

# Active tectonics, Quaternary stress regime evolution and seismotectonic faults in southern central Hispaniola: implications for the quantitative seismic hazard assessment

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## Abstract

Present-day convergence between Caribbean and North American plates is accommodated by subduction zones, major active thrusts and strike-slip faults, which are probably the source of the historical large earthquakes on Hispaniola. However, little is known of their geometric and kinematic characteristics, slip rates and seismic activity over time. This information is important to understand the active tectonics in Hispaniola, but it is also crucial to estimate the seismic hazard in the region.

Here we show that a relatively constant NE-directed shortening controlled the geometry and kinematics of main active faults in southern central Hispaniola, as well as the evolution of the Quaternary stress regime. This evolution included a pre-Early Pleistocene D1 event of NE-trending compression, which gave rise to the large-scale fold and thrust structure in the Cordillera Central, Peralta Belt, Sierra Martín García and San Juan-Azua basin. This was followed by a near pure strike-slip D2 stress regime, partitioned into the N-S to NE-SW transverse Ocoa-Bonao-La Guácara and Beata Ridge fault zones, as well as subordinate structures in related sub-parallel deformation corridors. Shift to D2 strike-slip deformation was related to indentation of the Beata Ridge in southern Hispaniola from the Early to Middle Pleistocene and continues today. D2 was locally coeval by a more heterogeneous and geographically localized D3 extensional deformation. Defined seismotectonic fault zones divide the region into a set of simplified seismogenic zones as starting point for a seismic hazard modeling. Highest peak ground acceleration values computed in the Ocoa Bay establish a very high seismic hazard.

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**Active tectonics, Quaternary stress regime evolution and seismotectonic faults**  
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**assessment**

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**Key Points:**

- Active faults in central southern Hispaniola are controlled by NE-directed shortening
- Quaternary stress regime evolution includes a compressional D1 followed by a strike-slip D2, locally coeval with an extensional D3
- Modeling establishes a very high seismic hazard zone centered in the Ocoa Bay

**Abstract**

Present-day convergence between Caribbean and North American plates is accommodated by subduction zones, major active thrusts and strike-slip faults, which are probably the source of the historical large earthquakes on Hispaniola. However, little is known of their geometric and kinematic characteristics, slip rates and seismic activity over time. This information is important to understand the active tectonics in Hispaniola, but it is also crucial to estimate the seismic hazard in the region.

Here we show that a relatively constant NE-directed shortening controlled the geometry and kinematics of main active faults in southern central Hispaniola, as well as the evolution of the Quaternary stress regime. This evolution included a pre-Early Pleistocene D1 event of NE-trending compression, which gave rise to the large-scale fold and thrust structure in the Cordillera Central, Peralta Belt, Sierra Martín García and San Juan-Azua basin. This was followed by a near pure strike-

slip D2 stress regime, partitioned into the N-S to NE-SW transverse Ocoa-Bonao-La Guácara and Beata Ridge fault zones, as well as subordinate structures in related sub-parallel deformation corridors. Shift to D2 strike-slip deformation was related to indentation of the Beata Ridge in southern Hispaniola from the Early to Middle Pleistocene and continues today. D2 was locally coeval by a more heterogeneous and geographically localized D3 extensional deformation. Defined seismotectonic fault zones divide the region into a set of simplified seismogenic zones as starting point for a seismic hazard modeling. Highest peak ground acceleration values computed in the Ocoa Bay establish a very high seismic hazard.

## **1. Introduction**

The present-day convergence between the Caribbean and North American plates is partially accommodated within Hispaniola Island (Dominican Republic and Haiti). As consequence, this region contains a number of major active faults with lengths of up to several hundred kilometres that stand out as geomorphological and tectonic features (Mann et al., 1991a, 2002; Hernaiz Huerta & Pérez-Estaún, 2002; Calais et al., 2016) and have been the source of some of the largest earthquakes documented in the past 250 years (Calais et al., 2010; Prentice et al., 2010; Bakun et al., 2012; Mercier de Lépinay, 2011). For example, the strike-slip system of the Enriquillo-Plantain Garden Fault Zone (EPGFZ) produced the 12 January 2010  $M_w$  7.0 Leogâne and 14 August 2012  $M_w$  7.2 Nippes earthquakes, as well as probably other historical earthquakes in 1770 (estimated  $M_w$  7.5; SISFRANCE Antilles, 2009) and 1860 (estimated  $M_w$  6.3; Bakun et al. 2012).

The current tectonics in this densely populated and rapidly developing region are broadly understood (e.g., Mann et al., 1995, 2002; Hernaiz Huerta & Pérez-Estaún, 2002; Pérez-Estaún et al., 2007; Granja Bruña et al., 2014; Calais et al., 2016; Courbeau et al., 2016; Wang et al., 2018; Saint Fleur et al., 2019; Escuder-Virujete & Pérez, 2020), with subduction zones, major active thrusts and oblique strike-slip faults identified in several clearly defined zones. However, little is known of detailed

geometric and kinematic characteristics, Quaternary offsets and slip rates on these active fault systems, and its seismic activity over time is poorly constrained by geochronological data. This information is essential for understanding the present-day tectonics of Hispaniola, but it is also crucial to establish an inventory of potential seismogenic sources, a seismotectonic zonation model, and an estimate of the seismic hazard in the region.

The focus of this study is threefold: (1) to characterize the active deformation pattern and the main seismotectonic faults in southern central Hispaniola, based on tectonic and geomorphologic field observations, complemented with regional gravimetric, magnetic and seismic data analysis; (2) to determine the Quaternary to present-day stress regime evolution by inversion of geologically determined slip vectors on minor and major faults throughout the zone; and (3) to divide the region into a set of simplified seismogenic zones as a starting point for a subsequent quantitative seismic hazard modeling in terms of the peak ground acceleration (PGA).

## **2. Geological setting**

### *2.1. From intra-oceanic subduction to arc-continent collision and subduction polarity reversal in the northern Caribbean plate*

Located on the northern edge of the Caribbean plate, the Hispaniola Island is a tectonic collage produced by the SW-dipping Cretaceous subduction to Eocene oblique collision of the Caribbean intra-oceanic arc with the southern continental margin of North America (Mann et al., 1991a; Draper et al., 1994; Pérez-Estaún et al., 2007; Escuder-Viruete et al., 2011a, 2011b, 2013, 2016b). The main consequence of the arc-continent collision were the blocking of the suture zone, the transfer of deformation to the back-arc region in southern Hispaniola and subduction polarity reversal, with renewed subduction beginning along a new NE-dipping subduction zone (e.g., Kroehler et al., 2011).

The geodynamic modeling software *GPlates* V2.3 (Müller et al., 2018; [www.gplates.org](http://www.gplates.org)), regional bibliographic data, and observations presented in this paper allows the reconstruction of this relative



79 movement between the Caribbean and North American plates, distinguishing three main stages of  
80 evolution from the lower Campanian to the present-day. In the lower Campanian (80 Ma), the Lower  
81 Cretaceous Caribbean island-arc is moving to the northeast at a rate of about 4-5 cm/yr (Fig. 1a). By  
82 this time, the Caribbean arc had overridden the Galapagos hotspot, giving rise to a period of vigorous  
83 submarine oceanic plateau volcanism that began as early as 139 Ma and was widespread by 88 Ma,  
84 building the Caribbean Large Igneous Province (CLIP). In a SW-dipping subduction zone, the intra-  
85 oceanic arc consumed through subduction the large area occupied by the proto-Caribbean oceanic  
86 crust in the current central part of the Caribbean, and obliquely collided with the Maya block of the  
87 southeastern margin of the North American plate (Mann et al., 2007).

88 By the middle Eocene (40 Ma) the collision and suturing of the Caribbean island-arc against North  
89 America had halted the forward motion of the oceanic arc (Fig. 1b). Previous convergence and arc-  
90 continent collision caused the extinction of the arc volcanism, the emplacement of supra-subduction  
91 zone ophiolites and developed a zone of NE-directed foreland thrusting on the lower plate (Mann et  
92 al., 1991a; Pindell & Kennan, 2009; Pérez-Estaún et al., 2007; Escuder-Viruete et al., 2013, 2016b).  
93 Because the arc can no longer subduct the more continental crust of North America, convergence  
94 was accommodated by back-thrusting and the initiation of a new, NE-dipping subduction zone  
95 (Kroehler et al., 2011). In southern Hispaniola, the arc-continent collision in the middle to upper  
96 Eocene resulted in a reversal of subduction polarity in the back-arc region, as well as deformation by  
97 back-thrusting at the resulting Peralta-Muertos accretionary prism in the upper plate (Witschard &  
98 Dolan, 1990; Mann et al., 1991b; Dolan et al., 1991; Hernáiz-Huerta & Pérez-Estaún, 2002). The  
99 new subduction zone separated the Venezuela basin from the upper island-arc crust of central  
100 Hispaniola.

101 Since the lower Miocene, the ENE movement of the Caribbean plate gave rise to the oblique  
102 collision/accretion of the northern sector of the Caribbean oceanic plateau of transitional crust with  
103 the Peralta accretionary prism, resulting in the formation of the SW-directed Haitian-Neiba fold-and-

thrust belt and the San Juan-Azua basins in the foreland (Fig. 1c; Mercier de Lépinay, 1987; Witschard & Dolan, 1990; Dolan et al., 1991; Heubeck & Mann, 1991; Pubellier et al., 2000; Hernáiz-Huerta & Pérez-Estaún, 2002; Granja Bruña et al., 2014; Escuder-Virujete et al., 2023). During the Pliocene and until the present-day, convergence gave rise to the accretion of the central sector of the Caribbean oceanic plateau of thickened crust, causing in southern Hispaniola the uplift and folding of the Massif de la Serre-Sierra Batoruco and the tectonic individualization of the Enriquillo-Cul de Sac basin. The blue star and line in the Fig. 1c represents the reconstructed location and path of a point of the Sierra Batoruco as it travels with the Caribbean plate. *GPlates* reconstruction shows that the point started traveling to the NE at a rapid rate of 3-5 cm/yr and changed to a ENE direction in middle Eocene times (45 Ma) with a slow rate of 0.7-2.1 cm/yr. This change coincides with the arc-continent collision in northern Hispaniola and the start of subduction by back-thrusting in the Peralta-Muertos belts of southern Hispaniola.

## *2.2. Present-day geodynamic configuration of southern Hispaniola*

Oblique collision has also led to the fragmentation of the northern Caribbean plate into several microplates, which are Septentrional (or North Hispaniola), Hispaniola-Puerto Rico and Gonâve (Fig. 2; Mann et al., 1995; Calais et al., 2016; Rodríguez-Zurrunero et al., 2019). Microplates are limited by major fault zones that, as reflected by the associated seismicity and modeled displacement rates, accommodate part of the relative movement between them (Mann et al., 2002; Manaker et al., 2008; Benford et al., 2012; Symithe et al., 2015; Calais et al., 2016; Corbeau et al., 2019).

In northern Hispaniola, the Septentrional microplate is a wedge-shaped tectonic forearc sliver, limited offshore to the north by the Northern Hispaniola Fault Zone (NHFZ, or western extension of the Puerto Rico Trench) and onshore to the south by the Septentrional Fault Zone (SFZ; Dolan et al., 1988; Mann et al., 2002; Escuder-Virujete & Pérez, 2020; Escuder-Virujete et al., 2020).

127 The Hispaniola-Puerto Rico microplate is an arc-crust block, limited to the north by the SFZ and to  
128 the south by the Muertos Trough (MT). To the west, the boundary between the Hispaniola block and  
129 Gonâve microplate is generally placed in the San Juan-Los Pozos fault zone (SJPFZ; e.g., Mann et  
130 al., 1991). From N to S (Fig. 2), the Hispaniola microplate includes the Cordillera Central domain  
131 and the Peralta-Muertos deformed belts. The Cordillera Central in the Dominican Republic and its  
132 prolongation in the Massif du Nord in Haiti is an Upper Jurassic-Upper Cretaceous basement  
133 composed mainly of Pacific-derived oceanic units, structurally limited to the northeast by the  
134 Hispaniola fault zone (HFZ) and to the southwest by the SJPFZ (Mann et al., 1991). It comprises  
135 fragments of oceanic lithosphere, plutono-volcanic complexes related to subduction, and basaltic  
136 units related to a mantle-plume magmatism (Escuder-Viruete et al., 2008). It also comprises the  
137 Quaternary intramountain basins of Cibao, Jarabacoa, Bonao, Constanza, Rancho Arriba and San  
138 José de Ocoa (Fig. 3).

139 The Peralta Belt (PB) is a NW-trending and SW-verging fold-and-thrust belt, structurally  
140 sandwiched between the San José-Restauración (SJRFZ) and the San Juan-Pozos fault zones (Mann  
141 et al., 1991b). The belt evolved from a back-arc basin for the Caribbean island-arc, with deposition  
142 of Campanian to Eocene turbidites (Trois Rivières and Las Palmas Formations), through a  
143 transpressive fold-and-thrust belt in the middle Eocene to early Miocene (Mercier de Lépinay, 1987;  
144 Witschard & Dolan, 1990; Dolan et al., 1991; Ramírez, 1995; Hernáiz-Huerta et al., 2007b). In the  
145 SE sector, the lithostratigraphic units of the Peralta Belt have been grouped into three thick  
146 sedimentary sequences, separated from each other by major unconformities: the Paleocene-Eocene  
147 Peralta Group; the middle Eocene-early Miocene Río Ocoa Group; and the middle Miocene-  
148 Pleistocene Ingenio Caei Group (Heubeck et al., 1991; Dolan et al., 1991; Díaz de Neira & Solé  
149 Pont, 2002; Hernaiz Huerta & Pérez-Estaún, 2002; Hernáiz-Huerta et al., 2007a; Pérez-Valera,  
150 2010).

151 The Muertos Trough (MT) marks the boundary between the subducting floor of the Caribbean plate  
152 and the overlying deformed belt south of eastern Hispaniola and Puerto Rico (Fig. 2). The thrust  
153 focal mechanism of the June 24, 1984 earthquake ( $M_w$  6.7) indicates subduction along the MT  
154 (Byrne et al., 1985). Seismic reflection profiles across the MT image an N-dipping low-angle thrust  
155 structure, with folded and faulted sedimentary rocks on top forming an accretionary prism (Driscoll  
156 & Diebold, 1998; Dolan et al., 1998; Mauffret & Leroy, 1999; Granja Bruña et al., 2009, 2014).  
157 Convergence across the Muertos Trough probably begins with the end of the arc-continent collision  
158 and the onset of back-thrusting. *GPlates* reconstruction shows that the amount of oblique subduction  
159 of the Caribbean plate beneath southern Hispaniola at the westernmost Muertos Trench has been  
160 about 100 km (Fig. 1).

161 The Gonâve microplate is limited to the north by the Oriente Fault Zone and to the south by the  
162 Enriquillo-Plantain Garden fault zone (EPGFZ). The nature of the crust of the Gonâve microplate is  
163 not well established, having been proposed as a Cretaceous-Eocene remnant arc (Heubeck et al.,  
164 1991), back-arc (Mann et al., 1991a) or the rifted crust of the eastern Cayman continental passive  
165 margin (Corbeau et al., 2017, 2019). From the lower Miocene onwards, the oblique  
166 collision/accretion processes have formed the Haitian-Neiba belt in the microplate. It consists of a  
167 NW-trending and SW-verging fold-and-thrust belt, bounded to the north by the NE-dipping SJPFZ  
168 (Mann et al., 1995; Ramírez, 1995; Pubellier et al., 2000; Hernáiz Huerta et al., 2007a). The thrust  
169 bounding the NE Montagnes Noires-Sierra de Neiba thrust sheet was activated during the lower-  
170 middle Miocene and the thrust bounding the SW Chaîne des Matheux-southern Sierra de Neiba  
171 thrust sheet was developed from the middle-upper Miocene (Pubellier et al., 2000). Tilting and  
172 faulting of Quaternary age deposits in the northern margin of the Enriquillo basin indicate that the  
173 thrust belt is still active (Hernáiz Huerta et al. 2007), which is consistent with the seismic activity  
174 recorded by Corbeau et al. (2019).

175 The Beata Ridge is a prominent NE-trending bathymetric high of about 450 km long built on the  
176 Caribbean plate (Fig. 2a; Mauffret & Leroy, 1999; Mauffret et al., 2001; Granja Bruña et al., 2014).  
177 It separates the Haiti basin to the west, from the Venezuelan basin to the east. The width of this  
178 structure in the zone of interaction with the Hispaniola-Puerto Rico microplate is about 120 km. The  
179 ridge comprises an unusually thick oceanic crust, essentially made up of mafic igneous rocks  
180 (gabbros, dolerites and basalts) of the Cretaceous CLIP (Sinton et al., 1998; Révillon et al., 2000;  
181 Dürkefalden et al., 2019). The Cretaceous basement of the Sierra Bahoruco and Hotte-Selle Massifs  
182 is made up of submarine volcanic rocks with similar petrological and geochemical characteristics,  
183 indicating that the Southern Peninsula of Hispaniola is an emerged fragment of the CLIP (Escuder-  
184 Viruete et al., 2016a).

185 Onshore/offshore seismic refraction and wide-angle reflection studies indicate a crustal thickness of  
186 26 to 32 km in the Sierra Bahoruco, 20 to 22 km at the Beata Ridge, 10 to 15 km in the Venezuelan  
187 basin and only 5 to 10 km in the Haitian basin (Nuñez et al., 2016, 2019; Kumar et al., 2020). The  
188 crustal thickness of the Beata Ridge is controlled mainly by NNE to NE-striking faults, such as the  
189 Beata Ridge fault zone (BRFZ) that follows its axial trend (Mauffret et al., 2001). According to  
190 Granja Bruña et al. (2014), this fault acted in pre-Neogene times as a normal fault. The BRFZ  
191 continues to the NE, abruptly limiting the Sierra Bahoruco and Sierra Martín García to the southeast,  
192 until it intersects the SJPFZ in the western Ocoa Bay.

193 The NE-directed collision and impingement of the thickened crust of the Beata Ridge with the  
194 Hispaniola microplate in the Neogene gave rise to a recess of the Muertos accretionary prism and the  
195 clockwise rotation of fold and thrust structures of the Peralta Belt east of the Ocoa Bay (Mercier de  
196 Lépinay et al. 1988; Mann et al., 1991b; Ramírez, 1995; Hernaiz Huerta & Pérez-Estaún, 2002;  
197 Granja Bruña et al., 2014). Part of the present-day stress-field induced by the Beata Ridge collision is  
198 accommodated by the Ocoa-Bonao-La Guacara Fault Zone (OBFZ) within the Hispaniola microplate  
199 (Escuder-Viruete et al., 2023). This fault zone is an active, NNE to NE-striking large-scale fault

system that crosses the southern central sector of Hispaniola along more than 250 km (Pérez-Estaún et al., 2007). In the eastern Ocoa Bay, the OBFZ bends the fold and thrust structures of the Peralta Belt, the SJPFZ and the lower Miocene to Early Pleistocene sedimentary fill of the Azua basin (Fig. 2). This fault zone has recently been proposed as the onland transition between oceanic subduction and arc-oceanic plateau collision (Escuder-Virue et al., 2023).

### 3. Methodology

#### 3.1. Bathymetric, gravimetric and magnetic data sets

Maps shown in this study were created using several topographic and bathymetric data sets: (1) digital elevation models (DEMs) from the Shuttle Radar Topography Mission-30 Plus (<https://www2.jpl.nasa.gov/srtm/>) with a resolution of 30 m (Tozel et al., 2019); (2) DEMs obtained from ALOS Global Digital Surface Model with a resolution of 25 m (Takaku et al., 2020); (3) satellite images from *Google Earth Pro*® (<http://earth.google.com>) with a resolution from 1 km to 10 m; and (4) topographic and bathymetric profiles obtained from the GMRT data-set (available with *GeoMapApp*; [www.geomapapp.org](http://www.geomapapp.org); Ryan et al., 2009).

The gravity data are from the compilation by the National Geospatial-Intelligence Agency (<https://www.nga.mil>). This compilation includes data from various campaigns conducted onland between 1939 and 1991 by IFREMER, Royal Astronomical Society, Cambridge University, Lamont-Doherty Geological Observatory and Woods Hole Oceanographic Institution. This compilation contains 3012 stations with a resolution of  $\pm 2$  mGal and a density reduction of  $2.67 \text{ g/cm}^3$ .

The magnetic data come from the magnetic and radiometric flight carried out between 1995 and 1997 by the Compagnie Générale de Géophysique (CGG) in the onland territory of the Dominican Republic, with 500 m of separation of lines and a nominal height of 120 m, in the context of the SYSMIN Program of the EU of *Cartografía geotemática de la República Dominicana* (García

Lobón & Rey Moral, 2004). Data were gridded at 250 m and several procedures were followed to ensure data quality, such as micro-levelling and filtering of frequencies greater than Nyquist.

### 3.2. Regional analysis of potential field data

In this work, lithological, structural, gravimetric and magnetic data were analyzed together to constrain the crustal structure underlying southern central Hispaniola, to establish the distribution of different crustal blocks with specific geophysical properties at the surface and to identify their tectonic boundaries (even when they are hidden by a recent sedimentary cover). The analysis of the potential field data was carried out with the *Geosoft Oasis Montaj*® software. Gravity data has been gridded using a minimum curvature algorithm, resulting a grid cell size of 250 x 250 m.

Magnetic grids were calculated from raw data provided by CGG. Qualitative interpretation of the magnetic data has been made using a reduction technique to the pole of the total magnetic field. This reduction to the pole was calculated assuming a local inclination of 48° N and declination of 10° W. This process removes the inclination effect of the total magnetic field by transforming the anomaly into the vertical component of the field produced as if the source were at the North magnetic pole - 90° inclination. The inclination and declination used in the reduction to the pole were calculated from the International Geomagnetic Reference Field subroutine in *Oasis Montaj*® based on the latitude and longitude. Gravimetric and magnetic grids were interpreted on variations in amplitude, wavelength character, lineament distribution, texture, and structural discontinuities. More details of the methodology of acquisition and analysis of the potential field data are included in García-Lobón & Rey-Moral (2004), García-Lobón & Ayala (2007) and Ayala et al., (2017).

### 3.3. Seismicity

Compilation of an earthquake catalog allowed us to determine the spatial and temporal distribution of earthquakes, their relationships with the main active fault zones, and the characterization of the seismotectonic structures in terms of their seismic parameters. For southern central Hispaniola, the

catalog was compiled by collecting and analyzing historical data (SISFRANCE-Antilles 2009; Bertil et al., 2010; Flores et al. 2011; McCann et al. 2011; ten Brink et al. 2011; Bakun et al., 2012; and references herein) and instrumental/experimental data (Russo & Villaseñor 1995; Alvarez et al. 1999; Tanner & Shepherd 1997; Terrier-Sedan & Bertil, 2021; ISC 2014; RSPR-Puerto Rico catalogue, and USGC-NEIC catalogue). Access the local network, constituted by the *Instituto Sismológico de la Universidad Autónoma de Santo Domingo* (ISU; [uasd.edu.do](http://uasd.edu.do)), the *Observatorio Sismológico Politécnico Loyola* (OSPL; [ospl.ipl.edu.do](http://ospl.ipl.edu.do)) and the Projet Ayiti-Seismes performed by Laboratoire Mixte International CARIBACT ([ayiti.unice.fr/ayiti-seismes](http://ayiti.unice.fr/ayiti-seismes); BME, UEH/FDS/URGéo, ENS and Géoazur partnership), enabled information on earthquakes in ghost areas of the global seismic network to be obtained. Following Bertil et al. (2010, 2015), the catalogue was revised for data quality, duplication, and foreshocks or aftershocks, and then homogenized the magnitudes using the moment magnitude scale ( $M_w$ ). The most contemporary sources describing the historical earthquakes were investigated to establish the focal characteristics of each seismic event.

Due to the high uncertainty related to the magnitude and location of historical earthquakes, the earthquakes in the catalogue were classified into three groups following Terrier-Sedan and Bertil (2021): historical data before 1750 of unreliable magnitude and location; historical data between 1750 and 1960 of intermediate quality; and instrumental data after 1960 generally of high quality. More than 12 epicentres of earthquakes of magnitude at least  $M_w$  6.0 are located in the study area, or the immediate surroundings, including the earthquakes of 1562 (destroyed La Vega and Santiago de los Caballeros), 1615 (destroyed Santo Domingo), 1684 (affected Azua and Santo Domingo), 1691 (destroyed Azua, affected Santo Domingo), 1751 (destroyed Azua) and 1911 (affected San Juan). The catalogue of earthquakes and focal mechanisms compiled for southern central Hispaniola is included in the Supporting Information S1.

### 3.4. U-Th geochronology



271 Fossil coral reefs provide a snapshot of the ocean environment and sea level during their growth. As  
272 uranium is incorporated into the coral calcium carbonate skeleton during its growth, the U-Th  
273 geochronology method is appropriate for dating geological and oceanographic events in the  
274 Quaternary (e.g., Hibbert et al., 2016). To limit the effects of element mobility during diagenesis  
275 and/or any chemical alteration, coral samples were collected in apparently unaltered sectors on top of  
276 fossil wave-cut platforms and paleo-cliffs of the topographically lower terraces, as well as on the  
277 sides of small narrow gorges cut through the higher terraces. Fragments of corals that preserve the  
278 pristine growth structure were selected in the field. Subsequently, thin sections of these fragments  
279 were examined under the petrographic microscope to characterize the crystalline habit, the  
280 mineralogy as indicated by staining for carbonate minerals, and the degree of neomorphism. Coral  
281 fragments with an excellent preservation of the original skeleton and primary pore spaces void of  
282 cement and sediment were selected. Fragments were ground and the resulting powder analyzed by  
283 X-ray diffraction analysis at the IGME Laboratories in Madrid. Samples selected for U-Th dating  
284 yielded more than 95% aragonite during diffraction analysis. This test eliminated most samples of  
285 the topographically higher coral reef terraces.

286 U-series measurements were carried out using a high-resolution multi-collector inductively coupled  
287 plasma mass-spectrometer (ThermoFisher Neptune® multi-collector equipped with an Aridus®  
288 desolvation nebuliser) at the National Centre for Isotope Geochemistry, University College Dublin.  
289 Samples were spiked with a mixed  $^{229}\text{Th}$ - $^{233}\text{U}$ - $^{236}\text{U}$  spike and were slowly dissolved in 7M  $\text{HNO}_3$ .  
290 Following overnight sample-spike equilibration, U and Th fractions were separated using anion ion-  
291 exchange columns using standard methods (Fankhauser et al., 2016).  $^{238}\text{U}/^{236}\text{U}$  and  $^{233}\text{U}/^{236}\text{U}$  isotope  
292 ratios were measured simultaneously using Faraday collectors, and the lower intensity of the  $^{234}\text{U}$  ion  
293 beam was measured using a Faraday collector equipped with a low-noise  $10^{13}$  Ohm resistor. A  
294 mass-fractionation correction was applied to the measured uranium and thorium isotope ratios based  
295 on the certified near-unity  $^{233}\text{U}/^{236}\text{U}$  ratio of the mixed spike (sufficient uranium carry over into the

thorium fraction permitted this for all thorium analyses). For thorium,  $^{230}\text{Th}/^{232}\text{Th}$  and  $^{230}\text{Th}/^{223}\text{Th}$  ratios were measured separately from the uranium runs, using the  $10^{13}$  Ohm equipped Faraday collector in the axial position for  $^{223}\text{Th}$ . A detrital (inherited initial non-radiogenic  $^{230}\text{Th}$ ) correction was applied to all U-Th analyses, assuming that the inherited component had a ( $^{230}\text{Th}/^{232}\text{Th}$ ) ratio of  $0.8 \pm 0.4$  (parentheses denote activity ratios). Whilst the choice of ( $^{230}\text{Th}/^{232}\text{Th}$ ) for the detrital component is somewhat arbitrary, the calculated ages for all are quite insensitive to this correction (Supporting Information S2). All ages were calculated using the following decay constants ( $\lambda^{234}\text{U} = 2.826\text{E-}6$ ,  $\lambda^{230}\text{Th} = 9.1577\text{E-}6$ ,  $\lambda^{238}\text{U} = 1.551\text{E-}10$  and  $\lambda^{232}\text{Th} = 4.9475\text{-}11$ ; Jaffey et al., 1971; Cheng et al., 2000). Laboratory blanks for  $^{238}\text{U}$  and  $^{232}\text{Th}$  were typically in the 1 to 10 pg range, and no blank corrections were applied.

### 3.5. Fault-slip data analysis

Stress or strain can be evaluated using brittle fault-slip data inversion methods (Angelier, 1994). It involves collecting faults data such as plane orientations, fault striae directions and sense of slip from kinematic indicators at the outcrop scale. The methodology of fault-slip data inversion to determine stress fields and to demonstrate temporal and spatial changes in the late Cenozoic stress states has been used in many active tectonic areas around the world over the past three decades (Angelier, 1994; Twiss & Unruh, 1998; C  lerier et al., 2012; and references therein). It consists of obtaining from a population of fault-slip data the principal stress axes that best fit the reduced stress tensor at a given measurement site. Inversion results include the orientation (azimuth and plunge) of the principal stress axes ( $\sigma_1 > \sigma_2 > \sigma_3$ ) of a reduced stress tensor as well as the stress ratio  $R = (\sigma_2 - \sigma_3) / (\sigma_1 - \sigma_3)$ , a parameter describing relative stress magnitudes. The stress tensor provides information on the stress regime, i.e., compressional (with  $\sigma_3$  vertical), strike-slip (with  $\sigma_2$  vertical), transpressional ( $\sigma_2$  or  $\sigma_3$  vertical and  $\sigma_2$  close to  $\sigma_3$  in magnitude), or extensional (with  $\sigma_1$  vertical).

319 Several inversion methods have been proposed in the literature (see reviews by Célérrier et al., 2012).  
320 Using multiple methods, hence different algorithms, increases the accuracy of the results by reducing  
321 the effect of systematic errors. In this study, the fault-slip data analysis and principal stress axes  
322 calculation were first performed with the kinematic right-dihedra method (RDM; Angelier, 1994),  
323 which shows the distribution of the percentage of compression or extension dihedra in stereographic  
324 projection. Next, fault-slip data were inverted with two independent methods: direct inversion (DIM)  
325 and numeric dynamic analysis (NDA). All inversion methods were performed with *TectonicsFP*  
326 v1.7.9 software of Reiter & Acs (2000) and Ortnet et al. (2002). Statistically stable stress tensors  
327 were obtained from 10 to 30 fault-slip data measured in each structural site, which have dimensions  
328 between 5 and 50 m in length and/or width. A further explanation of the fault-slip data acquisition  
329 and analysis is included in the Supporting Information S3.

330 The main goal of the fault-slip data analysis was to characterize the Quaternary to present-day stress  
331 regime evolution in southern central Hispaniola. Unfortunately, it is generally challenging to date the  
332 striations more precisely than by simply recording that they are younger than the rocks deformed by  
333 the faults. For this reason, we include results obtained from fault-slip data measured in igneous and  
334 sedimentary rocks of upper Cretaceous to Late Pleistocene age, which were constrained with  
335 geochronological data from faulted Middle to Late Pleistocene coral reef terraces and alluvial fan  
336 deposits.

337 On occasions, more than one set of striae are present on a fault plane at the same measurement site,  
338 which were formed by different stress tensors. For example, reverse dip-slip and oblique reverse  
339 striations caused by an NE-trending compression are crosscut by low-pitch angle striations resulting  
340 from a NE-trending  $\sigma_1$  strike-slip stress regime. The discrimination of striae belonging to each fault  
341 set was conducted by numerical checking of compatibility with an inversion method and geological  
342 arguments. Fault crosscutting relationships, the overprinting of several striations in the same fault

plane and the absolute age of the faulted rocks have also been used as geological criteria to date faults and thus discriminate paleostress tensors and tectonic events.

#### **4. Neotectonics of southern central Hispaniola**

##### *4.1. Large-scale structures*

The geology of southern central Hispaniola consists of five main elements: (1) an Upper Jurassic to Upper Cretaceous igneous and metamorphic basement of the Cordillera Central domain; (2) a group of latest Cretaceous-lower Eocene sedimentary rocks of the Peralta fold-and-thrust belt, that locally unconformably overlie the basement; (3) an unconformable cover of folded and faulted middle Eocene to lower Miocene sedimentary rocks of the Rio Ocoa Group; (4) a sequence of slightly faulted and tilted Neogene sediments of the San Juan-Azua and Enriquillo basins; and (5) an unconformable cover of Quaternary alluvial, fluvial and shallow marine deposits. The spatial distribution of these elements is included in the geological map and cross-sections of Fig. 2.

The neotectonic structures of southern central Hispaniola are compiled in the map of the Fig. 3, which results from integrating of new field data with the geologic map obtained by the SYSMIN Project in the Dominican Republic (Pérez-Estaún et al., 2007). This map covers the eastern half of the Cordillera Central, the Sierra Martín García, the southeastern part of the San Juan basin and the whole Azua basin including the Baní pediment. This fieldwork area was selected because it contains the main active structures of the arc-plateau collision in the southern Dominican Republic.

The neotectonic structures of this area show three main trends: NW to WNW-striking folds and thrusts; N to NE-striking right-lateral strike-slip faults; and ENE to E-striking left-lateral strike-slip faults. Combined detailed structural analysis, fault-slip data inversion, and geochronology show that these structures were generated during three main tectonic events (Table 1). The NW to WNW-trending D1 structures are parallel to the structural grain of the Cordillera Central, as the folds and thrusts of the Peralta belt. Located in the southern and eastern parts of the study area, the N-S to NE-

367 SW transverse D2 structures are the strike-slip fault segments of the Ocoa-Bonao-La Guácara and  
368 Beata Ridge fault zones, as well as second-order structural elements of the related sub-parallel  
369 deformation corridors. The ENE to E-striking faults are also transverse D2 structures concentrated in  
370 the northwestern sector of the studied area, which also deform the boundary with the Cibao basin.  
371 The previous structures appear locally cut by two families of D3 extensional faults, which have a  
372 different trend depending on their geographic location: WNW to W in the Ocoa Bay sector; and NE  
373 in the Cordillera Central sector.

374 The Ocoa-Bonao-La Guácara fault zone represents the structural transition between the Muertos  
375 oceanic accretionary prism and the Beata Ridge-Peralta Belt collision zone (Fig. 3; Escuder-Virue et al., 2023). This fault runs along a N to NNE-striking band, 2 to 12 km wide and 120 km long,  
376 extending from the western termination of the Muertos Trough in the south to the Cibao basin in the  
377 north. Northward, the fault zone connects to the Hispaniola fault zone and changes its trend, curving  
378 from N-S to W-E over a distance of ~30 km. Based on discontinuities in its trace (such as step-overs,  
379 relays and bends), the fault zone comprises four major segments which are, from south to north (Fig.  
380 3): Offshore (O-OBFZ), Southern (S-OBFZ), Central (C-OBFZ) and Western (W-OBFZ). These  
381 active segments are variably oblique with respect to the regional convergence direction and,  
382 therefore, they exhibit strike-slip, oblique reverse and thrust fault movements. Along the eastern  
383 margin of Ocoa Bay, the O-OBFZ segment produces offshore a clockwise rotation of the folded  
384 sediments of the Muertos accretionary prism, and bends and deforms onshore the fold and thrust  
385 structures of the Peralta Belt, the SJPFZ, the sedimentary fill of the Azua basin and the Late  
386 Pleistocene to Holocene alluvial fans (Escuder-Virue et al., 2023).

388 The Beata Ridge fault zone (BRFZ) is part of the Beata Ridge. Mauffret & Leroy (1999) and  
389 Mauffret et al. (2001) suggested that the Beata Ridge fault zone (BRFZ) is a recent major NE-  
390 striking right-lateral strike-slip fault that limits the Beata Ridge to the west and would continue to the  
391 NE along the Beata Ridge summit and the eastern termination of the Sierra Bahoruco. On the basis

of swath bathymetry and offshore seismic reflection data (Granja Bruña et al., 2014), the northern part of the BRFZ can be divided in two fault segments which are, from south to north (Fig. 3): Central (C-BRFZ) and Northern (N-BRFZ). The N-BRFZ segment shows a constant NE-SW trend along 55 km, parallel to the east coast of the Sierra Bahoruco and Martín García and extends onland through the Azua Basin. The linearity of the fault segment indicates that it is a nearly vertical structure that accommodates a predominantly strike-slip displacement. Field data obtained in onland sub-parallel faults (see below) indicate a left-lateral strike-slip displacement characterizes the N-BRFZ. This fault segment cuts the post-middle Miocene macrostructure of NW-trending folds in the Sierra Bahoruco and Martín García, as well as the post-Early Pleistocene folds and thrusts that deform the Arroyo Seco Formation in the Azua Basin. This segment connects at its northern end with the OBFZ.

#### *4.2. Regional gravimetric data*

The regional pattern of the Bouguer anomaly reflects the main lithological and structural characteristics of southern central Hispaniola (Fig. 4), even though the distribution of the stations of gravity data is sparse and the used grid cell size is 250 x 250 m. Areas with positive anomalies ranging from 40 to 134 mGal (yellow and orange tones) are present in the northern and eastern sectors of the Cordillera Central, the northern sector of Sierra Bahoruco, Sierra Martín García and the western Ocoa Bay, as well as the eastern Cordillera Septentrional. These anomalies have a general NW-SE to WNW-ESE trend and are interpreted as areas of variably thickened island arc crust of mafic to intermediate composition, with average densities around  $2.95 \text{ g/cm}^3$  (Ayala et al., 2007). Varying textures of the gravity field within these areas include smooth parallel lineaments associated with bands of foliated amphibolites, elongated gabbro-dioritic batholiths and CLIP-related basaltic units of the Upper Cretaceous basement. In contrast, long wavelength gravity anomalies between -60 and 14 mGal (dark and light blue tones) characterize the Cibao basin, the southern sector of the Cordillera Central (Peralta belt) and the San Juan-Azua basin. These areas comprise

silicilastic and carbonate rocks of the late Cenozoic sedimentary cover, with average densities around 2.55 g/cm<sup>3</sup> (Ayala et al., 2007). These negative anomalies present a smooth texture and are elongated in the NW-SE to WNW-ESE direction, subparallel to the positive anomalies of the Cretaceous basement.

Therefore, the regional gravity field defines an alternance of NW to WNW-striking long wavelength anomalies, reflecting the contrast between the high-density Cretaceous igneous and metamorphic basement of the northern Cordillera Central and the low-density Cenozoic sedimentary rocks of the southern Cordillera Central, Cibao and San Juan-Azua basins (Fig. 4). The positive gravity anomaly of the Cordillera Central is disturbed by intermediate anomalies between 20 and 50 mGal (light green tones), along the arcuate trace of the Ocoa-Bonao-La Guacara fault zone. This pattern seems to reflect a NE-directed thick-skinned thrust structure in the basement, composed of tonalitic batholiths and intermediate volcanic rocks of the Tiroo Group, over the Cibao and Bonao sedimentary basins.

The northern boundary of the Cibao Basin is linear and defined by a steep gradient in the gravity field towards the positive anomaly of the eastern Cordillera Septentrional, coinciding with the surface trace of the Septentrional fault zone. The southern limit of the Cibao basin with the Cordillera Central is also marked by a strong gradient and coincide with the northern branch of the Hispaniola fault zone. In turn, the southern boundary of the San Juan-Azua basin is marked by a transition towards a NE-striking long wavelength positive gravity anomaly, which suggests the existence of high-density basaltic rocks of the CLIP forming the basement of the Sierra Bahoruco, the Sierra Martín García and the Ocoa Bay.

#### *4.3. Regional magnetic data*

The large-scale structure of southern central Hispaniola can be deduced by the orientation of the regional magnetic field in NW to WNW-trending subparallel bands, which are often characterized by

441 a distinctive textural pattern (Ayala et al., 2017). In the reduced to the pole grid (Fig. 4a), short-  
442 wavelength magnetic signatures generally correspond to areas where the magnetic sources in the  
443 basement are located close to the surface. These areas are characterized by a pattern of NW to  
444 WNW-trending sub-parallel anomalies, related to the late Cretaceous regional structure of  
445 amphibolites, gabbro-tonalitic batholiths and ridges of basalts in the Cordillera Central. The strong  
446 ferromagnetic character of the gabbros produces positive anomalies between 100 and 250 nT, which  
447 delineate the intrusive contact of the Arroyo Caña, Jumunuco and El Río batholiths. In the eastern  
448 Cordillera Central, the Hispaniola fault zone defines a strong NW-trending positive magnetic  
449 anomaly between 150 and 270 nT, separating the Loma Caribe Peridotite outcrops and the mafic  
450 volcanic rocks of the Los Ranchos Formation. This strong anomaly is related to the titanomagnetite  
451 growth during serpentinization by tectonic exhumation and surficial alteration of the peridotites in  
452 the late Cenozoic. A part of the southern Cordillera Central is characterized by scattered short-  
453 wavelengths anomalies, roughly elongated along a NE-SW trend and related to the extrusion of  
454 mafic to intermediate calc-alkaline to alkaline volcanic rocks during the Quaternary.

455 By contrast, long-wavelength magnetic signatures correspond to areas where the magnetic source is  
456 located farther from the surface, or where the magnetic rock intensity in the basement is weak. In the  
457 reduced to the pole grid (Fig. 4b), the Cibao and San Juan-Azua sedimentary basins generally display  
458 long-wavelength negative magnetic anomalies between -10 and -350 nT. These negative anomalies  
459 are related to the 1 to 5 km-thick overburden of paramagnetic sediments that fill these basins and  
460 reduces the basement magnetic intensity. In turn, the pattern of NW-striking negative magnetic  
461 anomalies that characterize the southern sector of the Cordillera Central is associated with the fold  
462 and thrust structure of the Peralta belt sedimentary rocks. Nevertheless, the sedimentary carbonate  
463 massifs of the Sierra Baboruco, Sierra Martín García and coastal sectors of the Ocoa Bay shows  
464 long-wavelengths of relatively high magnetic intensity between 0 and 100 nT, suggesting a magnetic  
465 source that is shallower than the surrounding areas. A possible explanation for this high magnetic



intensity is that the basement in these areas comprises magnetite-rich basalts typical of the late Cretaceous CLIP.

In general, the main magnetic discontinuities in the reduced to the pole magnetic grid correlate well with the large-scale fault zones and the regional macrostructure of southern central Hispaniola (Fig. 4b). These discontinuities juxtapose areas of long and short-wavelength magnetic anomalies and define tectonic blocks bounded by the segments of the Septentrional, Hispaniola, Ocoa-Bonao-La Guácara and Beata Ridge fault zones (Fig. 4b). In detail, low magnetic values mark the surface trace of the main fault zones, suggesting that the deformed fault rocks are partially demagnetized. The reduced to the pole magnetic grid also shows the truncation and displacement of the NW to WNW-striking D1 structures in the Cordillera Central by the transverse D2 strike-slip fault system of the Ocoa-Bonao-La Guácara fault zone.

#### *4.4. Late Neogene and Quaternary lithostratigraphy*

In southern central Hispaniola, neotectonic activity is recorded by the sedimentary fill of the San Juan-Azua and Enriquillo basins, and by the growing of a system of coral reef terraces overlain by alluvial fans. The San Juan-Azua and Enriquillo basins have been defined as syn-tectonic flexural basins limited to the NE by SW-verging thrust systems, although the Enriquillo basin is also limited by opposite-vergence thrusts directed to the NE (Mann et al., 1991b, 1995; Hernáiz Huerta & Pérez-Estaún, 2002; Díaz de Neira, 2004; Hernaiz Huerta et al., 2007b). These basins form the southeastern extension of the Plateau Central and Cul-de-Sac basins in Haiti (Pubellier et al., 2000). The basins are filled by a ~4 km-thick upward shallowing and coarsening mega-sequence composed of Miocene to Early Pleistocene marine sediments (McLaughlin et al., 1991; Díaz de Neira & Solé Pont, 2002; Pérez-Valera, 2010).

The megasequence includes initial sedimentation on a regionally extensive open marine carbonate platform in the lower to middle Miocene (Sombbrero Formation). The deposition of turbiditic-type

490 clastic sediments in a deep marine environment occurred in the middle to upper Miocene (Trinchera  
491 Formation) and in a shallow-water platform in the lower Pliocene (Quita Coraza Formation). The  
492 regressive character of sedimentation gave rise in the Pliocene to shallow-marine clastic and coral  
493 reef deposits (Arroyo Blanco Formation) and, in the Enriquillo basin, the deposition of halite and  
494 gypsum evaporites (Angostura Formation), and fine-grained clastic sediments of a bay with excess  
495 salinity (Las Salinas Formation). The change to continental deposition took place in the late  
496 Pliocene-Early Pleistocene in the Azua basin, with the sedimentation of coarse-grained clastic  
497 sediments (Arroyo Seco Formation), and in the Early-Middle Pleistocene in the Enriquillo basin,  
498 with an upward change from reef limestones and supralittoral marls to alluvial-fan breccias and  
499 conglomerates (Jimaní Formation).

500 From the uppermost Pliocene, tectonic indentation of the southern margin of Hispaniola by  
501 northeastward displacement of the Beata Ridge (Heubeck & Mann, 1991; Mann et al., 1991b;  
502 Ramírez, 1995; Hernaiz Huerta & Pérez-Estaún, 2002), destroyed the foreland configuration  
503 developed in the southern margin, uplift and erosion of the Cordillera Central and emergence of the  
504 Azua and Enriquillo basins. After a sedimentary hiatus, uplift is recorded by the formation and  
505 progressive elevation of coral reef terraces in the southernmost coastal area during the Middle to  
506 Late Pleistocene. These coral terraces were probably organized in a staircase marine terrace system,  
507 which has only been partially preserved by later erosion and tectonics (see below).

508 Continued uplift and erosion were also accompanied by syn-tectonic sedimentation of coarse-grained  
509 alluvial fan systems directed towards Ocoa Bay (Díaz de Neira, 2000; Pérez-Valera, 2010). Recently,  
510 three alluvial fans systems developed at different topographic levels have been distinguished on the  
511 southern slope of the Cordillera Central during the Late Pleistocene to the Holocene (Fig. 3; Escuder-  
512 Viruete et al., 2023). The upper alluvial fan system forms small relict plateaus inclined towards the S  
513 and SE, over the surface of the intermediate alluvial fan system. The intermediate alluvial fan system  
514 forms a more extensive and better-preserved deposit, which connects the foot of the Cordillera

Central relief with the southern coast. The current lower alluvial fan system is spatially restricted to the Sabana Buey valley and small coastal plains of southwestern Baní (Fig. 3). Its incision in the intermediate alluvial fans implies a change in the geometry of the drainage network, which has been attributed to fluvial capture processes (Díaz de Neira, 2000; Pérez-Valera, 2010), triggered by the activity of the OBFZ (Escuder-Virue et al., 2023).

## **5. Fault-slip data inversion and late Cenozoic stress regimes**

This work calculated stress tensors from a population of 562 fault-slip data, measured in 32 sites covering an area of approximately  $100 \times 50 \text{ km}^2$  in southern central Hispaniola. Geological characteristics of these sites are reported in the Supporting Information S4. Stress inversion of the fault-slip data yields 44 stress tensors and their respective plots are shown in the Figs. 5 to 14. The orientation of their maximum horizontal stress, stress regime, kinematic type of faults and the immersion method used in each site, which are generally consistent with each other, are included in the Supporting Information S5.

### *5.1. Fault-slip data inversion in the Peralta Belt*

Sites 21JE94, 21JE95 and 21JE96 are located along the highway between San José de Ocoa and Nizao, within the Peralta Belt (Fig. 5). Both the mudstone-rich siliciclastic rocks of the lower Eocene Peralta Group and the turbiditic sediments and conglomerates of the late Eocene Ocoa Group exhibit 1 to 100 m-thick zones of intense stratal disruption, which were interpreted by Witschard & Dolan (1990) as thrust surfaces formed during the late Eocene in the Peralta accretionary prism. Disrupted zones are mesoscopically characterized by boudinaged sandstone beds and pinch-and-swell structures defining lozenged blocks. Mudstone interbeds and argillaceous cataclastic bands have a WNW to NW-striking pervasive scaly clay fabric (Sp) and occasionally contain tight to isoclinal folds with rootless limbs (Fig. 5). The planar fabric (Sp) often contains a N to NE-trending calcite stretching lineation (Lp) subparallel to the fold axes. In this Sp-Lp fabric, kinematic indicators such as the

539 asymmetry of the boudins or the S-C structures establish a general top-to-the W and SW shear sense.  
540 All these structures are typical of a block-in-matrix fabric of a *mélange*, essentially produced by syn-  
541 sedimentary deformation during the Late Eocene (see below). Stratally disrupted zones in the Ocoa  
542 Group rocks are cut by WNW-striking D1 thrust surfaces that typically dip 5° to 30° more deeply  
543 than the Sp fabric, when they are restored to the original horizontal position. Both the block-in-  
544 matrix fabric and the thrusts are cut by NNW to NE-striking strike-slip faults D2 and by high-dip  
545 angle normal faults D3 (Fig. 6).

546 The inversion of fault-slip data collected in several localidades throughout the Peralta Belt enables  
547 the separation of populations related to three distinct stress regimes (Fig. 6). The first population is  
548 represented by variably oblique reverse slip vectors in NW-striking and NE-dipping faults (sites  
549 20JE09, 21JE94, 22JE12 in Yayas de Viajama, and 22JE13 in Tabera Arriba). It corresponds to a  
550 compressional stress regime D1 characterized by a NE-SW to ENE-WSW trending  $\sigma_1$ . The second  
551 population contains strike-slip to oblique reverse left-lateral slip vectors in NE-striking subvertical  
552 faults (sites 22JE13 and 21JE93). It is related to a near pure strike-slip stress regime D2  
553 characterized by a NNE to NE-trending  $\sigma_1$ . The third population includes normal vectors along dip-  
554 slip and oblique normal striae in NE-striking high-angle faults (site 21JE96), and corresponds to an  
555 extensional stress regime D3 with a N136°E trending  $\sigma_3$  (Fig. 13).

556 Sites 21JE123 and 21JE125 are located in the southernmost outcrops of the Peralta Belt, on the road  
557 Azua-Baní in the northern sector of the Ocoa Bay (Fig. 7). In this sector, the macrostructure in the  
558 Peralta Belt consists of a fold system of the lower to middle Eocene limestones of the Peralta Group  
559 and the upper Eocene turbidites with olistoliths of the Ocoa Group, related to a D1 deformation in a  
560 fold-and-thrust belt of NW-SE trend and SW-directed vergence. Towards the SE, the NW-trending  
561 D1 folds and thrusts of the Peralta Belt turn progressively towards a N-S trend and are cut and  
562 displaced by the O-OBFZ right-lateral D2 strike-slip fault. In the geological cross-section of Fig. 7,  
563 the Eocene rocks of the Peralta Belt overthrust through the SJPFZ the lower to middle Miocene

564 limestones of the Sobrerito Formation, and the assemblage overthrust in turn the conglomerates of  
565 the upper Pliocene to Early Pleistocene Arroyo Seco Formation. Therefore, D1 thrusting continued  
566 in this sector until the Early Pleistocene times.

567 Site 21JE123 is located along the Loma Vieja frontal thrust, which corresponds to the thrust ramp  
568 associated with the NNW to NW-trending D1 Loma Vieja anticline (Fig. 7). The related D1  
569 structures in the limestones of the Sobrerito Formation consist of SW-verging asymmetric folds,  
570 associated with mid-dip angle faults inclined to the NE and SW, subparallel to the Loma Vieja thrust.  
571 The mesoscopic S-C structures and other kinematic indicators imprinted on the fault planes establish  
572 a top-to-the-SW reverse movement. Predominant dip-slip striae in reverse fault planes define a  
573 population compatible with a thrust faulting stress regime and a N204°E trending compressional axis  
574 (Fig. 7). On the other hand, the orientation of calcite veins and T-planes in clasts of the  
575 conglomerates of the Arroyo Seco Fm establish a consistent NNE trend of D3 subhorizontal  
576 extension.

577 Site 21JE125 is located at the northwest end of the Ocoa Bay, 3.5 km southeast of the town of Azua,  
578 on the southern flank of a W-trending D1 anticline build-up in the limestones of the Sombrerito  
579 Formation (Fig. 7). The anticline is affected by a system of high-dip angle ( $> 60^\circ$ ) strike-slip faults  
580 D2 composed by NE-striking left-lateral strike-slip faults and antithetic N-striking right-lateral  
581 strike-slip faults. The assemblage is truncated by E to WNW-trending D3 normal faults, exhibiting a  
582 high-dip angle towards the N and S. Fault-slip measurements define two contrasting subsets (Fig. 7).  
583 The first subset includes striations with a low-pitch angle in NE-striking strike-slip faults. Normal  
584 and oblique normal dip-slip vectors in WNW-striking and NE and SW-dipping conjugate faults  
585 represent the second subset. The first population corresponds to a purely strike-slip stress regime D2  
586 with a N200°E trending  $\sigma_1$  and the second population to an extensional stress regime D3  
587 characterized by a N020°E trending  $\sigma_3$ .

## 588 5.2. Fault-slip data inversion in the Ocoa Bay

589 The limestone outcrops of the Sombrerito Fm located on the eastern coast of Loma Vigía, about 6  
590 km southeast of Azua (Fig. 7) exhibit folds, reverse faults and thrusts related to D1 deformation. In  
591 the 21JE112 site, these D1 structures consist of NW to WNW-trending and SW-verging asymmetric  
592 folds, associated with low to mid-dip angle reverse faults inclined to the NE, subparallel to the  
593 SJPFZ. The mesoscopic S-C structures and other kinematic indicators imprinted on these fault planes  
594 establish a top-to-the-SW reverse movement. Striae on the fault planes define a reverse slip  
595 consistent with a thrust faulting stress regime and a N222°E trending compressional axis (Fig. 8). In  
596 the 21JE120 site, several WNW-striking and NE-verging asymmetric anticlines correspond to the  
597 folds associated with a D1 back-thrust (Fig. 8). Slip measurements on the limbs of a decameter-scale  
598 anticline define a population of WNW-striking and SW-dipping reverse faults. This population  
599 corresponds to a purely compressional stress regime with a N026°E trending  $\sigma_1$ .

600 Throughout the Loma Vigia sector, D1 structures appear truncated by NW to W-trending D3 normal  
601 faults, which exhibit mid to high-dip angles towards the NE and SW (Fig. 8). For instance, the SW  
602 limb of a D1 anticline is cut by a D3 normal fault, causing the dragging of the limestone layers in the  
603 hanging-wall block (site 21JE120), or the stratification appears displaced by a system of ENE to E-  
604 striking normal faults with associated intrusion of mafic magmas (site 21JE113). In general, dip-slip  
605 striae in these fault planes define normal faults populations compatible with extensional stress  
606 regimes D3, characterized by a SSW to SW trending  $\sigma_3$  (Fig. 8).

## 607 5.3. Fault-slip data inversion in the eastern Sierra Martín García

608 In the limestone outcrops of the Sombrerito Formation located on the southeast coast of the Sierra  
609 Martín García, the D1 structures are strongly obliterated by the D2 deformation. This deformation  
610 has given rise to fault zones with a general NE-SW direction and several kilometres in length,  
611 parallel to the trace of the Beata Ridge fault zone, which runs offshore about 1.5 km to the SE. Fault

612 zones are marked by the development of bands of fault-gouge and fine crush breccia several tens of  
613 meters thick (Fig. 9). Individual fault planes are very steep (dip > 70°) and have low-pitch angle (<  
614 20°) striations that indicate a dominant strike-slip displacement.

615 At the Playa Caobita (site 21JE126), the Miocene limestones appear to be unconformably overlain  
616 by a ~50 m-thick Quaternary sequence composed from bottom to top of (Fig. 9): reef limestones of a  
617 basal fossil coral terrace; poorly consolidated conglomerates of an intermediate alluvial fan, that  
618 includes reworked corals and cobbles at the base; and gravels and sands rich in limestone cobbles of  
619 a younger alluvial fan. A gastropod *Strombus* sp. collected at the highest stratigraphic levels of the  
620 coral terrace has provided a U-Th age of  $118.47 \pm 0.52$  ka (21JE126C), so its growth took place in  
621 the interglacial marine oxygen isotope stage (MIS) 5e, in the Middle to Late Pleistocene boundary.  
622 Two specimens of coral *Diploria* sp. collected towards the basal and intermediate levels of the  
623 terrace have provided U-Th ages of  $476.99 \pm 23.6$  ka (21JE126B) and  $331.92 \pm 4.51$  ka (21JE126A),  
624 respectively. These ages indicate growth of the coral terrace during MIS 9c and 11 stages in the  
625 Middle Pleistocene, although a reworking of the corals is not ruled out because the terrace  
626 intercalates clastic levels.

627 Observations made in outcrops along the coast indicate that D2 strike-slip faults cut both reef terrace  
628 limestones and intermediate alluvial fan conglomerates. These relationships establish a Late  
629 Pleistocene age at least for the D2 deformation. On the other hand, the whole sequence of the coral  
630 terrace and the overlying alluvial fans is tilted 10-18° towards the SE and appear deformed by a  
631 system of E to ENE-striking D3 normal faults, which exhibits a high-dip angle (>60°) towards the N  
632 and S (Fig. 9). Therefore, the D3 deformation has taken place in the Late Pleistocene to Holocene.  
633 Fault-slip data measurements allow discrimination of a first population characterized by striations  
634 with a low-pitch angle in subvertical fault planes and a second population of normal dip-slip vectors  
635 in high-dipping faults. These populations represent two successive stress regimes: the first  
636 corresponds to a near strike-slip stress regime D2 with a N020°E trending  $\sigma_1$  (21JE127); and the

second matches with a near purely extensional regime D3 with a N008°E trending  $\sigma_3$  (22JE15; Fig. 9).

#### 5.4. Fault-slip data inversion in the northeast Sierra Bahoruco

The northwestern end of the Sierra Bahoruco shows a very sharp NE-SW trending coastline, subparallel to the trace of the central segment of the Beata Ridge fault zone, which runs offshore about 2.5 km to the SE (Fig. 3). Swath bathymetry and seismic reflection data suggest that fault segment constitutes the active boundary between the 1500 m elevated mountains of the Sierra Bahoruco to the NW and the -2500 m submerged slope of the Dominican sub-basin to the SE (Mauffret & Leroy, 1999; Granja Bruña et al., 2014).

Sites 21JE137 and 22JE26 are located in Playa Azul, about 6 km southeast of Barahona (Fig. 10). Along the coast, the limestone outcrops of the Sombrerito Formation define the northern limb of a D1 thrusting anticline of WNW trend and kilometric scale, which constitutes the internal structure of the Sierra Bahoruco in this sector. Limestones are strongly karstified and are unconformably overlain by a fossil coral reef 6 to 10 m-thick that defines a discontinuous morphological terrace along the coast. A *Diploria* sp. collected in the stratigraphically highest levels of the coral terrace has provided a U-Th age of  $123.23 \pm 0.70$  ka (22JE26B), so its growth took place in the MIS 5e. The assemblage is also unconformably overlain by the poorly consolidated conglomerates and red clays of a >20 m-thick alluvial fan (Fig. 10).

Both the karstified limestones and the overlying coral terrace and clastic deposits are faulted and tilted by a system of N to NE-striking D2 strike-slip faults, subparallel to the trace of the Beata Ridge fault zone (see cross section in Fig. 10). Fault planes are generally very steep (dip > 65°) and the striae show left- and right-lateral conjugate motion with a low reverse component (pitch < 25°). Fault-slip data inversion reveals a strike-slip D2 stress-field with a N046°E trending  $\sigma_1$  stress axis (Fig. 10). Therefore, the central Beata Ridge fault segment has been active for at least the Late



Pleistocene. In turn, the internal structure in the alluvial fan deposit defines a system of open anticlines and synclines of WNW-ESE trend and decametric to hectometric wavelength, which appears faulted on the flanks by WNW to W-trending D3 normal faults. These folds are interpreted as roll-over structures related to the D3 extensional faulting. Striations measured in decametric-scale fault planes and T-planes open in clasts are compatible with an extensional D3 regime, characterized by a N010°E trending  $\sigma_3$  (Fig. 10).

#### 5.5. Fault-slip data inversion in the southwestern Sierra Martín García

In the southwestern sector of the Sierra Martín García, the limestones of the Miocene Sombrerito Formation overthrust the sandstones, marls and gypsum of the upper Pliocene La Salina Formation, and the ensemble regionally overthrusts the continental conglomerates and mudstones of the upper Pliocene to Early Pleistocene Arroyo Seco Formation. These WNW-trending structures constitute the SW-directed D1 frontal thrust, which juxtaposes the anticline macrostructure of the Sierra Martín García to the poorly consolidated alluvial, floodplain and deltaic Quaternary deposits of the Enriquillo basin (Fig. 3).

Site 21JE130 is located in the hanging-wall block, about 300 m northward of the frontal thrust, where the gypsum and marl beds of the La Salina Formation exhibit structures related to D1 deformation (Fig. 11). These structures consist of WNW-trending and SW-verging asymmetric folds, associated with mid-dip angle reverse faults inclined to the NE, subparallel to the basal thrust. The asymmetry of D1 folds and mesoscopic S-C structures establish a top-to-the-SW reverse movement. Oblique reverse striae measured in the reverse fault planes define a fault population compatible with a D1 thrust faulting stress regime and a N047°E trending compressional axis (Fig. 11). In the site 21JE129, the D1 frontal thrust is locally fossilized by a coral reef terrace, where a *Strombus* sp. collected near its stratigraphic base has provided a U-Th age of  $124.65 \pm 0.85$  ka. A *Diploria* sp also

collected in the same basal levels has given a similar U-Th age of 128.89 ka (Fig. 11), thus the coral terrace grew in the MIS 5e stage.

Both the D1 thrust-related folds and the coral terrace are affected by a system of high-dip ( $> 60^\circ$ ) strike-slip faults D2. This fault system comprises NE to ENE-striking left-lateral strike-slip faults filled by subvertical calcite veins and antithetical N to NNE-striking right-lateral strike-slip faults, establishing a strike-slip stress D2 regime characterized by a NE-trending  $\sigma_1$  axis (Fig. 11). Therefore, the thrust-related D1 deformation reaches the Middle Pleistocene and the strike-slip-related D2 deformation continues after the Middle to Late Pleistocene boundary. Finally, a set of E to ENE-trending and high-dip angle D3 normal faults locally truncate the assemblage. Normal dip-slip striae measured in these fault planes cutting the gypsum beds are compatible to an extensional regime with a N355°E trending  $\sigma_3$  axis (Fig. 11).

#### *5.6. Fault-slip data inversion in the San José de Ocoa basin*

The southern segment of the Ocoa-Bonao-La Guácara fault zone is a roughly 50 km-long, N-striking right-lateral strike-slip fault system, that cuts at a high-angle and clockwise rotate the NW-SE trending folds and thrusts of the Peralta Belt and the structure of the late Cretaceous basement of the Cordillera Central (Figs. 3, 5). The segment is geometrically characterized by separating the fault trace into two branches south of San José de Ocoa town, giving rise to a right-hand releasing-bend and the San José de Ocoa pull-apart sedimentary basin (Escuder-Virquete et al., 2023). Both branches rejoin about 35 km northward into the C-OBFZ segment in the Bonao basin.

Site 21JE102 is located near the eastern branch of the S-OBFZ segment, in the town of Naranjal, 2.5 km east of San José de Ocoa (Fig. 5). At outcrop scale, N to NNE-striking right-lateral D2 strike-slip faults cut D1 thrust surfaces inclined to the NE, defined by sheared bands of mudstones in the upper Eocene Ocoa Formation, characterized by a SW-directed penetrative S-C fabric. Fault-slip measurements define two contrasting subsets (Fig. 12): the first subset is reverse to oblique reverse

and is associated to striations with a high-pitch angle in NW-striking fault planes; the second subset has oblique reverse and strike-slip vectors in conjugate right- and left-lateral faults. The first population corresponds to a compressional stress regime D1 with a N243°E trending  $\sigma_1$  and the second population establishes a purely strike-slip stress regime D2 characterized by a N047°E trending  $\sigma_1$ .

Site 21JE100 is also located along the eastern strand of the S-OBFZ segment, 2 km north of San José de Ocoa, on the road to Constanza (Fig. 5). The fault segment tectonically juxtaposes the folded and faulted volcanic rocks of the Tiroo Group and the gravels and sands of the Quaternary fill of the San José de Ocoa basin. Volcanic rocks have developed hectometer-scale subvertical fault planes subparallel to the S-OBFZ, characterized by a 5 m-wide brecciated damage zone and striations with a low-pitch angle (Fig. 12). These striae define a vector population compatible with a D2 strike-slip stress regime and a N226°E trending compressional axis. South of San José de Ocoa town, the eastern strand juxtaposes the folded and faulted mudstones of the Numero Fm and the gravels and sands of the basin through a subvertical D2 fault system that also tilts westward the Quaternary deposits. Striations measured in NNW-striking oblique reverse right-lateral faults of the 20JE15 site are compatible with a D2 strike-slip regime and a N186°E  $\sigma_1$  axis (Fig. 12)

The D2 strike-slip deformation is also very penetrative along the western branch of the S-OBFZ segment, which is morphologically marked by an east facing 200 m-high scarp. This strand juxtaposes the folded and faulted rocks of the Peralta Belt to the W with poorly consolidated gravels and sands that fill the San José de Ocoa basin to the E (Fig. 5). In several sites (20JE14, 21JE97, 21JE98 and 21JE99), striations measured on metric to decametric-scale faults subparallel to the S-OBFZ segment define a population of pure strike-slip to oblique reverse slip vectors, compatible with a D2 strike-slip regime and a predominant NE trending  $\sigma_1$  stress axis (Fig. 12). The subvertical

orientation of calcite veins and T-planes in clasts of the Quaternary conglomerates consistently establishes a WNW to NW trend of subhorizontal extension.

However, the effects of D3 extensional deformation are limited in this sector and geographically localized to the right-hand releasing-bend that forms the San José de Ocoa basin. Site 21JE99 is located between the two branches of the S-OBFZ segment, where the mafic to intermediate tuffs of the Tireo Group are in tectonic contact with the well-bedded limestones of Maastrichtian age. Slip measurements show two different populations: oblique reverse right-lateral and dip-slip reverse vectors (Fig. 12); and oblique normal vectors (Fig. 13). The overlap relationships between striae establish that these two populations represent two successive stress fields: the first corresponds to a transpressional D2 stress regime with a N270°E trending  $\sigma_1$ ; and the second corresponds to a more purely extensional D3 regime characterized by a N121°E trending  $\sigma_3$ . Also related to local D3 extensional tectonics, NW to N-striking normal faults have also been observed at sites 21JE103 and 21JE108 (Fig. 5). Dip-slip striations measured on these fault planes establish an extensional D3 regime characterized by a W-trending  $\sigma_3$  (Fig. 13).

#### *5.7. Fault-slip data inversion in the Cordillera Central*

The Cordillera Central is characterized by the development of crustal shortening structures, topographic uplift and exposure of the late Cretaceous volcano-plutonic basement in the core of large-scale WNW to NW-trending D1 anticlines (Fig. 3). In the Constanza-Sabana Quéliz area, the macrostructure consists of D1 folds and thrusts of WNW-ESE trend and usually SW-directed vergence, built up in a sequence composed by Cenomanian limestones and siliceous-rich sediments, Turonian to early Campanian mafic tuffs of the Constanza Formation, late Campanian mudstones of the El Convento Formation and Maastrichtian shallow-water platform limestones, as well as plutonic rocks (Escuder-Viruete et al., 2008). This macrostructure is locally cut and displaced by a system of high-dip angle strike-slip D2 faults, which includes NW to NNE-striking right-lateral strike-slip

755 faults and conjugate NE to ESE-striking left-lateral strike-slip faults. NE to E-trending D3 normal  
756 faults locally truncated the assemblage exhibiting a high-dip angle towards the NW and SE.

757 This deformative sequence is well recorded at the 21JE80 site, located ~10 km east of the Constanza  
758 town in the Tireo river valley (Fig. 5). Fault-slip data measurements and cross-cutting relationships  
759 show three contrasting subpopulations (Fig. 14): reverse and oblique reverse left-lateral faults;  
760 subvertical faults with low-pitch angle striations; and normal and oblique normal right-lateral faults.  
761 These subsets represent three successive stress regimes: the first corresponds to a compressional  
762 stress regime D1 characterized by a N213°E trending  $\sigma_1$ ; the second adjusts to a near strike-slip  
763 stress regime D2 with a N321°E trending  $\sigma_1$ ; and the third matches with a purely extensional regime  
764 D3 with a N183°E trending  $\sigma_3$  (Fig. 14).

765 Field observations at various points in the Cordillera Central show that D3 extensional faults overlap  
766 D1 thrusts. For example, the various kinematic types of faults that deform green mafic tuffs at sites  
767 21JE78 and 21JE79, where the fault-slip data inversion enables the separation of populations related  
768 to two distinct stress regimes (Figs. 13, 14). The first population is represented by dip-slip to variably  
769 oblique reverse slip vectors in low to mid-dip faults; the second population includes normal slip  
770 vectors along dip-slip striae in predominantly NNE to NE-striking high-dip faults. Both fault types  
771 define a NE-SW to W-E compressional D1 regime and a NW-SE extensional D3 regime,  
772 respectively.

773 A similar deformative sequence is recognized in the southeastern Cordillera Central, located east of  
774 the S-OBFZ segment (Fig. 3). In this sector, the S-OBFZ exhibits a fresh N-S scarp and represents  
775 the boundary between the >2000 m elevated mountains of the Cordillera Central to the W and the  
776 650 m elevated Rancho Arriba valley to the E. Along this morphological step, the fault zone  
777 produces triangular facets, perched valleys, and dextral offsets of streams that testify its recent  
778 activity. Deformation related to this segment produced the local clockwise rotation of the D1 fold

and thrust structure and the formation of the Rancho Arriba intramountain basin, which is filled by the Quaternary alluvial fans fed by the erosion of the surrounding relief.

Site 21JE70 is located at the Rancho Arriba basin southern bound. The basin exhibits an ENE-WSW elongated rectangular shape controlled by a southern active master fault (Fig. 5). Fault-slip measurements on hornblende-bearing tonalities of the Arroyo Caña batholith define two contrasting subsets (Figs. 13, 14): oblique reverse to strike-slip vectors in NE to ENE-striking faults; and oblique normal slip vectors in conjugate ENE and WSW-striking high-angle faults. The first set corresponds to a strike-slip stress regime D2 with a N022°E trending  $\sigma_1$  and the second set establishes a near pure extensional stress regime D3 characterized by a N023°E trending  $\sigma_3$ , related to the formation of the basin. Observations carried out in the volcanic rocks of the Tiroo Group at sites 21JE72 and 21JE73 allow us to establish similar temporal relationships: younger NE to E-striking D3 normal faults cut older D2 strike-slip and D1 reverse faults. Striations measured in decametric-scale fault planes subparallel to the NW to WNW-trending regional macrostructure in these sites are compatible with a compressional D1 regime, characterized by a general NNE to NE-trending  $\sigma_1$  (Fig. 14).

The NW-SE to W-E directed D3 extensional tectonics has also been detected in other sites located close to the OBFZ. Site 21JE99 is located between the two branches of the S-OBFZ segment, north of San José de Ocoa town (Fig. 5), where Maastrichtian limestones exhibit a penetrative brecciation associated with decametric-scale NE-striking fault surfaces, dipping a high-angle to the NW and SE. Slip measurements establish an oblique normal left-lateral movement on these planes and their inversion establishes an extensional D3 regime characterized by a N121°E trending  $\sigma_3$  (Fig. 13). Site 21JE103 is located ~1.5 km east of the S-OBFZ (Fig. 5), where decametric-scale fault planes measured in basalts of the Tiroo Group contain striations of an oblique normal kinematics in NW to NE-trending and E-dipping faults. Inversion of these fault-slip data yields an extensional stress regime characterized by a N256°E trending  $\sigma_3$  (Fig. 13). Finally, slip measurements in N- to NNW-

striking fault planes at site 21JE108 (Fig. 5) indicates a predominance of dip-slip normal vectors, compatible with a purely extensional regime with a N069°E trending  $\sigma_3$  (Fig. 13).

## 6. Discussion

### 6.1. Relations between neotectonics and Quaternary sedimentation in southern central Hispaniola

Several researchers have previously described the uplift, folding and thrusting of the Cordillera Central basement and the Peralta Belt during the lower Miocene to upper Pliocene time interval (Heubeck & Mann, 1991; Mann et al., 1991a, b; Ramírez, 1995; Pubellier et al., 2000). The style of this D1 deformation consisted of a SW-directed fold-and-thrust belt, which was developed above a basal detachment horizon following a forward propagating sequence (Heubeck & Mann, 1991; Hernaiz Huerta & Pérez-Estaún, 2002; this work). D1 is superimposed on a deformation produced during the upper Eocene in the Peralta accretionary prism (Witschard and Dolan, 1990). Data obtained in this study indicate that a large part of this deformation is syn-sedimentary and characterized by the formation of a block-in-matrix fabric typical of a heterogeneous mélange. Syn-sedimentary deformation was very probably produced by submarine gravitational processes, triggered by instability in the basin slope and/or tectonics (e.g., Alonso et al., 2014; Escuder-Viruete & Baumgartner, 2014; Suárez Rodríguez et al., 2017).

In this tectonic context, the San Juan-Azua basin constitutes the foreland flexural basin developed in front of the SW-directed Peralta fold-and-thrust belt, and bounded to the NE by SW-directed thrusts (Hernaiz Huerta & Pérez-Estaún, 2002; Díaz de Neira, 2004; this work). These basins were filled by shelf marine, turbiditic, and continental sediments giving rise to a 4-km thick, shallowing and coarsening upward regressive megasequence of lower Miocene to Lower Pleistocene age. D1 deformation propagated to the Azua basin sediments and continued until Early to Middle Pleistocene times, which is the age of the Arroyo Seco Formation deformed in the footwall of the Peralta frontal thrust (Díaz de Neira & Solé Pont, 2002). As documented in the present work, these coarse-grained

827 continental deposits were folded and thrust in the footwall of the San Juan-Pozos fault zone,  
828 current frontal thrust of the Peralta Belt.

829 After a sedimentary hiatus, uplift and erosion of the resulting relief fed several staircase alluvial fan  
830 systems, developed at different topographic levels on the southern slopes of the Cordillera Central  
831 probably since the Early Pleistocene (Díaz de Neira, 2000; Pérez-Valera, 2010). The ages obtained  
832 by optically stimulated luminescence geochronology between  $43.8 \pm 8.3$  ka and  $5.9 \pm 0.6$  ka for the  
833 intermediate and young alluvial fan systems indicate the uplift continued, at least, during the Late  
834 Pleistocene to the Holocene (Escuder-Virquete et al., 2023). The older alluvial fan is not yet been  
835 dated, but probably also records uplift and erosion during the Middle Pleistocene.

836 In both the eastern and western sectors of the Ocoa Bay, these alluvial fan systems unconformably  
837 overlie a fossil coral reef terrace developed during the MIS 5e (118-128 ka) and 5c (~105 ka) stages  
838 in the Last Interglacial. Therefore, the fluvial incision and the progradation/aggradation of the  
839 intermediate and young alluvial fans were also controlled by the global sea level fall and its  
840 variations during the Last Glacial. The coral terrace also includes reworked corals grown during MIS  
841 7 and 9 stages. However, no remains have been found of coral terraces built during these stages in  
842 the Ocoa Bay, so erosion might eliminated them.

843 On the other hand, the Quaternary coral reef terrace and the alluvial fans unconformably overlie the  
844 D1 large-scale structure of NW-trending and SW-verging folds and thrusts that affect the NE margin  
845 of the Azua basin. Therefore, the regional pervasive D1 deformation ends, at least in southern central  
846 Hispaniola, during the Lower Pleistocene. Late Pleistocene to Holocene alluvial fan and pediment  
847 deposits are all offset by strike-slip faults of the D2 event, implying that the more recent tectonic  
848 activity is mainly partitioned in the Ocoa-Bonao and Beata Ridge fault zones, as well as by faulting  
849 in related structures. Finally, both the coral reef terraces of the MIS 5e and 5c stages and the



overlapping alluvial fan systems of Late Pleistocene to Holocene age appear deformed by the normal faults of the D3 event in the Ocoa Bay sector.

## *6.2. Evolution of the Quaternary stress regime in southern central Hispaniola*

The geometric characteristics of the faults at all scales, the striae sets in their planes, the compatible stress tensors and their chronological succession, constrained by offset relations with dated lithostratigraphic units, define three types of neotectonic fault kinematics in southern central Hispaniola (Table 1). These types are: thrust and reverse faulting with horizontal  $\sigma_1$  and  $\sigma_2$ ; strike-slip faulting with horizontal  $\sigma_1$  and  $\sigma_3$ ; and extensional faulting with vertical  $\sigma_1$  (Fig. 15; see also Supporting Information S6). Each type of fault kinematics is related to a different deformation event, characterized by a specific state of stress, whose development changes geographically throughout southern central Hispaniola (Fig. 16).

Throughout the Cordillera Central domain, Sierra Martín García and San Juan-Azua basin there is evidence of slip on thrust and reverse faults related to a NNE to NE-trending compressional D1 stress regime (Fig. 16). These NW-trending D1 structures deform the middle Eocene-lower Miocene sedimentary rocks of the Río Ocoa Group, as well as its substrate of the Peralta Belt accretionary prism, the early/middle Miocene to middle Pliocene marine turbiditic and platform sediments of the San Juan-Azua foreland basin, and the upper Pliocene to Early Pleistocene clastic continental deposits of the Arroyo Seco Formation (Heubeck & Mann, 1991; Díaz de Neira & Solé Pont, 2002; Hernáiz Huerta & Pérez-Estaún, 2002; among others). A similar structural relationship has been described along the southern border of the Chaîne des Matheux in Haiti, where SW-directed thrusts deform rocks of middle to upper Miocene age (Pubellier et al., 2000). Upper Miocene to lower Pliocene onset of uplift in the Cordillera Central is simultaneous with uplift in the Sierra de Neiba, with both ranges providing the clastic sedimentary source for the eastern San Juan and Azua (piggyback) basins. The pre-Early Pleistocene folding and thrusting D1 event has been related to the

874 collision between CLIP-related units of the Southern Hispaniola Peninsula and the island-arc crust of  
875 the Cordillera Central (Heubeck & Mann, 1991; Mann et al., 1991b).

876 Stress tensors obtained for D1 event indicate a general NE-trending  $\sigma_1$  axis, ranging between  
877 N017°E and N070°E directions, with a relatively well-constrained maximum value in the count-  
878 contours stereoplot at N202°E, plunging 06° to the SW (Fig. 15). A similar NE to ENE-directed  
879 shortening was obtained by Ramírez (1995) from the analysis of the faults that deform the Arroyo  
880 Blanco Fm. In the calculations, the orientation of  $\sigma_1$  has been restored at sites in the eastern Ocoa  
881 Bay that have undergone a post-D1 clockwise rotation. The vertical arrangement of  $\sigma_3$  axis in the  
882 computed stress tensors is consistent with deformation by thrusting. The R stress ratio has values of  
883  $0.50 \pm 0.25$ , indicating stress tensors close to pure compression. Therefore, the regional D1 stress  
884 regime consisted of a NE-trending subhorizontal compression. However, R ratio presents some  
885 values of 0.87 and 0.92 (21JE73 and 21JE79 sites), which means that the maximum and minimum  
886  $\sigma_H$  axes are close in magnitude and higher than  $\sigma_V$  ( $\sigma_1 \approx \sigma_2$ ). These R values suggest that a D1  
887 deformation locally characterized by a radial compression regime. On the other hand, the WSW-  
888 directed GPS motions measured in JUAN, CONS and BARA stations, located away from important  
889 D2 structures, are suborthogonal to the strike of the San Juan-Pozos and Barahona fault zones, as  
890 well as the thrust structures in the Cordillera Central and Peralta Belt domains. Therefore, present-  
891 day reverse faulting dominates active D1 structures (Fig. 16). This argument agrees with thrust  
892 kinematics and reverse stress regime obtained for D1 from field permanent deformation.

893 The NNE to NE-trending strike-slip D2 stress regime has been deduced near the Ocoa-Bonao-La  
894 Guácara fault zone, particularly in the eastern sector of the Ocoa Bay and in the margins of the San  
895 José de Ocoa and Rancho Arriba intramountain basins (Fig. 16). This strike-slip deformation has also  
896 been observed in the vicinity of the Beata Ridge fault zone, in the western sector of the Ocoa Bay,  
897 and in relation to NE to ENE-striking left-hand strike-slip faults that cut and displace D1 thrusts in

898 the Cordillera Central. Therefore, D2 strike-slip faults cut and displace the D1 fold-and-thrust  
899 structures, as described in previous works (Ramírez, 1995; Hernáiz Huerta & Pérez-Estaún, 2002;  
900 Pérez-Valera, 2010).

901 D2 strike-slip faults deform the Quaternary deposits in several places, as the fossil coral reef terrace  
902 built during the MIS 5e and 5c stages in the SW and NE flanks of the Sierra Martín García and  
903 Batoruco. Recent field-data collected in the eastern sector of Ocoa Bay (Escuder-Viruet et al.,  
904 2023), indicate that D2 strike-slip faults subparallel to the O-OBFZ segment deform an alluvial fan  
905 of latest Pleistocene age, but are locally fossilized by a younger alluvial fan of Holocene age.  
906 Nonetheless, the Ocoa-Bonao-La Guácara fault zone cuts geomorphic features along its path, such as  
907 Late Pleistocene-Holocene alluvial fans, fluvial terraces and streams in the San José de Ocoa and  
908 Rancho Arriba intramountain basins, which collectively evidence active D2 deformation. Therefore,  
909 the D2 stress-field continues through the latest Pleistocene and most likely today. This interpretation  
910 is consistent with the moment tensor solutions of strike-slip faulting calculated for the 1977 and 1988  
911 earthquakes (CMT catalogue; Ekström et al., 2012), whose epicenters are located very close to the  
912 OBFZ trace.

913 The change from a D1 compressional to a D2 strike-slip stress regimes took place during the Early  
914 and Middle Pleistocene (or sometime in the Early Pleistocene). This change could be related to the  
915 indentation of the eastern sector of the Beata Ridge in southern Hispaniola, as part of the Caribbean  
916 oceanic plateau collision in the back-arc region of the Hispaniola microplate. This processes could  
917 produced the large-scale drag of the pre-existing D1 structures in the eastern Ocoa Bay and  
918 generated large-scale conjugate D2 strike-slip faults on the margins of the aseismic ridge indenter. In  
919 this sense, the geometric characteristics of the D2 deformation, such as landward deflection of the  
920 Muertos trench axis and development of conjugate strike-slip faults and back-thrusts, are analogous  
921 to those produced above a subducted conical seamount or submarine aseismic ridge (Ranero and von

922 Huene, 2000; Sak et al., 2009; Gardner et al., 2013), as it has been faithfully reproduced by sandbox  
923 analogue models of such convergent margins (Dominguez et al., 2000).

924 Stress tensors of D2 event indicate a general NE-trending  $\sigma_1$  axis, comprised between N015°E and  
925 N055°E directions, with a relatively well-constrained maximum value at N045°E, plunging 01° to  
926 the NE (Fig. 15). The dispersion of  $\sigma_1$  axes is due to fault-slip measures in sites located in the San  
927 José de Ocoa Basin, where the S-OBFZ trends N-S, or probably due to the reorientation of the  
928 regional stress field that occurs around this large-scale strike-slip fault. The subvertical arrangement  
929 of the  $\sigma_2$  axes indicates that the D2 deformation took place in a strike-slip regime, characterized by a  
930 NE-trending subhorizontal shortening for R-values ranging between 0.21 and 0.60, with prevalence  
931 of values close to 0.5. These R-values suggest a deformation mode close to pure strike-slip,  
932 consistent with the low-pitch angle of the displacement vectors observed in D2 strike-slip faults and  
933 some focal mechanism solutions. Note that the SW-directed GPS motion measured in the TORT  
934 station is parallel to the NE-SW strike of the nearby Beata Ridge fault segment, implying a present-  
935 day pure strike-slip faulting along this D2 structure (Fig. 16).

936 D3 faults can be grouped into two geometric families: D3a and D3b. The first family includes NE-  
937 SW to NNE-SSW trending normal and oblique normal D3a faults. Although the number of  
938 computed stress tensors is limited, inversion of fault-slip data establishes a general NW-SE trending  
939 extensional stress regime D3a (Fig. 15). This extensional stress regime has only been detected in the  
940 western Cordillera Central-Peralta Belt domains and in sectors close to the S-OBFZ segment, as the  
941 San José de Ocoa and Rancho Arriba intramountain basins (Fig. 16). The orientation of  $\sigma_3$  axes has a  
942 high dispersion, ranging between N103°E and N168°E, for values of the R ratio very close to 0.5,  
943 establishing stress tensors of pure extension. The age of the D3a event could not be precisely  
944 established. However, these normal faults affect the volcanic rocks of Pleistocene age of the Valle  
945 Nuevo area whose K-Ar ages are  $0.5 \pm 0.3$  and  $0.3 \pm 0.2$  Ma (Vespucchi, 1988; Kamenov et al., 2011;

and references herein) and, therefore, its age is at least post-Middle Pleistocene. The alignment of the emissive centers of the Quaternary calc-alkaline to mafic alkaline volcanism in the region follows a NE-SW strike (Fig. 16), so the NW-directed extension during D3a could have favoured the opening of NE-SW striking planes and the rise of magmas. Alternatively, the rise of magmas could have taken place in favour of releasing bends formed during D2 deformation. Hot springs and associated travertine deposits in the volcanic region indicate that geothermal activity continues today.

The second family includes WNW to W-striking dip-slip normal faults that established an N to NNE-trending extensional D3b stress regime. Its development is geographically limited to the surroundings of Ocoa Bay (Fig. 16). D3b normal faults deform the coral reef terraces of the MIS 5e and 5c stages (22JE26, Playa Azul; 21JE126, Playa Las Caobitas; and 20JE04, Palmar de Ocoa sites) and the overlapping alluvial fan of Late Pleistocene to Holocene? age (21JE127 and 22JE15 sites). Inversion of 12 fault-slip data sets gives rise to NNE-trending  $\sigma_3$  axes, comprised between N003°E and N023°E directions, with a relatively well-constrained maximum value at N022°E, plunging 6° to the NE (Fig. 15). The subvertical arrangement of the  $\sigma_1$  axes and the obtained R-values close to 0.5 indicate stress tensors close to pure extension.

In summary, a relatively constant NE-trending horizontal shortening, parallel to the current direction of plate convergence established by GPS measurements, controlled the evolution of the Quaternary stress tensor in southern central Hispaniola. This evolution included a D1 event of compression followed by a D2 near pure strike-slip event, which was locally coeval by a more heterogeneous and geographically localized D3 regime of pure extension. The successive changes in the stress regime were related to a permutation of  $\sigma_3$  by  $\sigma_2$  vertical stress between D1 and D2 events, and to a permutation of  $\sigma_2$  by  $\sigma_1$  vertical stress between D2 and D3 events (Fig. 15).

### *6.3. Origin of the extensional tectonics localized in the Ocoa Bay*

969 Figure 13 shows the coast of southern central Hispaniola intermittently covered by the fossil coral  
970 reef terrace that grew as a fringing reef during MIS 5e/5c stage and currently ranges in elevation  
971 from 0 to 20 m above sea level. The MIS 5e stage corresponds to a relatively long period of high-sea  
972 level that took place during the 118–128 ka interval in the Last Interglacial, when the sea level was  
973 between 2 and 6 m above the current level (e.g., Schellmann & Radtke, 2004). The existence of coral  
974 reef terraces of the MIS 5e/5c stage at elevations between 1 and 25 m above the current sea level can  
975 be explained by active tectonics that has uplifted in variable amounts the southeastern margin of  
976 Cordillera Central, the Sierra Martín García and the Sierra Bahoruco. Assuming a *Diploria* sp.  
977 growth depth of ~2 m and an average sea level height during the MIS 5e of +2.0 m (see details in  
978 Escuder-Virue et al., 2020), the minimum uplift rates for the terrace range between 0.01 and 0.22  
979 m/ka. The absence of the coral terrace, as well as of other coral terraces presumably built up during  
980 older MIS stages, can be explained as due to erosion or to being located at elevations below present  
981 sea level as a consequence of later subsidence. In our case, the absence of the coral terrace of the  
982 MIS 5c/5e stage coincides with the reentrant of the Ocoa Bay affected by D3b extensional tectonics.

983 Figure 13 also shows two outcrops of the modern fringing reef raised between 0.5 and 3 m above sea  
984 level, which are currently subjected to strong marine erosion. An specimen of coral *Diploria* sp.  
985 collected in the uppermost levels of the terrace at Playa Caracoles (21JE124 site) have provided an  
986 U-Th ages of  $0.730 \pm 0.01$  ka. This age indicates growth of the coral terrace during MIS 1 stage in  
987 the Middle Holocene. The calculated minimum uplift rate for the terrace is 4.11 m/ka.

988 Thus, the spatial distribution of the topography in the Ocoa Bay, the D3b extensional deformation,  
989 outcrops of the coral reef terraces and uplift rates are, to the first order, controlled by the subducting  
990 Beata Ridge of the Caribbean plate. This is because the Muertos accretionary prism that makes up  
991 the margin wedge on the Ocoa Bay must deform around this indenting bathymetric feature. The  
992 topographically low area of the Ocoa Bay lines up along the relative plate motion vector with the  
993 Beata Ridge indenter. The highest uplift rates are recorded along the central northern coast of the

994 Ocoa Bay and decrease parallel to the trench and perpendicular to the Beata Ridge to the E  
995 (southeast coast of the Cordillera Central) and the W (Sierra Martín García).

996 This pattern in the distribution of the D3b deformation and vertical motions can be interpreted as a  
997 wave of rapid uplift followed by subsidence that propagates parallel to the current plate motion  
998 vector in the Ocoa Bay, as a response to the subduction of bathymetric irregularities in the Beata  
999 Ridge. Following the deformation mechanism proposed for seamount subduction in the New  
1000 Hebrides and Solomon Arcs (e.g., Taylor et al. 2005), the passage of a subducting bathymetric high  
1001 produces a history of rapid uplift and subsequent collapse and subsidence in a block of similar  
1002 dimensions of the upper plate.

1003 Along the Ocoa Bay, where the Beata Ridge indenter currently entering the Muertos Trench,  
1004 macroscale WNW to W-striking dip-slip normal faults and mesoscale fault-slip data, are consistent  
1005 with SSW to S-directed extension perpendicular to the margin (Fig. 16). These steeply-dipping  
1006 normal faults may be a consequence of the collapse of the margin, constituted by relatively soft  
1007 sedimentary rocks of the Muertos accretionary prism, in the wake of the stronger mafic igneous  
1008 rocks of the subducting Beata Ridge that produce a basal erosion of the upper plate (Ramero & von  
1009 Huene, 2000; Sak et al., 2009; Gardner et al., 2013; Vannucchi et al., 2013). The subvertical and NE-  
1010 striking faults formed during D2, such as the N-BRFZ and O-OBFZ segments, are arranged at high  
1011 angles to the margin and may have accommodated differential displacements during D3, acting as  
1012 transfer faults. Ocoa Bay would therefore be an embayment of ~20 km wide and ~25 km length,  
1013 formed by the passage of a bathymetric high in the subducting Beata Ridge. A similar mechanism  
1014 has been proposed along the northern Hispaniola-Puerto Rico margin, where subduction/collision of  
1015 bathymetric highs built on the Atlantic oceanic crust has resulted to basal erosion in the upper plate  
1016 and forearc collapse (Grindlay et al., 2005; Escuder-Viruete and Pérez, 2020).

This interpretation is favored by the absence of the coral terrace of the MIS 5e/5c inboard of the Beata indenter (Fig. 13), which must have been uplifted and eroded or sunk by D3b extensional tectonics. In turn, the gravimetric and magnetic anomaly produced by the CLIP magnetic basalts that characterizes the surroundings of Ocoa Bay (Fig. 4) suggests a thinning of the upper plate wedge in the submarine portion of the margin. In this tectonic context, the high uplift rates detected in the uplifted modern coral terrace indicate a new episode of rapid uplift and future subsidence of the margin during continued subduction of the Beata Ridge.

Although the role of seamounts/aseismic ridge subduction in the plate-margin seismicity is disputed, e.g. enhancing seismic coupling (e.g., Bangs et al., 2015) versus decreasing the degree of coupling and limiting the lateral propagation of megathrust-earthquake rupture planes (e.g., Bonnet et al., 2019), the 1751 earthquake occurred in the vicinity of the Ocoa Bay and destroyed the city of Azua. The presence of the subducting Beata Ridge in high-resolution bathymetric swath maps of the Caribbean-Hispaniola plate boundary region and in seismic reflection profiles across the Muertos Trench (e.g., Granja Bruña et al., 2014), together with the proposed location of the Azua earthquake intensity center in the Ocoa Bay (Bakun et al., 2012), suggest that this  $M_w = 7.5$  event may have resulted from the rupture of a ridge asperity in the subduction and can therefore be repeated in the future.

#### *6.4. Definition of seismotectonic structures in southern central Hispaniola*

The characterization of seismotectonic structures has been previously carried out in some sectors of Hispaniola Island (Frankel et al., 2010; Bertil et al., 2015; Escuder-Viruete et al., 2020; Terrier-Sedan and Bertil, 2021). Used criteria in the definition were: geodynamic situation; geological, stratigraphic and/or tectonic evidence of recent activity; kinematic type of fault and rate of displacement; and associated historical and/or instrumental seismicity. The definition of these large-scale active structures is critical since it allows for establishing seismotectonic zonation models of



the area under consideration, which is fundamental for the quantitative seismic hazard assessment. However, this approach requires updating the inventory of active structures as tectonic and seismological knowledge increases.

The previous works and the new data presented in this work allow updating the main seismotectonic fault zones structures in southern central Hispaniola, including its adjacent offshore sector. They are: San Juan-Pozos (Mann et al., 1991a); Matheux-Sierra de Neiba (MAFZ, Hernaiz Huerta et al., 2007); Batoruco (BAFZ or Barahona Thrust; Rodríguez et al., 2018); Beata Ridge (Mauffret and Leroy, 1999); Muertos Trough (Granja Bruña et al., 2009; 2014); Ocoa-Bonao-La Guacara (Pérez-Estaún et al., 2007; Escuder-Virue et al., 2023); and probably the eastern end of the Enriquillo-Plantain Garden (EPGFZ; Mann et al., 1995).

The attribute compilation of these seismotectonic structures is outlined in Table 2, including the fault or segment name, deformation mechanism, estimated maximum magnitude ( $M_w$ ), known biggest earthquake (magnitude in  $M_w$ ), estimated displacement rate, length, strike and dip, and width of the fault zone. Following Bertil et al. (2015), the maximum magnitude of rupture for each fault/segment was estimated using the major recorded historical earthquake (Terrier-Sedan and Bertil, 2021), or its geometric parameters (e.g. Wells and Coppersmith, 1994), taking into account its degree of uncertainty. Displacement rates for each seismogenic structure were derived from geological mapping (SYSMIN Project, 2010; Escuder-Virue et al., 2022; Escuder-Virue et al., 2023; this work), geodetic GPS measurements (Calais et al., 2002; Manaker et al., 2008), or calculated in the present study through empirical relationships (e.g. Wells and Coppersmith, 1994).

From this information, a simplified seismotectonic model was built that considers two types of seismic sources: subduction and strike-slip fault zones in which the seismicity of greater magnitude is concentrated; and superficial/upper crustal areas located between them where the seismicity is of

moderate to low magnitude and is more homogeneously distributed. This seismotectonic model is the starting point for the seismic hazard assessment in southern central Hispaniola presented below.

#### *6.5. Seismic hazard assessment in southern central Hispaniola*

The seismic hazard assessment carried out in this work follows a probabilistic approach. Following the Cornell-McGuire methodology, the R-CRISIS calculation code (Ordaz et al., 2014) builds a probabilistic model that takes into account the spatial distribution of seismogenic sources, the occurrence and magnitude of earthquakes in time, and the attenuation characteristics of the strong motion in the ground. Thus, the R-CRISIS code computes the seismic hazard in terms of the probability of exceeding a peak ground acceleration (PGA) value for a specific site in a given period.

For the seismic hazard assessment in southern Central Hispaniola, the previously described seismotectonic zonation model composed of subduction zones, main strike-slip fault zones and diffuse seismogenic zones located between them, was geometrically built as simplified 3-D polygonal surfaces of specific width. In these seismic sources, earthquakes can occur at any point of the source with equal probability (Ordaz et al., 2014). Functions selected to describe the attenuation of the ground motion acceleration with distance from the source are the same as those used by Frankel et al. (2010) and Benito et al. (2012) to evaluate the seismic hazard in Haiti after the 12 October, 2010 earthquake. Parameters used to establish the seismic activity of each source and their uncertainties are included in Escuder-Virue et al. (2020). The hazard analysis performed in this work did not take into account possible amplifications of the seismic parameters by site effects. Seismic hazard was computed in a rectangular point grid, framed by the coordinates: 17.960°N and 18.835°N of latitude; -70.175°W and -71.250°W of longitude (WGS-84 projection system), with a point spacing of 0.025° in both directions. Results were converted in a continuous surface using an inverse distance squared weighted interpolation algorithm.

Figure 17 includes the results of the seismic hazard assessment in southern central Hispaniola, expressed as PGA intervals (values in  $\text{cm/s}^2$ ) and for a return period of 475 years (i.e., for a probability of exceedance of 10% in 50 years). Modeled minimum and maximum PGA values are 86  $\text{cm/s}^2$  and 857  $\text{cm/s}^2$ , respectively, with a mean (and standard deviation) of 386.7 (148.0). Therefore, the regional seismic hazard values range from low to very high. PGA zoning defines an elongated pattern sub-parallel to the main fault zones. Seismic hazard is higher in the Ocoa Bay, along the San José de Ocoa valley in the southern Cordillera Central, and in the topographic transition between the northern Cordillera Central and the eastern Cibao Valley. It decreases towards the NW and SE regions. This seismic hazard zoning contrasts with the previously proposed by Frankel et al. (2010), Benito et al. (2012) and Bertil et al. (2010, 2015), which did not take into account all the seismic sources included in the present work, particularly the Ocoa-Bonao-La Guácara fault zone.

With this seismotectonic model, the highest PGA values ( $> 600 \text{ cm/s}^2$ ) are associated with the N-BRFZ, O-OBFZ and C-OBFZ segments. Branches of the S-OBFZ segment set intermediate to high values (500-650  $\text{cm/s}^2$ ). The C-OBFZ segment also has associated high PGA values ( $>650 \text{ cm/s}^2$ ), which decrease to intermediate and low values westward along the W-OBFZ segment (250-450  $\text{cm/s}^2$ ). The superficial trace of the O-MT segment of the Muertos Trough also has associated high PGA values (600  $\text{cm/s}^2$ ), which extend into depth following the subduction interface. The BAFZ has associated intermediate PGA values in its connection to the BRFZ, which also decrease westward. The S-JPFZ presents high PGA values at its SE end (500-700  $\text{cm/s}^2$ ), where it is displaced by the N-BRFZ, decreasing towards the NW. The Hispaniola fault zone has also been included in the modeling, giving rise to intermediate values of the PGA that become high in its central segment. The E-EPGFZ segment also has associated intermediate to low PGA values (400-500  $\text{cm/s}^2$ ).

In summary, the highest PGA values in southern central Hispaniola are associated with N-BRFZ, O-OBGFZ and A-MT segments, whose traces frame a triangular zone centered on Ocoa Bay where the seismic hazard is very high. As described, this sector includes the intensity center estimated for the

1751 Azua earthquake ( $M_w$  7.5), which according to some historical accounts gave rise to a tsunami in the bay. For Bakun et al., (2012), this event could be produced by the Enriquillo-Platain Garden fault zone or, as an alternative source, the Muertos deformation belt. The results of the PGA modeling rule out the EPGFZ as a source and indicate that the BRFZ, the O-MT and, less likely, the O-OBFZ, could have generated this earthquake.

Historical records indicate that the greatest effects of the 1615, 1684, 1691 and 1911 earthquakes ( $M_w$  from 6.0 to 7.5) were in the southern sector of the Central Cordillera. However, the determination of its seismic source could not be confidently established with the available data, since these strong earthquakes did not produce known surface ruptures. Future neotectonic and paleoseismological studies, together with the seismic hazard modeling, may shed light on the active fault zones that generated large earthquakes in the region and therefore allow updating the spatial distribution of the seismic hazard.

## **7. Conclusions**

The characterization of active tectonics and the main seismotectonic faults in southern central Hispaniola was developed from tectonic and geomorphologic field observations, the regional gravimetric and magnetic data interpretation, and the inversion of fault-slip data. The main conclusions of this work are:

(1) A relatively constant NE-directed shortening controlled the geometry and kinematics of main active faults in central southern Hispaniola, as well as the evolution of the Quaternary stress regime.

(2) Quaternary stress regime evolution includes a compressional D1 event, which gave rise to the large-scale fold and thrust structure in the Cordillera Central, Peralta Belt, Sierra Martín García and San Juan-Azua basin.

(3) D1 was followed by a strike-slip D2 stress regime, partitioned into the N-S to NE-SW transverse Ocoa-Bonao-La Guácara and Beata Ridge fault zones, which are related to indentation of the Beata Ridge in southern Hispaniola from the Early to Middle Pleistocene and continues today.

(4) D2 was coeval by a more heterogeneous and geographically localized D3 extensional deformation.

(5) Several seismotectonic fault zones divide the region into a set of simplified seismogenic zones.

(6) Modeled highest peak ground acceleration values establish a very high seismic hazard. in the Ocoa Bay.

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## **Open Research**

The data for this paper are contained in the text, figures and supporting information and can also be found in the data repository DIGITAL.CSIC which is the institutional repository of the Spanish National Research Council (Escuder-Viruete et al., 2023).

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## 1508 **Figure Captions**

1509 Fig. 1. Geodynamic reconstructions made with *Gplates* 2.3.0 software for arc-continent and arc-  
 1510 oceanic plateau collisions in northern and southern Hispaniola, respectively. Different tectonic  
 1511 elements are color coded and described in the key. (a) Reconstruction at 80 Ma (lower Campanian):  
 1512 the large white area in the present-day location of the central Caribbean represents the proto-  
 1513 Caribbean oceanic crust that will be subducted by the NE motion of the intra-oceanic Caribbean  
 1514 island-arc. Dark grey areas represent zones of thickened Caribbean oceanic plateau (CLIP;  
 1515 Caribbean Large Igneous Province). (b) Reconstruction at 40 Ma (middle Eocene): at this time the  
 1516 Caribbean island-arc collided with the southern North American continental margin and is moving  
 1517 eastward. The small white area adjacent to southern Hispaniola will start to subduct due to back-  
 1518 thrusting in the Peralta-Muertos accretionary prism. (c) Present-day: the ENE-directed convergence  
 1519 led to Caribbean plate subduction and Beata Ridge collision in southern central Hispaniola. The blue  
 1520 line represents the path of a point (blue star) on the Sierra Batoruco in the Caribbean plate since 80

Ma. *GPlates* reconstructions show that the direction of motion of the central Caribbean plate abruptly changes from northeastward to east-northeastward in the middle Eocene. This means that the direction of subduction/collision is highly oblique along the E-W-striking Peralta-Muertos deformed belt in southern Hispaniola. AVR, Aves Ridge; BR, Barbados Prism; BE, Beata Escarpement; BOFZ, Boconó Fault Zone; BOFZ, Bowin Fault Zone; BFZ, Bucaramanga fault zone; CAT, Cayman Trough; CHO, Chortis block; CGFZ, Cerro Golden Fault Zone; CR, Costa Rica; CU, Cuba; EPFZ, El Pilar Fault Zone; EPGFZ, Enriquillo-Platain Garden Fault Zone; GRB, Grenada basin; H, Hispaniola (Dominican Republic and Haiti); HP, Haiti Plateau; HE, Hess Escarpement; HB, Hispaniola Basin; LA, Lesser Antilles; MB, Maracaibo basin; MES, Mesquito terranes; MAT, Middle America Trench; MC, Mona Canyon; MP, Mona Passage; MDB, Muertos Deformed Belt; MT, Muertos Trough; MPFZ, Motaguá Polochic fault zone; NHDB, North Hispaniola Deformed Belt; NHFZ, North Hispaniola Fault Zone; NPFB, Northern Panamá Fold Belt; OFZ, Oriente fault zone; P, Panamá; PB, Peralta Belt; PR, Puerto Rico; PRT, Puerto Rico Trench; SFZ, Septentrional Fault Zone; SCDB, South Caribbean Deformed Belt; SFB, South Florida basin; TB, Tobago basin; VDB, Venezuela Deformed Belt; VI, Virgin Islands; YUC, Yucatán block.

Fig. 2. (a) Map of the northeastern margin of the Caribbean Plate showing the location of plate and microplate boundaries, as well as the main tectonic structures. The red arrow defines the movement vector of 18-20 mm/a in the direction N070°E of the Caribbean Plate with respect to the North American Plate (mod. Mann et al., 2002). Relief in color scale has been made from the GMRT synthesis data set (Ryan et al., 2009) with *GeoMapApp* ([www.geomapapp.org](http://www.geomapapp.org)). The discontinuous red rectangle defines the situation of the study area in the Dominican Republic. (b) Schematic geologic map of southern Hispaniola (mod. from the SYSMIN Project; Pérez-Estaún et al., 2007). The discontinuous red line marks location of Fig. 3. (c) Geological cross-section of the study area. The location is shown in (b). Abbreviations as in Fig. 1, in addition: CFZ, Camú fault zone; HFTB;

Haitian fold-and-thrust belt; OBFZ, Ocoa-Bonao-La Guacara fault zone; SFZ, Septentrional fault zone; SJPFZ, San Juan-Los Pozos fault zone; TBFZ; Trois Baies fault zone.

Fig. 3. Neotectonic map of southern central Hispaniola. Shaded relief in grayscale has been made from the GMRT synthesis data set (Ryan et al., 2009) with *GeoMapApp* ([www.geomapapp.org](http://www.geomapapp.org)). The neotectonic structures and late Neogene and Quaternary lithostratigraphic units compiled in the map result from integrating new field data with the geologic map obtained by the SYSMIN Project in the Dominican Republic (Pérez-Estaún et al., 2007). The main neotectonic structures are: Offshore (O-OBFZ) and Southern (S-OBFZ) segments of the Ocoa-Bonao-La Guacara fault zone; Central (C-SFZ) segment of the Septentrional fault zone; Southern (S-HFZ), Central (C-HFZ) and Northern (N-HFZ) segments of Hispaniola fault zone; Southern (S-JPFZ) and Central (C-JPFZ) segments of the San Juan-Los Pozos fault zone; and Northern (N-BRFZ) and Central (C-BRFZ) segments of the Beata Ridge fault Zone. Other relevant structures are the Bahoruco fault zone (BAFZ) and the Ocoa segment of the Muertos Trust (O-MT). Abbreviations as in Fig. 1, in addition: AZ, Azua; BA, Barahona; BN, Baní; BO, Bonao; CO, Constanza; JA, Jánico; JB, Jarabacoa; LV, La Vega; MA, Mao; MN, Manabao; RA, Rancho Arriba; SB, Sabana Buey; SC, Santiago de los Caballeros; SJ, San José de Ocoa; SQ, Sábana Quéliz; VA, Villa Altagracia.

Fig. 4. (a) Bouguer gravity anomaly map of southern central Hispaniola showing major tectonic features delineated by sharp gravity gradients (source of gravity data: <https://www.nga.mil>). Areas of exposed or shallow igneous and metamorphic basement of Cretaceous age in the northern Cordillera Central and Sierra Bahoruco show higher anomalies (red to yellow tones), whereas areas with high sediment accumulation in the Cibao, San Juan and Azua basins are expressed by low anomalies (dark to light blue tones). (b) Reduced to the magnetic pole map showing major lithologic and tectonic features of southern central Hispaniola (mod. from García Lobón & Rey Moral, 2004; Ayala et al., (2017). The reduction to the pole of the whole magnetic field improves the spatial position of



anomalies that help to define areas with long and short wavelengths, regional trends and tectonic boundaries between magnetic provinces. See explanation in the text.

Fig. 5. Structural analysis of the Peralta Belt. (a) Structural map of the eastern Peralta Belt in southern central Hispaniola (see location in Fig. 3), showing the tectonic domains, main Quaternary lithostratigraphic units, neotectonic structures, and sites of fault-slip data measurements. OBFZ, Ocoa-Bonao-La Guacara fault zone; SJ, San José de Ocoa intramountain basin; RA, Rancho Arriba intramountain basin. (b) Stereoplots of the non-coaxial Sp-Lp fabric produced by syn-sedimentary deformation during the upper Eocene in the Peralta and Ocoa Groups. (c) Field aspect of the syn-sedimentary deformation developed heterogeneously in the turbidites of the Ocoa Group. (d) Zone of stratal disruption in the Ocoa Group, characterized by boudinage of sandstone beds, tight to isoclinal folds with rootless limbs, and a pervasive scaly clay fabric (Sp) in mudstone interbeds. (e) Detail of the boudinaged sandstone beds and the S-C structures in the mudstones.

Fig. 6. Structural analysis of the Peralta Belt. (a) Stereoplots of the principal stress axes obtained from inversion of fault-slip data. n, number of data. Inversion methods: PTM, P-T Method; RDM, Right Dihedra Method; DIM (Direct Inversion Method); and NDA, Numerical Dynamic Analysis Method. (b) Asymmetric D1 folds of SW vergence, associated with mid-dip angle reverse faults, developed in the limestones of the lower Miocene Sobrerito Formation. (c) D1 reverse faults cutting the block-in-matrix fabric (Sp) and calcite veins in the Peralta Group.

Fig. 7. Structural analysis of the Peralta Belt. (a) Stereoplots of the principal stress-axes obtained from inversion of fault-slip data. (b) Structural map of the northern Ocoa Bay in southern central Hispaniola (see location in Fig. 3), showing main lithostratigraphic units, neotectonic structures, and sites of fault-slip data measurements. (c) Geological cross-section of the frontal part of the Peralta Belt. (d, e) Field aspect of the Loma Vieja frontal thrust (site 21JE123, see location in the cross-section), where the lower to middle Miocene limestones of the Sobrerito Formation overthrust the

conglomerates of the late Pliocene to lower Pleistocene Arroyo Seco Formation. Acronyms as in Fig. 3.

Fig. 8. Structural analysis in the Loma Vigia sector of the Ocoa Bay. (a) Stereoplots of the principal stress axes obtained from inversion of fault-slip data. (b) D1 fold and reverse fault truncated by D3 normal faults, dragging the limestone layers in the hanging-wall block to the SW (site 21JE120). (c, d) Limestone beds of the Sombrerito Formation displaced by a system of ENE to E-striking normal faults along whose planes intrude locally mafic magmas (site 21JE113).

Fig. 9. Structural analysis in the Sierra Martín García. (a) Stereoplots of the principal stress axes obtained from inversion of fault-slip data. (b) Structural map of the eastern Sierra Martín García in southern central Hispaniola (see location in Fig. 3), showing main lithostratigraphic units, neotectonic structures, and sites of fault-slip data measurements. (c) Stratigraphic logs in Cañada Arenazo and Playa Caobita sites. (d) Aspect of the bands of fault-gouge and fine crush breccia several tens of meters thick associated with the strike-slip D2 deformation. View width is 40 m. (e) D3 normal faults deforming the alluvial fans deposits of Late Pleistocene to Holocene age?. View width is 18 m. (f) Coral-reef terrace of Late Pleistocene boundary age (MIS 5c) deformed by D3 normal faults. (g) Field appearance of the calcite fill in the planes of the D3 extensional faults that deform the coral terrace. Gm, matrix-supported, muddy-sandy conglomerate; Gp, clast-supported, sandy conglomerate; Sc, medium-to-thick bedded coarse-grained sandstone and microconglomerates; Sm, thin-bedded, medium-to-fine-grained sandstone; Sf, fine-grained sandstone, siltstone and laminated mudstone; Ms, massive variocolored mudstone; Ls, coral reef limestone.

Fig. 10. Structural analysis in northeast Sierra Bahoruco. (a) Stereoplots of the principal stress axes obtained from inversion of fault-slip data. (b) Geological cross-section of the Playa Azul site (see location in Fig. 3). (c) Panoramic view of the geological cross-section outcrop. (d) Field relationships

of the limestones of the Sombrerito Formation, coral-reef terrace of Middle to Late Pleistocene boundary age (MIS 5e stage) and gravels and red mudstones of the alluvial fan faulted by D2 strike-slip faults. (e) Field aspect of the gravels and red mudstones of the alluvial fan faulted by D3 extensional faults.

Fig. 11. Structural analysis in the Sierra Martín García. (a) Stereoplots of the principal stress axes obtained from inversion of fault-slip data. (b) Structural map of the southwestern Sierra Martín García (see location in Fig. 3), showing main lithostratigraphic units, neotectonic structures, and sites of fault-slip data measurements. (c) WNW-trending and SW-verging asymmetric D1 folds developed in the the marls and gypsum beds of the upper Pliocene La Salina Formation. (d) SW-verging asymmetric D1 folds associated with mid-dip angle reverse faults inclined to the NE developed in the gypsum deposits. The asymmetry of D1 folds and mesoscopic S-C structures establish a top-to-the-SW reverse movement. (e) Field relationships of superposition of D3 normal fault striations (e2) over D2 strike-slip striations (e1) deforming stratified gypsum beds (S0). (f) Accumulation of *Diploria* sp. forming the reef-platform facies of a coral terrace grown during the MIS 5e stage that fossilizes the D1 frontal thrust of the Sierra Martín García.

Fig. 12. Structural analysis in the San José de Ocoa basin. (a) Stereoplots of the principal stress axes obtained from inversion of fault-slip data. (b) Folded and faulted mudstones of the Numero Formation juxtaposed to tilted and faulted San José de Ocoa basin deposits through subvertical oblique reverse right-lateral D2 faults of the eastern branch of the S-OBFZ segment. (c) Low-pitch angle striations developed in subvertical fault planes subparallel to the S-OBFZ segment deforming the volcanic rocks of the Tireo Group. (d) Brecciated damage fault zone with low-pitch angle striations developed in subvertical D2 fault planes sub-parallel to the S-OBFZ segment. (e) Field overlapping relationships between low-pitch angle strike-slip striaes deformed by dip-slip normal striaes in the same fault plane, establishing a temporal order between D2 and D3 deformations.

Fig. 13. Structural analysis of the extensional D3 deformation in the Cordillera Central. (a) Stereoplots of the principal stress-axes obtained from inversion of fault-slip data. (b) Map of distribution of fossil coral reef terraces of the MIS 5e/5c and 1 stages, as well as the area affected by the D3b extensional event. See explanation in text. Shaded relief as in Fig.3.

Fig. 14. Structural analysis in the Cordillera Central. (a) Stereoplots of the principal stress-axes obtained from inversion of fault-slip data. (b) Late Cretaceous green tuffs of the Tiroo Group imbricated by WSW-directed D1 thrusts. (c) Detail of D3 extensional faults overlapping D1 thrusts in the green tuffs. (d) Panoramic view of Constanza mountainous area showing a D1 macrostructure of NW-striking thrusts built up on late Cretaceous volcano-plutonic rocks overthrusting the sedimentary fill of a Quaternary intramontaneous basin. Note the development of perched valleys and triangular facets in the hanging-wall block. (e) Detail of the striae associated to the development of a D2 sinistral strike-slip fault in late Cretaceous tonalitic rocks.

Fig. 15. (a) Stereoplots of the principal stress-axes obtained from fault-slip data inversion grouped by deformation events and (b) their classification according to its kinematic regime (see Supporting Information S6). The orientation of  $\sigma_1$  has been restored at sites that have undergone a post-D1 rotation. Contours in stereoplots at 1.0, 5.0, 10.0, 15.0, 20.0 and 25.0%. Computed mean stress axes for each deformation event are expressed in % and trend/plunge angles.

Fig. 16. Neotectonic diagrams showing the evolution of the Quaternary stress regime in southern central Hispaniola set up for (a) D1 reverse, (b) D2 strike-slip, and (c) D3 extensional events (Table 1). Maximum horizontal stress axis trends were derived from fault-slip data inversion. Shaded relief in grayscale has been made as in Fig. 3, indicating the overlay colour different geological domains: green, the Cretaceous basement of the Cordillera Central; pink, the middle Eocene-lower Miocene Peralta fold-and-thrust belt and the overlying uplifted forearc basin; yellow, the Tertiary limestones of the Sierras de Neiba, Martín García y Bahoruco; and pale blue, the Neogene San Juan-Azua

1665 Basin. Vectors of GPS velocities relative to the Caribbean plate at stations located in the study area  
1666 are from Symithe et al. (2015), Calais et al. (2016) and UNAVCO (<https://coconet.unavco.org>).  
1667 Acronyms as in Fig. 3. See text for explanation.

1668 Fig. 17. Result of the probabilistic seismic hazard modeling in southern central Hispaniola, as well as  
1669 adjacent offshore sectors of the Ocoa Bay, expressed as iso-PGA (Peak Ground Acceleration) zones  
1670 (in  $\text{cm/s}^2$ ) for a return period of 475 years (i.e., an exceedance probability of 10% in 50 years). The  
1671 modelled area is the same of Fig. 3 and framed by a discontinuous red box in Fig. 2. Figure also  
1672 includes the trace of the main seismogenic structures described in Table 2. MDT. Shaded relief and  
1673 acronyms as in Fig. 3. See text for explanation.

Figure1.

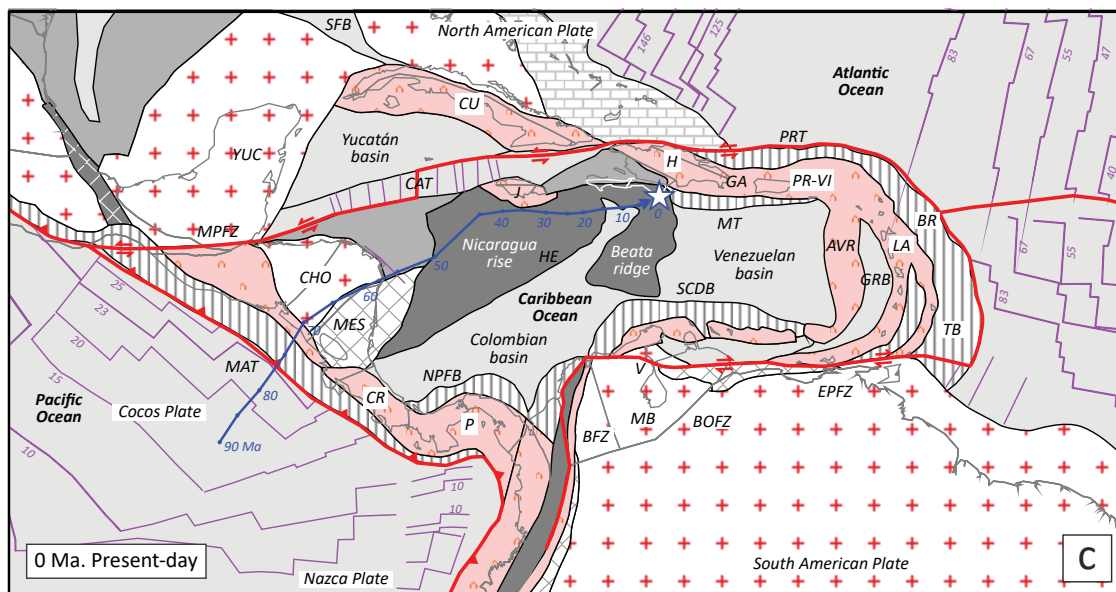
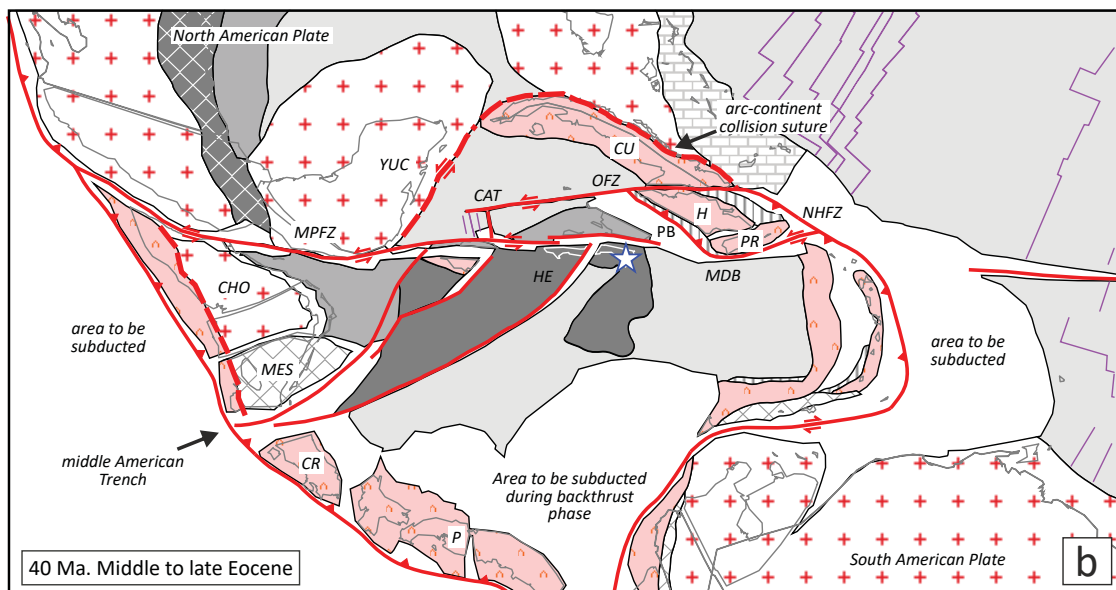
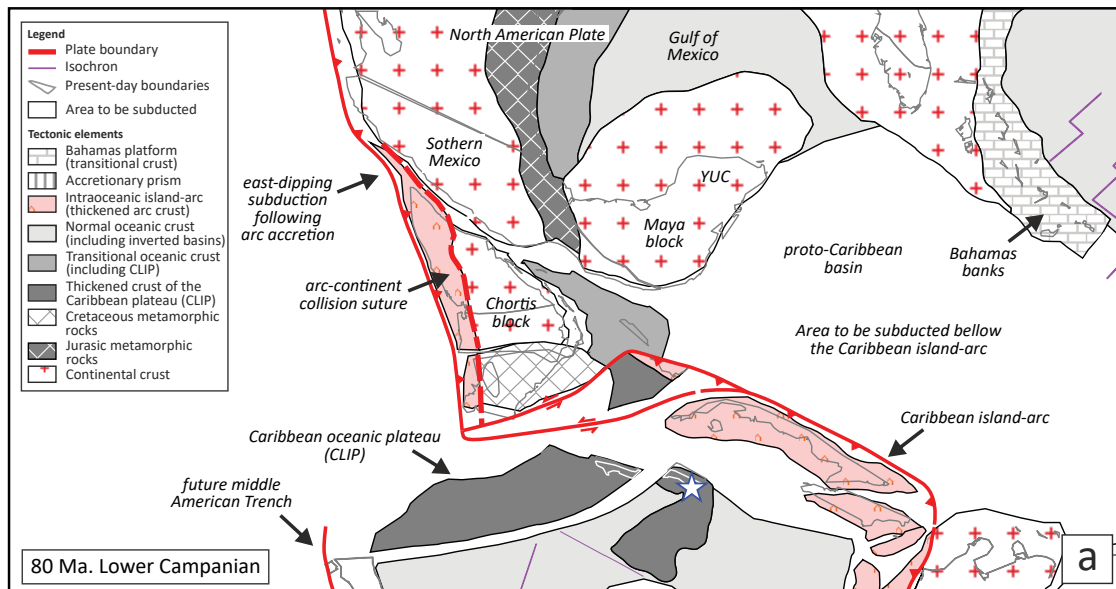
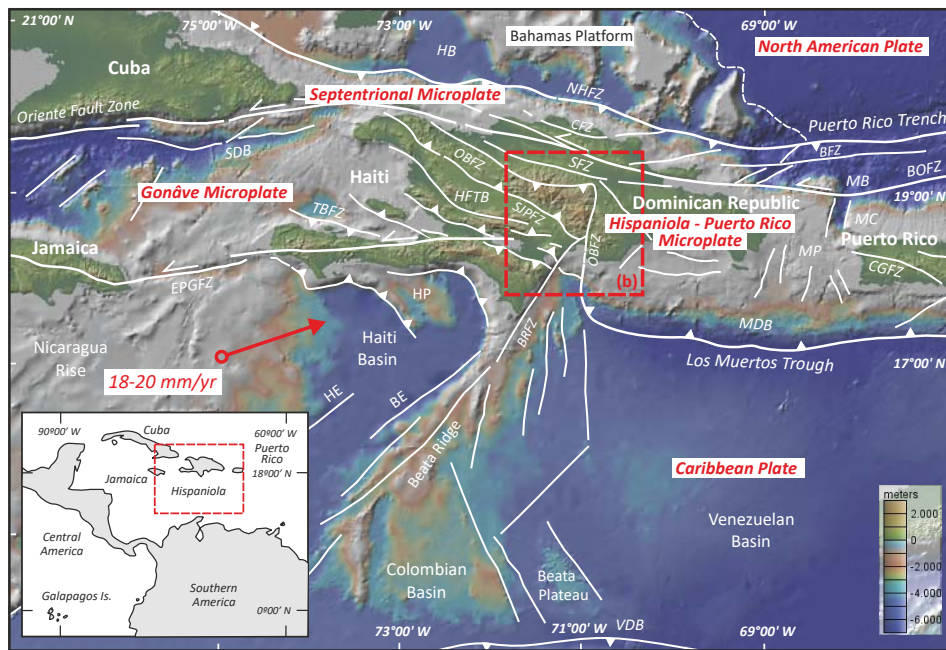
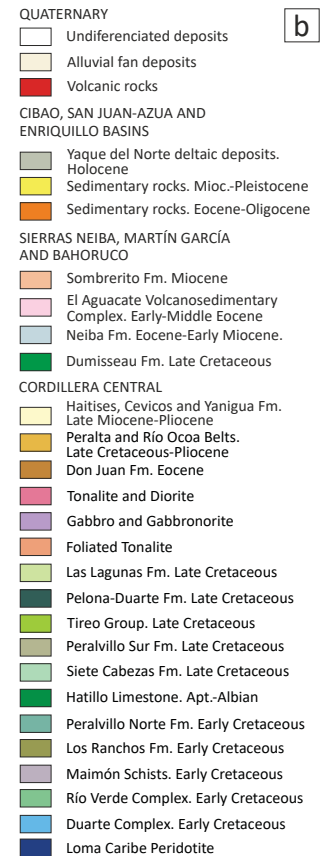
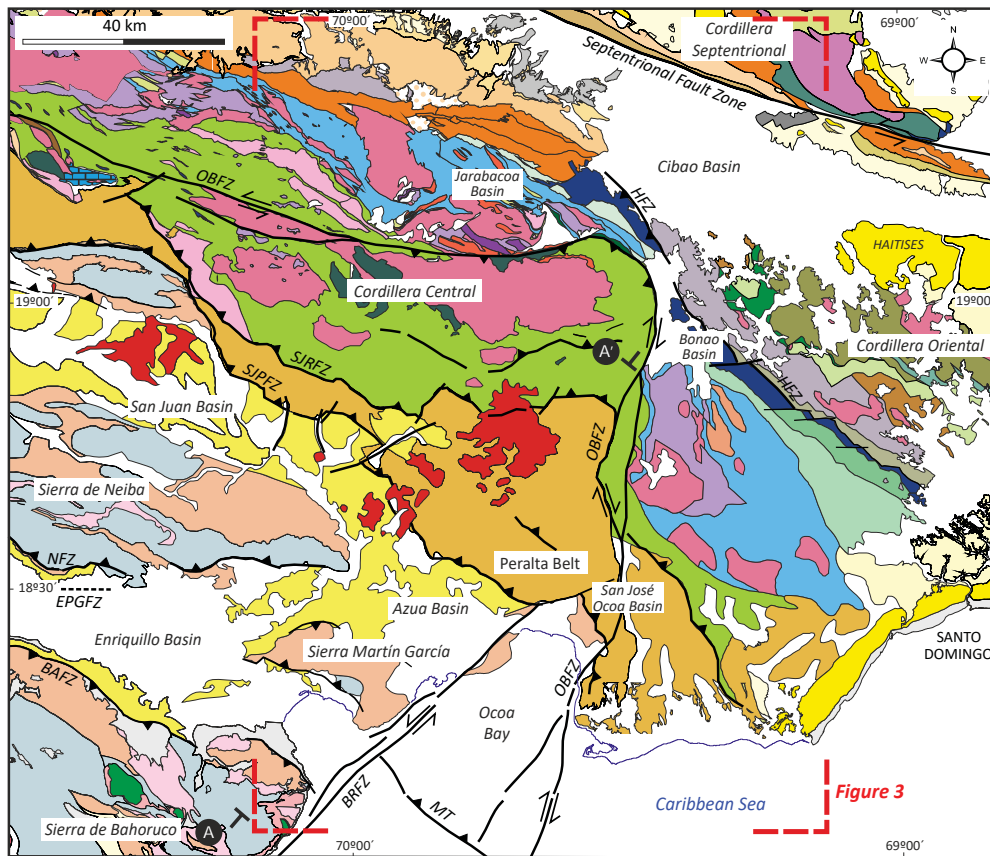


Figure2.

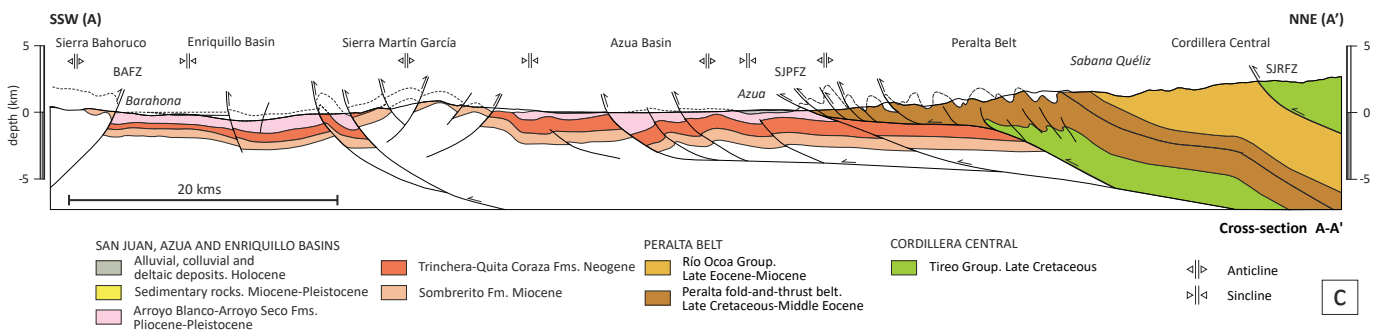




a



b



c

Figure3.



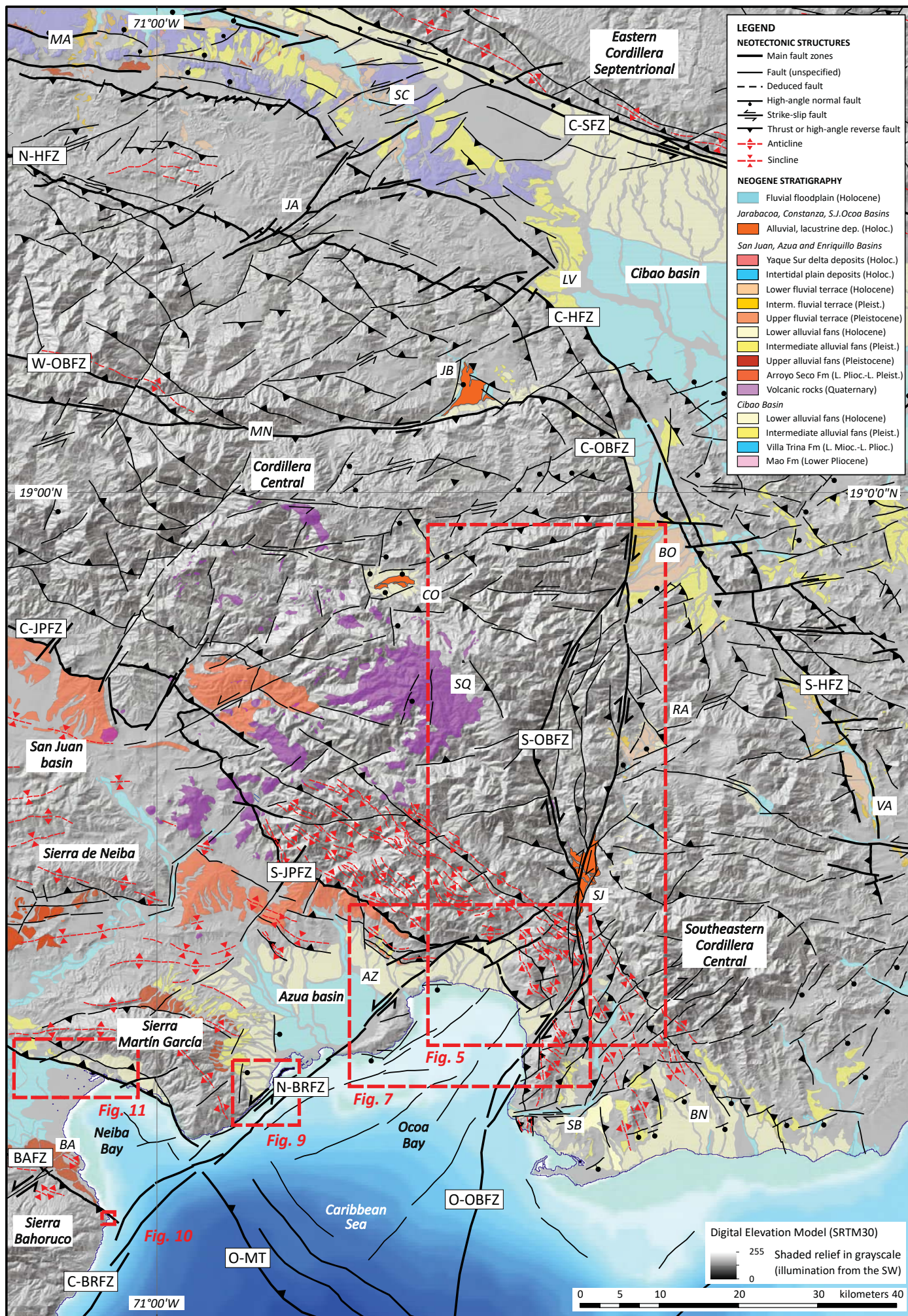


Figure4.



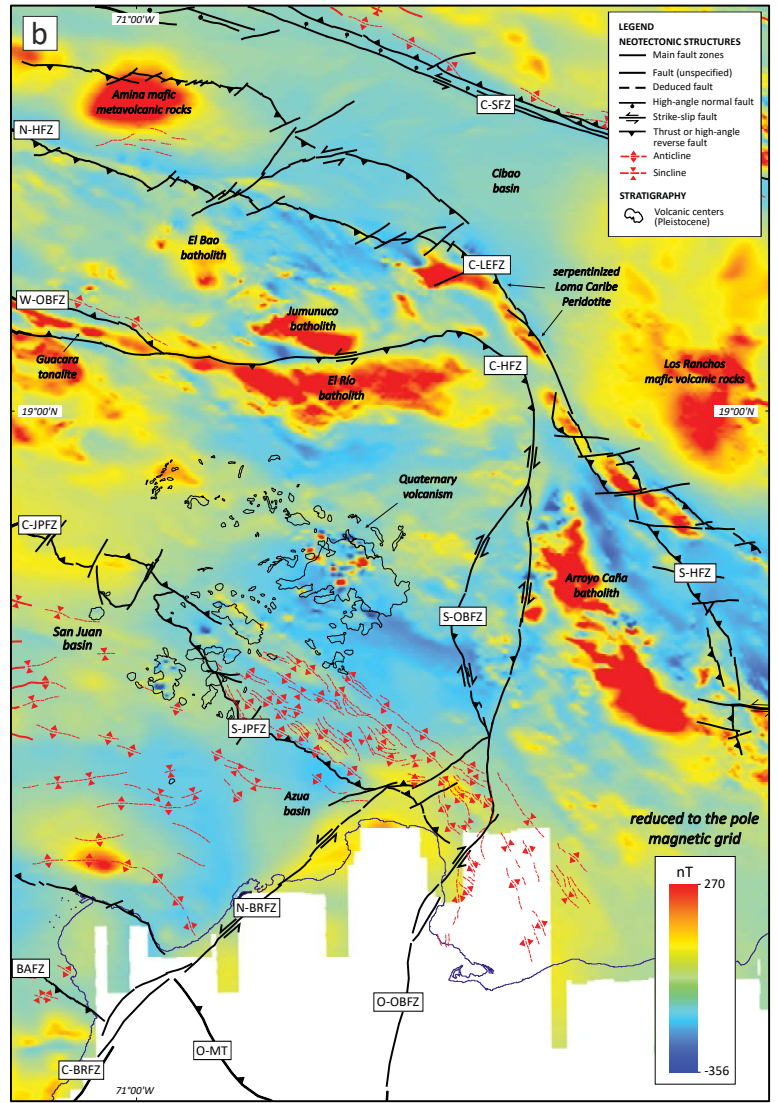
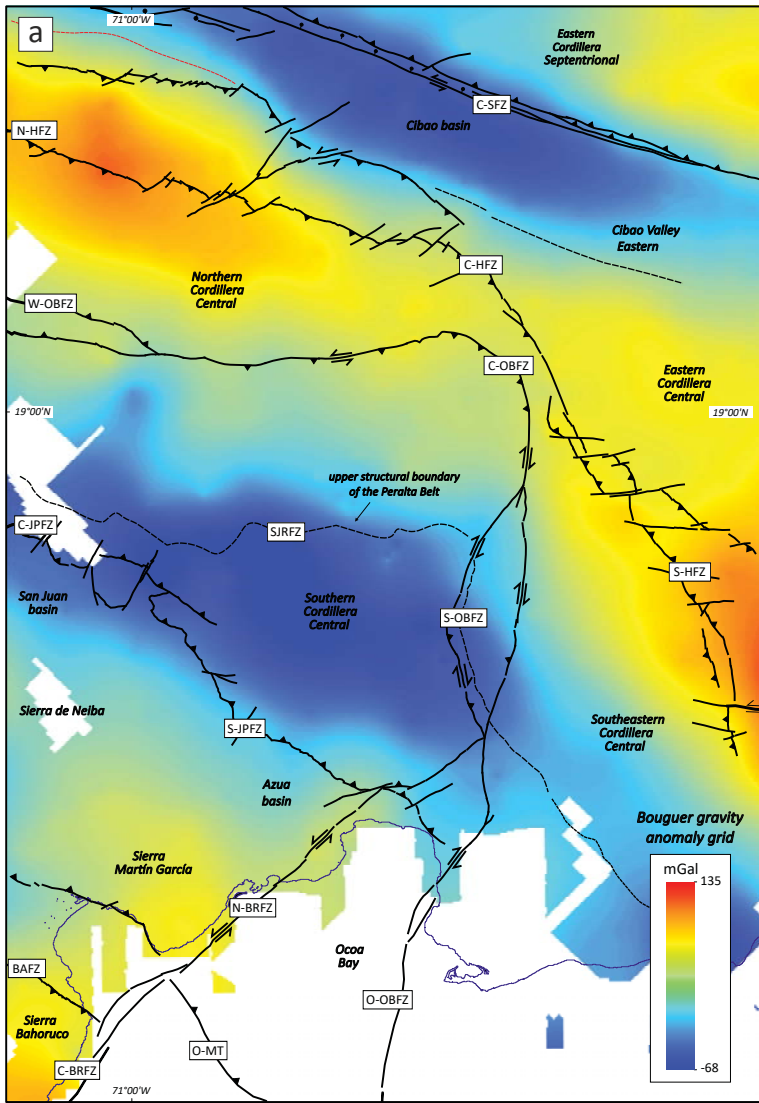
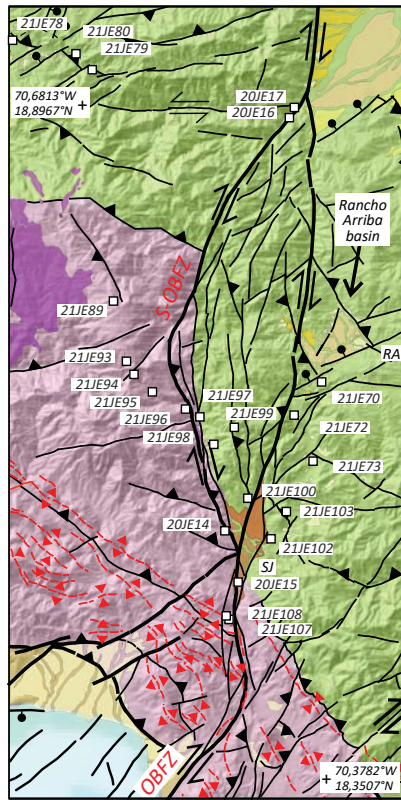


Figure5.

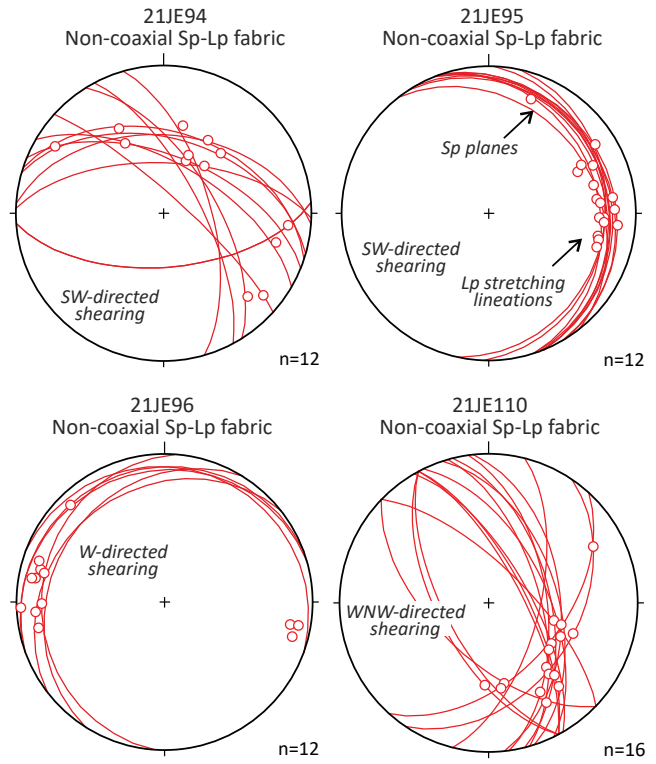




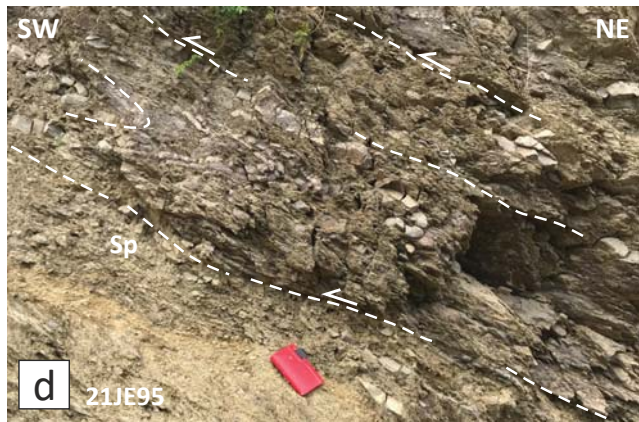
- Volcanic rocks**
- Mafic and intermediate lavas (Quaternary)
- San José Ocoa Basin**
- Alluvial and lacustrine deposits (Holocene)
- Azua Basin**
- Lower fluvial terrace (Holocene)
  - Intermediate fluvial terrace (Pleistocene)
  - Lower alluvial fans (Holocene)
  - Intermediate alluvial fans (Pleistocene)
- Peralta Belt**
- Peralta and Ocoa Groups
- Cordillera Central**
- Volcanic and plutonic rocks (Cretaceous)

a

### syn-sedimentary deformation, Peralta Belt



b



d

e

Figure6.



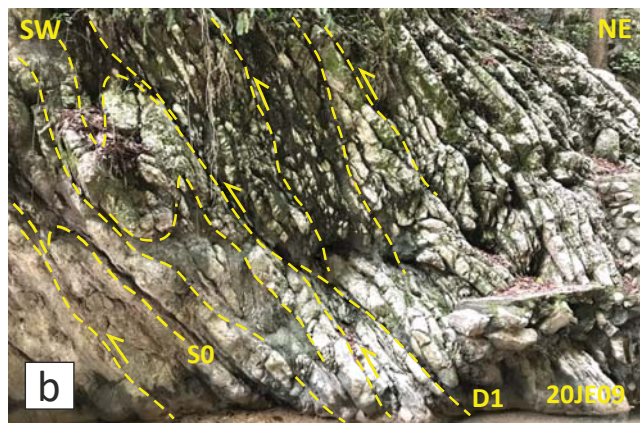
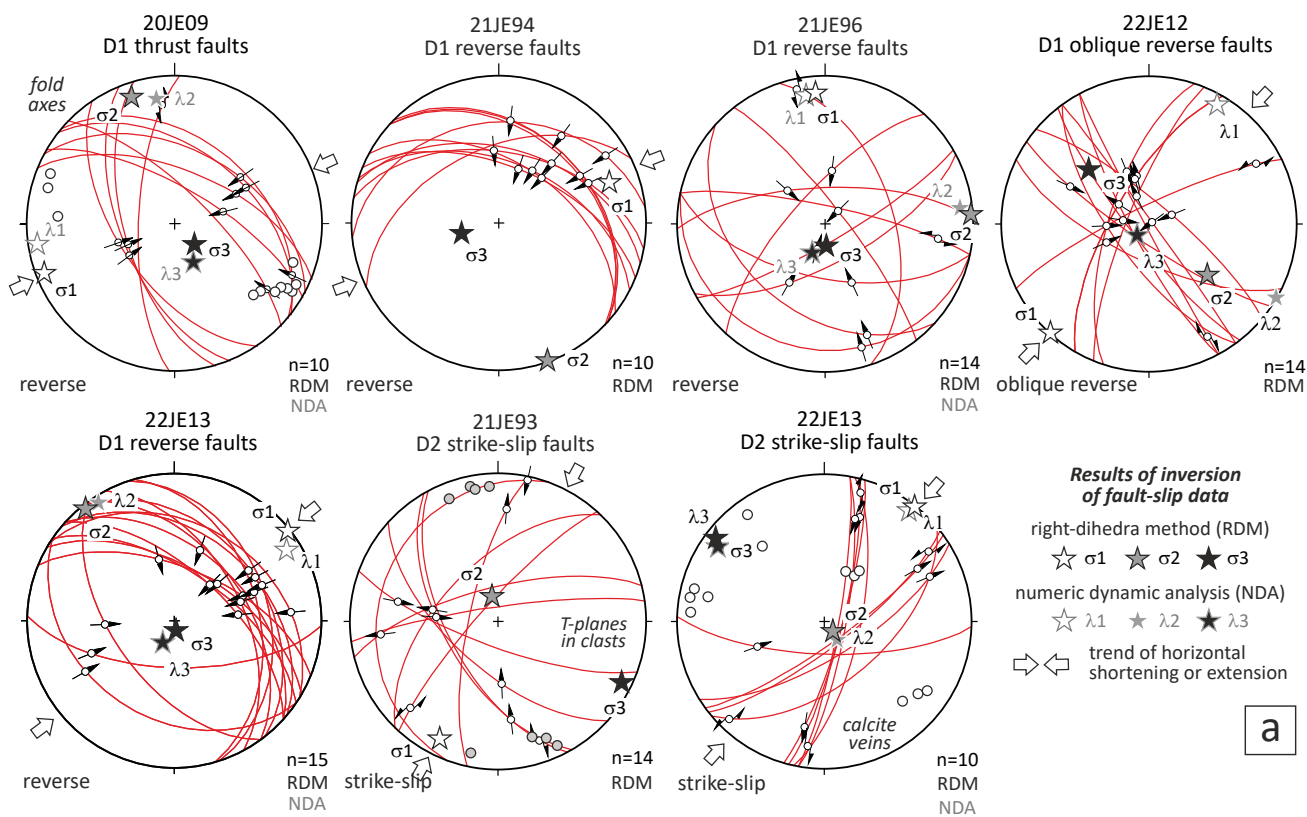


Figure7.

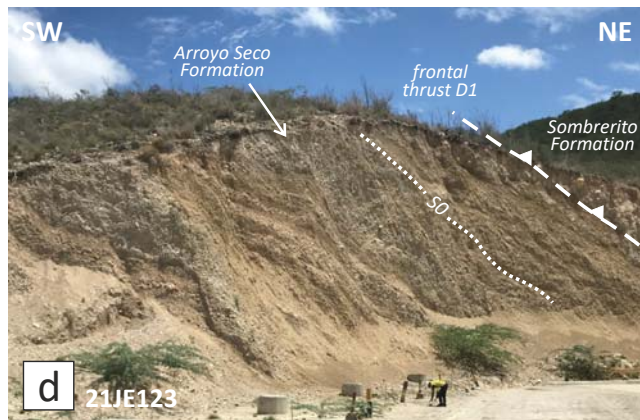
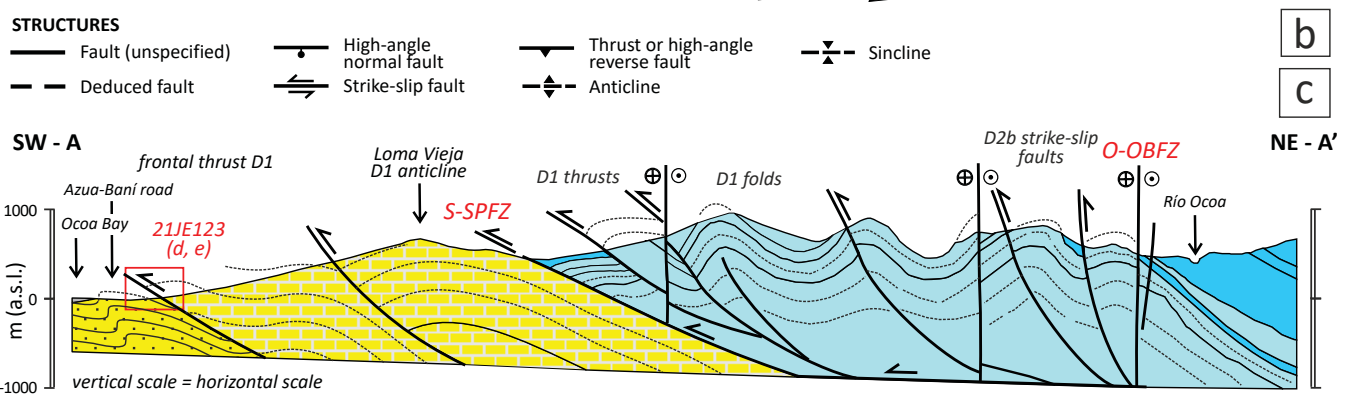
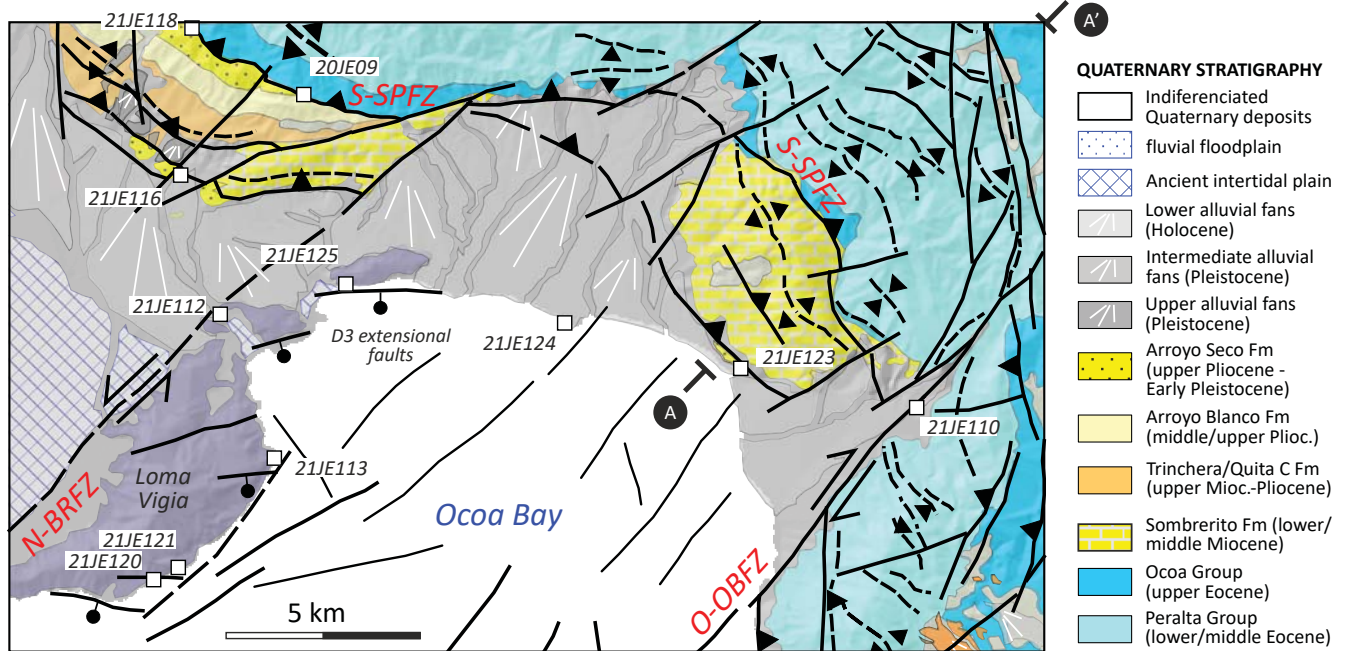
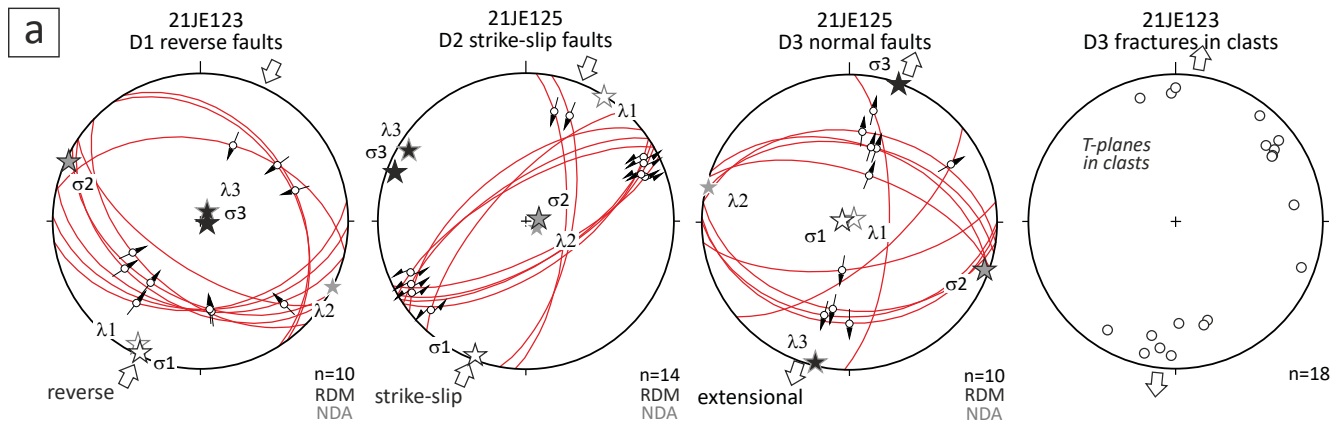


Figure8.



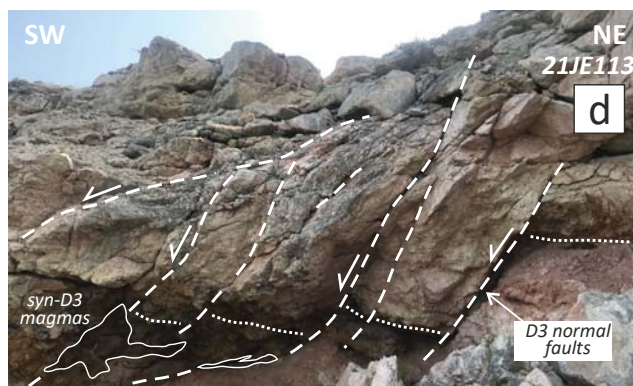
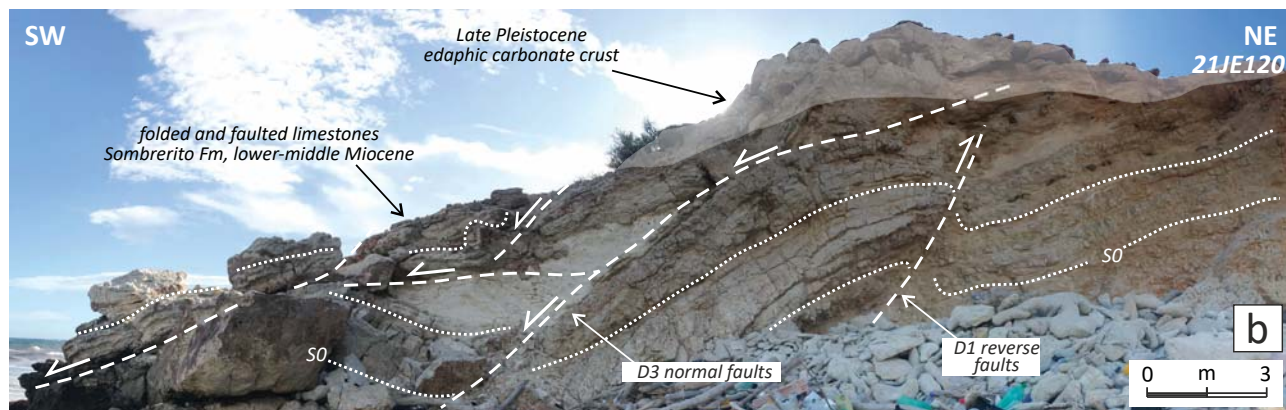
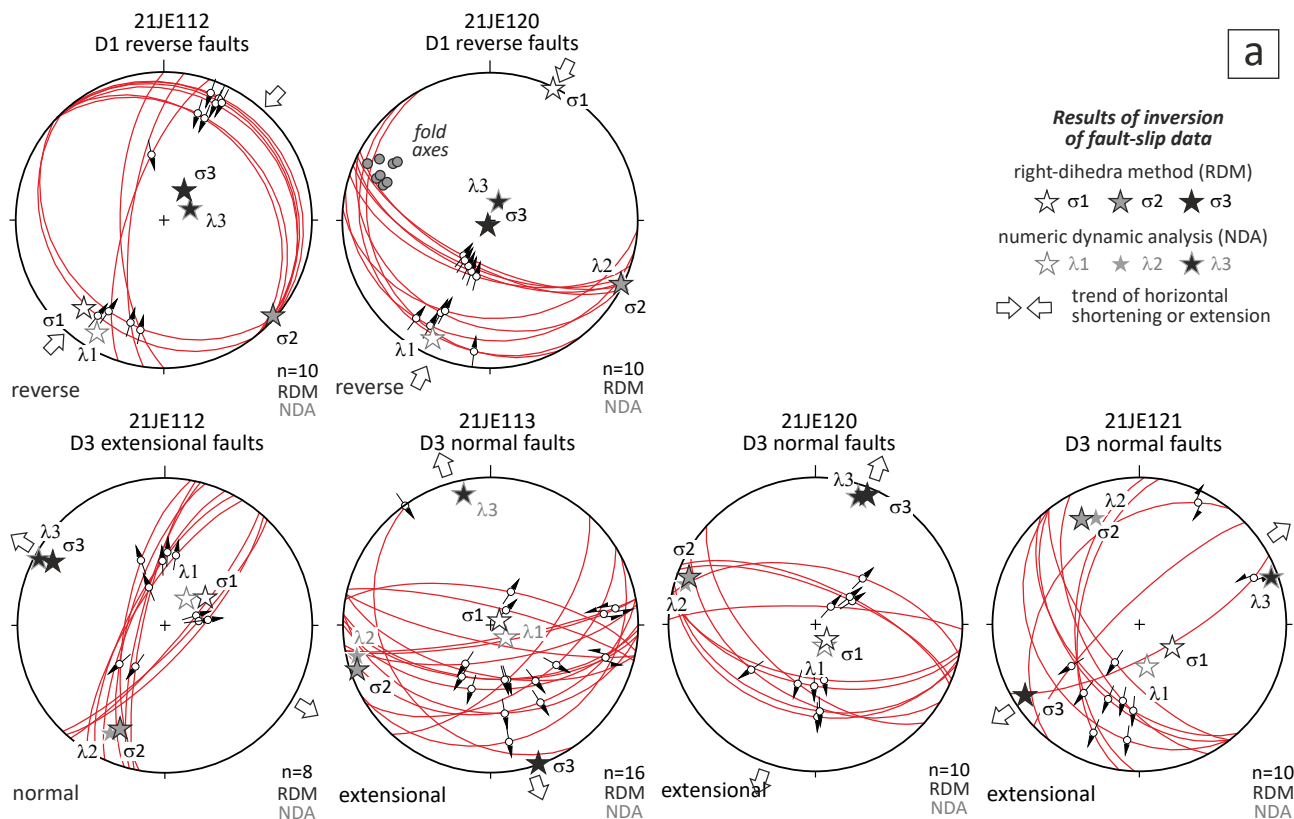


Figure9.



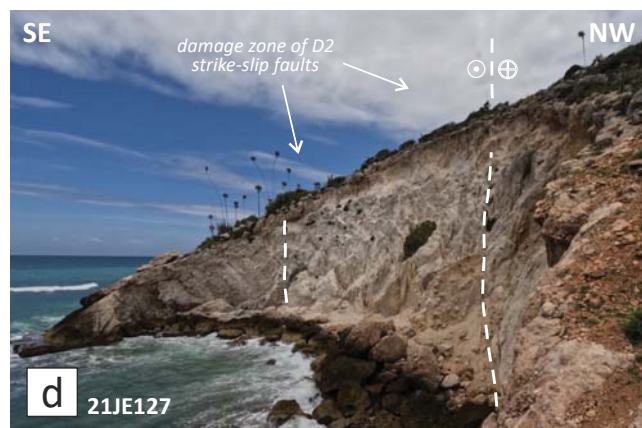
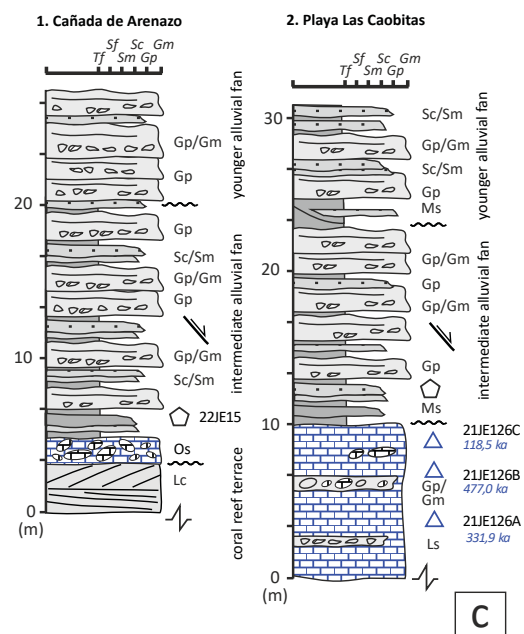
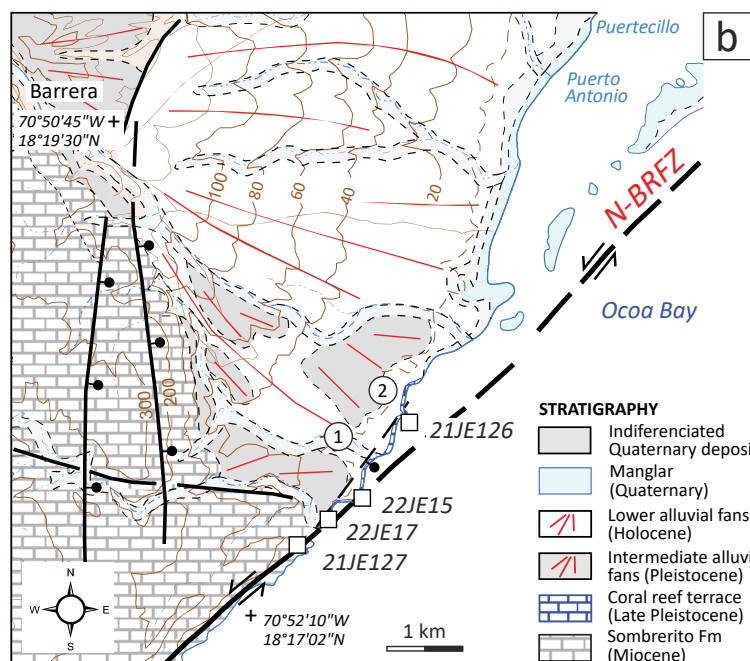
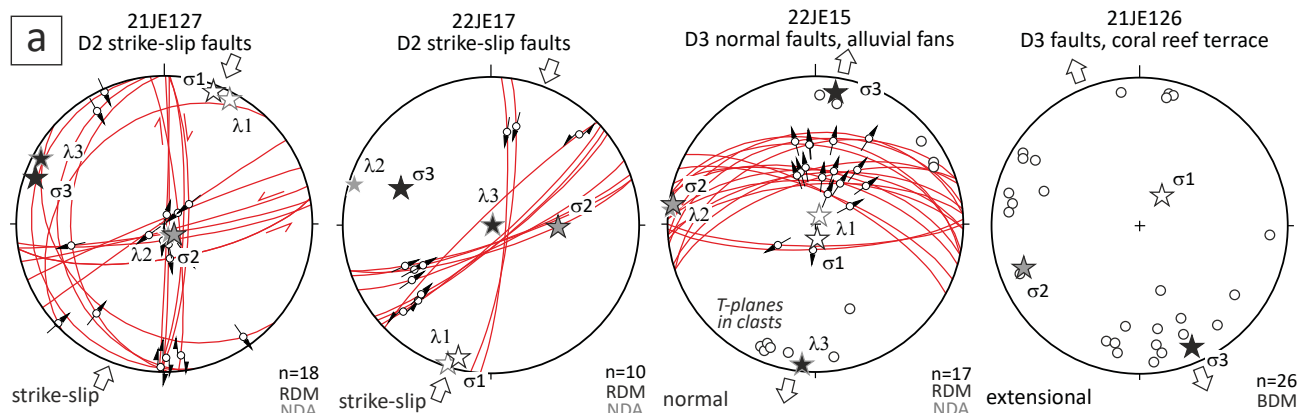
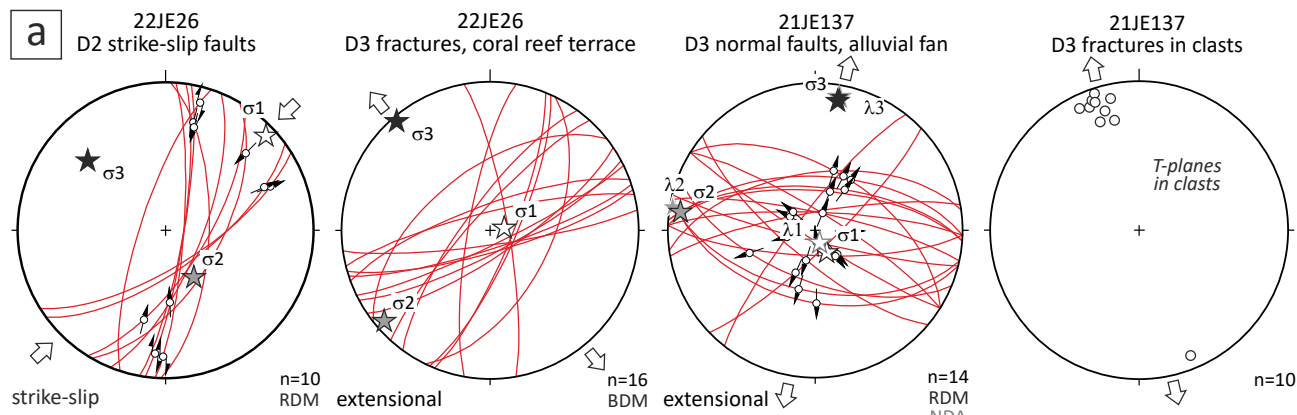


Figure10.





**SSW**

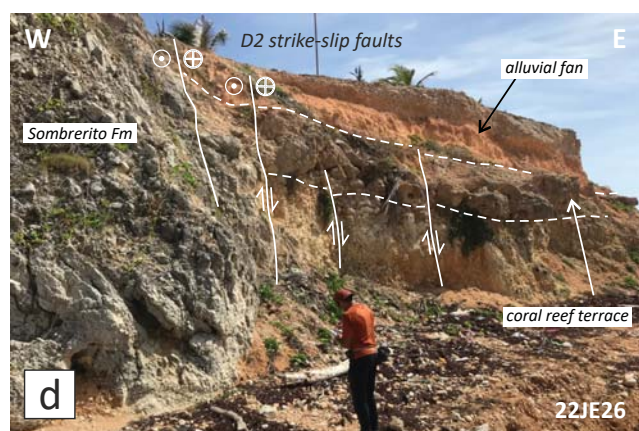
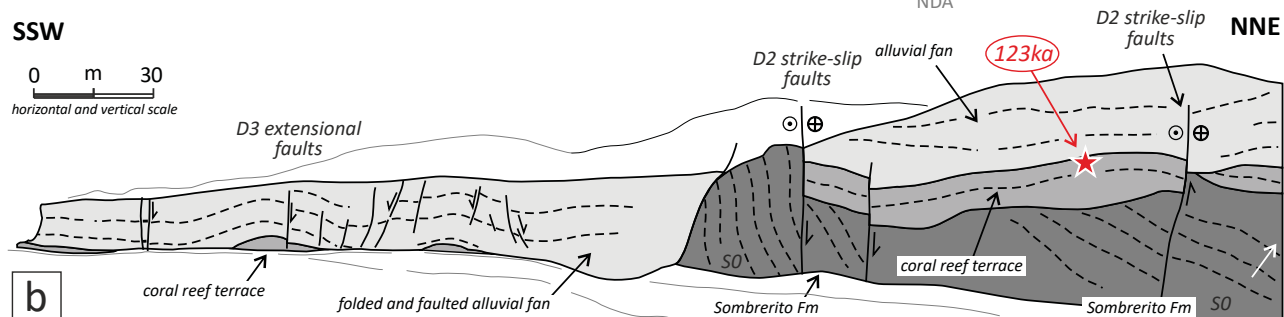
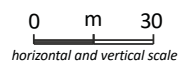


Figure11.



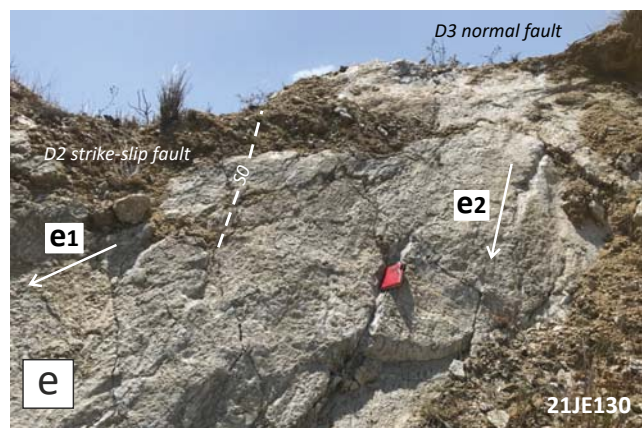
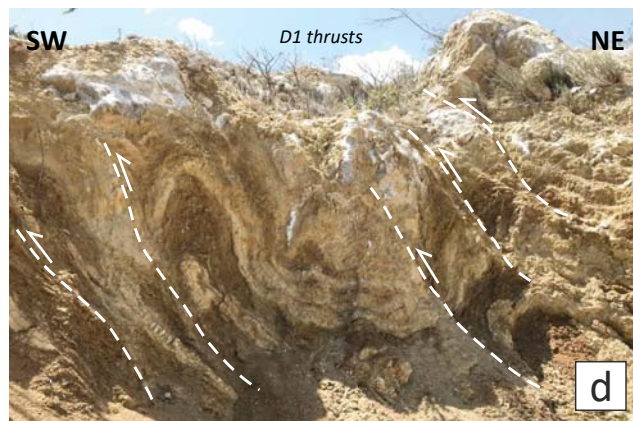
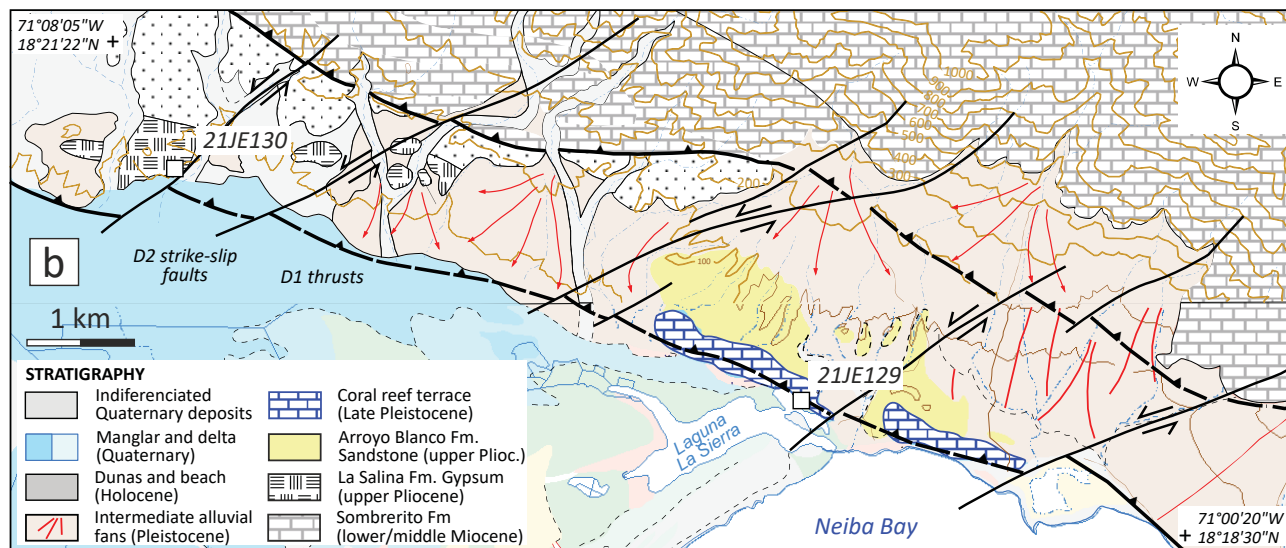
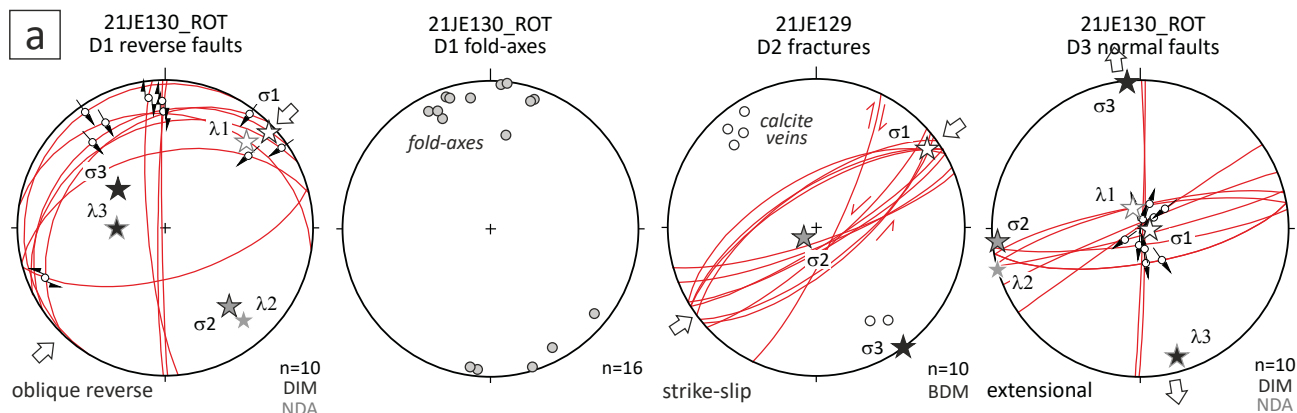


Figure12.



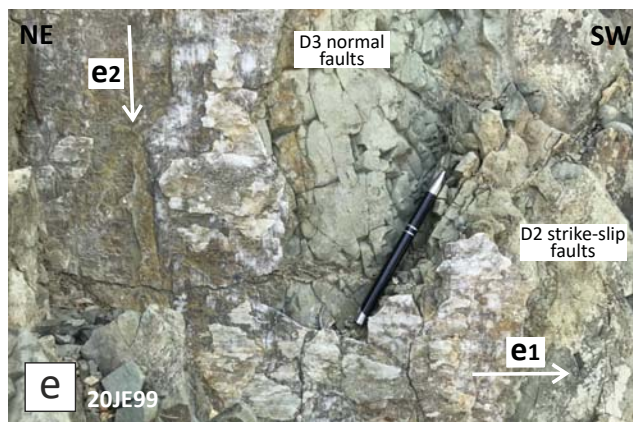
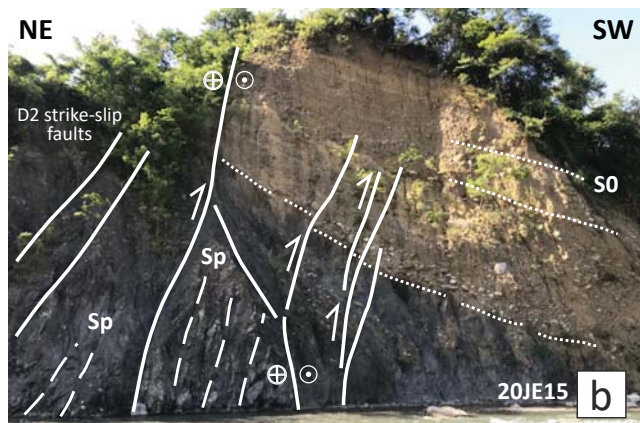
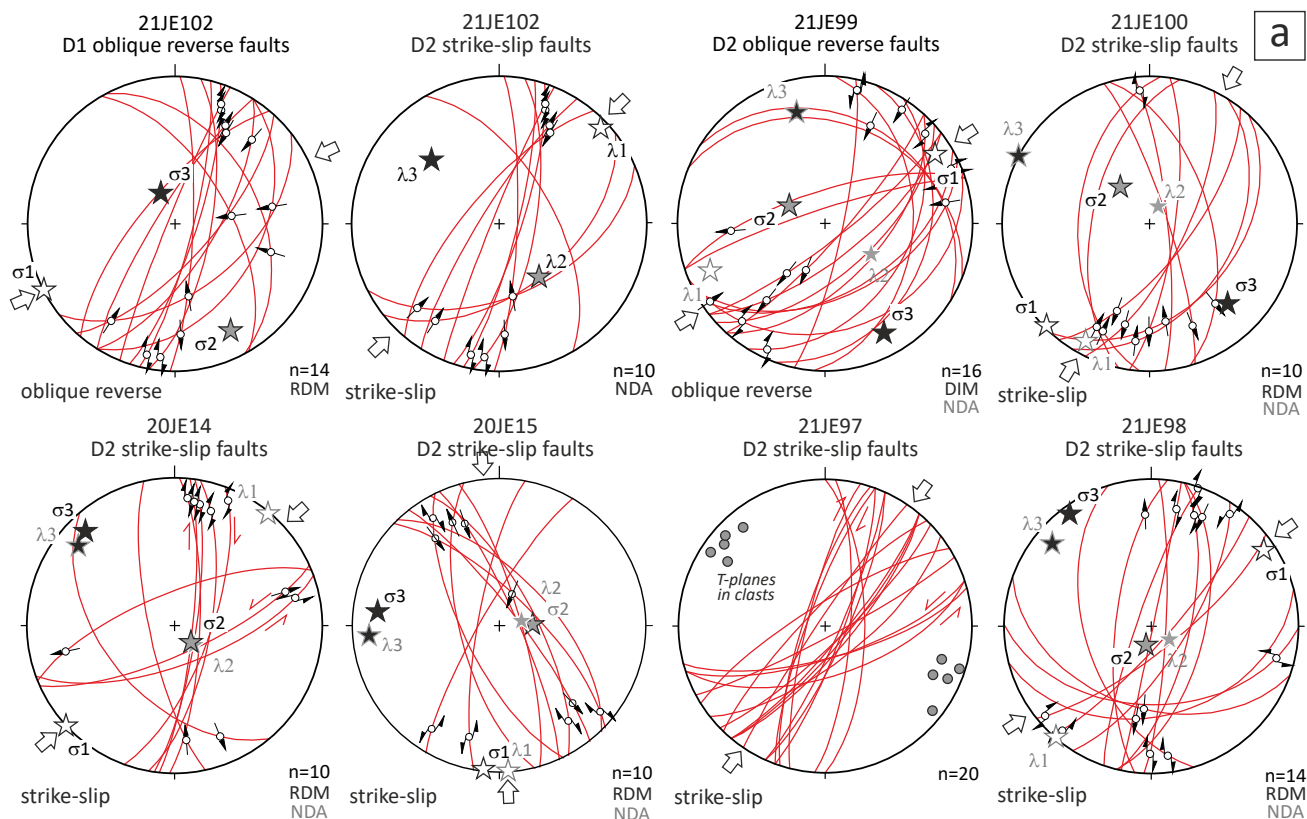


Figure13.

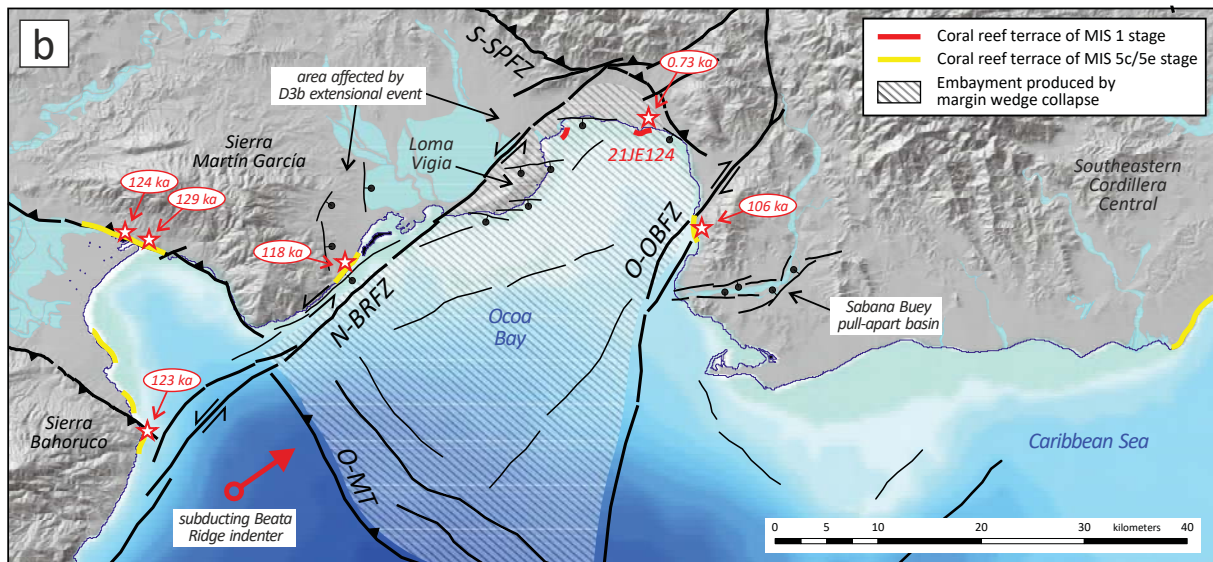
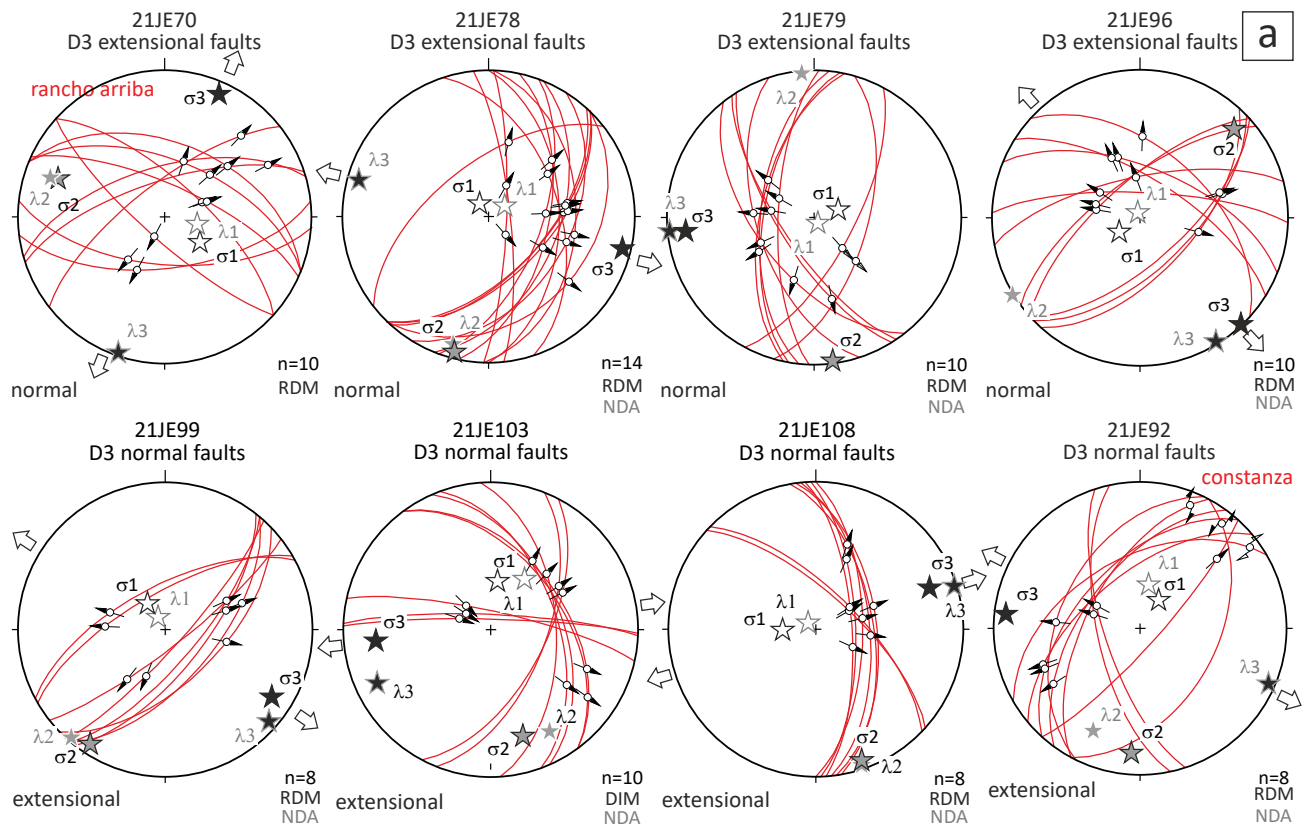


Figure14.



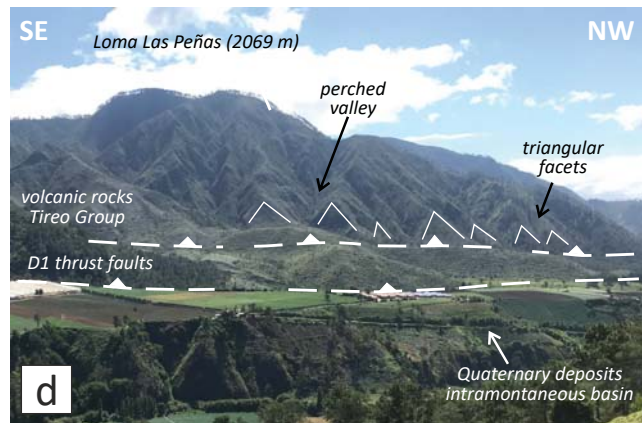
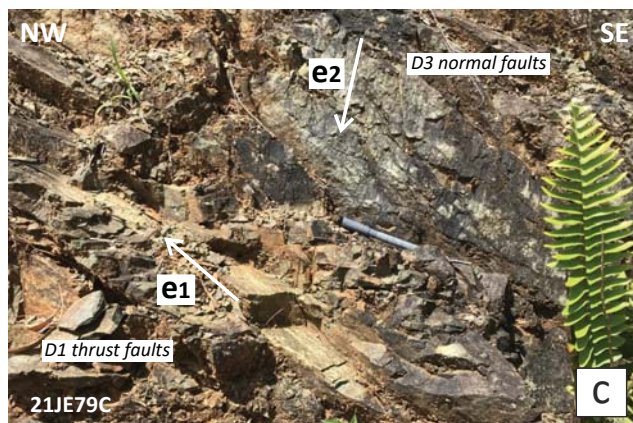
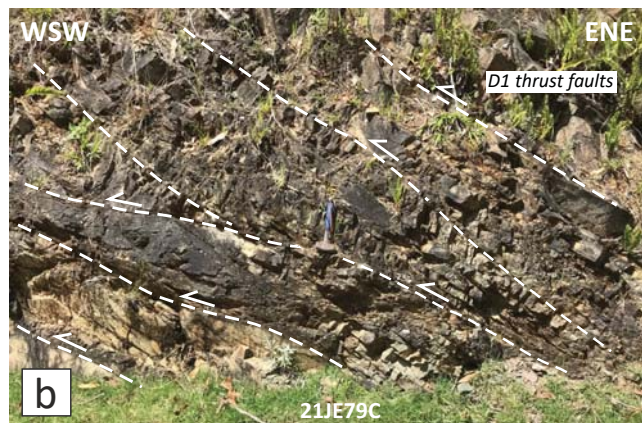
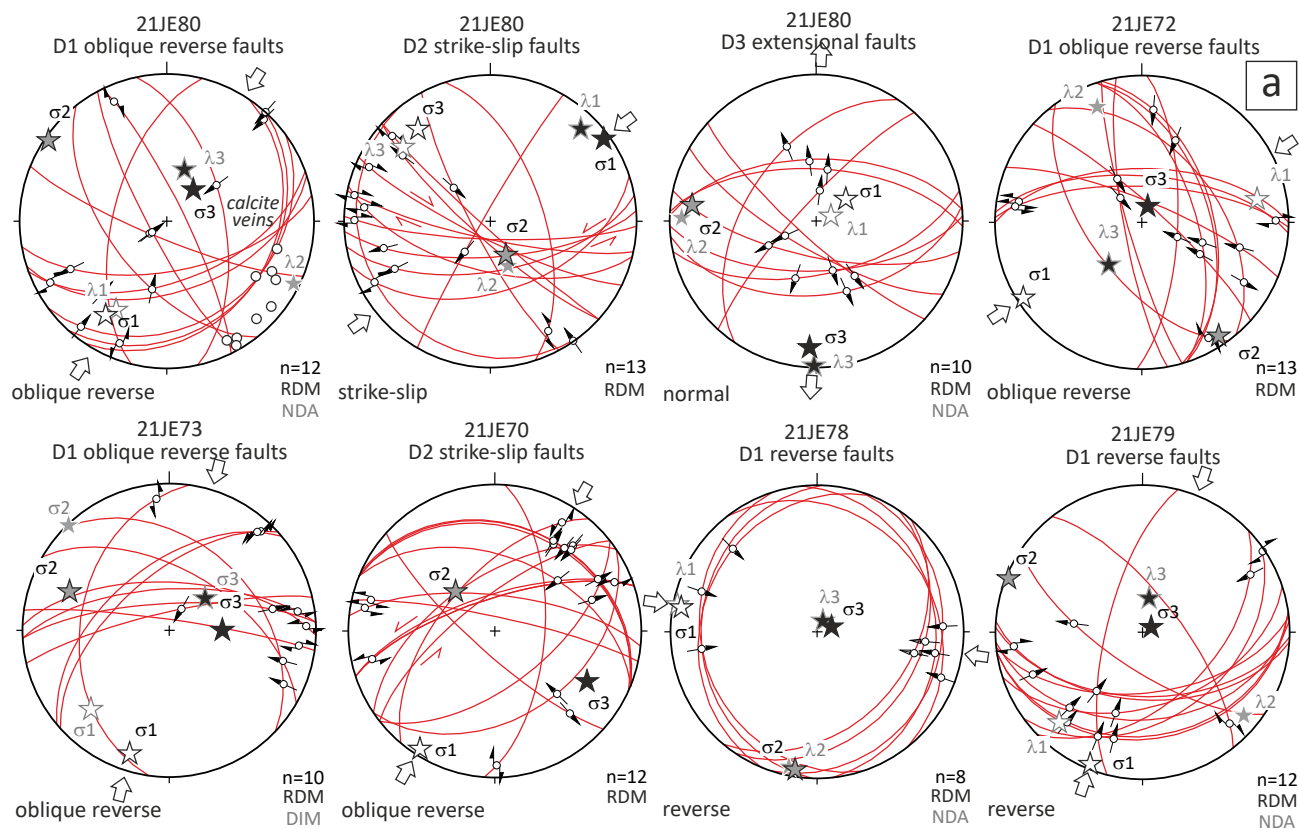
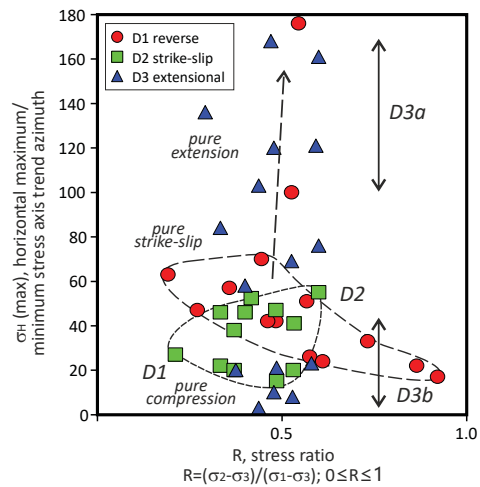
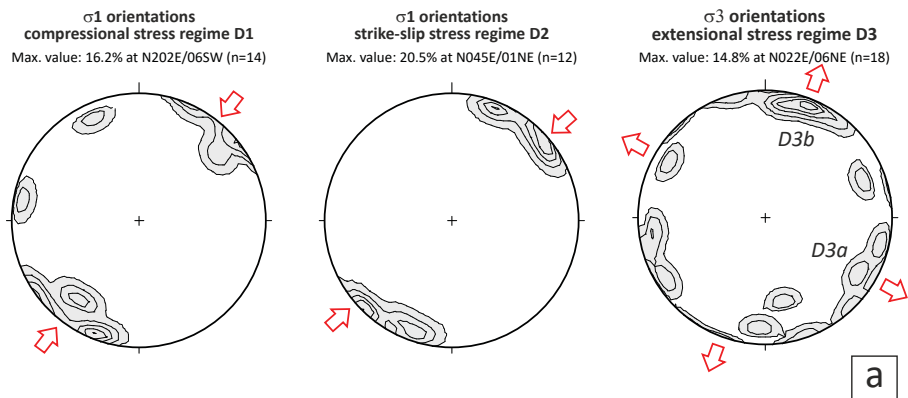


Figure15.

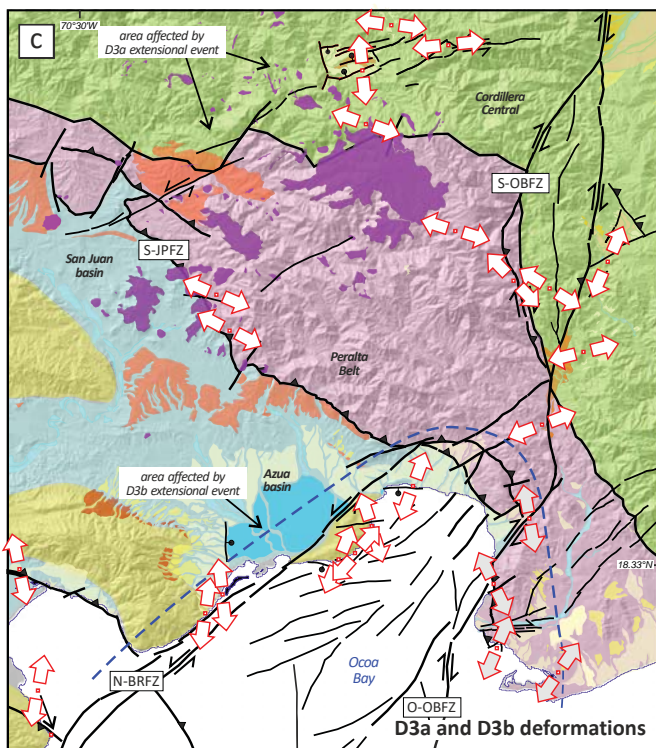
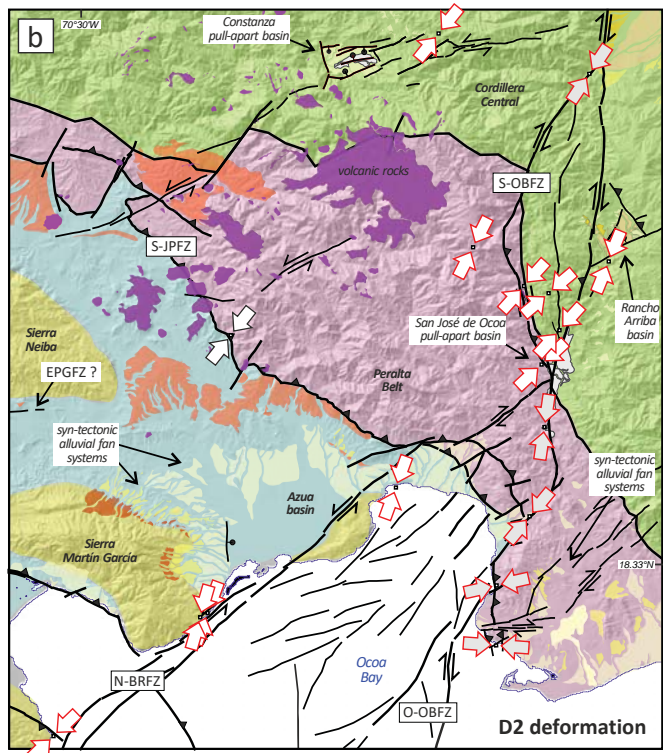
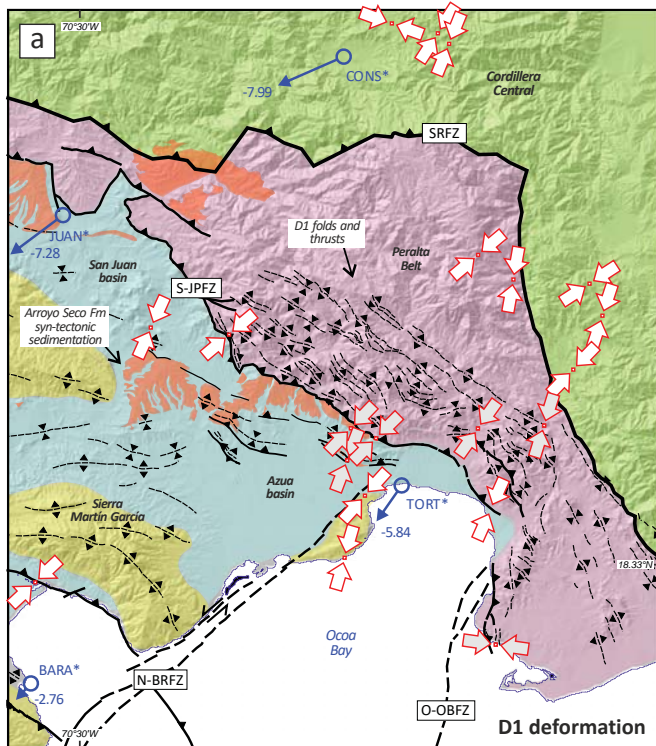


- Evolution of the Quaternary stress-field regime**
1. NE/ENE-trending pure compression D1 ( $R \approx 0.5$ ) to radial compression ( $R \gg 0.5$ ;  $\sigma_1 \approx \sigma_2$ )  $\sigma_2/\sigma_3$  permutation
  2. NE-trending pure strike-slip D2 ( $R \approx 0.5$ )  $\sigma_1/\sigma_2$  permutation
  3. NW-trending pure extension D3a ( $R \approx 0.5$ ), or
  4. NNE/N-trending extension D3b ( $R \approx 0.5$ )

b

Figure16.





## LEGEND

- Maximum horizontal stress axis obtained from fault-slip data
- Id. obtained by Escuder-Viruete et al. (2023)
- 5 mm/yr GPS velocity vector with respect to Caribbean plate (Calais et al., 2016)

## NEOTECTONIC STRUCTURES

- Main fault zones
- Fault (unspecified)
- Deduced fault
- High-angle normal fault
- Strike-slip fault
- Thrust or high-angle reverse fault
- Anticline
- Syncline

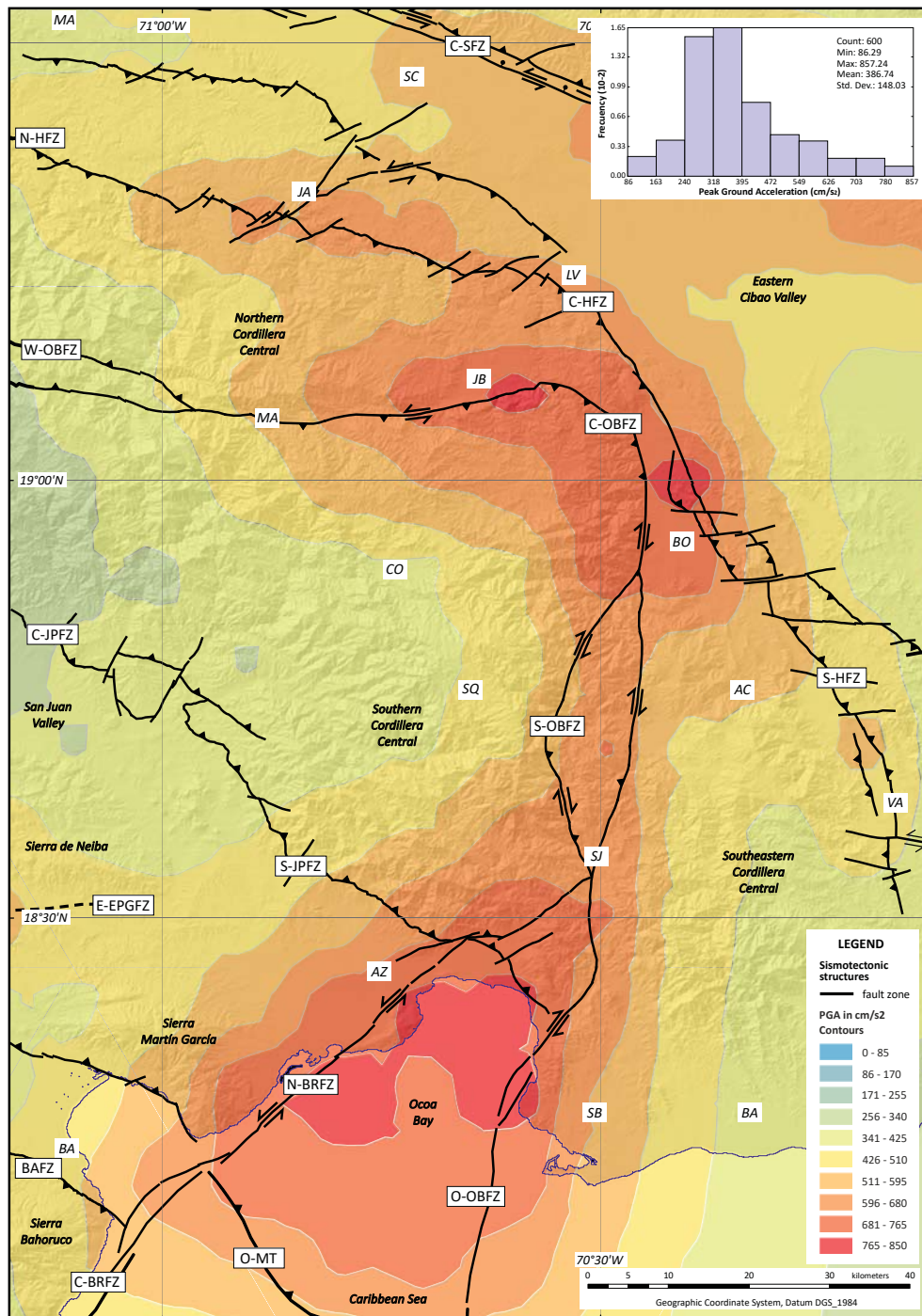
## GEOLOGICAL DOMAINS

- Gónave microplate**
  - Sierra de Naiba, Martín García and Bahoruco (Tertiary)
- Peralta Belt**
  - Peralta and Ocoa Groups (Paleogene)
- Cordillera Central**
  - Volcanic and plutonic rocks (Cretaceous)

## NEOGENE STRATIGRAPHY

- Fluvial floodplain (Holocene)
- Constanza, San José de Ocoa and Rancho Arriba intramontaneous Basins**
  - Alluvial and lacustrine dep. (Holoc.)
- San Juan, Azua and Enriquillo Basins**
  - Yaque Sur delta deposits (Holoc.)
  - Intertidal plain deposits (Holocene)
  - Lower fluvial terrace (Holocene)
  - Intermediate fluvial terrace (Pleist.)
  - Upper fluvial terrace (Pleistocene)
  - Lower alluvial fans (Holocene)
  - Intermediate alluvial fans (Pleist.)
  - Upper alluvial fans (Pleistocene)
  - Arroyo Seco Fm (L. Plioc.-L. Pleist.)
  - Volcanic rocks (Quaternary)

Figure17.



**Table 1. Characteristics of neotectonic strictires in southern central Hispaniola**

<i>Deformative event</i>	<i>Main structures</i>	<i>Stress regime</i>	<i>Age</i>
D1	NW to WNW-striking folds and thrusts. Large-scale structure of the Cordillera Central and Peralta belt.	Compression. NE-trending $\sigma_1$ axis.	Lower Miocene to Early Pleistocene
D2	N-S to NE-SW strike-slip faults and fault segment. Transverse structures related to Ocoa-Bonao-La Guácara and Beata Ridge fault zones. ENE to E-striking left-lateral strike-slip faults. Transverse structures in the NW sector.	Near pure strike-slip. NE-trending $\sigma_1$ axis.	Early-Middle Pleistocene to Holocene
D3	Extensional faults. D3a, NE to NNE striking faults in the Cordillera Central sector; D3b, WNW to W striking faults in the Ocoa Bay sector. Alignment of the Quaternary emissive centers.	D3a, NW-directed pure extension; D3b, N to NNE-directed extension.	Late Pleistocene-Holocene



**Table 2. Characteristics of the main seismotectonic structures in southern central Hispaniola**

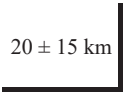
<i>Code</i>	<i>Fault/segment name</i>	<i>Deformation mechanism</i>	<i>Maximum estimated magnitude (<math>M_w</math>)</i>	<i>Major known earthquake and magnitude (<math>M_w</math>)</i>	<i>Estimated displacement rate<sup>a</sup></i>	<i>Length<sup>b</sup></i>	<i>Azimuth and dip of fault zone<sup>b</sup></i>
S-OBFZ	Southern Ocoa-Bonao-La Guacara Fault Zone	right-lateral strike-slip	$M_{\max}$ 6,5 ( $\pm$ 0,5)	unknown	0.4 ( $\pm$ 0.2) mm/yr	60 ( $\pm$ 10) km	N010°E to N030°E, subvertical dip
C-OBFZ	Central Ocoa-Bonao-La Guacara Fault Zone	oblique reverse	$M_{\max}$ 6,5 ( $\pm$ 0,5)	unknown	0.4 ( $\pm$ 0.2) mm/yr	50 ( $\pm$ 10) km	N160°E to N010°E, subvertical dip to 40° ( $\pm$ 15) W
W-OBFZ	Western Ocoa-Bonao-La Guacara Fault Zone	left-lateral oblique reverse to strike-slip	$M_{\max}$ 6,0 ( $\pm$ 1,0)	1562 earthquake?	0.2 ( $\pm$ 0.1) mm/yr	90 to 120 km	N100°E to N150°E, 45° ( $\pm$ 15) SW to subvertical dip
S-HFZ	Southern Hispaniola Fault Zone	right-lateral strike-slip?	$M_{\max}$ 7.1 ( $\pm$ 0.3)	unknown	< 0.2 mm/yr	50 to 70 km	N140°E ( $\pm$ 15), subvertical dip
E-MAFZ	Eastern Sierra de Neiba-Matheux thrust	left-lateral oblique reverse to thrust	$M_{\max}$ 5.8 ( $\pm$ 0.3)	1942 (Mw 5.8)	0.2 mm/yr	segments of 30 ( $\pm$ 5) km	N120°S ( $\pm$ 15), 65° ( $\pm$ 10) N
S-JPFZ	Southern San Juan - Pozos Fault Zone	left-lateral oblique reverse to thrust	$M_{\max}$ 6.5 ( $\pm$ 0.5)	1911 (Mw 6.7)	1.5 ( $\pm$ 1.5) mm/yr	>250 km, including segments and imbrications	N130°E ( $\pm$ 15), 45° ( $\pm$ 15) NE
BAFZ	Bahoruco Fault Zone or Barahona Thrust	reverse	$M_{\max}$ 7.0 ( $\pm$ 0.6)	1963, Mw 5.7; FD 9 km	0.2 mm/yr	50 to 85 km, including segments and imbrications	N100°E ( $\pm$ 15), 50° ( $\pm$ 15) S
E-EPGFZ	Eastern Enriquillo-Plantain Garden Fault Zone	left-lateral strike-slip to oblique reverse	$M_{\max}$ 7.3 ( $\pm$ 0.3)	1751 (Mw 7.5)?, 2010 (Mw 7.0)	8.3 ( $\pm$ 2.0) mm/yr	80 ( $\pm$ 10) km	N080°E ( $\pm$ 10), subvertical dip
N-BRFZ	Northern Beata Ridge Fault Zone	left-lateral strike-slip to reverse	$M_{\max}$ 6.5 ( $\pm$ 0.1), recorded: 4.3 (12/09/2005) and 4.1 (12/10/2013)	1751 (Mw 7.5-8.0)?, 1691?	0.2 mm/yr, probably	60 ( $\pm$ 5) km	N040°E ( $\pm$ 10), subvertical dip
C-BRFZ	Central Beata Ridge Fault Zone	left-lateral strike-slip to normal dip-slip	$M_{\max}$ 6.5 ( $\pm$ 0.1), recorded: 4.3 (12/09/2005) and 4.1 (12/10/2013)	1751 (Mw 7.5-8.0)?, 1691?	0.2 mm/yr, probably	100 ( $\pm$ 5) km	N030°E ( $\pm$ 10), 70° ( $\pm$ 15) E to subvertical dip
W-MT	Western Muertos Trough	reverse to oblique reverse	$M_{\max}$ 7.5 ( $\pm$ 0.5)	1673 (Mw 7.3)?; 1691 (Mw 7.7)?; 1751 (Mw 7.5-8.0)?	6.2 ( $\pm$ 1.0) mm/yr	170 km	N100°E ( $\pm$ 5), 12° ( $\pm$ 5) northward dip
W-DEEP-MT	Western Muertos Trough, deep part	reverse to oblique reverse	$M_{\max}$ 7.5 ( $\pm$ 0.5)	1673 (Mw 7.3)?; 1691 (Mw 7.7)?; 1751 (Mw 7.5-8.0)?	6.2 ( $\pm$ 1.0) mm/yr	170 km	N100°E ( $\pm$ 5), 40° ( $\pm$ 5) northward dip

O-MT	Muertos Trough, Ocoa Bay part	reverse to oblique reverse	$M_{\max}$ 7.5 ( $\pm$ 0.5)	1751 (Mw 7.5- 8.0)?	6.2 ( $\pm$ 1.0) mm/yr	40 km	N135°E ( $\pm$ 5), 15° ( $\pm$ 5) northward dip
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<sup>a</sup>Derived from geological mapping (SYSMIN Project, 2010; Escuder-Virue, 2022; this work), geodetic GPS measurements (Calais

<sup>b</sup>Derived from geological mapping (SYSMIN Project, 2010; Bertil et al., 2015; Terrier-Sedan & Bertil, 2021; Escuder-Virue, 2022;

<i>Width of fault zone<sup>b</sup></i>
10 ± 5 km
10 ± 5 km
5 ± 2 km
10 ± 5 km
15 ± 5 km
7 ± 5 km
7 ± 5 km
10 ± 5 km
10 ± 5 km
10 ± 5 km
20 ± 15 km
20 ± 15 km



$20 \pm 15$  km

et al., 2002; Manaker et al., 2008) and empirical relationships (e.g. Wells and Coppersmith, 1994). Escuder-Virue et al., 2023; this work))



Supporting Information for

**Active tectonics, Quaternary stress regime evolution and seismotectonic faults in southern central Hispaniola: implications for the quantitative seismic hazard assessment**

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**Contents of this file**

Supporting Information S1. Catalogue of earthquakes and focal mechanisms compiled for southern central Hispaniola.

Supporting Information S3. Methodology of dynamic fault-slip analysis.

**Additional Supporting Information (Files uploaded separately)**

Supporting Information S2. U-Th geochronological results of coral samples.

Supporting Information S4. Sites used to obtain stress tensors from the analysis of fault-slip data sets.

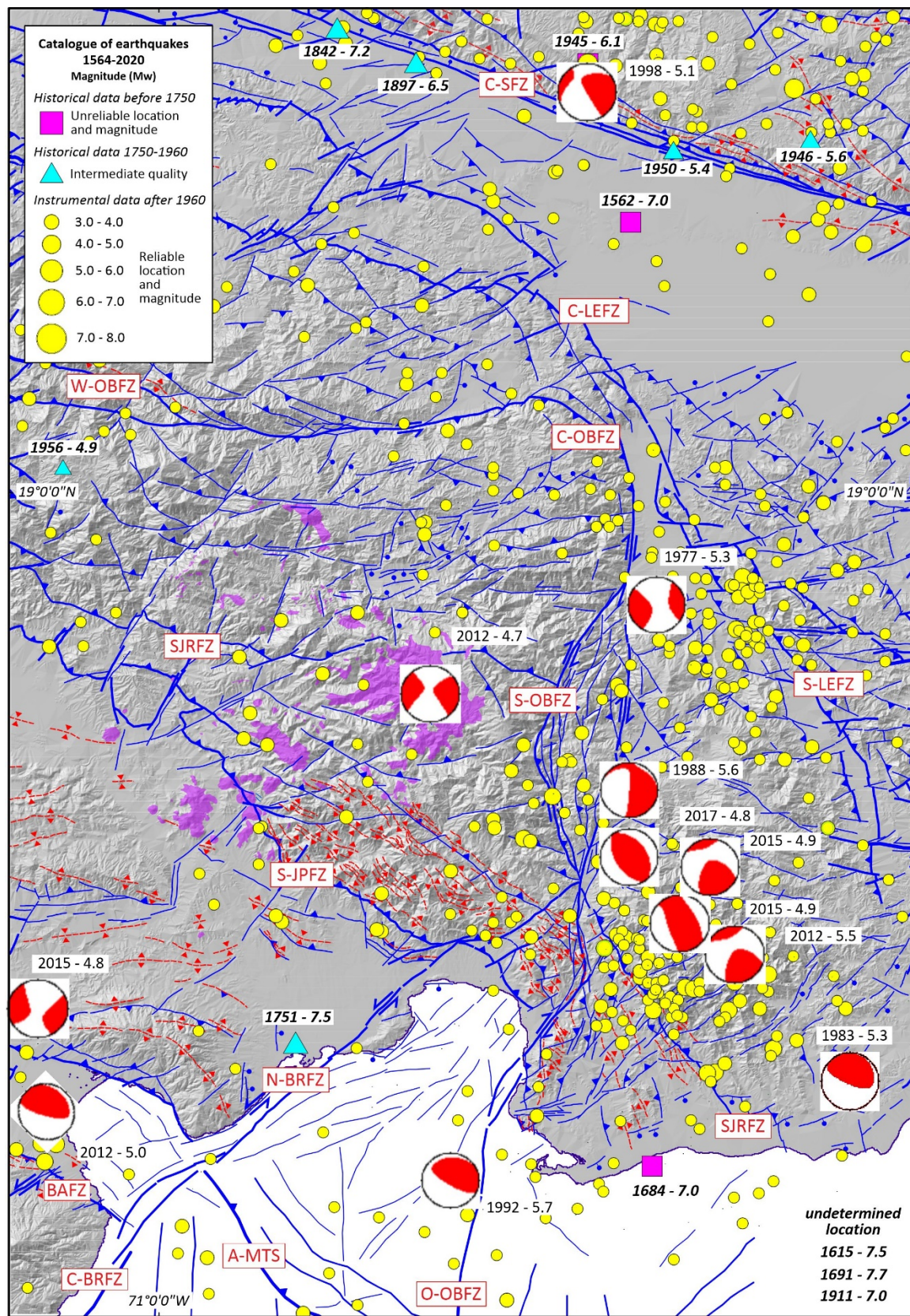
Supporting Information S5. Stress tensors obtained by methods of inversion of fault-slip data.

Supporting Information S6. Maximum horizontal stresses and tectonic regimes obtained by inversion methods.

**Introduction**



This supporting information provides a Figure with the graphic representation of the catalogue of earthquakes and focal mechanisms compiled for southern central Hispaniola (S1), a Table including the U-Th geochronological results of coral samples (S2), a Text describing the methodology of dynamic fault-slip analysis (S3), a Table describing the sites used to obtain stress tensors from the analysis of fault-slip data sets (S4), a Table including the stress tensors obtained by methods of inversion of fault-slip data (S5), and a Table showing the maximum horizontal stresses and tectonic regimes obtained by inversion methods (S6).



## Figure Supporting Information S1.

Catalogue of earthquakes and focal mechanisms compiled for southern central Hispaniola. White symbols are historical and instrumental earthquakes of  $M_w > 3.0$  from 1564 to 2020 A.D. Due to the high uncertainty related to the magnitude and location of historical earthquakes, they are classified into three groups following Terrier-Sedan & Bertil (2021): historical data before 1750 of unreliable magnitude and location; historical data between 1750 and 1960 of intermediate quality; and instrumental data after 1960 generally of high quality. More than 12 epicentres of earthquakes of magnitude at least  $M_w 6.0$  are located in the study area, or the immediate surroundings. They include the earthquakes of 1562 (destroyed La Vega and Santiago de los Caballeros), 1615 (destroyed Santo Domingo), 1684 (affected Azua and Santo Domingo), 1691 (destroyed Azua, affected Santo Domingo), 1751 (destroyed Azua) and 1911 (affected San Juan). Well-constrained fault plane solutions from 1976-2021 are taken from the Global Centroid-Moment-Tensor (CMT) Catalog (Ekström et al., 2012). Outcrops of Quaternary volcanic rocks are indicated in purple. Details on the data sources can be found in Escuder-Virue (2022) and Escuder-Virue et al. (2023).

## Text Supporting Information S3. Methodology of dynamic fault-slip analysis

The reconstruction in time and space of the different tectonic paleostress tensors in the studied area involved the use of different inversion methods. This approach is key when working with fault-slip inversion techniques in order to account for the critical assumptions and limitations of each method (see reviews by Célérier et al., 2012; Sperner & Zweig, 2010; Twiss & Unruh, 1998). Using multiple methods, hence different algorithms, will allow us to increase accuracy in the results by reducing the effect of systematic errors.

**Input data.** Fault-slip data were collected at several localities along southern central Hispaniola and used to determine the stresses causing the observed brittle deformations. A fault-slip data is composed of the fault plane orientation (dip azimuth and dip angle), the slip orientation (striae, bearing and plunge), and the sense of slip (normal, reverse, sinistral, dextral). Mesoscopic kinematic indicators used to infer the sense of slip on the fault-plane along the striae direction were the growth of fibrous crystals (slickensides) in steps, R and R' shears, grooves and half-moon, among others (e.g. Petit, 1987).

Carey and Brunier (1974) have demonstrated that from numerous fault-slip data obtained from outcrops, it is possible to investigate the stresses responsible for the fault movements and for the observed deformation. Each state of stress is expressed as a reduced stress, or paleostress, tensor, which includes the orientations of the three principal stress axes ( $\sigma_1$ ,  $\sigma_2$  and  $\sigma_3$ ) and the ratio of principal stress differences [ $R = (\sigma_2 - \sigma_3)/(\sigma_1 - \sigma_3)$ ]. Fault crosscutting relationships, the overprinting of several striations in the same fault plane and the absolute age of the faulted rocks have been used as primary criteria to date faults and thus paleostress tensors and tectonic phases.

In natural conditions, fault-slip data sets measured at one location can be polyphasic, that is, include several fault subpopulations consistent with different stress tensors. The separation of the fault subpopulations referring to each tensor was performed prior to the inversion method for stress, calculating the pressure (P or compression axis) and tension (T or tensional axis) axes for each fault-slip data. Appearance of several clusters of P or T axes in stereographic projection was interpreted as indicative of the existence of fault-slip data related to different paleostress tensors, allowing their separation in fault subpopulations.

The different paleostress tensors and stress regimes were attributed to the lithostratigraphic units where the major and minor faults were measured. The available absolute ages of each unit enabled assigning an age to each faulting event. This information, combined with a relative chronology of fault offsets based on field observations, allowed to distinguish the main tectonic events that have characterized the study area from the early Miocene to the Holocene.

Sites of measurements located in the oldest lithostratigraphic units allow recognizing the faults of different tectonic events, where multiple generations of structures are linked to successive tectonic phases. The systematic lacking of structures belonging to a given tectonic phase in a lithostratigraphic unit can indicate the existence of deformation partitioning in space or constrain the oldest age of the deformation.

Software Tectonics FP. The software Tectonics FP 1.7.9 (open version; Ortner et al., 2002; Reiter & Acs, 2000) allows using several inversion methods of fault slip data: the P–T Method (Turner, 1953); the Right Dihedra Method (Angelier & Mechler, 1977); the Direct Inversion Method (Angelier & Goguel, 1979); and the Numerical Dynamic Analysis (Spang, 1972).

The P–T Method (PTM) calculates for each fault plane a compression/maximum shortening axis (P-axis) and an extension/ maximum stretching axis (T-axis). For a set of kinematically related fault planes, the mean vectors of the P- and T-axes represent an approximation to the theoretical strain axes. P- and T-axes define clusters in stereographic projection, based on which kinematically consistent subsets of fault planes can be established from a heterogeneous fault population. According to the Mohr–Coulomb fracture criterion, the PTM adopts a fracture angle ( $\theta$ ) between P and the fault plane. This angle is considered uniform to all fault population, since potentially varying material properties are not taken into account by the method.

The Right Dihedra Method (RDM) uses a grid search technique. The method calculates the compressive dihedral for each given fault plane, by considering all the possible locations of the compression axis on the stereographic projection grid. The eigenvectors of the superposed compressive dihedra will then identify the locus of the stress ellipsoid axes. The RDM calculates the orientations of the three principal stress axes, but not the  $\varphi$  ratio, which is one of the four critical variables of the reduced stress tensor. As this method usually represents the best paleostress tensor, the  $\varphi$  ratio value and the forehead derived stress axes are calculated in a complementary way with other methods.

The classical Direct Inversion Method (DIM) calculates the reduced stress tensor in a single step from the whole data-set by applying a least-square minimization of the angles (misfit angles) between the theoretical and observed slip directions of each fault. The DIM



provides solutions in terms of paleostress tensors, notably the directions of principal paleostress axes ( $\sigma_1 > \sigma_2 > \sigma_3$ ) and the stress ratio (R).

The Numerical Dynamic Analysis (NDA) calculates a strain tensor for each fault-striae set, assuming a fixed angle between the maximum principal strain axis and fault plane. The sum of the resulting strain tensors for every plane, divided by the number of sets, yields the bulk strain tensor. The orientations and relative values of the principal strain axes are obtained from the eigenvalues and eigenvectors of the bulk strain tensor. Therefore, the NDA calculates a reduced strain tensor, composed of the directions of the three principal reciprocal strains ( $\lambda_1 > \lambda_2 > \lambda_3$ ) and the strain ratio (E). The angle  $\theta$  between  $\lambda_1$  and the maximum resolved shear strain, which is believed to be parallel to the mineral lineation on the shear fault, was calculated by iteration using a best fitting approach. However, it is necessary to verify that any selected  $\theta$  value does not condition the results of the stress inversion. The values of the angle  $\theta$  obtained experimentally range 17° and 40° (Jaeger & Cook, 1979).

Procedure of analysis. The first step in the analysis was to check the consistency of the fault-slip data and correct it so that each striation lay within the associated fault plane. Measurements with very large deviations were not included in the calculations

The second step in the analysis consisted of splitting of inhomogeneous fault-slip data sets into kinematically homogeneous subsets basing on mechanical consistency. The definition of subsets was made through field observations, such as the existence of more than one set of striations on the same fault plane, or with the help of graphical procedures, such as tangent-lineation plotting method or P- and T-axes clusters in stereographic projection. Separation of homogeneous subsets from heterogeneous fault-slip data was also performed numerically, by applying an inverse method and obtaining more than one paleostress tensor to groups of faults that integrate data from sites spatially close.

The third step in the analysis was to run the Tectonics FP routine to the defined homogeneous subsets of single outcrops and obtain four solutions, one for each inversion technique considered (PTM, RDM, DIM and NDA). These calculations eliminated fault-slip data that are incompatible with any stress tensor. Statistically stable stress or strain tensors were obtained from 10 to 20 fault-slip data measured in each structural site.

The choice of the best fitting method for a given fault population relied on: (1) the negative expected value (nev) which is the number of fault planes showing opposite senses of slip when the measured values are compared with those theoretically determined; (2) the best fit of principal axes orientations with respect to fault orientation; (3) the accuracy of the slip sense (confidence level assigned during field measurements); and 4) the value of the fluctuation parameter (F). F is calculated as the arithmetic mean of the angles between the theoretical slip directions (according to the "Wallace-Bott hypothesis"; Bott, 1959) and the measured ones. A deviation  $>20^\circ$  points to a heterogeneous data set (Ortner et al., 2002). Hence, we will regard a fault-slip datum with the misfit  $<20^\circ$  as being compatible with the state of stress evaluated.

Supporting Information S2

U-Th geochronological results of coral samples

Sample	<sup>238</sup> U (ppm)	<sup>238</sup> U 2σ error	<sup>230</sup> Th/ <sup>238</sup> U	<sup>230</sup> Th/ <sup>238</sup> U 2s error	<sup>234</sup> U/ <sup>238</sup> U	<sup>234</sup> U/ <sup>238</sup> U 2s error	δ <sup>234</sup> U	δ <sup>234</sup> U (initial)	<sup>230</sup> Th/ <sup>232</sup> Th	<sup>230</sup> Th/ <sup>232</sup> Th 2s error	<sup>234</sup> U/ <sup>238</sup> U (initial)	<sup>234</sup> U/ <sup>238</sup> U (error)	Age (ka)	Age detr. corr. (ka)	Age d.c. 2σ > error	Age d.c. 2σ < error	<sup>230</sup> Th (ppm)	<sup>230</sup> Th sample (ppm)
20JE04A	3,352	0,000	0,709	0,001	1,1234	0,0008	123,41	166,52	732,3	0,90	1,167	0,006	106,20	106,00	0,54	-0,53	3,89E-05	9,92E-03
20JE04D	5,506	0,001	1,179	0,002	1,2738	0,0007	273,77	521,23	2811,0	1,65	1,521	0,013	228,04	227,85	1,88	-1,85	1,06E-04	7,05E-03
21JE39	2,832	0,001	1,031	0,004	1,0795	0,0010	79,53	183,96	42800,1	140,29	1,184	0,030	296,76	296,73	7,46	-6,98	4,78E-05	2,09E-04
21JE64	2,653	0,000	0,792	0,001	1,1224	0,0004	122,35	176,12	5449,9	2,18	1,176	0,004	128,93	128,89	0,40	-0,40	3,44E-05	1,18E-03
21JE124	2,711	0,001	0,008	0,000	1,1570	0,0005	157,04	157,36	189,7	1,93	1,157	0,016	0,73	0,73	0,01	-0,01	3,42E-07	3,36E-04
21JE126A	2,345	0,000	1,072	0,002	1,0944	0,0007	94,37	241,09	35721,6	25,68	1,241	0,017	331,96	331,92	4,51	-4,33	4,11E-05	2,15E-04
21JE126B	2,709	0,001	1,121	0,002	1,0968	0,0007	96,77	372,54	9159,3	3,29	1,373	0,068	477,60	476,99	23,63	-19,64	1,12E+00	2,43E-03
21JE126C	1,254	0,000	0,824	0,001	1,2117	0,0008	211,71	295,89	10615,2	13,28	1,296	0,006	118,48	118,47	0,52	-0,51	1,69E-05	2,97E-04
21JE129	5,573	0,002	0,833	0,003	1,1924	0,0007	192,39	273,64	9942,4	17,22	1,274	0,009	124,67	124,65	0,85	-0,84	7,59E-05	1,43E-03
22JE26B	2,850	0,000	0,770	0,002	1,1193	0,0009	119,32	169,03	4863,5	4,05	1,169	0,007	123,27	123,23	0,70	-0,69	3,59E-05	1,38E-03

<sup>238</sup>U concentration in coral sample (ppm). 2σ (95% confidence limit) measurement error on <sup>238</sup>U concentration in coral sample (ppm)

<sup>230</sup>Th/<sup>238</sup>U is measured activity ratio. <sup>230</sup>Th/<sup>238</sup>U 2σ error is measurement error on activity ratio.

<sup>234</sup>U/<sup>238</sup>U is measured activity ratio. <sup>234</sup>U/<sup>238</sup>U 2σ error is measurement error on activity ratio.

d<sup>234</sup>U is measured, present day d<sup>234</sup>. d<sup>234</sup>U (initial) is calculated initial d<sup>234</sup>.

<sup>230</sup>Th/<sup>232</sup>Th is measured activity ratio. <sup>230</sup>Th/<sup>232</sup>Th 2σ error is measurement error on activity ratio.

<sup>234</sup>U/<sup>238</sup>U (initial) is calculated ratio on the basis of the present-day measured <sup>234</sup>U/<sup>238</sup>U and the calculated age.

<sup>234</sup>U/<sup>238</sup>U (error) is uncertainty on calculated initial <sup>234</sup>U/<sup>238</sup>U taking into account uncertainty on measured ratio and calculated age.

Age (ka) is calculated U-Th age in ka

Age 2σ error is age uncertainty in ka on (>) high and (<) low age sides (age errors are not symmetrical because ratios change in a non-linear way with age).

Age detr. Corr. (ka) is calculated U-Th age in ka, corrected for detrital contamination (assuming <sup>230</sup>Th/<sup>232</sup>Th = 0.8 ± 0.4).



## Supporting Information S5

### Stress tensors obtained by methods of inversion of fault-slip data

Data		RDM										DIM						NDA									
		σ <sub>1</sub>			σ <sub>2</sub>			σ <sub>3</sub>				σ <sub>1</sub>		σ <sub>2</sub>		σ <sub>3</sub>		λ <sub>1</sub>		λ <sub>2</sub>		λ <sub>3</sub>					
Site	Num.	Eigen	Dir	Inm	Eigen	Dir	Inm	Eigen	Dir	Inm	R	Dir	Inm	Dir	Inm	Dir	Inm	Dir	Inm	Dir	Inm	Dir	Inm	Dir	Inm	E	
20JE14	10	0,49	227	0	0,33	137	77	0,18	317	13	0,48	78	4	252	86	348	0	40	1	136	75	310	15	0,58			
21JE70A	16	0,4	22	7	0,32	313	59	0,28	118	30	0,33	190	13	295	47	89	40	44	12	292	61	140	26	0,34			
21JE70B	8	0,48	130	64	0,35	289	24	0,17	23	8	0,58	41	57	283	17	184	28	98	71	289	19	198	3	0,49			
21JE72	14	0,41	237	1	0,32	147	6	0,27	337	83	0,36	62	44	269	43	166	14	79	23	341	17	219	61	0,37			
21JE73	10	0,38	197	8	0,37	292	29	0,25	93	60	0,92	224	26	315	2	50	64	267	18	7	28	149	56	0,68			
21JE78A	8	0,484	280	7	0,34	189	6	0,18	60	81	0,53	183	1	274	25	90	65	281	2	191	5	35	85	0,51			
21JE78B	14	0,5	324	81	0,32	194	5	0,18	103	7	0,44	327	49	199	28	94	27	54	79	195	9	286	7	0,43			
21JE79A	12	0,39	202	4	0,37	292	3	0,24	61	85	0,87	249	10	151	36	352	52	223	18	129	12	8	68	0,66			
21JE79B	10	0,5	70	77	0,3	173	3	0,2	264	13	0,33	267	67	100	22	8	5	135	87	355	3	265	2	0,38			
21JE80A	8	0,5	53	69	0,32	277	16	0,18	183	14	0,44	118	64	309	26	217	5	66	81	272	8	181	4	0,39			
21JE80B	12	0,4	213	24	0,36	304	2	0,25	39	66	0,73	287	67	117	23	26	4	210	30	117	5	19	59	0,46			
21JE80C	13	0,46	53	5	0,32	157	69	0,22	321	20	0,42	273	21	14	26	149	56	210	30	117	5	19	59	0,46			
21JE93A	10	0,45	207	13	0,3	350	74	0,26	115	9	0,21	197	38	33	51	293	8	213	28	341	50	108	27	0,41			
21JE94A	8	0,38	70	21	0,33	161	2	0,29	255	69	0,44	105	22	227	52	2	29	50	35	311	13	204	52	0,47			
21JE96A	10	0,44	356	12	0,34	86	1	0,22	180	78	0,55	165	1	256	17	71	73	351	13	83	9	206	74	0,37			
21JE96B	8	0,47	236	76	0,3	46	13	0,23	136	2	0,29	18	66	173	22	266	9	330	86	239	1	149	4	0,28			
21JE97	11	0,41	221	3	0,34	128	46	0,26	314	44	0,53	182	16	7	74	272	1	202	1	111	24	294	56	0,64			
21JE98	13	0,43	55	9	0,35	201	79	0,23	324	6	0,60	93	11	303	78	184	6	221	1	125	77	311	13	0,65			
21JE99B	8	0,46	326	73	0,35	213	7	0,19	121	16	0,59	176	62	29	24	293	14	332	83	221	3	131	7	0,59			
21JE100	10	0,44	226	3	0,32	321	64	0,24	135	26	0,40	261	4	358	58	169	31	208	11	28	79	298	0	0,36			
21JE102A	13	0,46	243	1	0,29	153	19	0,25	336	71	0,19	180	2	270	20	84	70	47	5	144	52	312	37	0,37			
21JE103	9	0,45	349	45	0,35	162	45	0,2	256	4	0,60	33	56	164	24	265	22	9	63	149	21	246	16	0,56			
21JE108B	8	0,52	269	71	0,34	161	6	0,14	69	18	0,53	61	62	298	16	202	22	311	84	162	5	72	3	0,50			
21JE112A	9	0,49	222	21	0,33	131	4	0,18	32	69	0,48	11	7	102	5	226	81	210	12	302	9	68	74	0,40			
21JE112B	10	0,45	55	62	0,33	204	24	0,22	300	13	0,48	337	64	205	18	109	18	39	71	207	18	298	3	0,51			
21JE113	16	0,45	65	85	0,35	251	5	0,2	161	1	0,60	98	19	302	70	191	8	132	78	256	7	348	10	0,59			
21JE120A	10	0,49	26	2	0,35	116	1	0,16	226	88	0,58	205	12	295	1	28	78	206	13	116	1	22	77	0,50			
21JE120B	10	0,51	147	80	0,33	291	8	0,16	21	6	0,49	25	52	285	8	189	37	155	77	288	9	19	9	0,48			
21JE121	10	0,47	124	67	0,32	331	20	0,22	238	9	0,40	135	60	319	30	228	2	169	66	338	23	70	4	0,28			
21JE123A	10	0,5	204	1	0,36	294	3	0,14	99	87	0,61	4	7	113	68	271	21	207	7	117	0	24	83	0,64			
21JE125A	10	0,49	200	4	0,34	82	83	0,17	291	6	0,53	36	20	179	65	301	14	32	1	128	82	302	8	0,54			
21JE125B	10	0,51	279	86	0,31	110	4	0,19	20	1	0,38	2	65	174	25	266	3	59	88	284	1	194	1	0,33			
21JE127	10	0,48	20	4	0,31	137	80	0,21	290	9	0,37	34	4	304	2	183	85	28	4	151	82	298	6	0,50			

21JE130A	10	0,41	78	28	0,34	183	27	0,25	310	49	0,56	47	6	141	34	309	56	28	4	151	82	298	6	0,50
21JE130B	10	0,42	336	58	0,33	75	5	0,25	168	32	0,47	109	86	265	4	355	2	346	78	254	0	164	12	0,36
21JE137	18	0,45	154	76	0,33	278	8	0,22	10	11	0,48	126	87	340	3	250	2	155	83	279	4	9	6	0,42
22JE12	13	0,4	222	0	0,33	132	45	0,27	312	45	0,46	-	-	-	-	-	-	29	8	120	2	225	81	-
22JE13A	15	0,51	51	2	0,35	321	3	0,14	185	86	0,57	90	12	359	4	248	78	58	12	327	6	212	76	0,51
22JE13B	10	0,48	38	2	0,31	146	83	0,21	308	7	0,37	264	10	167	33	9	55	37	4	148	78	306	11	0,55
22JE15	17	0,5	169	82	0,34	278	3	0,16	8	8	0,53	353	72	155	17	247	5	23	85	276	1	185	5	0,50
22JE17	10	0,51	195	10	0,33	91	54	0,16	292	35	0,49	196	2	286	2	66	87	196	2	286	2	66	87	0,54
22JE26	10	0,45	46	7	0,31	149	60	0,24	312	29	0,33	26	11	291	23	140	64	37	10	148	64	303	23	0,58

Num; fault-slip data number. RDM; Right-Dihedra Method. DIM; Direct Immersion Method. NDA; Numerical Dynamical Analysis.

$\sigma_1$ ,  $\sigma_2$  and  $\sigma_3$ ; principal stress axes ( $\sigma_1 > \sigma_2 > \sigma_3$ ). Eigen; eigenvalue. Dir; azimuth. Inm; immersion angle. R; stress ratio ( $R = (\sigma_2 - \sigma_3) / (\sigma_1 - \sigma_3)$ ;  $0 \leq R \leq 1$ ). E; strain ratio

## Supporting Information S6

### Maximum horizontal stresses and tectonic regimes obtained by inversion methods

Site	$\sigma_H$ (máx)		Stress regime			
	Dir	Inm	R	Fault type	Method	Oher consistent methods
21JE72	237	1	0,36	reverse D1	RDM	<i>DIM, NDA</i>
21JE73	197	8	0,92	reverse D1	RDM	<i>DIM, NDA</i>
21JE78A	280	7	0,53	reverse D1	RDM	<i>DIM, NDA</i>
21JE79A	202	4	0,87	reverse D1	RDM	<i>DIM, NDA</i>
21JE80B	213	24	0,73	reverse D1	RDM	<i>DIM, NDA</i>
21JE94A	70	21	0,44	reverse D1	RDM	<i>DIM, NDA</i>
21JE96A	356	12	0,55	reverse D1	RDM	<i>DIM, NDA</i>
21JE102A	243	1	0,19	reverse D1	RDM	<i>DIM, NDA</i>
21JE112A	222	21	0,48	reverse D1	RDM	<i>DIM, NDA</i>
21JE120A	26	2	0,58	reverse D1	RDM	<i>DIM, NDA</i>
21JE123A	204	1	0,61	reverse D1	RDM	<i>DIM, NDA</i>
21JE130	47	6	0,27	reverse D1	DIM	-
22JE12	222	0	0,46	reverse D1	RDM	<i>DIM, NDA</i>
22JE13A	51	2	0,57	reverse D1	RDM	<i>DIM, NDA</i>
20JE14	227	0	0,48	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE70A	22	7	0,33	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE80C	53	5	0,42	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE93A	207	13	0,21	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE97	221	3	0,53	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE98	55	9	0,60	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE100	226	3	0,40	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE125A	200	4	0,53	strike-slip D2	RDM	<i>DIM, NDA</i>
21JE127	20	4	0,37	strike-slip D2	RDM	<i>DIM, NDA</i>
22JE13B	38	2	0,37	strike-slip D2	RDM	<i>DIM, NDA</i>
22JE17	195	10	0,49	strike-slip D2	RDM	<i>DIM, NDA</i>
22JE26	46	7	0,33	strike-slip D2	RDM	-
21JE70B	289	24	0,58	extensional D3	RDM	<i>DIM, NDA</i>
21JE78B	194	5	0,44	extensional D3	RDM	<i>DIM, NDA</i>
21JE79B	173	3	0,33	extensional D3	RDM	<i>DIM, NDA</i>
21JE80A	277	16	0,44	extensional D3	RDM	<i>DIM, NDA</i>
21JE96B	46	13	0,29	extensional D3	RDM	<i>DIM, NDA</i>
21JE99B	213	7	0,59	extensional D3	RDM	<i>DIM, NDA</i>
21JE103	162	45	0,60	extensional D3	RDM	<i>DIM, NDA</i>
21JE108B	161	6	0,53	extensional D3	RDM	<i>DIM, NDA</i>
21JE112B	204	24	0,48	extensional D3	RDM	<i>DIM, NDA</i>
21JE113	251	5	0,60	extensional D3	RDM	<i>DIM, NDA</i>
21JE120B	291	8	0,49	extensional D3	RDM	<i>DIM, NDA</i>
21JE121	331	20	0,40	extensional D3	RDM	<i>DIM, NDA</i>
21JE125B	110	4	0,38	extensional D3	RDM	<i>DIM, NDA</i>
21JE130B	75	5	0,47	extensional D3	RDM	<i>DIM, NDA</i>
21JE137	278	8	0,48	extensional D3	RDM	<i>DIM, NDA</i>
22JE15	278	3	0,53	extensional D3	RDM	<i>DIM, NDA</i>

$\sigma_H$  (max); máximo horizontal stress. Dir; azimuth. Inm; immersion angle.

R; stress ratio ( $R=(\sigma_2-\sigma_3)/(\sigma_1-\sigma_3)$ ;  $0 \leq R \leq 1$ ).

RDM; Right-Dihedra Method, DIM; Direct Immersion Method, NDA; Numerical Dynamical Analysis.

# Supporting Information S4

## Sites used to obtain stress tensors from the analysis of fault-slip data sets

Site	Longitude (X)	Latitude (Y)	Geological Domain	Quadrangle (E. 1:50k)	Stratigraphic Unit	Unit Age	Main Lithology
20JE14	-70,51720	18,56506	San Jose de Ocoa basin	San José de Ocoa	Numero Fm, Fluvial Terrace	Late Eocene, Quaternary	Dark mudstone, gravel and clay
21JE70	-70,45311	18,67916	Cordillera Central	Arroyo Caño	Arroyo Caña Batholith	Late Cretaceous	Tonalite
21JE72	-70,47095	18,65749	Cordillera Central	Sabana Larga	Arroyo Caña Batholith	Late Cretaceous	Gabbro
21JE73	-70,45346	18,62003	Cordillera Central	Sabana Larga	Arroyo Caña Batholith	Late Cretaceous	Leucotonalite
21JE78	-70,68596	18,94094	Cordillera Central	Constanza	Tireo Group, Constanza Fm	Late Cretaceous	Green tuff
21JE79	-70,62475	18,91709	Cordillera Central	Constanza	Tireo Group, Constanza Fm	Late Cretaceous	Green tuff and dolerite
21JE80	-70,63772	18,92659	Cordillera Central	Constanza	Tireo Group, Constanza Fm	Late Cretaceous	Green tuff
21JE93	-70,59447	18,69213	'eralta fold-and-thrust be	Sabana Quéliz	Ocoa Fm	Late Eocene	Conglomerate
21JE94	-70,58965	18,68086	'eralta fold-and-thrust be	Sabana Quéliz	Numero Fm	Middle to Late Eocene	Dark mudstone and sandstone
21JE96	-70,54467	18,65448	'eralta fold-and-thrust be	San José de Ocoa	Ocoa Fm	Late Eocene	Conglomerate
21JE97	-70,53869	18,65274	'eralta fold-and-thrust be	San José de Ocoa	Ocoa Fm	Late Eocene	Conglomerate
21JE98	-70,52779	18,63098	'eralta fold-and-thrust be	San José de Ocoa	Ocoa Fm	Late Eocene	Dark mudstone and sandstone
21JE99	-70,51323	18,64305	'eralta fold-and-thrust be	San José de Ocoa	Tireo Group	Maastrichtian	Limestone
21JE100	-70,50453	18,58901	'eralta fold-and-thrust be	San José de Ocoa	Ocoa Fm	Late Eocene	Dark mudstone and sandstone
21JE102	-70,48251	18,56104	'eralta fold-and-thrust be	Sabana Larga	Tireo Group, Constanza Fm	Late Cretaceous	Green tuff and sandstone
21JE103	-70,47375	18,57525	'eralta fold-and-thrust be	Sabana Larga	Tireo Group, Constanza Fm	Late Cretaceous	Green tuff and sandstone
21JE108	-70,51455	18,49968	'eralta fold-and-thrust be	Azua	Jura Fm, Ocoa Fm	Middle to late Eocene	Red limestone and mudstone
21JE112	-70,71758	18,42512	Azua basin	Azua	Sobrerito Fm	Early Miocene	Limestone
21JE113	-70,70741	18,39128	Azua basin	Azua	Sobrerito Fm	Early Miocene	Limestone
21JE120	-70,73770	18,35106	Azua basin	Azua	Sobrerito Fm	Early Miocene	Limestone
21JE121	-70,72878	18,35725	Azua basin	Azua	Sobrerito Fm	Early Miocene	Limestone
21JE123	-70,56557	18,40412	'eralta fold-and-thrust be	Azua	Arroyo Seco	Late Plioc.-Early Pleistoc.	Conglomerate and clay
21JE125	-70,67905	18,43106	Azua basin	Azua	Sobrerito Fm	Early Miocene	Limestone
21JE126	-70,87761	18,29942	Sierra Martin Garcia	Barrero	Coral reef terrace	Late Pleistocene	Limestone
21JE127	-70,88598	18,28986	Sierra Martin Garcia	Barrero	Upper Alluvial Fan	Late Pleistocene	gravel and silt/clay
21JE129	-71,06235	18,32466	Sierra Martin Garcia	Barahona	Coral reef terrace	Late Pleistocene	Limestone
21JE130	-71,13364	18,34960	Sierra Martin Garcia	Vicente Noble	Angostura Fm	Early Pliocene	Gypsum, mudstone and sandstone
21JE137	-71,05840	18,15422	Sierra Bahoruco	La Ciénaga	Coral reef terrace, alluvial fan	Late Pleistocene	Limestone, gravel and mudstone
22JE12	-70,94551	18,60346	'eralta fold-and-thrust be	Vayas de Viajama	Trinchera Fm	Middle to late Miocene	Sandstone, gravel and silt/clay
22JE13	-70,86600	18,59378	'eralta fold-and-thrust be	Vayas de Viajama	Ventura Fm, Jura Fm	Lower to middle Eocene	Dark mudstone, sandstone and limestone
22JE15	-70,88191	18,29371	Sierra Martin Garcia	Barrero	Upper Alluvial Fan	Late Pleistocene	gravel and silt/clay
22JE17	-70,88424	18,29217	Sierra Martin Garcia	Barrero	Upper Alluvial Fan	Late Pleistocene	gravel and silt/clay
22JE26	-71,05918	18,15463	Sierra Bahoruco	La Ciénaga	Coral reef terrace, alluvial fan	Late Pleistocene	Limestone, gravel and mudstone