change in the tectonic stresses and in the deformation style on both sides of the collision area. The deformation style in southern Hispaniola evolves from southward-thrusting along the Muertos margin, to collision and uplifting in the Bahoruco peninsula to transpression and uplifting in the Southern peninsula of Haiti (Fig. 17f, g). In the south-central Hispaniola the left-lateral shear on E–W oriented Enriquillo Plantain-Garden fault zone coexists with the NE–SW oriented compression associated with the oblique collision of the Beata ridge and the island arc (Hernáiz-Huerta et al., 2007).

6. Summary and conclusions

The morphotectonic interpretation of the offshore southern slope of the Dominican Republic has provided new constraints on the oblique collision between the thickened oceanic crust of the Beata ridge and the inactive island arc. A combined interpretation of the swath bathymetry and seismic reflection data allowed us to identify four morphotectonic provinces: the Dominican sub-basin, the Muertos margin, the Beata ridge and the Haiti sub-basin.

- The Dominican sub-basin has a seafloor characterized by frequent isolated igneous mounds, deep water sediment waves and a wedge of turbidites along a fault controlled and deeper Muertos trough. Locally there are active E–W-trending normal faults resulting from the bending of the Caribbean plate's interior and the differential overburden of the Muertos thrust belt.
- The Muertos margin has several distinct slope regions resulting from the differential rate of convergence accommodation between the Caribbean plate's interior and the island arc. The lower slope is occupied by the active Muertos thrust belt that is growing southward over the Dominican sub-basin and the eastern flank of the Beata ridge. This thrust belt includes several active out-of-sequence thrust

faults which, if they rupture along their entire length, could generate earthquakes of large magnitude. The convergence between of the thrust belt and the Beata ridge has yielded a huge recess and the imbricate system is replaced by a narrow deformed belt at the base of the Punta Salinas slope. The upper slope has extensive thick slope deposits (e.g., San Pedro basin) burying the rear zone of the Muertos thrust belt and old tilted island arc blocks. Extensional tectonics and slumping are frequent processes caused by the vertical growth of the thrust belt and the consequent over-steepening of the insular slope.

- The northern Beata ridge consists of an uplifted and highly-faulted block of oceanic crust showing an asymmetrical E–W cross-section. The eastern flank shows an alternation of subsidiary hills and perched basins that obliquely spread away from the main ridge. The hills are old faulted and uplifted blocks of oceanic crust which have experienced several tectonic stages of extension resulting in a complex configuration. The only evidence of active tectonics is a 10 km-wide band of stepped seafloor along the summit of the ridge suggesting active NW–SE oriented extension.
- The Haiti sub-basin is bounded northward by a narrow compressive deformed belt that locally emerges and forms a deep terrace. This deformed belt is buried by horizontal sedimentary layers suggesting that the compressive deformation in not active.

Our results suggest that the shallower structure and morphology of the northern Beata ridge can be mainly explained by a mechanism of extensional unloading from the Upper Cretaceous which is still active residually along the summit of the ridge. Active reverse strike–faults and transpression caused by the oblique convergence between the Beata ridge and the island arc are not supported by our structural interpretation. The eastern Bahoruco slope is the exposed scarp of an old normal fault that acts as a passive tear fault accommodating the sharp along-strike transition from thrusting in the Muertos margin to collision and uplifting in the Bahoruco peninsula.

Future studies integrating the onshore and offshore data will help to complete and constrain the timing and kinematics of the deformation and provide a better understanding of the tectonic implications in the southern Hispaniola. More systematic acquisition of swath bathymetry and closely-spaced seismic profiles are needed to improve our knowledge of the complex history of the Beata ridge which is intimately related to the evolution of the Caribbean plate. The study of the rupture history and potential of the active thrust faults in the Muertos margin need to be considered in future studies on the seismic and tsunamigenic hazard assessment for the region.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.tecto.2014.02.001.

Acknowledgments

This work was funded by the Spanish Ministry of Innovation and Science (Projects: CTM2006-13666 and CGL2010-17715), by the U.S. Geological Survey Coastal and Marine Geology Program and by the Dirección General de Minería of the Dominican Republic. Particular thanks to Santiago Muñoz, Director of the Servicio Geológico Nacional of Dominican Republic, and to Pacual Prota, President of the Autoridad Nacional de Asuntos Marítimos of Dominican Republic, for their engagement and contribution to the CARIBENORTE project. We are indebted to the crew and technicians of the R/V Hespérides, R/V Ron Brown and the Dominican Navy Ship Orion for their professional help at sea. The professional work of the technicians of the Unidad de Tecnología Marina is greatly appreciated. Most of the seismic profiles were provided by The Academic Seismic Portal (UTIG), the National Archive of Marine Seismic Surveys (USGS), the Marine Geoscience Data Management System (LDEO), the Woods Hole Oceanographic Institution (WHOI), the National Geophysical Data Center (NOAA) and the Institut Français de Recherche pour l'Exploitation de la Mer (IFREMER). We thank the National Aeronautics and Space Administration (NASA) for the SRTM90 data used in the onshore regions. We are grateful to Manuel Roman Alpiste and Claudia Flores for technical assistance and seismic data processing, and to Bill Danforth and Brian Andrews for swath data processing and mapping. We thank to two anonymous reviewers for their helpful comments on the manuscript.

References

Andrews, B.D., ten Brink, U.S., Chaytor, J.D., Danforth, W., Granja Bruña, J.L., 2010. Marine geology and tsunami hazard assessment of the northeast Caribbean plate boundary. Annual Meeting of the Association of American Geographers 2010. Poster Session 4445: Coastal, Cryosphere, energy, Global Change, Hazards. Association of American Geographers, Washington D.C.

Bakun, W.H., Flores, C.H., ten Brink, U.S., 2012. Significant Earthquakes on the Enriquillo Fault System, Hispaniola, 1500–2010: implications for seismic hazard. Bull. Seismol. Soc. Am. 102, 18–30.

Benford, B., DeMets, C., Calais, E., 2012. GPS estimates of microplate motions, northern Caribbean: evidence for a Hispaniola microplate and implications for earthquake hazard. Geophys. J. Int. 191, 481–490.

Bien-Aime Momplaisir, R., 1986. Contribution à l'étude géologique de la partie orientale du massif de la Hotte (Presqu'île du sud d'Haïti). Synthèse structurale des marges de la Presqu'île à partir de données sismiques. (Thèse) de l'Université Paris 6 (223 pp.).

Biju-Duval, B., Bizon, G., Mascle, A., Muller, C., 1982. Active margin processes; field observations in southern Hispaniola. Am. Assoc. Pet. Geol. Mem. 34, 325–344.

Bowland, C.L., 1993. Depositional history of the western Colombian Basin, Caribbean Sea, revealed by seismic stratigraphy. Geol. Soc. Am. Bull. 105, 1321–1345.

Brown, K.M., Westbrook, G.K., 1987. The tectonic fabric of the Barbados ridge accretionary complex. Mar. Pet. Geol. 4, 71–81.

Byrne, D.B., Suarez, G., McCann, W.R., 1985. Muertos Trough subduction, microplate tectonics in the northern Caribbean? Nature 317, 420–421.

Calais, E., Mazabraud, Y., Mercier de Lepinay, B., Mann, P., Mattioli, G., Jansma, P., 2002. Strain partitioning and fault slip rates in the north-eastern Caribbean from GPS measurements. Geophys. Res. Lett. 29, 1856. http://dx.doi.org/10.1029/2002GL015397.

Calais, E., Freed, A., Mattioli, G., Amelung, F., Jónsson, S., Jansma, P., Hong, S.-H., Dixon, D., Prépetit, C., Momplaisir, R., 2010. Transpressional rupture of an unmapped fault during the 2010 Haiti earthquake. Nat. Geosci. 3, 794–799. http://dx.doi.org/10.1038/NGEO992.

Calassou, S., Larroque, C., Mallavieille, J., 1993. Transfer zones of deformation in thrust wedges: an experimental study. Tectonophysics 221, 325–344.

Carbó-Gorosabel, A., Córdoba, D., Martín Dávila, J., ten Brink, U., Herranz, P., Von Hilldebrandt, C., Payero, J., Muñoz Martín, A., Pazos, A., Catalán, M., Granja Bruña, J.L., Gómez, M., GEOPRICO-DO Working Group, 2005. New marine geophysical survey explores active tectonics at north-eastern Caribbean Plate Boundary. EOS Trans. Am. Geophys. Union 86 (51), 537–540 (20Dec2005).

Carbó-Gorosabel, A., Córdoba-Barba, D., Martín-Dávila, J., Granja-Bruña, J.L., Llanes Estrada, P., Muñoz-Martín, A., ten Brink, U.S., 2010. Exploring Active Tectonics in the Dominican Republic. (ISSN: 0096-3941) 91 (30), 261–268 (27July2010).

Case, J.E., Holcombe, T.L., 1980. Geologic-Tectonic map of the Caribbean region: Reston, Virginia, USA, U.S. Geological Survey Miscellaneous Investigations Series Map I–1100, scale: 1:2500000.

Case, J.E., MacDonald, W.D., Fox, P.J., 1990. Caribbean crustal provinces; seismic and gravity evidence. In: Dengo, G., Case, J.E. (Eds.), The Geology of North America. The Caribbean Region, H. Geol. Soc. Am. Boulder, Colorado, pp. 15–36.

Chaytor, J.D., ten Brink, U.S., 2010. Extension in Mona Passage, Northeast Caribbean. Tectonophysics 493 (1–2), 74–92.

Cooey, J.C., 1982. Structure and stratigraphy of the offshore margin of the southern Dominican Republic. (M.S. thesis) University of Southern Mississippi, Hattiesburg, Mississippi (53 p.).

Dahlen, F.A., 1990. Critical taper model of fold-and-thrust belts and accretionary wedges. Annu. Rev. Earth Planet. Sci. 18, 55–99.

DeMets, C., Gordon, R.G., Argus, D.F., 2010. Geologically current plate motions. Geophys. J. Int. 181, 1–80. http://dx.doi.org/10.1111/j.1365-246X.2009.04491.x.

Diebold, J.B., 2009. Submarine volcanic stratigraphy and the Caribbean LIP's formational environment. In: James, K.H., Lorente, M.A., Pindell, J.L. (Eds.), The Origin and Evolution of the Caribbean Plate. Geological Society, London, Special Publications, 328, pp. 799–808. http://dx.doi.org/10.1144/SP328.0.

Diebold, J.B., Driscoll, N.W., EW9501 Team, 1999. New insights on the formation of the Caribbean basalt province revealed by multichannel seismic images of volcanic structures in the Venezuelan Basin, in: Hsü, K.J., (Series Ed.). Sedimentary Basins of the World, 4. Caribbean Basins. Mann, P. (Ed), Elsevier Science, N. Y. 561–589.

Diebold, J.B., Stoffa, P.L., Buhl, P., Truchan, M., 1981. Venezuelan Basin crustal structure. J. Geophys. Res. 86 (B9), 7901–7923.

Dillon, W.P., Edgar, N.T., Scanlon, K.M., Coleman, D.F., 1996. A review of the tectonic problems of the strike–slip northern boundary of the Caribbean Plate and examination by Gloria. Geology of the United States' Seafloor: The view from GLORIA., 9 135–164.

Dolan, J., Mann, P., De Zoeten, R., Heubeck, C., Shiroma, J., 1991. Sedimentologic, stratigraphic, and tectonic synthesis of Eocene-Miocene sedimentary basins, Hispaniola and Puerto Rico. In: Mann, P., Draper, G., Lewis, J.F. (Eds.), Geologic and Tectonic Development of the North America–Caribbean Plate Boundary in Hispaniola. Geological Society of America, Special Papers, 262, pp. 17–264.

Dolan, J.F., Mullins, H.T., Wald, D.J., 1998. Active tectonics of north-central Caribbean: oblique collision, strain partitioning, and opposing subducted slabs. In: Dolan, J., Mann, P. (Eds.), Active Strike–slip and Collisional Tectonics of the Northern Caribbean Plate Boundary Zone. Geological Society of America, Special Paper, 326, pp. 1–62.

Donnelly, T.W., 1973. Magnetic anomaly observations in eastern Caribbean Sea. In: Edgar, N.T., Saunders, J.B., et al. (Eds.), Initial Reports of de DSDP 15. U. S. Government Printing Office, Washington, pp. 1023–1029.

Driscoll, N.W., Diebold, J.B., 1999. Tectonic and stratigraphic development of the eastern Caribbean: new constraints from multichannel seismic data, in: Hsü, K.J., (Series Ed.). Sedimentary Basins of the World, 4. Caribbean Basins. Mann, P. (Ed), Elsevier Science, N. Y. 591–626.

Driscroll, N.W., Diebold, J.B., 1998. Deformation of the Caribbean region: one plate or two? Geology 26 (11), 1043–1046.

Edgar, N.T., Ewing, J.I., Hennion, J., 1971. Seismic refraction and reflection in Caribbean Sea. Am. Assoc. Pet. Geol. Bull. 55, 833–870.

Edgar, N.T., Saunders, J.B., et al., 1973. Initial reports of the Deep Sea Drilling Project. D. C. U. S. Govt, Washington (v. 15, 1137 p.).

Ewing, J., Antonie, J., Ewing, M., 1960. Geophysical measurements in the western Caribbean Sea and in the Gulf of Mexico. J. Geophys. Res. 65, 4087–4126.

Ewing, J., Talwani, M., Ewing, M., Edgar, T., 1967. Sediments of the Caribbean. Miami Univ. Stud. Trop. Oceanogr. 5, 88–102.

Fox, P.J., Heezen, B.C., 1975. Geology of the Caribbean crust. In: Nairn, A.E.M., Stehli, T.G. (Eds.), The Ocean Basins and Margins, V 3, The Gulf of Mexico and the Caribbean. Plenum Press, New York, pp. 421–426.

Fox, P.J., Ruddiman, W.F., Ryan, W.B.F., Heezen, B.C., 1970. The geology of the Caribbean crust, 1: Beata Ridge. Tectonophysics 10, 495–513.

Granja Bruña, J.L., ten Brink, U.S., Carbó-Gorosabel, A., Muñoz-Martín, A., Gómez Ballesteros, M., 2009. Morphotectonics of the central Muertos thrust belt and Muertos Trough (northeastern Caribbean). Mar. Geol. (ISSN: 0025-3227) 263, 7–33. http://dx.doi.org/10.1016/j.margeo.2009.03.010.

Granja Bruña, J.L., Muñoz-Martín, A., ten Brink, U.S., Carbó-Gorosabel, A., Llanes Estrada, P., Córdoba-Barba, D., Martín-Dávila, J., Catalán Morollón, M., 2010. Gravity modeling of the Muertos Trough and tectonic implications (NE Caribbean). Mar. Geophys. Res. (ISSN: 0025-3235) 31 (4), 263–283. http://dx.doi.org/10.1007/s11001-010-9107-8.

Hernáiz-Huerta, P.P., Diaz de Neira, J.A., García-Senz, J., Deschamps, I., Genna, A., Nicole, Lopera, E., Escuder Viruete, J., Ardévol Oró, Ll, Pérez-Estaún, A., 2007. La estructura del suroeste de la República Dominicana: un ejemplo de deformación en régimen transpresivo. Bol. Geol. Min. 118 (2), 337–357.

Heubeck, C., Mann, P., 1991. Structural geology and Cenozoic tectonic history of the southeastern termination of the Cordillera Central, Dominican Republic. Geol. Soc. Am. Spec. Pap. 262, 315–336.

Heubeck, C., Mann, P., Dolan, J., Monechi, S., 1991. Diachronous uplift and recycling of sedimentary basins during Cenozoic tectonic transpression; northeastern Caribbean plate margin. Sediment. Geol. 70, 1–32.

Holcombe, T.L., Ladd, J.W., Westbrook, G., Edgar, T., Bowland, C.L., 1990. Caribbean marine geology and basins of the plate interior. In: Dengo, G., Case, J.E. (Eds.), The Geology of North America, Vol. H, The Caribbean Region. Geological Society of America, Boulder, Colorado, pp. 231–260.

Hornbach, M.J., Braudy, N., Briggs, R.W., Cormier, M.-H., Davis, M.B., Diebold, J.B., Dieudonne, N., Douilly, R., Frohlich, C., Gulick, S.P.S., Johnson III, H.E., Mann, P., McHugh, C., Ryan-Mishkin, K., Prentice, C.S., Seeber, L., Sorlien, C.C., Steckler, M.S., Symithe, S.J., Taylor, F.W., Templeton, J., 2010. High tsunami frequency as a result of combined strike–slip faulting and coastal landslides. Nat. Geosci. 3, 783–788.

Jansma, P.E., Mattioli, G.S., 2005. GPS results from Puerto Rico and the Virgin Islands: constraints on tectonic setting and rates of active faulting. In: Mann, P. (Ed.), Active Tectonics and Seismic Hazards of Puerto Rico, the Virgin Islands, and Offshore Areas. Spec. Pap. Geol. Soc. Am., 385, pp. 13–30.

Jany, I., 1989. Neotectonique au sud des Grandes Antilles, collision (ride de Beata, Presqu'ile de Bahoruco); subduction (fosse de Muertos); transtension (passage d'Anegada). (Ph.D. thesis) Université Pierre-et-Marie-Curie (Paris-VI), Paris (306 p.).

Kroehler, M.E., Mann, P., Escalona, E., Christeson, G.L., 2011. Late Cretaceous-Miocene diachronous onset of back thrusting along the South Caribbean deformed belt and its importance for understanding processes of arc collision and crustal growth. Tectonics 30 (6), 1–31. http://dx.doi.org/10.1029/2011TC002918.

Ladd, W.J., Worzel, J.L., Watkins, J.S., 1977. Multifold Seismic Reflections Records from the Northern Venezuela Basin and the North Slope of Muertos Trench. Marine Science Institute, University of Texas 41–56.

Ladd, J.W., Shih, Tai-Chang, Tsai, C.J., 1981. Cenozoic tectonics of central Hispaniola and adjacent Caribbean Sea. Am. Assoc. Pet. Geol. Bull. 65, 466–489.

Larue, D.K., Ryan, H.F., 1998. Seismic reflection profiles of the Puerto Rico Trench: shortening between the North American and Caribbean plates. In: Lidiak, E.G., Larue, D.K. (Eds.), Tectonics and Geochemistry of the Northeastern Caribbean. Geol. Soc. Am., Special Paper, 322, pp. 193–210.

Leroy, S., 1995. Structure et origine de la plaque Caribe. Implications geodynamiques. (These) de l'Universite Paris 6 (249 p.).

Leroy, S., Mauffret, A., Patriat, P., Mercier de Lepinay, B., 2000. An alternative interpretation of the Cayman trough evolution from a re-identification of magnetic anomalies. Geophys. J. Int. 141, 539–557. http://dx.doi.org/10.1046/j.1365-246x. 2000.00059.x.

Llanes Estrada, P., ten Brink, U., Granja Bruña, J.L., Carbó-Gorosabel, A., Flores, C., Villaseñor, A., Martín Dávila, J., Pazos, A., 2012. True subduction vs. underthrusting of the Caribbean plate beneath Hispaniola, northern Caribbean. American Geophysical Union Fall Meeting EOS, Fall Meet. Suppl., Abstract T13A-2567.

Manaker, D., Calais, E., Freed, A., Ali, T., Przybylski, P., Mattioli, G.S., Jansma, P.E., Pre'petit, C., deChabalier, J.B., 2008. Interseismic plate coupling and strain partitioning in the Northeastern Caribbean. Geophys. J. Int. 174 (3), 889–903.

Mann, P., Burke, K., 1984. Neotectonics of the Caribbean. Rev. Geophys. Space Phys. 22 (4), 309–362.

Mann, P., Draper, G., Lewis, J., 1991. Overview of the geologic and tectonic development of Hispaniola. In: Mann, P., Draper, G., Lewis, J.F. (Eds.), Geologic and Tectonic Development of the North America-Caribbean Plate Boundary in Hispaniola. Geol. Soc. Am., Special Paper, 262, pp. 1–28.

Mann, P., Taylor, F.W., Edwards, R.L., Ku, T.L., 1995. Actively evolving microplate formation by oblique collision and sideways motion along strike–slip faults: an example from north-western Caribbean plate margin. Tectonophysics 246, 1–69.

Mann, P., Calais, E., Ruegg, J.-C., DeMets, C., Jansma, P.E., 2002. Oblique collision in the north-eastern Caribbean from GPS measurements and geological observations. Tectonics 2 (6), 1057. http://dx.doi.org/10.1029/2001TC0011304,2002.

Marshak, S., 2004. Salients, recesses, arcs, oroclines, and syntaxes. A review of ideas concerning the formation of map-view curves in fold-thrust belts. In: McClay, K.R. (Ed.), Thrust Tectonics and Hydrocarbons Systems. American Association of Petroleum Geologists Memoir, 82, pp. 131–156.

Masson, D.G., Scanlon, K.M., 1991. The neotectonic setting of Puerto Rico. Geol. Soc. Am. Bull. 103, 144–154.

Matthews, J., Holcombe, T., 1976. Possible Caribbean underthrusting of the Greater Antilles along Muertos Trough. Transactions of the 7th Caribbean Geological Conference (Guadeloupe, 1974), pp. 235–242.

Matthews, J., Holcombe, T., 1985. Venezuelan Basin of the Caribbean Sea—stratigraphy and sediment distribution. Mar. Geol. 68, 1–23.

Mauffret, A., Leroy, S., 1997. Seismic stratigraphy and structure of the Caribbean igneous province. Tectonophysics 283 (1–4), 61–104.

Mauffret, A., Leroy, S., 1999. Neogene intraplate deformation of the Caribbean plate at the Beata Ridge, in: Hsü, K.J., (Series Ed.). Sedimentary Basins of the World, 4. Caribbean Basins. Mann, P. (Ed.), Elsevier Science, N. Y. 627–669.

Mauffret, A., Leroy, S., Vila, J.-M., Hallot, E., Mercier de Lepinay, B., Duncan, R.A., 2001. Prolonged magmatic and tectonic development of the Caribbean Igneous Province revealed by a diving submersible survey. Mar. Geophys. Res. 22, 17–45.

Maurasse, F., Husler, G., Georges, G., Schmitt, R., Damond, P., 1979. Upraised Caribbean Sea Floor below acoustic reflector B″ and the Southern Peninsula of Haïti. Geol. Minj. 8, 71–83.

McClay, K.R., 1992. Glossary of thrust tectonics terms. In: McClay, K.R. (Ed.), Thrust Tectonics. Chapman and Hall, pp. 419–434.

McHugh, C., Seeber, L., Braudy, N., Cormier, M.-H., Davis, M.B., Diebold, J.B., Dieudonne, N., Douilly, R., Gulick, S.P.S., Hornbach, M.J., Johnson III, H.E., Ryan-Mishkin, C.S., Sorlien, C.C., Steckler, M.S., Symithe, S.J., Templeton, J., 2011. Offshore sedimentary effects of the 12 January 2010 Haiti earthquake. Geology 39 (8), 723–726.

Mercier de Lépinay, B., Mauffret, A., Jany, I., Bouysse, P., Mascle, A., Renard, V., Stephan, J.F., Hernandez, E., 1988. Une collision oblique sur la bordure nord-caraïbe à la jonction entre la ride de Beata et la fosse de Muertos. C. R. Acad. Sci. Paris 307 (Serie II), 1289–1296.

Mercier de Lépinay, B., Deschamps, A., Klingelhoefer, F., Mazabraud, Y., Delouis, B., Clouard, V., Hello, Y., Crozon, J., Marcaillou, B., Graindorge, D., Vallée, M., Perrot, J., Bouin, M.-P., Saurel, J.-M., Charvis, P., St-Louis, M., 2011. The 2010 Haiti earthquake: a complex fault pattern constrained by seismologic and tectonic observations. Geophys. Res. Lett. 38, l22305. http://dx.doi.org/10.1029/2011gl049799.

Moore, G.T., Falhquist, D.A., 1976. Seismic profile tying Caribbean DSDP Sites 153, 151, and 152. Geol. Soc. Am. Bull. 87, 1609–1614.

Pindell, J.L., Kennan, L., 2009. Tectonic evolution of the Gulf of Mexico, Caribbean and northern South America in the mantle reference frame: an update. In: James, K.H., Lorente, M.A., Pindell, J.L. (Eds.), The Origin and Evolution of the Caribbean Plate. Geological Society of London, Special Publications, 328, pp. 1–55.

Prentice, C.S., Mann, P., Crone, A.J., Gold, R.D., Hudnut, K.W., Briggs, R.W., Koehler, R.D., Jean, P., 2010. Seismic hazard of the Enriquillo–Plantain Garden fault in Haiti inferred from palaeoseismology. Nat. Geosci. 3, 789–793. http://dx.doi.org/10.1038/NGEO991.

Pubellier, M., Mauffret, A., Leroy, S., Vila, J., Amilcar, H., 2000. Plate boundary readjustment in oblique convergence: example of the Neogene of Hispaniola, Greater Antilles. Tectonics 19, 630–648.

Ranero, C.R., von Huene, R., 2000. Subduction erosion along the Middle America convergent margin. Nature 404, 748–752.

Révillon, S., Hallot, E., Ardnt, N.T., Chauvel, C., Duncan, R.A., 2000. A complex history for the Caribbean plateau: petrology, geochemistry and geochronology of the Beata Ridge, southern Hispaniola. J. Geol. 108, 641–661.

Rodgers, B.K., 1991. Tech service job No. CD50708. Biostratigraphic, ecologic and lithologic descriptions for the S. A. Petrolera Las Mercedes San Pedro 1 and San Andres 1 wells, Dominican Republic. Mobil Exploration and Producing Services, Inc.(report).

Roemer, L.B., 1973. Geology and geophysics of the Beata Ridge, Caribbean Sea [MS thesis]: College Station. Texas A&M University, Texas (93 p.).

Scholl, D., von Henue, R., Vallier, T., Howell, D., 1980. Sedimentary masses and concepts about tectonic processes at underthrust ocean margins. Geology 8, 564–568.

Seely, D., 1979. The evolution of structural highs bordering major forearc basins. In: Watkins, J.S., Montadert, L., Dickerson, P.W. (Eds.), Geologic and Geophysical Investigations of Continental Margins. Am. Assoc. Pet. Geol. Mem., 29, pp. 45–260.

Sen, G.R., Hickey-Vargas, R., Waggoner, G., Maurasse, F., 1988. Geochemistry of basalts from the Dumisseau Formation, southern Haiti: implication for the origin of the Caribbean crust. Earth Planet. Sci. Lett. 87, 423437.

Shipley, T., Gahagan, L., Johnson, K., Davis, M. (Eds.), 2005. Seismic Data Center. Institute for Geophysics, University of Texas (http://www.ig.utexas.edu/sdc/).

Silver, E.A., Galewsky, J., McIntosh, K.D., 1995. Variation in structure, style, and driving mechanism of adjoining segments of the North Panama deformed belt. Geol. Soc. Am. Spec. Pap. 295, 225–233.

Smith, W.H.F., Sandwell, D.T., 1997. Global seafloor topography from satellite altimetry and ship depth soundings. Science 277, 1957–1962.

Smith, G., McNeill, L., Henstock, T.J., Bull, J., 2012. The structure and fault activity of the Makran accretionary prism. J. Geophys. Res. 117, B07407. http://dx.doi.org/10.1029/2012JB009312.

Stoffa, E.L., Mauffret, A., Truchan, M., Buhl, E., 1981. Sub-B″ layering in the southern Caribbean: the Aruba Gap and Venezuela Basin. Earth Planet. Sci. Lett. 53, 131–146.

Stolt, R.H., 1978. Migration by Fourier transform. Geophysics 43, 23–48.

ten Brink, U.S., López-Venegas, A., 2012. Plate interaction in the NE Caribbean subduction zone from continuous GPS observations. Geophys. Res. Lett. 39, L10304. http://dx.doi.org/10.1029/2012GL051485.

ten Brink, U.S., Marshak, S., Granja Bruña, J.L., 2009. Bivergent thrust wedges surrounding island arcs: insights from observations and sandbox models of the northeastern Caribbean plate. Geol. Soc. Am. Bull. 121 (11/12), 1522–1536. http://dx.doi.org/10.1130/B26512.1. ISSN 0016-7606.

ten Brink, U.S., Bakun, W.H., Flores, C.H., 2011. Historical perspective on seismic hazard to Hispaniola and the northeast Caribbean region. J. Geophys. Res. 116, b12318. http://dx.doi.org/10.1029/2011JB008497.

van den Berghe, B., 1983. Evolution sedimentaire et structurale depuis le Paleocene du secteur Massif de la Selle (Haiti)-Bahoruco (Republique Dominicaine)-Nord de la Ride de Beata dans l'orogene nord Caraibe (Hispaniola-Grandes Antilles). (Ph.D.) Université Pierre-et-Marie-Curie (No. 83-36, 205 p.).

Vila, J.-M., Jany, I., Lepvrier, C., Feinberg, H., Mauffret, A., 1990. Mise en évidence de l'âge post-pliocène inférieur de la collision entre la ride de Beata et l'orogène nord-caraïbe (Grandes Antilles). C. R. Acad. Sci. Paris 311 (Série II), 1359–1366.

von Huene, R., Lallemand, S., 1990. Tectonic erosion along the Japan and Peru convergent margins. Geol. Soc. Am. Bull. 102, 704–720.

Wessel, P., Smith, W.H.F., 1998. New improved version of Generic Mapping Tools released. EOS Trans. Am. Geophys. Union 79 (47), 579.

White, P.M., 1993. Structural evolution of the San Pedro basin: implications for the paleogeographic position of the Dominican Republic and Puerto Rico in a plate-tectonic framework. University of Texas, Arlington (M.S. thesis: 46 p.).

Wynn, R.B., Stow, D.A.V., 2002. Classification and characterization of deep-water sediment waves. Mar. Geol. 192, 7–22.