Volume 3

Chapter 15

https://doi.org/10.32685/pub.esp.37.2019.15
Published online 30 April 2020

Isthmian Bedrock Geology: Tilted, Bent, and Broken

Camilo MONTES[1]* and Natalia HOYOS[2]

1 camilomontes@uninorte.edu.co
Universidad del Norte
Km 5 vía al mar
Barranquilla, Colombia

2 nbotero@uninorte.edu.co
Universidad del Norte
Km 5 vía al mar
Barranquilla, Colombia

* Corresponding author

Abstract A review of the bedrock geology of the Isthmus of Panama highlights tectonic deformation—tilting, bending, and breaking—, as the major controlling factor in the sites and modes of Cenozoic sedimentation. Deformation in Paleocene – early Eocene times folded and faulted a basement complex composed of plateau basalts, pelagic and hemipelagic sequences, and an overprinted magmatic arc. This deformation episode brought parts of the isthmus from lower bathyal depths to subaerial exposure, bringing about basement cooling and eroding the plutonic bodies that make up the roots of a Campanian to Eocene arc. A clastic–carbonate, less deformed, upper Eocene and younger sedimentary sequence onlaps nonconformably the basement complex. Southward tilting of the isthmus controlled the accumulation of the clastic wedge, recording first shallow marine depositional environments, followed by deepening, and then by shoaling. This sequence resulted from basin tilting that simultaneously raised the San Blas Range, eroding it, while deepening the axis of the Chucunaque Basin. Bending and breaking of the isthmus took place as it was being detached from the trailing edge of the Caribbean Plate, and marked the start of left–lateral offset of the isthmus in late Eocene times.

Keywords: *Panama, isthmus, deformation.*

Resumen Una revisión de la geología del basamento del Istmo de Panamá muestra que la deformación tectónica —el basculamiento, la flexión y la ruptura— es el factor principal que controla los sitios y modos de sedimentación cenozoica. La deformación durante el Paleoceno–Eoceno temprano plegó y falló el complejo de basamento compuesto por basaltos de *plateau*, secuencias pelágicas y hemipelágicas, y un arco magmático sobreimpuesto. Este episodio de deformación trajo partes del istmo desde las profundidades batiales inferiores a exposición subaérea, provocando el enfriamiento de rocas del basamento y la erosión de los cuerpos plutónicos que forman las raíces del arco Campaniano–Eoceno. Una secuencia sedimentaria clástica–calcárea, menos deformada, del Eoceno superior y más joven cubre discordantemente el complejo de basamento. El basculamiento del istmo hacia el sur controló la acumulación de la cuña clástica, registrando primero ambientes deposicionales marinos poco profundos, seguidos por profundización y luego somerización. Esta secuencia resultó del basculamiento de la cuenca que levantó simultáneamente la cordillera de San Blas, erosionándola, mientras se profundizaba el eje de la Cuenca de Chucunaque. La flexión y la ruptura del istmo ocurrieron cuando este se despegó de la parte trasera de la Placa del Caribe, marcando el inicio del desplazamiento sinestral del istmo a finales del Eoceno.

Palabras clave: *Panamá, istmo, deformación.*

Citation: Montes, C. & Hoyos, N. 2020. Isthmian bedrock geology: Tilted, bent, and broken. In: Gómez, J. & Mateus–Zabala, D. (editors), The Geology of Colombia, Volume 3 Paleogene – Neogene. Servicio Geológico Colombiano, Publicaciones Geológicas Especiales 37, p. 451–467. Bogotá. https://doi.org/10.32685/pub.esp.37.2019.15

1. Introduction

The Isthmus of Panama is part of an intraoceanic volcanic arc that is actively colliding with northwestern South America (Kellogg & Vega, 1995), the last of several collisions (Cardona et al., 2011, 2012) recorded along this accretional margin (Cediel et al., 2003; Kennan & Pindell, 2009). Because the isthmian segment of the arc has enjoyed a longer history of geological exploration, better access, and logistical conditions than the segment already attached to South America, a review of its bedrock geology is justified in a volume about the Geology of Colombia. Better knowledge of our neighbor's geology may shed light on some of the many shared geological processes that are recorded in the westernmost Andes and the Choco Block. The geology of the isthmus—across national boundaries—is key to understanding the patterns of tropical biodiversity and faunal exchanges, the start of northern hemisphere glaciations, and the birth of the modern Caribbean Sea.

In very general terms, the Isthmus of Panama is an oceanic plateau onto which a young volcanic arc was built, and then subsequently broken, bent, and tilted. Deformation on the isthmus controlled the location and sedimentation mode of depocenters, where clastics and carbonates accumulated after middle/late Eocene times. Arc volcanism in the isthmus started in Late Cretaceous times (Buchs et al., 2010; Wegner et al., 2011), onto a thickened oceanic plateau (Kerr et al., 2003) that had been interacting during most of the Cenozoic with subducting plate asperities born in the Galapagos hotspot (Hoernle et al., 2004), including ridges and large intraoceanic volcanoes (Buchs et al., 2011a). The evolution of this originally near–linear (Rodríguez–Parra et al., 2017) intraoceanic arc was interrupted in late Eocene times by whole arc deformation, fracturing, oroclinal bending (Montes et al., 2012a; Recchi & Metti, 1975), affecting magmatism and causing widespread subaerial exposure. Later, in Oligocene times, land connections to North America (Bloch et al., 2016; Kirby & MacFadden, 2005; Rincón et al., 2015), and in middle Miocene times to South America, completed the land bridge between the Americas (León et al., 2018; Montes et al., 2015). It is the Neogene history of the Isthmus of Panama that remains the most controversial (Coates & Stallard, 2013; Jaramillo et al., 2017; Leigh et al., 2013; Molnar, 2017; O'Dea et al., 2016), in particular, the time of sill rise from lower bathyal to become a continuous land path between southernmost Central America and South America.

In this contribution we provide a general review of the bedrock geology of the Isthmus of Panama based on published literature and several years of field expeditions in the Canal Basin, eastern Panama, and the Azuero Peninsula. Since comprehensive reviews already exist examining the marine biota (Lessios, 2008), paleoceanographic (Molnar, 2008), biogeographic (Jaramillo, 2018; Leigh et al., 2013), and molecular data (Bacon et al., 2015) on isthmus closure, it is perhaps a good opportunity to provide a review of the bedrock geology and deformation of the isthmus, a subject often overlooked in the controversy over time of closure. The emphasis of this review is to show that deformation (bending, tilting, breaking) is a major ingredient in the geological evolution of the isthmus, we therefore focus our discussion around this point. We first discuss the geometry, composition, and evolution of the basement complex, we then discuss the cover sequences, while highlighting observations on isthmian deformation. We finish this review by evaluating possible avenues of future research on isthmian geology, which despite a long history of geological investigations, and much improved logistical access conditions, still features large tracts of virtually unexplored land.

2. Isthmus Geology

Geologic research in the isthmus had an early start in the late XIX century resulting from multiple commissions evaluating some eight possible interoceanic canal routes, most of them in today's Panama and Darien (e.g., Reclus & De Vaisseau, 1880; Verbrugghe, 1879). While these early exploration efforts were concerned with the engineering details and financial prospects of canal construction, they also gathered basic topographic and geologic information about the proposed interoceanic canal routes. Choosing the best route for the interoceanic canal was however, an exercise of political maneuvering, little concerned with geologic/geotechnical considerations. The French *Compagnie Universelle*, charged with the digging of the canal under the direction of Ferdinand DE LESSEPS, obtained a concession from the Colombian government in 1879, and started excavations with only vague geological insights in 1881 (Douville, 1898; Hill et al., 1898). Digging progressed for more than 20 years at great human and financial cost, completing a large percentage of the excavation needed (see photographic material in Hill et al., 1898) before bankruptcies, Panama independence in 1903, and an operational take over by the United States in 1904 (de Banville, 2004).

It was not until 1910, with the bulk of the excavation nearly finished and reservoir flooding on the way, that the opportunities afforded by the unprecedented man–made digs were first realized (see Vaughan, 1946). Paleontological and biostratigraphic studies of the Canal Basin quickly followed (e.g., Berry, 1914; MacDonald, 1919) from a fruitful cooperation between the Canal Commission, the U.S. Geological Survey, and the Smithsonian Institution. Then, massive landslides of the fossiliferous clays of the Culebra and Cucaracha Formations in 1915–1916 blocked the recently opened Canal (see Brown, 1920), prompting in–depth geological and geotechnical studies of the Canal to begin in earnest (Becker, 1917; Lutton & Banks, 1970; MacDonald, 1947). This cooperation yielded beautiful geologic maps of the Canal Basin, as well as a solid stratigraphic and structural framework of the central part of the isthmus, all with a paleontological emphasis (see for instance Woodring & Thompson, 1949; Woodring, 1973; Stewart et al., 1980).

Away from the Canal Basin, studies motivated by Alexander Agassiz's early observations (Hill et al., 1898), and by mineral prospecting, recognized most of the basic geological elements of the isthmus (Hershey, 1901). These studies were much later complemented by the UNDP (United Nations Development Program) geologic and resource maps in the Azuero Peninsula (Giudice & Recchi, 1969), and in other regions of Panama including Darien and Bocas del Toro (United Nations Development Program, 1972). Petroleum exploration efforts also motivated early expeditions to Darien and the Chucunaque–Tuira basins setting up the basic stratigraphic framework (Shelton, 1952; Terry, 1956) in use today. During the last years of World War II, and the beginning of the Cold War, expansion of the canal became a military strategic priority, prompting the review of alternative routes to the east, including the Darien region (binational by then, Tavelli, 1947), evaluating the use of nuclear cratering as the main excavation technique (Sheffey et al., 1969).

2.1. Isthmus Basement and Cover

A distinction between basement and cover is used throughout this contribution. We call cover sequences those sequences, mostly sedimentary in origin, but also volcanic and volcaniclastic, that are separated by a nonconformity from a mostly volcanic, volcaniclastic and plutonic basement that is typically more intensely deformed than the cover above. This nonconformity records a period of deformation, cooling, exhumation, and erosion of the basement sequences before the accumulation of sequences above, which typically start in middle to late Eocene times. The distinction between basement and cover is concealed and blurred by a younger magmatic arc that is present west of the Canal Basin and north of the Azuero Peninsula (Figure 1), and also by the presence of accreted terranes along the southwestern edge of the isthmus. We therefore restrict our discussion to the older arc (see Rooney et al., 2015 for a review of the younger arc).

2.2. Isthmus Basement

The basement of the Isthmus of Panama consists of a Campanian to Eocene magmatic arc (Hoernle et al., 2008; Montes et al., 2012a; Wegner et al., 2011) built onto the southwestern, trailing edge of the Caribbean Plate (Pindell & Kennan, 2009). The Caribbean Plate that served as basement for the construction of the arcs is a complex, thickened oceanic plateau, buoyant and shallower than normal oceanic crust, with no magnetic anomalies (Case et al., 1990). This basement complex—plateau plus overlapping arc—seems fairly uniform throughout the isthmus, except for a magmatic hiatus in the San Blas Range. The transition from plateau to arc volcanism may have been marked by a period of pelagic/hemipelagic sedimentation that thus may serve as a regional key marker. Regional Bouguer anomalies over the San Blas Range (Figure 1, >120 milligal; Case, 1974; Westbrook, 1990) confirm that basement ranges consist of raised blocks of oceanic crust that host granitic intrusions. Geophysical anomalies are remarkably parallel and continuous along the axis of the San Blas Range, the Chucunaque Basin, and the North Panama Deformed Belt (Westbrook, 1990), suggesting continuity of lithologic units and structure, as also shown by geologic maps (Coates et al., 2004; Ministerio de Comercio e Industrias, 1991; Shelton, 1952).

2.2.1. The Plateau

The proto–Caribbean province came to be a Large Igneous Province (LIP) as Galapagos hotspot plume activity thickened it through ~70 Ma of hotspot volcanism (from 139 Ma to 69 Ma; Hoernle et al., 2002, 2004) to form one or several plateaus (~89 to 75 Ma pulses; Lissina, 2005; Baumgartner et al., 2008; Buchs et al., 2011b) collectively grouped—perhaps loosely—within the Caribbean Large Igneous Plateau (CLIP). This plateau may contain a much more diverse collection of fragments incorporated by subduction or collisions. Once the thickened Caribbean Plate was on a collision course with the westward–drifting American plates, its higher buoyancy favored accretion and preservation of its frontal fragments on the American margins. Fragments detached from the colliding leading edge of this plate are found today as accreted blocks along the northwestern margin of South America and the Antilles (e.g., Nivia, 1996; see review in Kerr et al., 2003). The processes involved in the formation of this plateau consolidated a complex igneous basement that seemingly left little or no trace of the original proto–Caribbean oceanic plate onto which the plume vulcanism took place. The exception to this in the isthmus may be distorted beds of red cherts and radiolarites intermingled with lava flows found only in the Azuero Peninsula near Torio (Hershey, 1901), south of Malena (Coniacian; Kolarsky et al., 1995), and fringing the Chortis Block in Nicaragua (Upper Triassic – Cretaceous; Baumgartner et al., 2008), among others. These strata may represent remnants of the pre–Campanian ocean floor onto which long–lived plateau volcanism took place.

Once the plateau was established, and while the Caribbean Plate was ploughing through the Americas with west–dipping subduction at its leading edge (Kennan & Pindell, 2009), subduction also started along its trailing edge, giving birth to the Central American arc in Cretaceous times (~70 Ma, Lissina, 2005; ~75–73 Ma, migrating east, Buchs et al., 2011b; ~71 Ma, and migrating east, Wegner et al., 2011). The location of this early Central American arc should outline an edge where the Caribbean Plate was thick, buoyant, perhaps still hot, at the time of subduction initiation. It is however, difficult to discriminate

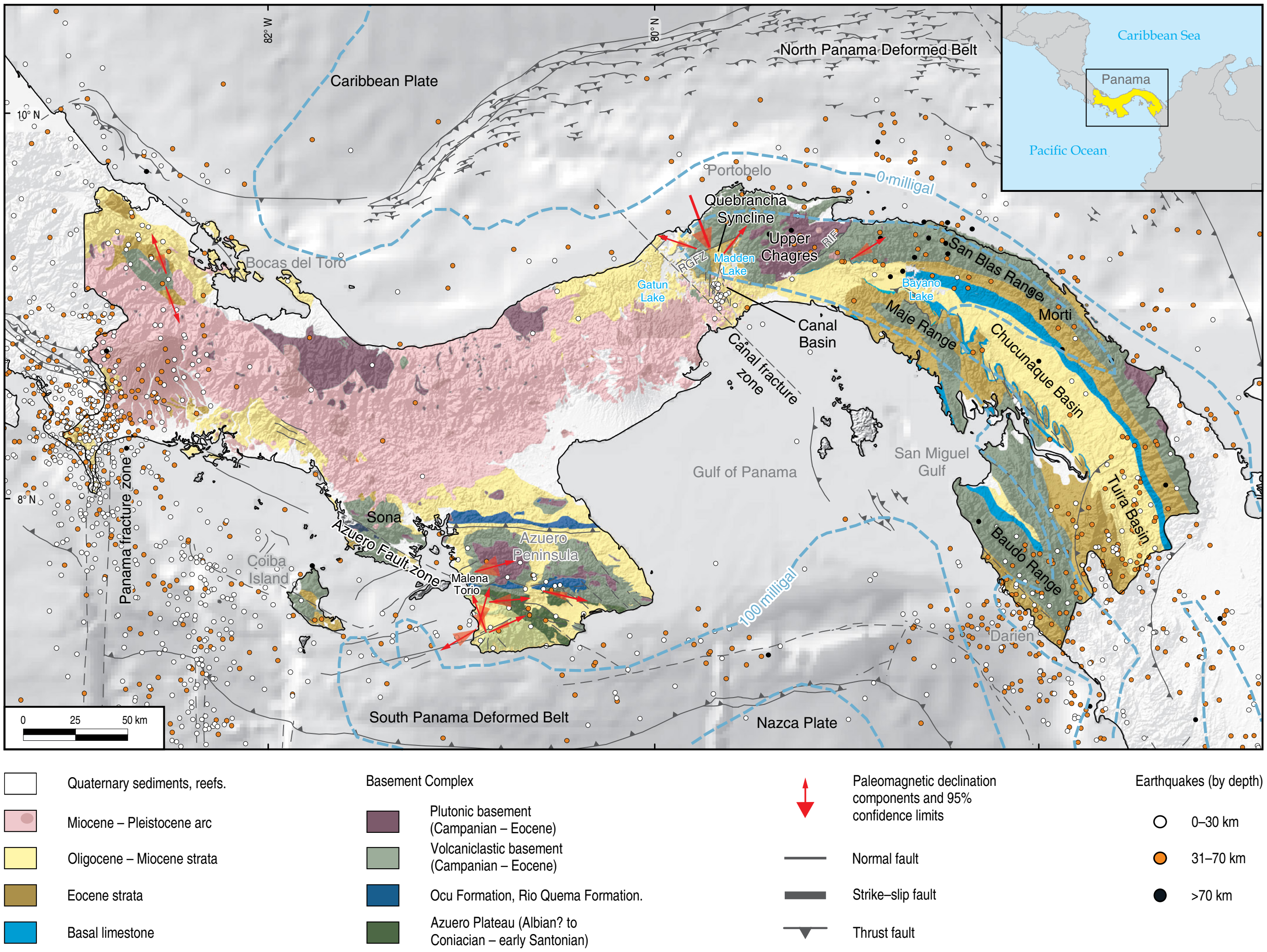


Figure 1. Geologic map of the Isthmus of Panama (modified from Buchs et al., 2010; Cowan et al., 1998; Di Marco et al., 1995; Giudice & Recchi, 1969; Kesler et al., 1977; Kirby et al., 2008; Ministerio de Comercio e Industrias, 1991; Montes et al., 2012b, 2015; Silver et al., 1990; Stewart et al., 1980; Rodríguez–Parra et al., 2017; Woodring, 1957). Topography and earthquakes from Amante & Eakins (2009), U.S. Geological Survey (2010, 2017). (RGFZ) Rio Gatun Fault zone; (RIF) Rio Indio Fault.

the magmatic products of the Galapagos hotspot from the first subduction–related magmatism, especially where Galapagos–modified crust is being subducted. Temporal and geochemical continuity between the last Galapagos–derived magmatism in the Caribbean Plate and the first subduction–related arc magmatism, open the possibility that subduction may have instead started as a result of the interaction between the hot plume and the cold lithosphere around it, thus hypothetically pushing the date of subduction initiation as far back as ~100 Ma (Whattam & Stern, 2015).

While the Caribbean Plate drifted away from the Galapagos hotspot, and became isolated by inward–dipping subduction zones along its leading and trailing edges, the hotspot continued its activity giving rise to plate asperities (Buchs et al., 2011a, 2016; Hoernle et al., 2002; Lissina, 2005) onto the Farallon Plate (Lonsdale, 2005; Lonsdale & Klitgord, 1978). These asperities—seamounts and ridges—were sequentially transported to the Central American trench where they eventually docked to a margin that thus contains the compressed history of the Galapagos hotspot. Docking of the more buoyant fragments, and subduction of the Galapagos–modified plate, left isotopic tracers that have been used to track margin–parallel asthenospheric flow as far west as Nicaragua (Gazel et al., 2011; Hoernle et al., 2008). The basement of the isthmus, has therefore always been interacting either directly or indirectly with the Galapagos hotspot. Directly, through the formation of large igneous provinces over nearly 70 Ma; indirectly, as plate asperities born in the hotspot have been brought to the trench, and have been either accreted, or assimilated in the asthenospheric flow.

2.2.2. Basement Composition

The basement of the isthmus is therefore made of thickened, long–lived Galapagos–derived oceanic plateau magmatic products (mostly submarine basalts) intercalated with two main sequences, also part of the basement complex: a Campanian – Maastrichtian pelagic and hemipelagic strata interbedded with submarine basalt flows and volcaniclastic materials, and a Campanian to Eocene subduction–related magmatic arc, volcanic and plutonic, and only partially subaerial. This of course, becomes a major source of ambiguity when mapping isthmus basement in the field, where distinct tectono–stratigraphic sequences are virtually identical basaltic successions. Geochemically also, most plateau and early arc and arc vulcanism are mafic with geochemical and isotopic signatures derived from, or contaminated by, Galapagos hotspot materials. Unambiguous geochemical discrimination of plateau versus arc vulcanism (e.g., Li et al., 2015), also remains a challenge in the isthmus.

2.3. Volcanic–Granitic Basement

The volcanic basement in most of the isthmus is composed of basalt flows, pillow basalt, diabase, interbedded chert and other siliceous sedimentary rocks intruded by granitoids and mafic dykes. Such sequences, with some variation, have been reported in the Morti River headwaters (Maury et al., 1995; Tavelli, 1947), near San Miguel Gulf (Bandy & Casey, 1973; Barat et al., 2014; Case, 1974), in the upper Chagres River (Wörner et al., 2005), and in a nearly 4 km thick sequence east of the Canal Basin (Montes et al., 2012b). Geologic explorations along the coastal transect of the northern flank of the San Blas Range confirm that its composition remains mostly basaltic and granitic, with granitoids between 59 and 39 Ma (Montes et al., 2015), but lacking the interbedded pelagic sediments ubiquitous in the southern flank of this range. This basement continues uniformly east to the Cuchillo Hills in the Darien region, perhaps with a greater thickness of the volcanic–volcaniclastic component with mafic tuffs, volcanic breccias, and cherts as young as middle Eocene (Barat et al., 2014), and magmatic activity as young as ~19 Ma (Whattam et al., 2012) correlative to the younger arc. Further east this sequence may continue into the Western Cordillera of Colombia (Case et al., 1971) with the Mande Batholith (Villagómez et al., 2011; Montes et al., 2015; see review in León et al., 2018).

A similar basement composition has been reported, and more thoroughly studied in the Azuero Peninsula, both in its eastern and western sides. In the eastern side of the peninsula, Corral et al. (2016) describe and map the ~1600 m thick Rio Quema Formation, a folded unit that sits on top of a basement that is composed of basalts, pillow basalts, and interbedded chert. The Rio Quema Formation contains a lower, crystal–rich sandy unit, a middle hemipelagic limestone unit, and siltstone upper unit, all intruded by dikes and a mineralized dacite. The Rio Quema Formation is covered by younger volcanic and volcaniclastic material with arc affinities. In the western side of the Azuero Peninsula, Buchs et al. (2011b) describe a basement sequence consisting of massive, columnar, and pillow basalts interbedded with small volumes red siliceous pelagic sediment with Coniacian – early Santonian radiolarians. This sequence is covered by hemipelagic limestones of the Ocu Formation and interbedded basalts, and then volcanic and plutonic rocks with arc affinities.

2.4. Hemipelagic and Pelagic Sequences

A key marker horizon that may help discriminate the different tectono–stratigraphic packages within the basement complex is defined by hemipelagic light–colored carbonates, and pelagic siliceous sequences. These pelagic/hemipelagic sequences may correspond to the B” horizon mapped, nearly reached in the ODP 999 in the Kogi Rise in the Caribbean Plate interior (Abrams & Hu, 2000; Bowland, 1993; Röhl & Abrams, 2000). Geologic maps of the eastern Azuero Peninsula by Corral et al. (2011, 2013), and by Montes et al. (2012b) in the southern flank of the San Blas Range, have delineated the outcrop pattern, and cross–cutting relationships of hemipelagic carbonates and pelagic mudstone and chert with the tectono–stratigraphic units above and below. Both studies use these sequences to separate plateau products from arc products, thought to represent the transition from plume vulcanism to subduction–related vulcanism.

Despite being a conspicuous unit (or units), the tectono–stratigraphic position of the hemipelagic sequences in the Azuero Peninsula remains unclear. Buchs et al. (2011b) report them resting on the CLIP, interbedded with the early arc system, and also within the accreted intraoceanic islands (Buchs et al., 2011a). These unusually thick hemipelagic limestones are characteristically light–colored, bioturbated, and contain planktonic foraminifera dated as late Campanian to Maastrichtian (Buchs et al., 2010; Corral et al., 2013; Fisher & Pessagno, 1965; Giudice & Recchi, 1969). Hemipelagic carbonates include the Torio Limestone and Ocu Formation, no more than 200 m thick in the Azuero Peninsula (Giudice & Recchi, 1969; Hershey, 1901), 350 m thick in Coiba Island (Kolarsky et al., 1995), ~1 km thick in eastern Azuero (Rio Quema Formation; Corral et al., 2016), and ~1 km thick in Bocas del Toro (Changuinola Formation; Fisher & Pessagno, 1965). Reports of hemipelagic limestones with similar ages also come from the San Miguel Gulf, the Portobelo Peninsula (Barat et al., 2014), extending the range of these limestones to the central and easternmost isthmus.

Still very poorly studied, the undifferentiated basalt sequences (Stewart et al., 1980) in the upper Chagres River catchment area are interbedded with thin pelagic beds of black siliceous siltstones, shales with thin sandstone stringers, and cherts. The siliceous pelagic sequence could be correlative to the hemipelagic limestone units above described, perhaps recording deeper accumulation environments. The age of this pelagic–hemipelagic sequence must be younger than the mafic volcanics on which it rests (pre–Campanian?), and older than the oldest granitoids that intrude them in the San Blas Range (~59 Ma; Montes et al., 2012b, 2015).

2.4.1. Basement Age

Several magmatic arcs are superimposed in the isthmus, with magmatic activity being nearly continuous west of the Canal Basin, and with magmatic gaps east of it. As stated above, only those arc–related rocks of Campanian to Eocene age are included within the basement complex. Those include intrusives in the eastern and western edges of the San Blas Range yielding hornblende and feldspar K/Ar ages between ~61 and 48 Ma (Kesler et al., 1977). Similarly, Ar/Ar dates in amphibole and plagioclase of unreported rock type, in its western half, range between ~66 and 41 Ma (Wegner et al., 2011). U/Pb zircon geochronological studies in granitoids of the basement complex exposed in the San Blas Range confirmed this age distribution with dates between ~59 and 39 Ma (Montes et al., 2012b, 2015; Ramírez et al., 2016), further constrained by zircon U/Pb detrital analyses in modern river sands, and in Eocene – Miocene strata (Montes et al., 2012b; Ramírez et al., 2016). Geochronological K/Ar, Ar/Ar, and U/Pb dates in the Azuero Peninsula range from ~71 to 41 Ma in basalts and granitoids (Corral et al., 2016; Giudice & Recchi, 1969; Kesler et al., 1977; Lissina, 2005; Montes et al., 2012a; Wegner et al., 2011). Two Ar/Ar step–heating plateau ages from the volcaniclastic sediments of the Rio Quema Formation are reported but discarded (143 ± 11 Ma and 105 ± 3 Ma; Corral et al., 2016), solely on the basis of being too old, therefore requiring further scrutiny as they still are within the age range of radiolarian determinations.

Lack of detailed geologic maps in the western side of the Azuero Peninsula hinders efforts to construct a reliable geochronological framework. Coastal transects along southwestern Azuero and Sona peninsulas report a large spread of ages with Ar/Ar step–heating plateaus between ~71 and 20 Ma (Hoernle et al., 2002). It has been noted though, that these ages are in conflict with the ages of overlying cover strata (Eocene Tonosi Formation; Kolarsky et al., 1995), so step–heating plateau ages may have suffered from Ar loss, and therefore may be unreliable (Buchs et al., 2011a). Since detailed geologic maps are yet to be produced in southwestern Azuero, the relationships of Eocene strata to sampled basalt sequences are still open to debate, as it is the tectono–stratigraphic affinity (plateau basement versus exotic accreted seamounts) of dated samples along coastal transects. Although geologic maps of southeastern Azuero show a nonconformable relationship of Eocene strata to isthmus basement (Mann & Kolarsky, 1995), such relationship may not be extrapolated to southwestern Azuero across the Azuero–Sona Fault zone, as this fault may represent the basement versus exotic boundary.

2.4.2. Basement Cooling Ages

The cooling history of the isthmus is recorded by numerous intermediate and felsic intrusive bodies exposed in the Azuero Peninsula and the San Blas Range all along the isthmus. Middle Eocene cooling events recorded by thermochronometers have been tied to exhumation and erosion by mapping the nonconformable relationship between the basement complex and upper Eocene and younger sedimentary sequences (Figure 2). Basement sequences below this nonconformity are pervasively deformed, tightly folded and faulted, while sequences above are simply tilted and folded (see for instance Figure 3c in Montes et al., 2012b). Cooling of the basement rocks in the San Blas Range is consistent with those nonconformities: apatite fission track and apatite and zircon U–Th/He analyses from 58–54 Ma granitoid bodies east of the Canal Basin record cooling from ~200 °C to ~70 °C (47–42 Ma), and cooling below ~40 °C between 12 and 9 Ma (Montes et al., 2012b). Apatite–zircon U–Th/He and fission–track thermochronology from the central part of the isthmus mark a cooling event between 22–28 Ma, with a peak at ~25 Ma, simultaneous with the onset of magmatism in the Canal Basin (Farris et al., 2011). Ramírez et al. (2016), using apatite–zircon U–Th/He, and fission tracks, show that most of the plutonic bodies of the isthmus were intruded and rapidly cooled to below ~200–110 °C by 30–40 Ma. The same authors, using modern sands from the Mamoni and Portogandi Rivers (draining the southern and northern flanks of the San Blas Range respectively) reveal a large spread of apatite–He ages (41–9 Ma) that together with the stratigraphic sequences onlapping the basement complex, suggest southward tilting of a crustal block (Figure 2, see below). In general, thermochronology of basement sequences shows a coherent history of early cooling of the roots of the magmatic arc (perhaps at shallow crustal levels), erosion, and development of a corresponding, regional nonconformity throughout the San Blas Range and the Azuero Peninsula.

2.4.3. Basement Deformation

The basement of the isthmus is intensely folded and faulted (Corral et al., 2016; Fisher & Pessagno, 1965; Montes et al., 2012b). Only locally this basement complex develops dynamic metamorphic lithologic types as foliated basalts and mylonites along the Azuero–Sona Fault zone (Buchs et al., 2011b; Hershey, 1901; Mann & Corrigan, 1990; Tournon et al., 1989), and along the Rio Gatun Fault zone (Wörner et al., 2005). The hemipelagic carbonate–basalt sequences described above serve as strain markers,

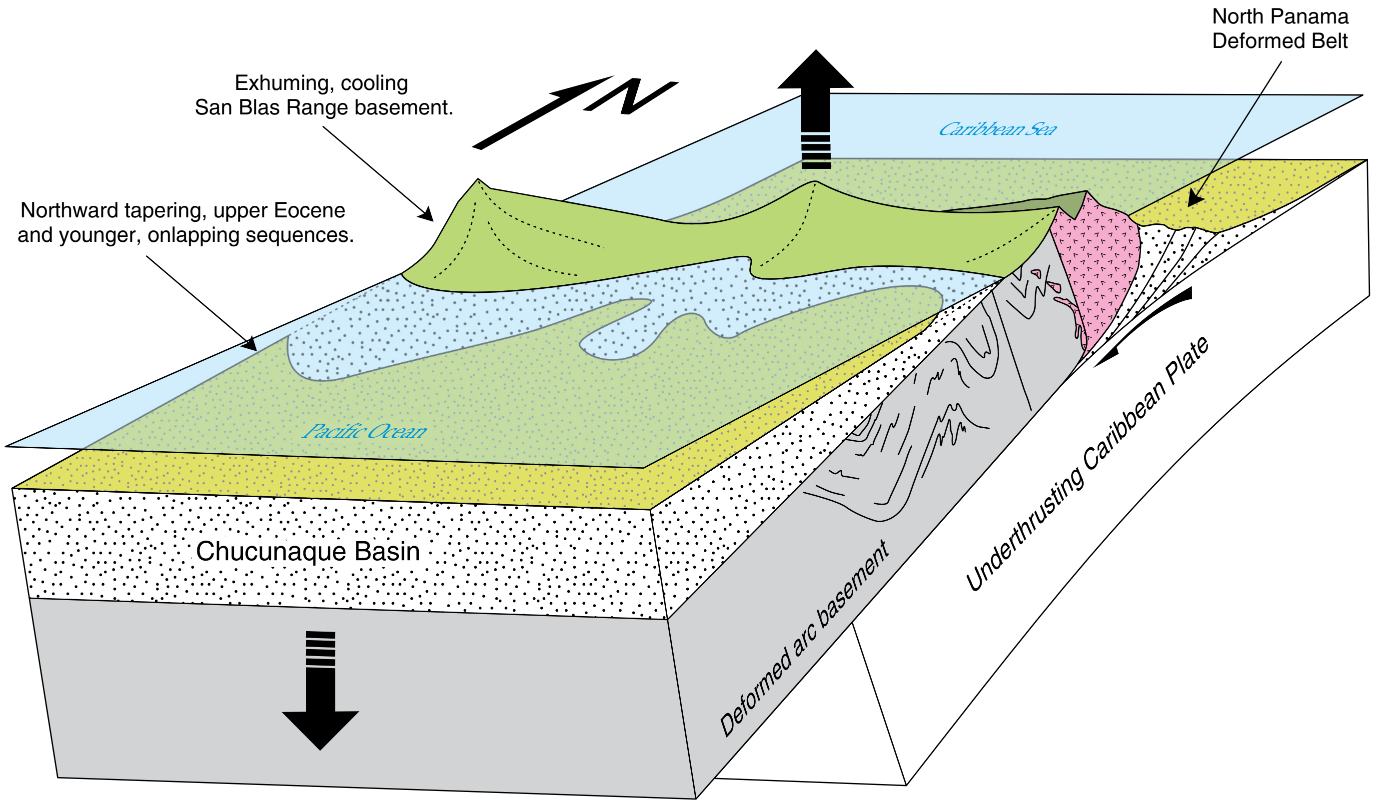


Figure 2. Conceptual diagram showing tilting of the Isthmus of Panama, exhumation and erosion of the basement complex and development of a northward–tapering, upper Eocene and younger sequence. See text for explanation.

revealing intense folding and faulting within the basement complex, marking a period of significant deformation after accumulation of these hemipelagic sequences, and before the accumulation of the first Eocene clastic/carbonate strata (Montes et al., 2012b).

Fault–related rocks in the Azuero Peninsula have been grouped as a melange (Buchs et al., 2010, 2011b) along the Azuero–Sona Fault zone. Vannucchi et al. (2006, 2007, 2013) in Costa Rica classify the Osa Melange as coherent, mappable blocks of fault–related rocks recording fragile to ductile conditions. Whether the sequences Vannucchi and co–workers refer to, are continuous into the Azuero Peninsula as coherent units, or they are a chaotic mixture of fault rocks and olistostromes formed in a subduction channel is still matter of debate. Detailed mapping should help locate the boundary, and the nature of the sequences involved, between the autochthonous plateau–related basalts, and the exotic, seamount–related basalt sequences.

In summary, the basement of the Isthmus of Panama consists of a fairly homogeneous deformed belt of submarine basalts and andesites, interlayered with pelagic and hemipelagic sequences of Campanian – Maastrichtian ages, and rare occurrences of red, older radiolarites, probably pre–Campanian in age. This complex is intruded by intermediate granitoids (68 to 39 Ma) that cooled quickly after intrusion, and were intensely deformed, exposed, and eroded before late Eocene times. A younger set of Galapagos–born oceanic asperities was accreted in early Paleogene times to the southwestern border of the isthmus and is therefore not considered part of the basement of the isthmus.

2.5. Isthmus Cover Sequences

We consider cover sequences those packages of rock accumulated after middle to late Eocene times, and discriminated from the basement by the absence of pervasive deformation. The absence of Paleocene – lower Eocene strata (Kolarsky & Mann, 1995; Woodring, 1957), marks a prominent, isthmus–wide hiatus, probably related to the accretion of plate asperities born in the Galapagos hotspot (Buchs et al., 2011a; Lissina, 2005), collisions with other plateaus (Kerr & Tarney, 2005), or very early interactions with South America (Barat et al., 2014).

Cover sequences are mostly clastic and carbonates in two main, and strikingly different onshore sedimentary basins: the Canal and the Chucunaque basins. Middle Eocene and younger sedimentary packages are also preserved in the Azuero Peninsula, in the Quebro, Mariato, and Tonosi valleys. Offshore, the North Panama Deformed Belt (Silver et al., 1990), and the Gulf of Panama to the south (Kolarsky et al., 1995), are sites of active sedimentation and contain mostly Neogene sedimentary packages (Barat et al., 2014). Cover sequences are influenced by volcanism where interbedded tuffs and volcano–sedimentary packages are abundant.

2.5.1. Canal Basin

The Canal Basin is defined by a complex collection of fault–bound compartments containing a very heterogeneous, but generally thin (~500 m), mostly clastic infill, intruded by large subvolcanic, mafic bodies and spotted by volcanic edifices. In a broader sense, the Canal Basin—centered around the Culebra Cut—also includes the Quebrancha Syncline, and the Alajuela (Madden) Basin to the north, as well as the Gatun Basin to the northwest. The Canal Basin sits at the westernmost tip of the San Blas Range, where not only the San Blas Range basement complex has its last outcrops (Stewart et al., 1980), but also where nearly continuous gravity anomalies following the axis of the range have their first break, and are found displaced to the southern Azuero Peninsula (Case, 1974; Westbrook, 1990). A fault, sometimes called the Canal fracture zone (Wolters, 1986), needed to explain the ~100 km left–lateral offset of the Campanian to Eocene magmatic arc from Azuero to the San Blas Range, must be located—but concealed—by younger volcanics west of the Canal (Montes et al., 2012a; Recchi & Metti, 1975). This fault may be continuous north of the North Panama Deformed Belt, as inferred by changes in seismic facies (Figure 16b of Bowland, 1993), and the texture in regional gravity anomaly maps (Figure 4 of Carvajal–Arenas & Mann, 2018). This northwest–trending fault, roughly parallel to the East Panama Deformed Belt (Mann & Kolarsky, 1995), was active at ~28 Ma (Montes et al., 2012a) and could be the master fault along which the isthmus reached its present northward–convex shape. As the isthmus was being offset along this fault, northeast–trending, strike–slip, right–stepping faults became active (Rio Gatun and Rio Indio faults) that, if both have a dextral character, must have defined an extensional step–over, with conjugate N–S trending, and NNE–trending faults (Azota and Pedro Miguel faults). Extension, subsidence, vertical–axis rotation, and a change in the character of volcanism from a hydrous subduction magmatism to extensional arc magmatism took place between 21–25 Ma as a result of this kinematic arrangement, and tectonic thinning of the crust of the isthmus (Farris et al., 2011, 2017; Montes et al., 2012a).

Sedimentation in the Canal compartment (see comprehensive description in Woodring, 1957) was nearly always punctuated by some volcanic activity (Farris et al., 2017). It starts with very coarse–grained, volcanic and volcaniclastic interfingering deposits of the middle Oligocene (~25 Ma; Rooney et al., 2011), Bohio and Bas Obispo Formations. These units are overlain by the volcaniclastic, tuffaceous, fossil–rich, lower Miocene (~21 Ma; Bloch et al., 2016) Las Cascadas Formation, grading to the east and northeast into a tuffaceous shallow–marine Caimito Formation. Shallow marine conditions were established by ~19 Ma (Montes et al., 2012b) in the Canal compartment with the accumulation of the Culebra Formation. This sequence is followed by the subaerial Cucaracha Formation, and then by the volcanic Pedro Miguel Formation (~18 Ma; Wegner et al., 2011). Younger deposits are present to the south of the Culebra Cut, but in general are poorly exposed.

Sedimentation in the Canal compartment is coeval with sedimentation in the other compartments, except for the presence of Eocene strata in the Quebrancha and Alajuela compartments (Coryell & Embich, 1937; Tripati & Zachos, 2002), predating the opening of the Canal compartment. Upper Eocene strata in the westernmost San Blas Range is probably related to the beginning of the Paleocene – Eocene tilting of the isthmus (see below). Rediscovered upper Miocene fossiliferous strata in the Alajuela compartment (MacFadden et al., 2017), suggests continuity of Gatun, Chucunaque, and Alajuela Formations across the isthmus at that time. The Gatun compartment, on the other hand, acted as the up–thrown block of the Rio Gatun Fault, with no record of sedimentation before the accumulation of the clastic and volcaniclastic wedges of the Gatun (Hidalgo et al., 2011; Rooney et al., 2015) and Chagres Formations (upper Miocene; Collins et al., 1996).

2.5.2. Chucunaque Basin

The Chucunaque Basin is an elongated, east–plunging, oroclinally curved (Montes et al., 2012a) trough whose north–northeastern flank is defined by a simple, south and southwest–dipping onlap onto the basement complex above described. The south–southwestern limb of this trough is more complex, where en–echelon left–stepping, left–lateral, north–south trending folds start near the Baudo Range, and culminate in the Maje Range (Mann & Corrigan, 1990; Mann & Kolarsky, 1995; Stephan et al., 1986). Cross–sections showing normal faults (Barat et al., 2014) are not supported by any geologic mapping, so they may not represent the structure of the basin. Onlapping strata are older (middle – upper Eocene), near the northern compartments of the Canal Basin, getting younger to the east, so that middle Miocene strata directly onlap volcanic and volcaniclastic rocks that may be considered basement in the Darien region (Coates et al., 2004; Shelton, 1952). This basin could be continuous southward into the Atrato Basin (Coates et al., 2004; Duque–Caro, 1990).

The sedimentary sequence in the Chucunaque–Tuira Basin starts with the middle – upper Eocene Gatuncillo Formation, which is better known in the Canal Basin, but has outcrops as far east as the Mamoni–Terable River, and other affluents of the Bayano River in the westernmost area of the Chucunaque Basin (Terry, 1956; Tripati & Zachos, 2002). The Gatuncillo Formation is a fining–upward mudstone, conglomerate–sandstone, and carbonate unit, of very variable thickness and facies, that nonconformably rests on the volcanic–plutonic basement complex (Woodring, 1957), deposited in fluvial to shallow–marine environments. To the east in the Darien region, the sedimentary sequence starts with ~400 m shale, limestone, and arkosic sandstone with large foraminifera (Shelton, 1952; lower bathyal, middle upper Oligocene Pocorna Formation, Coates et al., 2004; shallow marine, upper Oligocene, Barat et al., 2014), sitting nonconformably onto volcanic basement. This unit is conformably followed by hard, gray, tuffaceous limestone, locally nearly lithographic, more massive to the base (Shelton, 1952; lower bathyal, middle Miocene Clarita Formation, Coates et al., 2004,), directly onlapping the basement to the east (Shelton, 1952). This unit grades transitionally to a massive, uniform dark brown, calcareous foraminiferal shale with leaf remains and tuffs and thin sandstone beds (Arusa Formation of Shelton, 1952; or middle bathyal, middle to upper Miocene Tapaliza Formation of Coates et al., 2004; near–shore depths, Barat et al., 2014). This unit transitionally changes to a more conglomeratic and arkosic unit, with shales and dark brown/black carbonaceous material and lignite beds (Aquaqua Formation of Shelton, 1952; or neritic, upper Miocene Tuira Formation of Coates et al., 2004). This is in turn followed by the more regionally extensive, and lithologically more uniform Chucunaque Formation, correlative to the Gatun (Collins et al., 1996), and Alajuela (MacFadden et al., 2017) Formations.

Ages reported by Coates et al. (2004), for Cenozoic strata in the Chucunaque Formation have been confirmed using nannofossils and other microfossils (Barat et al., 2014). A major difference between these two studies, however, resides in paleobathymetric interpretations (see above), highlighting the need of a multi–proxy approach that includes—and prioritizes—sedimentological observations over microfossil inferences. Microfossils can be transported and reworked, severely limiting their usefulness as paleobathymetric indicators (see discussion in Jorissen et al., 2007). Multi–proxy approaches that include sedimentological, or ichnological analyses are conspicuously absent in paleobathymetric estimations in isthmian strata (Coates et al., 2004; Collins et al., 1996).

2.5.3. Offshore Deformed Belts, North, and South

Large negative gravity anomalies over the north and south Panama deformed belts outline the position of thick prisms of deformed, low–density strata (Case et al., 1990; Westbrook, 1990), thickened by Neogene convergence (Breen et al., 1988; Camacho et al., 2010; Reed & Silver, 1995; Reed et al., 1990; Silver et al., 1990, 1995). To the north, this deformed belt may contain a thick, Eocene and younger sedimentary sequence involved in an accretional belt with northward vergence (Reed & Silver, 1995; Rodríguez & Sierra, 2010; see seismic sections in Barat et al., 2014). To the south, surface sections, seismic sections, and boreholes have shown the presence of thick, middle Eocene and younger clastic sequences resting nonconformably on basement complex rocks (Kolarsky & Mann, 1995; Kolarsky et al., 1995; Mann & Kolarsky, 1995). These clastic sedimentary sequences may have resulted from erosion of the axis of the San Blas Range, and the Azuero Peninsula (Herrera et al., 2012; Krawinkel et al., 1999; Pérez–Consuegra et al., 2018) as they shed clastic materials to the north and south, including high–quartz clastics product of the erosion of intermediate and felsic intrusives, which were at the surface from latest Eocene times (Montes et al., 2012b; Ramírez et al., 2016). Deformation along the North Panama Deformed Belt may have started in middle Miocene times resulting from incipient south–dipping underthrusting/subduction of the Caribbean Plate under the Panama Block defining a Wadati–Benioff zone (Camacho et al., 2010; Wolters, 1986). A younging–eastward sequence directly overlapping the basement (see above) may record the direction of subduction/underthrusting initiation. Since penetration of the Caribbean Plate is ~150 km (Camacho et al., 2010), and assuming the current convergence rate of 11 mm/yr (Kellogg & Vega, 1995), a minimum age of underthrusting/subduction initiation would be middle Miocene. A slower convergence rate, likely during the initial stages of underthrusting of the Caribbean Plate, would push the age of initiation to Oligocene, and perhaps latest Eocene times (Figure 2).

2.6. The Isthmus Is Tilted, Bent, and Broken

An overall isthmus–wide southward tilting was recognized since the very first geological explorations of the isthmus (Hershey, 1901). This overall tendency can also be read in geological maps of the San Blas Range (Coates et al., 2004; Montes et al., 2012b; Shelton, 1952) that show a simple northward onlap of Eocene – Oligocene strata onto the deformed basaltic/granitic basement complex below, and a corresponding southward thickening of the same strata. The first post–hiatus strata (upper Eocene) record shallow marine, or even fluvial accumulation environments that later become deeper, punctuated by small hiatuses, that get progressively shallower towards the top. Although unencumbered by sedimentological data, paleobathymetric analyses (Coates et al., 2004) show first fluvial/coastal environments, followed by relative deepening, and then shoaling accumulation environments. Such sequence is better explained by an isthmus–wide tilting, erosion, and progressive filling of sedimentary basins, all while highlands to the north provided clastic materials to fill up the basin (Montes et al., 2012b). Southward tilting predicts that as the San Blas Range gained elevation, so the Chucunaque Basin gained accommodation space. It is therefore the interplay between subsidence rate and sediment availability—not a passive sill shoaling from ocean depths—that dictates changes in bathymetry. Detrital thermochronology in modern river sands on both flanks of the San Blas Range independently suggest that the wide distribution of cooling ages found (Ramírez et al., 2016), is better explained by tilting and exposure of the upper crustal section of the San Blas Range.

The isthmus is also bent. Paleomagnetic data in the isthmus show that the northward–convex shape of the isthmus may be the result of oroclinal bending (Montes et al., 2012a) that would have taken place after magmatic arc shut–down (~39 Ma), and was nearly completed around the time arc reinitiation, and Canal Basin opening in late Oligocene times. The incipient subduction/overthrusting mapped by Camacho et al. (2010), suggest that the isthmus detached from the trailing edge of the Caribbean Plate east of the Canal Basin. Approximately 150 km of subduction/overthrusting resulted from tightening the orocline that started forming following initial collision with western South America, starting at ~25 Ma (Farris et al., 2011). An under–thrusted, buoyant Caribbean plate provides geodynamic support for the tilted San Blas Range. The western half of the isthmus was affected by the collision of intraoceanic plate asperities born in the Galapagos hotspot (Buchs et al., 2011a), causing vertical–axis rotations in the Azuero Peninsula (Rodríguez–Parra et al., 2017), and shifting magmatic focus to the north during Paleocene – Eocene times.

The Canal Basin is broken at the point where the isthmus reaches its lowest topographic elevation, originally reaching ~90 m above sea–level, thus making it the best location for the construction of the interoceanic canal. The Canal Basin is also the point where the Campanian to Eocene magmatic arc is displaced nearly 100 km to the northwest (Lissina, 2005; Montes et al., 2012a; Recchi & Metti, 1975; Wolters, 1986), and where geophysical anomalies indicate changing basement types (Case, 1974; Westbrook, 1990). Tightening of the orocline, and left–lateral displacement of the arc may have contributed to thin the crust of the isthmus, changing magma sources, and open the Canal Basin (Farris et al., 2011, 2017).

3. A Way Forward for Isthmian Geology

This attempt to provide a review of current understanding of isthmian geology highlights just a few of the many issues in this topic. For instance, the stratigraphic location of the pelagic/hemipelagic strata within the basement sequences. Also, the bathymetry recorded by Neogene strata in the Chucunaque Basin, or the nature, or even the existence, of the Canal Basin Fault zone, and the cause of the Paleocene – Eocene deformation in the isthmus. Finally, the age of accretion and location of the boundary between the plateau and accreted sequences. A recurrent underlying problem that is common to most of these issues is rooted in the lack of consistent, standardized, detailed geologic maps at scales larger than 1:250 000. This problem is particularly acute in the eastern part of the isthmus, and even more pronounced east of the border in the Choco–Darien region. Another related problem is the reliance of single proxies to interpret paleobathymetric data, when a multi–proxy approach that includes sedimentological, ichnological, or other primary features, that cannot be transported, would provide more reliable results.

In general, except for the Canal Basin, most of the isthmus has only been through one generation of geologic mapping, and most of it has only been reconnaissance mapping at very general scales (1:250 000 or smaller). As noted by Woodring in 1957, even in the thickest jungle conditions, the availability of closely–spaced drainages and steep topography offer a very rich network of fresh outcrops where accurate geologic maps can be made. The geology of the isthmus is far too complex to attempt understanding it without basic geologic mapping to support analytical efforts.

Acknowledgments

We thank Smithsonian Tropical Research Institute (STRI), Ricardo Perez S.A., Ministerio de Industria y Comercio, the Panama Canal Authority, The University of Florida, and Universidad de los Andes for logistical and scientific support. Natalia HOYOS was partially funded by The Canadian Queen Elizabeth II Diamond Jubilee Scholarships (QES), a partnership among Universities in Canada, the Rideau Hall Foundation (RHF), Community Foundations of Canada (CFC). The QES–AS is made possible with financial support from IDRC and SSHRC. Natalia HOYOS was also partially supported by the Fulbright Visiting Scholar Program. This review benefited from the enthusiastic help and sharp observations of more than 100 undergraduate and graduate students, professional geologists, paleontologists, biologists that took part in field expeditions, Canal paleontological rescue, and university field camps. Thanks to Dr. Agustín CARDONA and David BUCHS for comments and suggestions.

References

Abrams, L.J. & Hu, M. 2000. Data report: Depth to volcanic basement at Site 999, Kogi Rise, Colombian Basin. In: Leckie, R.M., Sigurdsson, H., Acton, G.D. & Draper, G. (editors), Proceedings of the Ocean Drilling Program, Scientific Results 165, p. 219–224. College Station, USA. https://doi.org/10.2973/odp.proc.sr.165.028.2000

Amante, C. & Eakins, B.W. 2009. ETOPO1 1 Arc–minute global relief model: Procedures, data sources and analysis. NOAA Technical Memorandum NESDIS NGDC–24. National Oceanic and Atmospheric Administration, 25 p.

Bacon, C.D., Silvestro, D., Jaramillo, C., Smith, B.T., Chakrabarty, P. & Antonelli, A. 2015. Biological evidence supports an early and complex emergence of the Isthmus of Panama. Proceedings of the National Academy of Sciences of the United States of America, 112(19): 6110–6115. https://doi.org/10.1073/pnas.1423853112

Bandy, O.L. & Casey, R.E. 1973. Reflector horizons and paleobathymetric history, eastern Panama. Geological Society of America Bulletin, 84(9): 3081–3086. https://doi.org/10.1130/0016-7606(1973)84<3081:RHAPHE>2.0.CO;2

Barat, F., De Lépinay, B.M., Sosson, M., Müller, C., Baumgartner, P.O. & Baumgartner–Mora, C. 2014. Transition from the Farallon Plate subduction to the collision between South and Central America: Geological evolution of the Panama Isthmus. Tectonophysics, 622: 145–167. https://doi.org/10.1016/j.tecto.2014.03.008

Baumgartner, P.O., Flores, K., Bandini, A.N., Girault, F. & Cruz, D. 2008. Upper Triassic to Cretaceous radiolaria from Nicaragua and northern Costa Rica–The Mesquito composite oceanic terrane. Ofioliti, 33(1): 1–19.

Becker, G.F. 1917. Mechanics of the Panama Canal slides. In: White, D. (editor), Shorter contributions to general geology, 1916. Professional paper 98–N. U.S. Geological Survey, p. 253–261. Washington D.C. https://doi.org/10.3133/pp98N

Berry, E.W. 1914. Fossil plants in the Panama Canal zone. Science, 39(1001): p. 357. https://doi.org/10.1126/science.39.1001.357

Bloch, J.I., Woodruff, E.D., Wood, A.R., Rincón, A.F., Harrington, A.R., Morgan, G.S., Foster, D.A., Montes, C., Jaramillo, C., Jud, N.A., Jones, D.S. & MacFadden, B.J. 2016. First North American fossil monkey and early Miocene tropical biotic interchange. Nature, 533(7602): 243–246. https://doi.org/10.1038/nature17415

Bowland, C.L. 1993. Depositional history of the western Colombian Basin, Caribbean Sea, revealed by seismic stratigraphy. Geological Society of America Bulletin, 105(10): 1321–1345. https://doi.org/10.1130/0016-7606(1993)105<1321:DHOTWC>2.3.CO;2

Breen, N.A., Tagudin, J.E., Reed, D.L. & Silver, E.A. 1988. Mud–cored parallel folds and possible melange development in the north Panama thrust belt. Geology, 16(3): 207–210. https://doi.org/10.1130/0091-7613(1988)16[207:MPFAPM]2.0.CO;2

Brown, R.M. 1920. Five years of the Panama Canal: An evaluation. Geographical Review, 9(3): 191–198. https://doi.org/10.2307/207257

Buchs, D.M., Arculus, R.J., Baumgartner, P.O., Baumgartner–Mora, C. & Ulianov, A. 2010. Late Cretaceous arc development on the SW margin of the Caribbean Plate: Insights from the Golfito, Costa Rica, and Azuero, Panama, complexes. Geochemistry, Geophysics, Geosystems, 11(7): 1–35. https://doi.org/10.1029/2009GC002901

Buchs, D.M., Arculus, R.J., Baumgartner, P.O. & Ulianov, A. 2011a. Oceanic intraplate volcanoes exposed: Example from seamounts accreted in Panama. Geology, 39(4): 335–338. https://doi.org/10.1130/G31703.1

Buchs, D.M., Baumgartner, P.O., Baumgartner–Mora, C., Flores, K. & Bandini, A.N. 2011b. Upper Cretaceous to Miocene tectonostratigraphy of the Azuero area (Panama) and the discontinuous accretion and subduction erosion along the middle American margin. Tectonophysics, 512(1–4): 31–46. https://doi.org/10.1016/j.tecto.2011.09.010

Buchs, D.M., Hoernle, K., Hauff, F. & Baumgartner, P.O. 2016. Evidence from accreted seamounts for a depleted component in the early Galapagos plume. Geology, 44(5): 383–386. https://doi.org/10.1130/G37618.1

Camacho, E., Hutton, W. & Pacheco, J.F. 2010. A new look at evidence for a Wadati–Benioff zone and active convergence at the North Panama Deformed Belt. Bulletin of the Seismological Society of America, 100(1): 343–348. https://doi.org/10.1785/0120090204

Cardona, A., Valencia, V., Bayona, G., Duque, J., Ducea, M., Gehrels, G., Jaramillo, C., Montes, C., Ojeda, G. & Ruiz, J. 2011. Early–subduction–related orogeny in the northern Andes: Turonian to Eocene magmatic and provenance record in the Santa Marta Massif and Rancheria Basin, northern Colombia. Terra Nova, 23(1): 26–34. https://doi.org/10.1111/j.1365-3121.2010.00979.x

Cardona, A., Montes, C., Ayala, C., Bustamante, C., Hoyos, N., Montenegro, O., Ojeda, C., Niño, H., Ramírez, V., Valencia, V., Rincón, D., Vervoort, J.D. & Zapata, S. 2012. From arc–continent collision to continuous convergence, clues from Paleogene conglomerates along the southern Caribbean–South America Plate boundary. Tectonophysics, 580: 58–87. https://doi.org/10.1016/j.tecto.2012.08.039

Carvajal–Arenas, L.C. & Mann, P. 2018. Western Caribbean intraplate deformation: Defining a continuous and active microplate boundary along the San Andres Rift and Hess Escarpment fault zone, Colombian, Caribbean Sea. American Association of Petroleum Geologists Bulletin, 102(8): 1523–1563. https://doi.org/10.1306/12081717221

Case, J.E. 1974. Oceanic crust forms basement of eastern Panamá. Geological Society of America Bulletin, 85(4): 645–652. https://doi.org/10.1130/0016-7606(1974)85<645:OCFBOE>2.0.CO;2

Case, J.E., Durán, L.G., López, A. & Moore, W.R. 1971. Tectonic investigations in western Colombia and eastern Panama. Geological Society of America Bulletin, 82(10): 2685–2711. https://doi.org/10.1130/0016-7606(1971)82[2685:TIIWCA]2.0.CO;2

Case, J.E., Shagam, R. & Giegengack, R.F. 1990. Geology of the northern Andes: An overview. In: Dengo, G. & Case, J.E. (editors), The Caribbean region. Geological Society of America, p. 177–200. Boulder, USA. https://doi.org/10.1130/DNAG-GNA-H.177

Cediel, F., Shaw, R.P. & Cáceres, C. 2003. Tectonic assembly of the northern Andean Block. In: Bartolini, C., Buffler, R.T. & Blickwede, J. (editors), The circum–Gulf of Mexico and the

Caribbean: Hydrocarbon habitats, basin formation, and plate tectonics. American Association of Petroleum Geologists, Memoir 79, p. 815–848. Tulsa, USA.

Coates, A.G. & Stallard, R.F. 2013. How old is the Isthmus of Panama? Bulletin of Marine Science, 89(4): 801–813. https://doi.org/10.5343/bms.2012.1076

Coates, A.G., Collins, L.S., Aubry, M.P. & Berggren, W.A. 2004. The geology of the Darien, Panama, and the late Miocene – Pliocene collision of the Panama arc with northwestern South America. Geological Society of America Bulletin, 116(11–12): 1327–1344. https://doi.org/10.1130/B25275.1

Collins, L.S., Coates, A.G., Berggren, W.A., Aubry, M.P. & Zhang, J. 1996. The late Miocene Panama isthmian strait. Geology, 24(8): 687–690. https://doi.org/10.1130/0091-7613(1996)024<0687:TLMPIS>2.3.CO;2

Corral, I., Griera, A., Gómez–Gras, D., Corbella, M., Canals, Á., Pineda–Falconett, M. & Cardellach, E. 2011. Geology of the Cerro Quema Au–Cu deposit (Azuero Peninsula, Panama). Geologica Acta, 9(3–4): 481–498. https://doi.org/10.1344/105.000001742

Corral, I., Gómez–Gras, D., Griera, A., Corbella, M. & Cardellach, E. 2013. Sedimentation and volcanism in the Panamanian Cretaceous intra–oceanic arc and fore–arc: New insights from the Azuero Peninsula (SW Panama). Bulletin de la Société Géologique de France, 184(1–2): 35–45. https://doi.org/10.2113/gssgfbull.184.1-2.35

Corral, I., Cardellach, E., Corbella, M., Canals, Á., Gómez–Gras, D., Griera, A. & Cosca, M.A. 2016. Cerro Quema (Azuero Peninsula, Panama): Geology, alteration, mineralization, and geochronology of a volcanic dome–hosted high–sulfidation Au–Cu deposit. Economic Geology, 111(2): 287–310. https://doi.org/10.2113/econgeo.111.2.287

Coryell, H.N. & Embich, J.R. 1937. The Tranquilla Shale (upper Eocene) of Panama and its foraminiferal fauna. Journal of Paleontology, 11(4): 289–305.

Cowan, H.A., Dart, R.L. & Machette, M.N. 1998. Map of Quaternary faults and folds of Panama and its offshore regions. Scale 1:750 000. U.S. Geological Survey, 1 sheet. https://doi.org/10.3133/ofr98779

de Banville, M. 2004. Canal Français: L'aventure illustrée des français au Panama. 1880–1904. Éditeur Canal Valley, 194 p.

Di Marco, G., Baumgartner, P.O. & Channell, J.E.T. 1995. Late Cretaceous – early Tertiary paleomagnetic data and a revised tectonostratigraphic subdivision of Costa Rica and western Panama. In: Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 1–27. Boulder, USA. https://doi.org/10.1130/SPE295-p1

Douville, H. 1898. Sur l'âge des couches traversées par le canal de Panama. Bulletin de la Société Géologique de France, 3(26): 587–600.

Duque–Caro, H. 1990. Neogene stratigraphy, paleoceanography and paleobiogeography in northwest South America and the evolution of the Panama Seaway. Palaeogeography, Palaeoclimatology, Palaeoecology, 77(3–4): 203–234. https://doi.org/10.1016/0031-0182(90)90178-A

Farris, D.W., Jaramillo, C., Bayona, G., Restrepo–Moreno, S.A., Montes, C., Cardona, A., Mora, A., Speakman, R.J., Glascock, M.D. & Valencia, V. 2011. Fracturing of the Panamanian Isthmus during initial collision with South America. Geology, 39(11): 1007–1010. https://doi.org/10.1130/G32237.1

Farris, D.W., Cardona, A., Montes, C., Foster, D. & Jaramillo, C. 2017. Magmatic evolution of Panama Canal volcanic rocks: A record of arc processes and tectonic change. PLOS ONE, 12(5): 1–44. https://doi.org/10.1371/journal.pone.0176010

Fisher, S.P. & Pessagno, E.A. 1965. Upper Cretaceous strata of northwestern Panama. American Association of Petroleum Geologists Bulletin, 49(4): 433–444. https://doi.org/10.1306/A6633630-16C0-11D7-8645000102C1865D

Gazel, E., Hoernle, K., Carr, M.J., Herzberg, C., Saginor, I., van den Bogaard, P., Hauff, F., Feigenson, M.D. & Swisher III, C. 2011. Plume–subduction interaction in southern Central America: Mantle upwelling and slab melting. Lithos, 121(1–4): 117–134. https://doi.org/10.1016/j.lithos.2010.10.008

Giudice, D. & Recchi, G. 1969. Geología del área del proyecto minero de Azuero. Programa para el desarrollo de las Naciones Unidas, 53 p.

Herrera, F., Manchester, S.R. & Jaramillo, C. 2012. Permineralized fruits from the late Eocene of Panama give clues of the composition of forests established early in the uplift of Central America. Review of Paleobotany and Palynology, 175: 10–24. https://doi.org/10.1016/j.revpalbo.2012.02.007

Hershey, O.H. 1901. The geology of the central portion of the Isthmus of Panama. University of California, Bulletin of the Department of Geology, 2(8): 231–267.

Hidalgo, P.J., Vogel, T.A., Rooney, T.O., Currier, R.M. & Layer, P.W. 2011. Origin of silicic volcanism in the Panamanian Arc: Evidence for a two–stage fractionation process at El Valle Volcano. Contributions to Mineralogy and Petrology, 162(6): 1115–1138. https://doi.org/10.1007/s00410-011-0643-2

Hill, R.T., Dall, W.H., Bagg, M., Vaughan, T.W., Wolff, J.E., Turner, H.W. & Sjögren, A. 1898. The geological history of the Isthmus of Panama and portions of Costa Rica: Based upon a reconnoissance made for Alexander Agassiz. Bulletin of the Museum of Comparative Zoology at Harvard College, 28(5): 149–285.

Hoernle, K., van den Bogaard, P., Werner, R., Lissinna, B., Hauff, F., Alvarado, G. & Garbe–Schönberg, D. 2002. Missing history (16–71 Ma) of the Galapagos hotspot: Implications for the tectonic and biological evolution of the Americas. Geology, 30(9): 795–798. https://doi.org/10.1130/0091-7613(2002)030<0795:MHMOTG>2.0.CO;2

Hoernle, K., Hauff, F. & van den Bogaard, P. 2004. 70 m.y. history (139–69 Ma) for the Caribbean Large Igneous Province. Geology, 32(8): 697–700. https://doi.org/10.1130/G20574.1

Hoernle, K., Abt, D.L., Fischer, K.M., Nichols, H., Hauff, F., Abers, G.A., van den Bogaard, P., Heydolph, K., Alvarado, G., Protti, M. & Strauch, W. 2008. Arc–parallel flow in the mantle wedge beneath Costa Rica and Nicaragua. Nature, 451(7182): 1094–1097. https://doi.org/10.1038/nature06550

Jaramillo, C. 2018. Evolution of the Isthmus of Panama: Biological, paleoceanographic, and paleoclimatological implications. In: Hoorn, C., Perrigo, A. & Antonelli, A. (editors), Mountains, climate and biodiversity. Wiley–Blackwell, p. 323–338. Chichester, UK.

Jaramillo, C., Montes, C., Cardona, A., Silvestro, D., Antonelli, A. & Bacon, C.D. 2017. Comment (1) on "Formation of the Isthmus of Panama" by O'Dea et al. Science Advances, 3(6): 1–8. https://doi.org/10.1126/sciadv.1602321

Jorissen, F.J., Fontanier, C. & Thomas, E. 2007. Paleoceanographical proxies based on deep–sea benthic foraminiferal assemblage characteristics. In: Hillaire–Marcel, C. & de Vernal, A. (editors), Proxies in late Cenozoic paleoceanography. Elsevier Science, p. 263–325. https://doi.org/10.1016/S1572-5480(07)01012-3

Kellogg, J.N., & Vega, V., 1995. Tectonic development of Panama, Costa Rica, and the Colombian Andes: Constraints from Global Positioning System geodetic studies and gravity. In: Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 75–90. Boulder, USA. https://doi.org/10.1130/SPE295-p75

Kennan, L. & Pindell, J.L. 2009. Dextral shear, terrane accretion and basin formation in the northern Andes: Best explained by interaction with a Pacific–derived Caribbean Plate? In: James, K.H., Lorente, M.A. & Pindell, J.L. (editors), The origin and evolution of the Caribbean Plate. Geological Society of London, Special Publication 328, p. 487–531. https://doi.org/10.1144/SP328.20

Kerr, A.C. & Tarney, J. 2005. Tectonic evolution of the Caribbean and northwestern South America: The case for accretion of two Late Cretaceous oceanic plateaus. Geology, 33(4): 269–272. https://doi.org/10.1130/G21109.1

Kerr, A.C., White, R.V., Thompson, P.M.E., Tarney, J. & Saunders, A.D. 2003. No oceanic plateau–no Caribbean Plate? The seminal role of an oceanic plateau in Caribbean Plate evolution. In: Bartolini, C., Buffler, R.T. & Blickwede, J. (editors), The circum–Gulf of Mexico and the Caribbean: Hydrocarbon habitats, basin formation, and plate tectonics. American Association of Petroleum Geologists, Memoir 79, p. 126–168. Tulsa, USA.

Kesler, S.E., Sutter, J.F., Issigonis, M.J., Jones, L.M. & Walker, R.L. 1977. Evolution of porphyry copper mineralization in an oceanic island arc: Panama. Economic Geology, 72(6): 1142–1153. https://doi.org/10.2113/gsecongeo.72.6.1142

Kirby, M.X. & MacFadden, B. 2005. Was southern Central America an archipelago or a peninsula in the middle Miocene? A test using land–mammal body size. Palaeogeography, Palaeoclimatology, Palaeoecology, 228(3–4): 193–202. https://doi.org/10.1016/j.palaeo.2005.06.002

Kirby, M.X., Jones, D.S. & MacFadden, B.J. 2008. Lower Miocene stratigraphy along the Panama Canal and its bearing on the Central American Peninsula, PLOS ONE, 3(7): 1–14. https://doi.org/10.1371/journal.pone.0002791

Kolarsky, R.A. & Mann, P. 1995. Structure and neotectonics of an oblique–subduction margin, southwestern Panama. In Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 131–157. Boulder, USA. https://doi.org/10.1130/SPE295-p131

Kolarsky, R.A., Mann, P. & Monechi, S. 1995. Stratigraphic development of southwestern Panama as determined from integration of marine seismic data and onshore geology. In Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 159–200. Boulder, USA. https://doi.org/10.1130/SPE295-p159

Krawinkel, H., Wozazek, S., Krawinkel, J. & Hellmann, W. 1999. Heavy–mineral analysis and clinopyroxene geochemistry applied to provenance analysis of lithic sandstones from the Azuero–Soná Complex (NW Panama). Sedimentary Geology, 124(1–4): 149–168. https://doi.org/10.1016/S0037-0738(98)00125-0

Leigh, E.G., O'Dea, A. & Vermeij, G.J. 2013. Historical biogeography of the Isthmus of Panama. Biological Reviews, 89(1): 148–172. https://doi.org/10.1111/brv.12048

León, S., Cardona, A., Parra, M., Sobel, E.R., Jaramillo, J.S., Glodny, J., Valencia, V.A., Chew, D., Montes, C., Posada, G., Monsalve, G. & Pardo–Trujillo, A. 2018. Transition from collisional to subduction–related regimes: An example from Neogene Panama–Nazca–South America interactions. Tectonics, 37(1): 119–139. https://doi.org/10.1002/2017TC004785

Lessios, H.A. 2008. The great American schism: Divergence of marine organisms after the rise of the Central American Isthmus. Annual Review of Ecology Evolution, and Systematics, 39: 63–91. https://doi.org/10.1146/annurev.ecolsys.38.091206.095815

Li, C., Arndt, N.T., Tang, Q. & Ripley, E.M. 2015. Trace element indiscrimination diagrams. Lithos, 232: 76–83. https://doi.org/10.1016/j.lithos.2015.06.022

Lissina, B. 2005. A profile through the Central American landbridge in western Panama: 115 Ma interplay between the Galápagos hotspot and the Central American subduction zone. Doctoral thesis, Christian–Albrechts–Universität zu Kiel, 102 p. Kiel, Germany.

Lonsdale, P. 2005. Creation of the Cocos and Nazca Plates by fission of the Farallon Plate. Tectonophysics, 404(3–4): 237–264. https://doi.org/10.1016/j.tecto.2005.05.011

Lonsdale, P. & Klitgord, K.D. 1978. Structure and tectonic history of the eastern Panama Basin. Geological Society of America Bulletin, 89(7): 981–999. https://doi.org/10.1130/0016-7606(1978)89<981:SATHOT>2.0.CO;2

Lutton, R.J. & Banks, D.C. 1970. Study of clay shale slopes along the Panama Canal. Report 1: East Culebra and west Culebra slides and the model slope. U.S. Army Engineer Waterways Experiment Station. Technical report, 385 p. Vicksburg, USA.

MacDonald, D.F. 1919. The sedimentary formations of the Panama Canal zone, with special reference to the stratigraphic relations of the fossiliferous beds. In: Vaughan, T.W. (editor), Contributions to the geology and paleontology of the Canal zone, Panama, and geologically related areas in Central America and the West Indies. Smithsonian Institution, Unites States National Museum, Press Bulletin 103, p. 525–545. Washington D.C.

MacDonald, D.F. 1947. Panama Canal slides. Department of operation and maintenance, Special Engineering Division, Balboa Heights Canal Zone, 73 p.

MacFadden, B.J., Jones, D.S., Jud, N.A., Moreno–Bernal, J.W., Morgan, G.S., Portell, R.W., Pérez, V.J., Moran, S.M. & Wood, A.R. 2017. Integrated chronology, flora and faunas, and paleoecology of the Alajuela Formation, late Miocene of Panama. PLOS ONE: 12(1): 1–27. https://doi.org/10.1371/journal.pone.0170300

Mann, P. & Corrigan, J. 1990. Model for late Neogene deformation in Panama. Geology, 18(6): 558–562. https://doi.org/10.1130/0091-7613(1990)018<0558:MFLNDI>2.3.CO;2

Mann, P. & Kolarsky, R.A. 1995. East Panama Deformed Belt: Structure, age, and neotectonic significance. In: Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 111–130. Boulder, USA. https://doi.org/10.1130/SPE295-p111

Maury, R.C., Defant, M.J., Bellon, H., de Boer, J.Z., Stewart, R.H. & Cotten, J. 1995. Early tertiary arc volcanics from eastern Panama. In: Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 29–34. Boulder, USA. https://doi.org/10.1130/SPE295-p29

Ministerio de Comercio e Industrias. 1991. Mapa Geológico de Panamá. Scale 1:500 000. Ministerio de Comercio e Industrias–División General de Recursos Minerales, 1sheet.

Molnar, P. 2008. Closing of the Central American Seaway and the ice age: A critical review. Paleoceanography and Paleoclimatology, 23(2): 1–15. https://doi.org/10.1029/2007PA001574

Molnar, P. 2017. Comment (2) on "Formation of the Isthmus of Panama" by O'Dea et al. Science Advances, 3(6): 1–4. https://doi.org/10.1126/sciadv.1602320

Montes, C., Bayona, G., Cardona, A., Buchs, D.M., Silva, C.A., Morón, S., Hoyos, N., Ramírez, D.A., Jaramillo, C. & Valencia, V. 2012a. Arc–continent collision and orocline formation: Closing of the Central American Seaway. Journal of Geophysical Research: Solid Earth, 117(B4): 1–25. https://doi.org/10.1029/2011JB008959

Montes, C., Cardona, A., McFadden, R., Moron, S.E., Silva, C.A., Restrepo–Moreno, S., Ramírez, D.A., Hoyos, N., Wilson, J., Farris, D.W., Bayona, G., Jaramillo, C., Valencia, V., Bryan, J. & Flores, J.A. 2012b. Evidence for middle Eocene and younger land emergence in Central Panama: Implications for isthmus closure. Geological Society of America Bulletin, 124(5–6): 780–799. https://doi.org/10.1130/B30528.1

Montes, C., Cardona, A., Jaramillo, C., Pardo, A., Silva, J.C., Valencia, V., Ayala, C., Pérez–Ángel, L.C., Rodríguez–Parra, L.A., Ramírez, V. & Niño, H. 2015. Middle Miocene closure of the Central American Seaway. Science, 348(6231): 226–229. https://doi.org/10.1126/science.aaa2815

Nivia, Á. 1996. The Bolivar mafic–ultramafic complex, SW Colombia: The base of an obducted oceanic plateau. Journal of South American Earth Sciences, 9(1–2): 59–68. https://doi.org/10.1016/0895-9811(96)00027-2

O'Dea, A., Lessios, H.A., Coates, A.G., Eytan, R.I., Restrepo–Moreno, S.A., Cione, A.L., Collins, L.S., de Queiroz, A., Farris, D.W., Norris, R.D., Stallard, R.F., Woodburne, M.O., Aguilera, O., Aubry, M. –P., Berggren, W.A., Budd, A.F., Cozzuol, M.A., Coppard, S.E., Duque–Caro, H., Finnegan, S., Gasparini, G.M., Grossman, E.L., Johnson, K.G., Keigwin, L.D., Knowlton, N., Leigh, E.G., Leonard–Pingel, J.S., Marko, P.B., Pyenson, N.D., Rachello–Dolmen, P.G., Soibelzon, E., Soibelzon, L., Todd, J.A., Vermeij, G.J. & Jackson, J.B.C. 2016. Formation of Isthmus of Panama. Science Advances, 2(8): 1–11. https://doi.org/10.1126/sciadv.1600883

Pérez–Consuegra, N., Góngora, D.E., Herrera, F., Jaramillo, C., Montes, C., Cuervo–Gómez, A.M., Hendy, A., Machado, A., Cárdenas, D. & Bayona, G. 2018. New records of Humiriaceae fossil fruits from the Oligocene and early Miocene of the western Azuero Peninsula, Panamá. Boletín de la Sociedad Geológica Mexicana, 70(1): 223–239. http://dx.doi.org/10.18268/BSGM2018v70n1a13

Pindell, J.L. & Kennan, L. 2009. Tectonic evolution of the Gulf of Mexico, Caribbean and northern South America in the mantle reference frame: An update. In: James, K.H., Lorente, M.A. & Pindell, J.L. (editors), The origin and evolution of the Caribbean Plate. Geological Society of London, Special Publication 328, p. 1–55. https://doi.org/10.1144/SP328.1

Ramírez, D.A., Foster, D.A., Min, K., Montes, C., Cardona, A. & Sadove, G. 2016. Exhumation of the Panama basement complex and basins: Implications for the closure of the Central American Seaway. Geochemistry, Geophysics, Geosystems, 17(5): 1758–1777. https://doi.org/10.1002/2016GC006289

Recchi, G. & Metti, A. 1975. Lámina 17. In: Molo, J.C. (editor), Atlas Nacional de Panamá. Instituto Geográfico Nacional Tommy Guardia, p. 71. Panamá.

Reclus, A. & De Vaisseau, L. 1880. Explorations aux Isthmes de Panama et de Darien. Le Tour du Monde, 21(1): 321–400.

Reed, D.L. & Silver, E.A. 1995. Sediment dispersal and accretionary growth of the North Panama Deformed Belt. In: Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 213–223. Boulder, USA. https://doi.org/10.1130/SPE295-p213

Reed, D.L., Silver, E.A., Tagudin, J.E., Shipley, T.H. & Vrolijk, P. 1990. Relations between mud volcanoes, thrust deformation, slope sedimentation, and gas hydrate, offshore north Panama. Marine and Petroleum Geology, 7(1): 44–54. https://doi.org/10.1016/0264-8172(90)90055-L

Rincón, A.F., Bloch, J.I., Macfadden, B.J. & Jaramillo, C. 2015. New early Miocene protoceratids (Mammalia, Artiodactyla) from Panama. Journal of Vertebrate Paleontology, 35(5): 1–22. https://doi.org/10.1080/02724634.2015.970688

Rodríguez, G. & Sierra, M.I. 2010. Las Sedimentitas de Tripogadí y las Brechas de Triganá: Un registro de volcanismo de arco, corrientes de turbidez y levantamiento rápido Eoceno en el noroccidente de Sur América. Geología Colombiana, 35: 74–86.

Rodríguez–Parra, L.A., Gaitán, C., Montes, C., Bayona, G. & Rapalini, A. 2017. Arc–seamount collision: Driver for vertical–axis rotations in Azuero, Panama. Studia Geophysica et Geodaetica, 61(2): 199–218. https://doi.org/10.1007/s11200-016-1173-1

Röhl, U. & Abrams, L.J. 2000. High–resolution, downhole, and nondestructive core measurements from Sites 999 and 1001 in the Caribbean Sea: Application to the late Paleocene Thermal Maximum. In: Leckie, R.M., Sigurdsson, H., Acton, G.D. & Draper, G. (editors), Proceedings of the Ocean Drilling Program, Scientific Results 165, p. 191–203. College Station, USA. https://doi.org/10.2973/odp.proc.sr.165.009.2000

Rooney, T.O., Franceschi, P. & Hall, C.M. 2011. Water–saturated magmas in the Panama Canal region: A precursor to adakite–like magma generation? Contributions to Mineralogy and Petrology, 161(3): 373–388. https://doi.org/10.1007/s00410-010-0537-8

Rooney, T.O., Morell, K.D., Hidalgo, P. & Franceschi, P. 2015. Magmatic consequences of the transition from orthogonal to oblique subduction in Panama. Geochemistry, Geophysics, Geosystems 16(12): 4178–4208. https://doi.org/10.1002/2015GC006150

Sheffey, J.P., Woodbury, H.G., Noble, C.C., Groves, R.H. & McGregor, E.W. 1969. Summary of geology and rock physical properties. U.S. Army Engineer Nuclear Cratering Group & Lawrence Radiation Laboratory, University of California, 1027 p. California.

Shelton, B.J. 1952. Geology and petroleum prospects of Darien, southeastern Panama. Master thesis, Oregon State College, 62 p. Corvallis, USA.

Silver, E.A., Reed, D.L., Tagudin, J.E. & Heil, D.J. 1990. Implications of the north and south Panama thrust belts for the origin of the Panama orocline. Tectonics, 9(2): 261–281. https://doi.org/10.1029/TC009i002p00261

Silver, E.A., Galewsky, J. & McIntosh, K.D. 1995. Variation in structure, style, and driving mechanism of adjoining segments of the North Panama Deformed Belt. In: Mann, P. (editor), Geologic and tectonic development of the Caribbean Plate boundary in southern Central America. Geological Society of America, Special Paper 295, p. 225–233. Boulder, USA. https://doi.org/10.1130/SPE295-p225

Stephan, J.F., Blanchet, R. & Mercier de Lepinay, B. 1986. Northern and southern Caribbean festoons (Panama, Colombia–Venezuela and Hispaniola–Puerto Rico), interpreted as pseudosubductions induced by the east–west shortening of the pericaribbean continental frame. In: Wezel, F.C. (editor), The origin of arcs. Elsevier, Developments in Geotectonics 21, p. 401–422. Amsterdam, the Netherlands. https://doi.org/10.1016/B978-0-444-42688-8.50022-9

Stewart, R.H., Stewart, J.L. (compilers) & Woodring, W.P. (collaborator). 1980. Geologic map of the Panama Canal and vicinity, Republic of Panama. Scale 1:100 000. U.S. Geological Survey, Miscellaneous Investigations Series, Map I–1232, 1 sheet. https://doi.org/10.3133/i1232

Tavelli, J.A. 1947. Geologic Explorations, Caledonia Bay, Route 17, Geology and Topography. Panama Canal Company, 21 p.

Terry, R.A. 1956. A geological reconnaissance of Panama. California Academy of Science, Occasional Papers 23–25, 91 p. San Francisco, USA.

Tournon, J., Triboulet, C. & Azema, J. 1989. Amphibolites from Panama: Anticlockwise P–T paths from a pre–Upper Cretaceous metamorphic basement in Isthmian Central America. Journal of Metamorphic Geology, 7(5): 539–546. https://doi.org/10.1111/j.1525-1314.1989.tb00616.x

Tripati, A. & Zachos, J.C. 2002. Late Eocene tropical sea surface temperatures: A perspective from Panama. Paleoceanography and Paleoclimatology, 17(3): 4-1–4-14. https://doi.org/10.1029/2000PA000605

United Nations Development Program. 1972. Reconnaissance geochemical survey of Bocas del Toro, Maje, Pirre and San Blas–Darien. United Nations, Technical report 2, 79 p. Panamá.

U.S. Geological Survey. 2010. Global multi–resolution terrain elevation data 2010 (GMTED2010). https://topotools.cr.usgs.gov/gmted_viewer/viewer.htm (consulted in July 2018).

U.S. Geological Survey. 2017. Search earthquake catalog. https://earthquake.usgs.gov/earthquakes/search/ (consulted in November 2018).

Vannucchi, P., Fisher, D.M., Bier, S. & Gardner, T.W. 2006. From seamount accretion to tectonic erosion: Formation of Osa Mélange and the effects of Cocos Ridge subduction in southern Costa Rica. Tectonics, 25(2): 1–19. https://doi.org/10.1029/2005TC001855

Vannucchi, P., Fisher, D.M. & Gardner, T.W. 2007. Reply to comment by David M. Buchs & Peter O. Baumgartner on "From seamount accretion to tectonic erosion: Formation of Osa Mélange and the effects of the Cocos Ridge subduction in southern Costa Rica". Tectonics, 26(3): 1–2. https://doi.org/10.1029/2007TC002129

Vannucchi, P., Sak, P.B., Morgan, J.P., Ohkushi, K. & Ujiie, K. 2013. Rapid pulses of uplift, subsidence, and subduction erosion offshore Central America: Implications for building the rock record of convergent margins. Geology, 41(9): 995–998. https://doi.org/10.1130/G34355.1

Vaughan, T.W. 1946. Initiation of geological investigations in the Panama Canal zone. Science, 104(2696): 209. https://doi.org/10.1126/science.104.2696.209

Verbrugghe, L. 1879. A travers l'isthme de Panama. Tracé interocéanique de L.N.B. Wyse et A. Reclus, Imprimerie de A. Quantin, 476 p. Paris.

Villagómez, D., Spikings, R., Magna, T., Kammer, A., Winkler, W. & Beltrán, A. 2011. Geochronology, geochemistry and tectonic evolution of the Western and Central Cordilleras of Colombia. Lithos, 125(3–4): 875–896. https://doi.org/10.1016/j.lithos.2011.05.003

Wegner, W., Wörner, G., Harmon, R.S. & Jicha, B.R. 2011. Magmatic history and evolution of the Central American land bridge in Panama since the Cretaceous times. Geological Society of America Bulletin, 123(3–4): 703–724. https://doi.org/10.1130/B30109.1

Westbrook, G.K., compiler. 1990. Gravity anomaly map of the Caribbean region. Scale 1:5 000 000. In: Dengo, G. & Case, J.E. (editors), The Caribbean region. Geological Society of America, plate 7, p. 537. Boulder, USA. https://doi.org/10.1130/DNAG-GNA-H

Whattam, S.A. & Stern, R.J. 2015. Late Cretaceous plume–induced subduction initiation along the southern margin of the Caribbean and NW South America: The first documented example with implications for the onset of plate tectonics. Gondwana Research, 27(1): 38–63. https://doi.org/10.1016/j.gr.2014.07.011

Whattam, S.A., Montes, C., McFadden, R.R., Cardona, A., Ramírez, D. & Valencia, V. 2012. Age and origin of earliest adakitic–like magmatism in Panama: Implications for the tectonic evolution of the Panamanian magmatic arc system. Lithos, 142–143: 226–244. https://doi.org/10.1016/j.lithos.2012.02.017

Wolters, B. 1986. Seismicity and tectonics of southern Central America and adjacent regions with special attention to the surroundings of Panama. Tectonophysics, 128(1–2): 21–46. https://doi.org/10.1016/0040-1951(86)90306-9

Woodring, W.P. 1957. Geology and description of tertiary mollusks (gastropods: Trochidae to Turritellidae). Geology and paleontology of Canal zone and adjoining parts of Panama. U.S. Geological Survey, Professional Paper 306–A, 135 p.

Woodring, W.P. 1973. Description of tertiary mollusks (additions to gastropods, scaphopods, pelecypods: Nuculidae to Malleidae). Geology and paleontology of Canal zone and adjoining parts of Panama. U.S. Geological Survey, Professional Paper 306–E: 453–531.

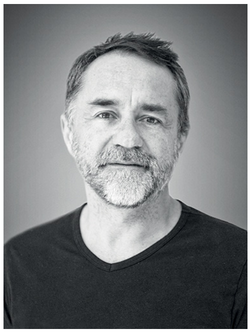
Woodring, W.P. & Thompson, T.F. 1949. Tertiary formations of Panama Canal zone and adjoining parts of Panama. American Association of Petroleum Geologists Bulletin, 33(2): 223–247.

Wörner, G., Harmon, R.S., Hartmann, G. & Simon, K. 2005. Igneous geology and geochemistry of the upper río Chagres Basin. In: Harmon, R.S. (editor), The río Chagres, Panama: A multidisciplinary profile of a tropical Watershed. Springer, p. 65–82. Dordrecht, the Netherlands.

Explanation of Acronyms, Abbreviations, and Symbols:

CLIP	Caribbean Large Igneous Plateau
LIP	Large Igneous Province
ODP	Ocean Drilling Program
STRI	Smithsonian Tropical Research Institute
UNDP	United Nations Development Program

Authors' Biographical Notes



Camilo MONTES is currently professor of geology at the Universidad del Norte in Barranquilla. He obtained his undergraduate degree in geology from Universidad Nacional de Colombia, and his MS and PhD at the University of Tennessee. Camilo's main research interests are the late Mesozoic – Cenozoic tectonic evolution of the Caribbean and the northwestern corner of South America, and its implications in climate and biotas. Basic geologic mapping is at the basis of all of his research and educational efforts.

Natalia HOYOS obtained her undergraduate degree in geology from Universidad EAFIT, Colombia (1996) and completed her master´s degree in environmental science at the University of New Haven, U.S. (2000), and PhD in geography at the University of Florida, U.S. (2004). She moved to Panama in 2010 to work as a postdoc at the Smithsonian Tropical Research Institute, and returned to Colombia in 2011 to work with Corporación Geológica ARES. Since 2014, Natalia HOYOS is affiliated with Universidad del Norte (Colombia). Her research interests include the use of remote sensing data and geographic analyses for geologic mapping and the study of human–environment interactions.

Photography Natalia HOYOS: Lina Pérez–Angel