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- Restored Cretaceous-Paleocene motion paths of northeastern and eastern fragments of the Manihiki Plateau
- Collision and partly obduction of northeastern fragment of the Manihiki Plateau with the South American craton during the Paleocene
- Subduction of eastern fragment of the Manihiki Plateau at the Gondwana convergent margin of West Antarctica during the Early Cretaceous

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Collision of Manihiki Plateau fragments to accretional margins of northern Andes and Antarctic Peninsula

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Abstract The Manihiki Plateau, a large igneous province (LIP), was emplaced in the Early Cretaceous as a single plateau together with the Ontong Java Plateau and the Hikurangi Plateau. Additional to the present Manihiki Plateau, fragments to its northeast and east have been formed. Plate kinematic reconstructions suggest the capturing of these fragments by the Farallon Plate and the Phoenix Plate, respectively. By tracing these fragments, we report a Paleocene collision of the northeastern Manihiki Plateau fragment with the northern South American craton. The northern Andes exhibit multiple terranes of LIP origin. We infer, based on geophysical, petrological, and geochemical data, that the Piñón formation consists of crustal units of the former Manihiki Plateau. An Early Cretaceous collision of the eastern Manihiki Plateau fragment is reconstructed for West Antarctica. The subduction of this fragment in the Palmer Land region can be associated with the so-called Palmer Land Event and a flattening of the subduction slab. By reconstructing the dispersal of the fragments of the Manihiki Plateau, we provide a deeper insight in the possible subduction scenarios and buildup of the accretional margins of the northern Andes and the Antarctic Peninsula.

1. Introduction: Large Igneous Provinces in the Plate Circuit

Oceanic plateaus and oceanic large igneous provinces (LIPs) [*Coffin and Eldholm*, 1994; *Ridley and Richards*, 2010; *Bryan and Ferrari*, 2013] (Figure 1) play an important role in the plate tectonic circuit, since they alter or radically transform the behavior of the subducting oceanic plate within the subduction zone. In the Pacific realm, a wide variety of interaction of oceanic plateaus with subduction zones can be observed, which range from the accretion of terranes (e.g., Malaita terranes) [*Musgrave*, 1990; *Ishikawa et al.*, 2005] and subsequent blocking of the subduction zone at the Ontong Java Plateau [*Coleman and Kroenke*, 1981; *Petterson et al.*, 1999; *Mann and Taira*, 2004; *Miura et al.*, 2004; *Taira et al.*, 2004] to the complete subduction of the oceanic plateau below the Americas [*Gutscher et al.*, 1999; *Liu et al.*, 2010] (Figure 1). Furthermore, the collision of LIPs with continental margins has been associated with the evolution of mountain ranges such as the Laramide orogeny [*Liu et al.*, 2010] or the Southern Alps of New Zealand [*Reyners et al.*, 2011]. The Pacific subduction margins of North and South America illustrate the interaction with oceanic plateaus in various stages and time frames. Oceanic plateaus, which subducted beneath North America, include, for example, the conjugate of the Shatsky Rise in the middle Cretaceous [*Liu et al.*, 2010] and the conjugate of the Hess Rise in the Late Cretaceous (Figure 1) [*Liu et al.*, 2010]. The Inca Plateau influenced the subduction at the South American trench during the Miocene [*Gutscher et al.*, 1999] and the Nazca Ridge within recent times (Figure 1). Unlike the LIPs mentioned above, fragments of the Manihiki Plateau have not been related to any onshore geology.

Overthickened oceanic crust is often invoked as the origin of accreted mafic terranes [e.g., *Tejada et al.*, 1996; *Mamberti et al.*, 2003]. The main challenges for connecting onshore terranes of oceanic plateau origin with their marine conjugates are the great distance between the emplacement areas of the LIP and today's location, considering uncertainties of plate tectonic reconstructions, and the alteration and overprint that these terranes experienced during the accretionary process.

Previous studies of the breakup of the "Super"-LIP Ontong Java Nui and the subsequent development of the Pacific Ocean focused only on the main units and subplateaus of Ontong Java Nui (Ontong Java, Hikurangi, and Manihiki), omitting smaller fragments (Figure 2). The northeastern and eastern fragments of the Manihiki Plateau were not integrated in the Pacific plate circuit. We trace the plate tectonic motion of these fragments and connect highly complex geology on the continental margins with the presence of LIP fragments. This study gives an insight into the role of LIP fragments and terranes in the plate circuit of the Pacific Ocean as well as possible information on deeper crustal layers of the oceanic plateau.

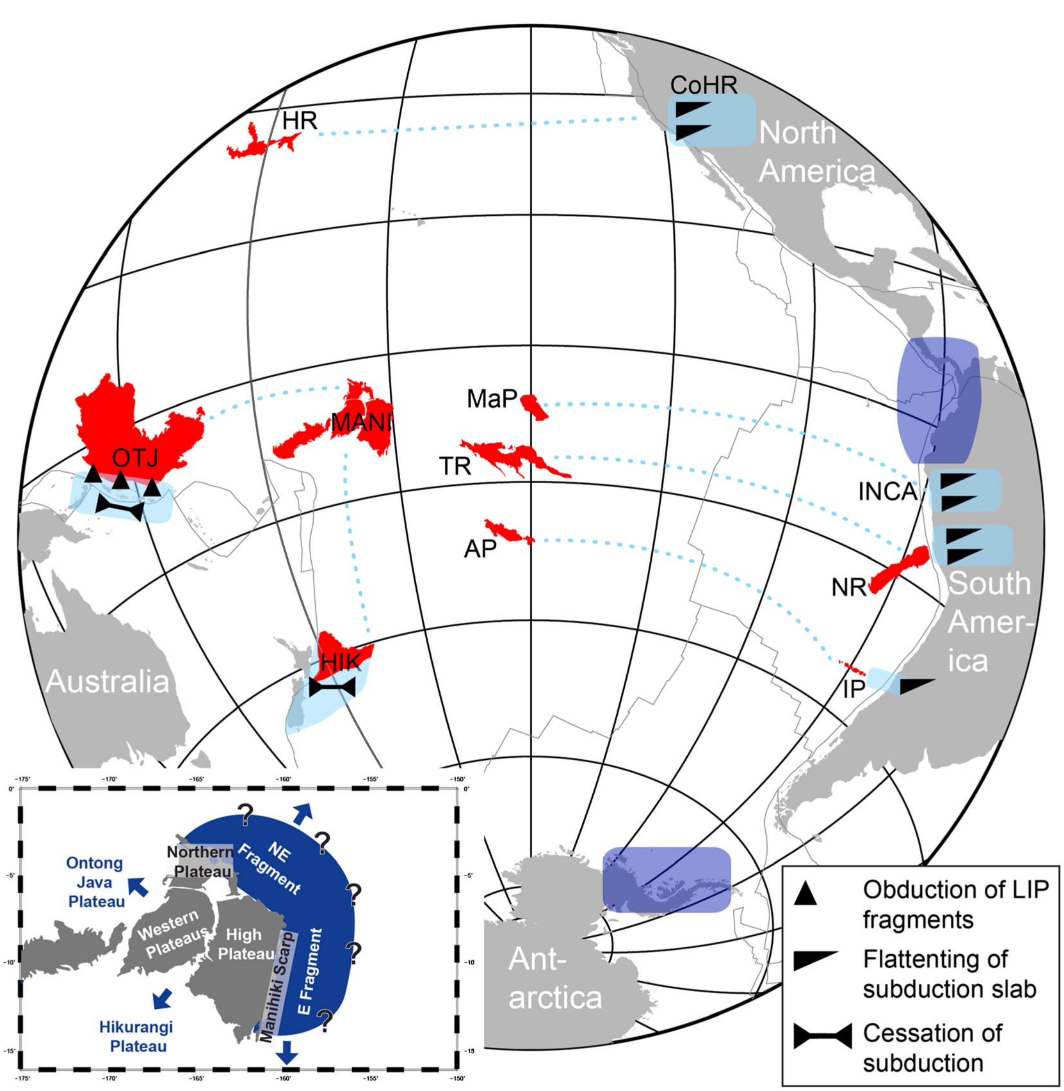


Figure 1. Overview of the some LIPs of the Pacific Ocean and their subducted remnants and accretion/subduction mode. Present LIPs—Hess Rise (HR), Ontong Java Plateau (OTJ), Manihiki Plateau (MANI), Marquesas Plateau (MaP), Tuamoto Ridge (TR), Austral Plateau (AP), Nazca Ridge (NR), Iquique Plateau (IP), and Hikurangi Plateau (HIK)—are marked in red, areas of subduction are in light blue, which includes the conjugate of the Hess Rise (CoHR) and the Inca Plateau (INCA). The dark blue areas indicate the possible areas of subducted former northeastern (NE) and eastern (E) fragments of the Manihiki Plateau at the northwestern South America and West Antarctica margins. Previously identified joined emplaced LIPs are connected by dashed light blue lines. Present plate boundaries are marked in gray. The insert map shows the different subprovinces of the Manihiki Plateau and the position of the parts of Ontong Java Nui encircling the Manihiki Plateau.

2. Geological and Plate Tectonic Setting: Ontong Java Nui and Its Remnants

Taylor [2006] proposed the emplacement of three major LIPs of the western Pacific region (Figure 1)—Ontong Java Plateau, Manihiki Plateau, and Hikurangi Plateau—as a single "Super"-LIP, named Ontong Java Nui. Shortly after its emplacement within the Early Cretaceous [*Hoernle et al.*, 2010], this LIP broke apart (Figure 2) [*Viso et al.*, 2005; *Taylor*, 2006; *Davy et al.*, 2008; *Hochmuth et al.*, 2015]. The Hikurangi Plateau collided with and partially subducted below the Chatham Rise (~100 Ma) [*Davy et al.*, 2008; *Reyners et al.*, 2011] (Figure 1). The Ontong Java Plateau blocks the subduction of the Pacific Plate below the Australian Plate and caused a change in subduction direction [e.g., *Mann and Taira*, 2004]. Prior to the cessation of subduction, multiple small fragments were obducted as part of the island arc such as the Malaita terranes at the Solomon Trench [*Musgrave*, 2013].

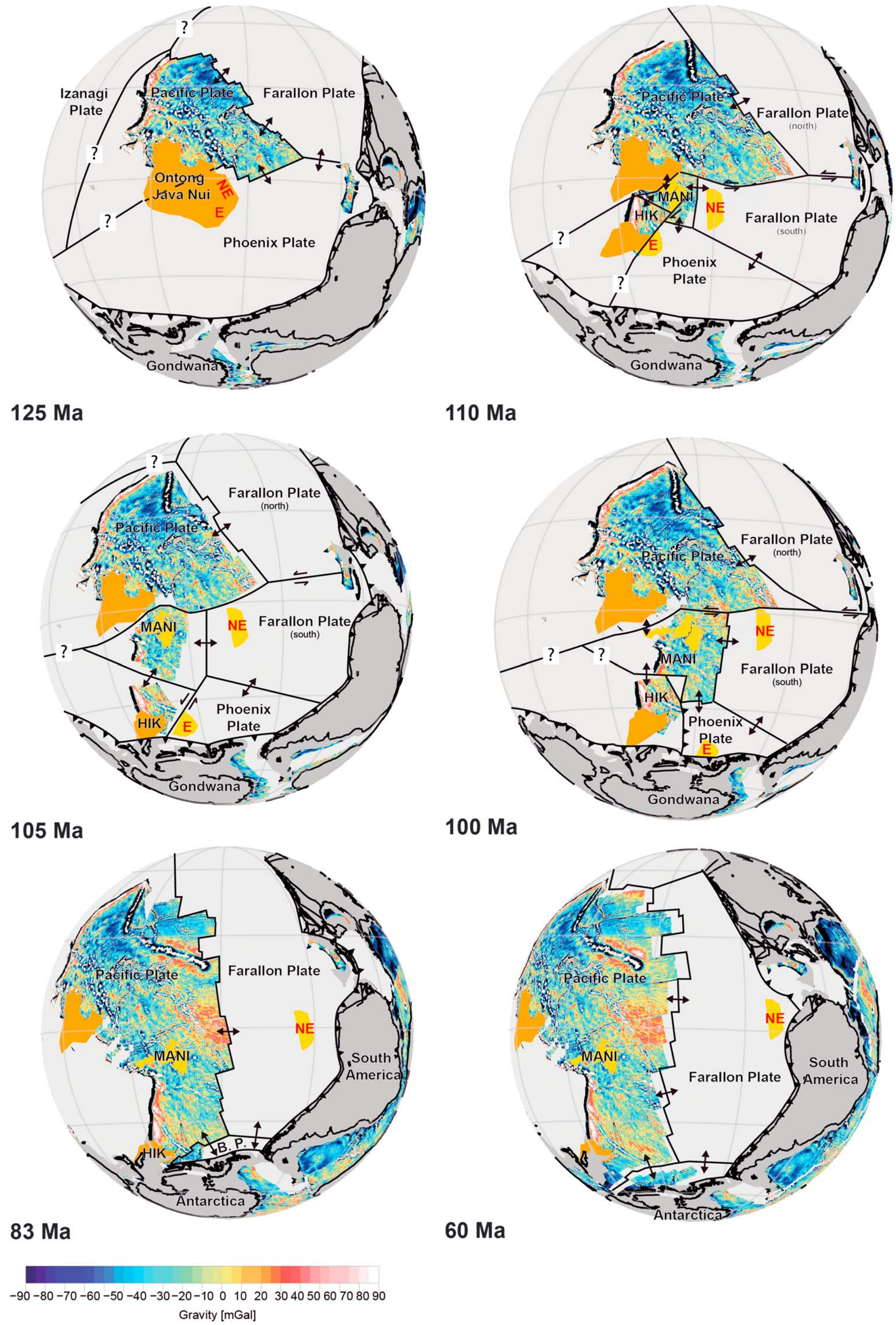


Figure 2. Plate kinematic reconstruction of the Pacific Ocean from 125 to 60 Ma; The northeastern (NE) and eastern (E) fragment of the Manihiki Plateau (MANI) are shown in yellow. Other parts of Ontong Java Nui are shown in orange such as the Hikurangi Plateau (HIK). Existing seafloor is shown with the gravity anomaly map by *Sandwell et al*. [2014].

In addition to the three largest parts of Ontong Java Nui, smaller fragments to the northeast and the east of the Manihiki Plateau have been proposed (Figure 1) [*Viso et al*., 2005; *Hochmuth et al*., 2015; *Pietsch and Uenzelmann-Neben*, 2015]. The current locations of the northeastern and eastern fragments of the Manihiki Plateau can be derived by carefully reconstructing their plate kinematic paths across the Pacific region and by geophysical and geological observations from their predicted areas of subduction or collision.

3. Results of Plate Kinematic Modeling of the Pacific Ocean (Cretaceous-Paleocene)

3.1. Northeastern Manihiki Plateau Fragment

A former northeastern continuation of the Manihiki Plateau beyond its present extent has been proposed from bathymetric and gravimetric observations by multiple authors [*Larson et al*., 2002; *Viso et al*., 2005]. Its size is constrained by the presence of the Clipperton Fracture Zone (FZ) to the north, which was presumably active during the breakup of Ontong Java Nui [*Taylor*, 2006]. The northeastern fragment was separated from the Manihiki Plateau by a fast clockwise rotation and captured by the Farallon Plate at 110 Ma (Figure 2) [*Viso et al*., 2005; *Hochmuth et al*., 2015]. Its motion across the Pacific is constrained by the spreading rate between the Pacific Plate and Farallon Plate and the rotation parameters established by *Seton et al*. [2012] (Figures 2 and 3a) for times after 110 Ma (Table 1). The Clipperton FZ, the Galapagos FZ, and the Marquesas FZ, which are still traceable on today's Pacific Plate, mimic the evolution of the spreading between the Pacific and the Farallon Plate and can therefore be used to track the motion of the fragment (Figure 3a). In accordance with these boundary conditions, the northeastern fragment of the Manihiki Plateau was transported toward the South American craton, where it arrived at the subduction trench at today's northern Andes of Ecuador and Colombia during the late Paleocene/early Eocene (~60–55 Ma) (Figures 2 and 3).

3.2. Eastern Manihiki Plateau Fragment

Seismic reflection and refraction seismic data show strong evidence of the former presence of an eastern fragment of the Manihiki Plateau [e.g., *Pietsch and Uenzelmann-Neben*, 2015]. The breakup between the High Plateau of the Manihiki Plateau and the eastern fragment involved a large shear zone at the Manihiki Scarp (Figure 1) and short-lived spreading centers [*Larson et al*., 2002; *Viso et al*., 2005; *Hochmuth et al*., 2015]. The plate boundary between the Manihiki Plateau and the Phoenix Plate is defined by the Tongareva triple junction trace (Figures 2 and 4a). We can, therefore, limit the possible collision area between the eastern fragment and the eastern Gondwana margin of present West Antarctica to the area west of the Tongareva triple junction and east of the Chatham Rise where the Hikurangi Plateau collided and partly subducted (Figures 2 and 4). The entire motion between the eastern fragment and the Manihiki Plateau took place during the Cretaceous Normal Superchron (CNS). As recognizable magnetic spreading anomalies are absent for this time period, we reconstruct the motion using seafloor fabric such as fracture zones [*Sandwell et al*., 2014] (Figure 4a). We also used major tectonic events as constraints, such as the collision of the Hikurangi Plateau with the Chatham Rise between 105 and 100 Ma [*Davy*, 2014], ceasing the southward subduction at the Chatham Rise margin. Plate kinematic reconstructions predict a similar time frame for the interaction of the eastern fragment with the subduction zone along the present Bellingshausen Sea margin of West Antarctica. Collision and subduction must have occurred within the Charcot Island, Ellsworth Land, and Palmer Land regions (Figures 2 and 4b).

4. Discussion

4.1. Plate Tectonic Reconstruction and Manihiki Plateau Fragments

4.1.1. On the Plate Kinematic Reconstruction

Plate tectonic reconstructions of the Pacific Ocean for the time frame of the Cretaceous to the Paleocene suffer from large uncertainties, because of lacking corresponding magnetic spreading anomalies, especially during the CNS, and seafloor fabric of the Farallon Plate and the Phoenix Plate due to subduction. We used a combination of two plate kinematic reconstructions for the motion paths of the fragments of the Manihiki Plateau. For the initial breakup (118–105 Ma), we used the rotation poles of *Hochmuth et al*. [2015]. The rotation poles are mainly based on the seafloor fabric of the region, but also rare dated events. For the subsequent development of the Pacific Ocean, the well-established rotation poles by *Seton et al*. [2012] were used. A complete list of the rotation poles of both fragments is provided as Table 1. By mirroring fracture zones to the subducted part of the Farallon plate, the motion path of the fragment can be tested. This is based on the assumption of a constant and symmetric spreading behavior as it can be observed at the East Pacific Rise, the descendent of the Farallon-Pacific spreading [*Seton et al*., 2012]. The motion of the Phoenix Plate subducting below the Gondwana margin can only be constrained by undated seafloor fabric and the estimated age of the Tongareva Triple Junction [*Davy*, 2014]. This constrains an uncertainty for the plate tectonic model of about ±3 Ma.

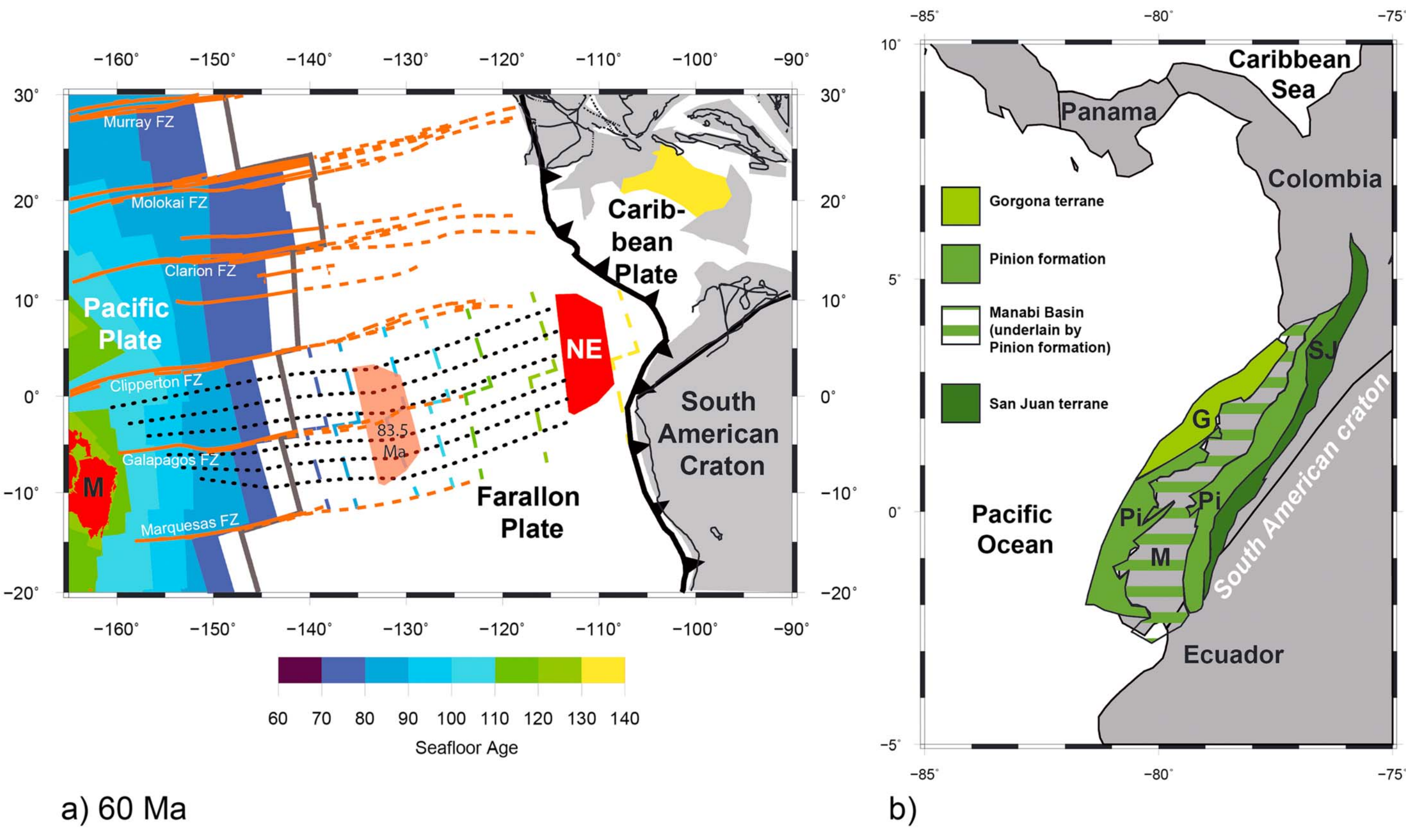


Figure 3. Paleocene collision of the northeastern fragment of the Manihiki Plateau (NE) with the South American craton: (a) Plate tectonic setting at 60 Ma with the motion path of the fragment (dotted black lines) and Pacific-Farallon spreading center (dark gray line) and fracture zones [*Matthews et al*., 2011; *Sandwell et al*., 2014] of the Pacific Plate are depicted in orange; subducted conjugate fracture zones on the Farallon Plate are shown in the dashed orange lines. The motion paths that display the motion between the High Plateau of the Manihiki Plateau and the northeastern fragment have been calculated with GPlates with a starting age of 120 Ma and a time interval of 5 Ma. All motion is depicted in regard to a fixed Pacific Plate. The dashed colored lines are the extrapolated seafloor isochrones of the subducted Farallon Plate. The shaded fragment marks the position of the northeastern fragment of the Manihiki Plateau at 83.5 Ma. Today's continental outlines are shown in black. (b) Simplified geological map of the northern Andes after *Mamberti et al*. [2003] with marked Cretaceous LIP remnants (Gorgona terrane (G), Pinion formation (P), Manabi Basin (M), and San Juan terrane (SJ)).

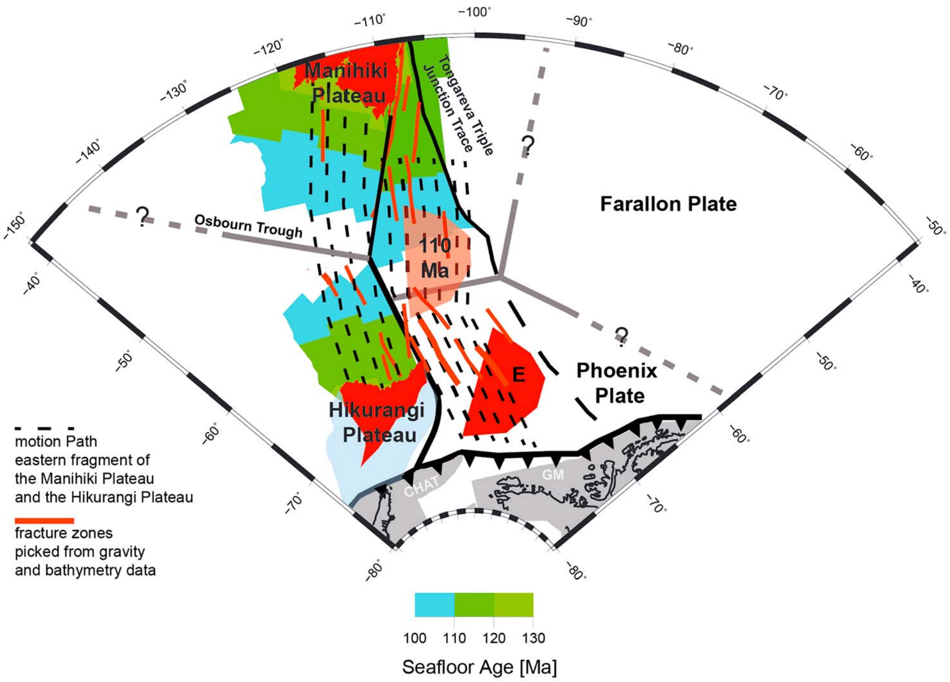
4.1.2. Size of the Manihiki Fragments

A second factor, which alters the estimated collision times and behavior of the subduction zones, is given by the size of the fragments. To estimate the approximate emplacement area of the fragments of the Manihiki Plateau, we considered the extent of the traceable subducted Hikurangi Plateau to be 800 km below the South Island of New Zealand (Figure 1) [*Reyners et al*., 2011] and the undisturbed transition of LIP crust to oceanic crust at the northern Ontong Java Plateau as analogues [*Mochizuki et al*., 2005]. The seafloor fabric gives additional constraints. The maximum extension of the northeastern fragment is marked by the presence of the Clipperton FZ, which was active during the emplacement and breakup of Ontong Java Nui. This allows for an up to 500 km northward continuation of the overthickened LIP crust with decreasing crustal thickness toward the plateaus edges. The size of the eastern fragment can be assumed to be at least of similar size. Intrabasement reflections of the eastern part of the High Plateau of the Manihiki Plateau show no decrease in the initial crustal thickness toward the eastern fragment [*Pietsch and Uenzelmann-Neben*, 2015].

Table 1. Relative Stage Rotation Poles Used in This Study for the Fragments of the Manihiki Plateau (Fixed to the Pacific Plate)

	t_1 (Ma)	t_2 (Ma)	Latitude	Longitude	ω (deg)	Reference
Northeastern fragment (individual plate)	118	110	−70.61	−18.19	129.52	*Hochmuth et al.* [2015]
Northeastern fragment (fixed to Farallon Plate)	110	83.5	78.35	113.06	−127.24°	*Seton et al.* [2012]
	83.5	67.7	81.35	119.76	−93.37	*Seton et al.* [2012]
	67.7	55.9	82.93	125.82	−80.17	*Seton et al.* [2012]
	55.9	47.9	84.28	135.44	−71.47	*Seton et al.* [2012]
Eastern fragment (individual plate)	118	105	−18.07	−24.77	49.50	*Hochmuth et al.* [2015]

a) 105 Ma



b)

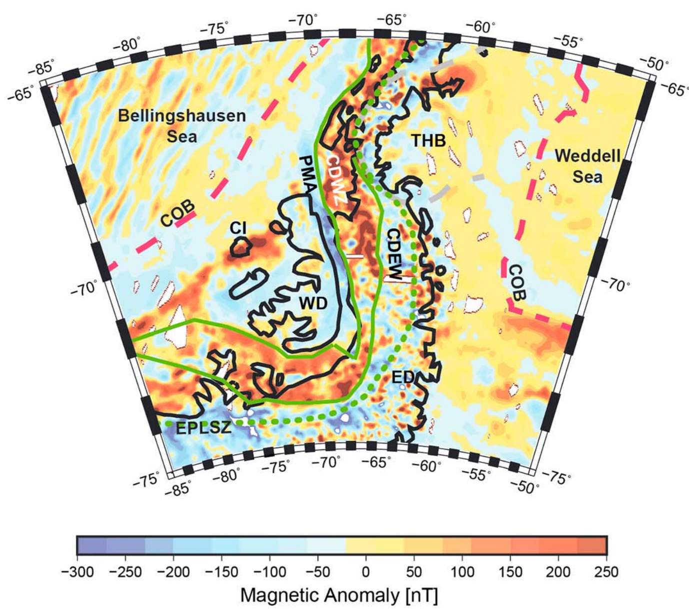


Figure 4. Mid-Cretaceous collision of the eastern fragment of the Manihiki Plateau (E) with the Gondwana margin (GM) of West Antarctica: (a) Plate tectonic setting at 105 Ma; dotted black lines mark the motion path of the eastern fragment along mapped fracture zones (orange). The motions paths have been calculated with GPlates with a starting age of 120 Ma and a time interval of 5 Ma. The motions paths show the motion between the High Plateau of the Manihiki Plateau and the Hikurangi Plateau and eastern fragment of the Manihiki Plateau, respectively. All motion is depicted in regard to a fixed Pacific Plate. Dark gray lines indicate spreading centers. The shaded fragment marks the position of the eastern fragment of the Manihiki Plateau at 110 Ma. Today's continental outlines are marked in black. (b) Magnetic anomaly map after *Maus et al*. [2009]. The green lines indicate magnetic lineaments, which separate the different terranes of the Antarctic Peninsula (Charcot Island CI, Western Domain WD, Central Domain Western Zone CDWZ, Central Domain Eastern Domain CDEW, Eastern Domain ED, and Target Hill Block THB) [*Ferraccioli et al*., 2006; *Vaughan et al*., 2012]. The eastern Palmer Land Shear Zone (EPLSZ) marks the boundary between CDEW and ED. Pink dashed lines indicate the position of the continent-ocean boundary.

If a steady decrease of basaltic volume similar to the northern Ontong Java Plateau is assumed [*Mochizuki et al.*, 2005], the eastern fragment extends at least 500 km to the east with decreasing crustal thickness. The size estimation of the fragments of the Manihiki Plateau is comparable to that of other parts of Ontong Java Nui such as the Hikurangi Plateau.

4.1.3. How can the Subduction of an Oceanic Plateau be Recognized?

By interfering time frames for the collision of a LIP fragment with a subduction zone, it is crucial to consider that the more of the LIP crust is subducted, the thicker the LIP crust becomes [*Tetreault and Buiter*, 2012]. Therefore, the initial contact of thin LIP crust with the subduction zone (soft-docking) leaves little to no observable traces. The subducting slab flattens with increasing crustal thickness, which eventually leads to partial obduction of terranes and the cessation of the subduction at the trench (hard docking). Thus, considerable amounts of LIP crust can be subducted before significant alteration of subduction mechanisms occurs (Figure 1).

The age of the accreted overthickened LIP crust plays, along with the igneous volume and the crustal thickness, a crucial role for the interaction with the subduction zone [*Cloos*, 1993]. Whereas a young LIP resists subduction due to its buoyancy, older LIPs are less buoyant and are rather prone to be subducted. The fragments of the Manihiki Plateau have a maximum crustal thickness of 20 km and arrive 20 Myr or 60 Myr after their emplacement at the subduction zone and, therefore, should subduct [*Cloos*, 1993]. Geodynamic modeling revealed that the subduction mode itself is also closely associated with the buildup and the physical properties of the oceanic plateau. Whereas a 20 km thick oceanic plateau of homogenous brittle strength subducts easily without any accreted crustal units, oceanic plateaus with a weak basal layer experience accretion [*Tetreault and Buiter*, 2012].

A possible analog for this interaction is the subduction of the conjugate of the Shatsky Rise below Southern California, which led to a flattening of the subduction slab and triggered, after the basalt-eclogite transformation, the buildup of the Laramide orogen [*Liu et al.*, 2010]. Another similar scenario can be observed with the Inca Plateau in the Peruvian Andes (Figure 1), where a flat subduction slab is present [*Gutscher et al.*, 1999]. The counterpart of the Inca Plateau, the Marquesas Plateau, has a crustal thickness of 17 km [*Caress et al.*, 1995], similar to that of the Manihiki Plateau. The Caribbean large igneous province (CLIP) with a crustal thickness of 16 km currently subducts at a very low angle below South America [*Mann*, 2007; *Bernal-Olaya et al.*, 2016]. An additional indicator of a flattened slab is also the presence of adakites [*Gutscher et al.*, 2000].

But can flattening of the subduction slab also be observed in the northern Andes or the Antarctic Peninsula? Seismological studies of the Colombian northern Andes indicate a flattening of the slab and east-west striking slab tear [*Vargas and Mann*, 2013], which is attributed to the subduction of thicker oceanic crust. Between 140 and 40 Ma, the northern Andes were an amagmatic arc [*Jaillard et al.*, 2009], which also included, along with the subduction motion, a north-south shearing component. Therefore, subduction mechanisms at the northern Andes Trench differed from the magmatic arc of the central Andes. This leads to the presence of oceanic plateau slivers within the northern Andes. At the Antarctic Peninsula, the presence of a flattened subduction slab cannot be used as an indicator for LIP subduction, as the plate tectonic setting changed dramatically from an active subduction to a passive margin since the Cretaceous.

4.2. Oceanic Terranes of the Northern Andes

4.2.1. Terrane Aggregation at the Northern Andes Margin

The northern Andes and the adjacent Caribbean regions, the projected area of collision of the northeastern Manihiki Plateau fragment during the Paleocene, are a mosaic of terranes of oceanic and continental origin [see, e.g., *Maloney et al.*, 2013; *Boschman et al.*, 2014, and references therein] (Figure 3b). The CLIP was emplaced at about 90 Ma and can be attributed to the Galapagos hot spot [e.g., *Hill*, 1993]. After the main emplacement phase of the CLIP, very young and therefore still buoyant LIP crust collided with the South American craton, resulting in terrane aggregation [*Mamberti et al.*, 2003; *Kerr and Tarney*, 2005; *Jaillard et al.*, 2009] and the reversal of the subduction trench [*Mann*, 2007]. Other oceanic LIP terranes of the region are older than the CLIP event and therefore not related to the Galapagos hot spot. The San Juan terranes and the Piñón formation in Ecuador and Colombia [*Reynaud et al.*, 1999; *Mamberti et al.*, 2003; *Jaillard et al.*, 2009] and the Gorgona Plateau are all oceanic features, which have been dated to be of Cretaceous origin (Figure 3b) [*Kerr and Tarney*, 2005] and have been accreted to the South American craton between the Late Cretaceous and the Paleocene.

Table 2. Comparison of the Terranes of the Northern Andes With the Manihiki Plateau[a]

	Manihiki Plateau		San Juan Formation	Piñón Formation	Gorgona Plateau
Emplacement age	>125 Ma (initial formation) 125–116 Ma (expansion phase) 100–65 Ma (secondary volcanic phase) [*Hoernle et al.*, 2010; *Timm et al.*, 2011; *Pietsch and Uenzelmann-Neben*, 2015]		123 ± 13 Ma (Sm-Nd isochron) [*Lapierre et al.*, 2000]	123 Ma [*Reynaud et al.*, 1999]	88.9 ± 1.2 Ma [*Kerr and Tarney*, 2005]
Collision age	Late Paleocene/early Eocene (~60–55 Ma) (this paper)		85–80 Ma [*Mamberti et al.*, 2003]	66–56 Ma [*Reynaud et al.*, 1999]	66–56 Ma [*Kerr and Tarney*, 2005]
Calculated paleolatitude at emplacement	25–30°S [*Cockerham and Jarrard*, 1976; *Hochmuth et al.*, 2015]		No data	No data	26–30°S [*Kerr and Tarney*, 2005]
Petrological description	Upper crust [*Hoernle et al.*, 2010; *Timm et al.*, 2011]: initial formation and expansion phase; Thoellitic magmas secondary volcanic phase; alkalitic magmas Lower crustal layers [*Hussong et al.*, 1979; *Hochmuth et al.*, 2015]: peridotite, gabbro		Cumulate peridotites and gabbros intruded by mafic and felsic dykes [*Mamberti et al.*, 2004]	Thoellitic massive and pillowed mafic lava flows, tuffs, and greywackes intruded by shallow level gabbros and dolerites [*Reynaud et al.*, 1999; *Mamberti et al.*, 2004]	Mafic to ultramafic basalt, peridotite, and gabbros [*Revillon et al.*, 2000]
Geochemical data	High TiO_2 group (expansion phase): MgO 3.33–9.1 wt% Nb 1.48–5.34 ppm Ta 0.264–0.51 ppm Th 0.087–0.599 ppm Zr 35.4–95.6 ppm Hf 1.14–2.63 ppm $^{87}Sr/^{86}Sr_i$ 0.704338–0.705657 [*Timm et al.*, 2011]	Low TiO_2 group (expansion phase): MgO 2.30–13.7 wt % Nb 4.83–12.3 ppm Ta 0.444–1.22 ppm Th 0.250–0.916 ppm Zr 5.77–37.7 ppm Hf 0.169–0.666 ppm $^{87}Sr/^{86}Sr_i$ 0.702559–0.703747 [*Timm et al.*, 2011]	MgO 5.77–46.8 wt % Nb 0.02–0.55 ppm Ta 0.01–0.09 ppm Th 0.01–0.4 ppm Zr 0.40–19.6 ppm Hf 0.01–0.63 ppm [*Mamberti et al.*, 2004]	MgO 6.78–15 wt % Nb 0.3–10.75 ppm Ta 0.03–0.67 ppm Th 0.11–1.44 ppm Zr 22–105 ppm Hf 0.57–2.8 ppm $^{87}Sr/^{86}Sr_i$ ratio 0.7032–0.7048 [*Reynaud et al.*, 1999; *Mamberti et al.*, 2004]	MgO 7.53–34.2 wt % Nb 0.57–4 ppm Ta 0.01–0.17 ppm Th 0.01–1.44 ppm Zr 8–42 ppm Hf 0.25–1.26 ppm [*Revillon et al.*, 2000]
Possible fragment of Manihiki Plateau?			Remnant of another Cretaceous oceanic plateau	Comparable to the high TiO_2 group of the Manihiki Plateau	Possibly secondary volcanism of the Manihiki Plateau

[a]Green-shaded cells are in agreement with data from the Manihiki Plateau, orange-shaded cells indicate a possible connection to secondary volcanism of the Manihiki Plateau, and red-shaded cells indicate discrepancies between the terrane and the Manihiki Plateau.

4.2.2. Petrological and Geochemical Data From Northern Andes Terranes

All these terranes are possible candidates to be a remnant of the Manihiki Plateau fragment. Multiple authors report strong petrological and geochemical similarities of the LIPs emplaced during the Ontong Java Nui event with the terranes [*Reynaud et al.*, 1999; *Mamberti et al.*, 2003; *Kerr and Tarney*, 2005; *Jaillard et al.*, 2009]. In the field, the mafic terranes consist of basalts, peridotites, and gabbros (Table 2), which have been partly altered by intrusions. This petrological evidence can be associated to the Manihiki Plateau, where thoellitic magmas have been emplaced during the initial formation and the expansion phase [*Hoernle et al.*, 2010; *Timm et al.*, 2011]. By comparing geochemical data and element concentrations (Table 2), we can relate the Piñón formation to the low TiO_2 group of the Manihiki Plateaus basalts [*Timm et al.*, 2011]. The San Juan terranes show overall different geochemical patterns. The Gorgona Plateau shows some similarities with the geochemical data published from the Manihiki Plateau, but larger discrepancies, for example, in the MgO content are recognizable. The data from Gorgona Island were measured on intrusive rocks rather than extrusive rocks as the data from the Manihiki Plateau. Therefore, to test the comparability of Gorgona Island and the Manihiki Plateau, samples from the deeper crust are needed.

Geochemical analysis also provides emplacement ages for the different oceanic terranes. The San Juan Formation and the Piñón formation are both emplaced at 123 Ma [*Reynaud et al*., 1999; *Lapierre et al*., 2000]. This is well in the approximated age range of the emplacement of Ontong Java Nui (>125 Ma initial formation, 125–116 Ma expansion phase) [*Hoernle et al*., 2010; *Timm et al*., 2011; *Pietsch and Uenzelmann-Neben*, 2015]. The Gorgona Plateau is of a younger age (88.9 Ma) [*Kerr and Tarney*, 2005], which falls in the time of the secondary volcanism observed on the Manihiki Plateau.

Judging from the available petrological data, the San Juan Formation seems to be a rather unlikely candidate to consist of remnants of the eastern fragment of the Manihiki Plateau. The Piñón Formation and the Gorgona Plateau show strong similarities in the petrological and geochemical data to be possibly related to the initial and/or expansion phase or the secondary volcanism, respectively. But how well do these petrological observations taken from the literature correspond to the presented geophysical data, as well as terrane aggregation scenarios within the northern Andes?

4.2.3. Plate kinematic and Geophysical Data of the Northern Andes Terranes

Our plate kinematic reconstruction points to a collision age of the northeastern fragment of the Manihiki Plateau of the late Paleocene/early Eocene (~60–55 Ma) (Figures 2 and 3a). *Reynaud et al*. [1999] postulate two major accretional phases in the northern Andes, during the Campanian and during the Paleocene. The San Juan formation was accreted to the South American craton in the early Campanian during a first major accretion phase [*Reynaud et al*., 1999]. This would call for an unrealistic fast and asymmetric spreading between the Pacific and the Farallon Plate, transporting the northeastern fragment toward the South American craton. Together with the evidence presented above, the San Juan formation has to be formed as part of another oceanic plateau and not as part of Ontong Java Nui.

The Gorgona Plateau, which is obducted at Gorgona Island (Figure 3b) and can possibly be related to the secondary volcanism of the Manihiki Plateau [*Pietsch and Uenzelmann-Neben*, 2015], was accreted to the South American craton in the Paleocene (Figure 3b) [*Kerr and Tarney*, 2005]. Paleolatitude calculations provide an emplacement latitude between 26°S and 30°S [*Kerr and Tarney*, 2005], which is also the emplacement latitude of the northern Manihiki Plateau [*Cockerham and Jarrard*, 1976; *Chandler et al*., 2013; *Hochmuth et al*., 2015]. The age and petrology correlates to the secondary magmatic phases of the Manihiki Plateau, but it is uncertain whether this magmatic phase occurred on all fragments, although both the Hikurangi Plateau and the Ontong Java Plateau experienced multiple phases of magmatic activity [*Inoue et al*., 2008; *Hoernle et al*., 2010].

As seen above the Piñón formation might be the most likely candidate to be originated during the Ontong Java Nui event with an accretion age at the South American craton during the second accretional phase between ~60 and 55 Ma matches the plate kinematic reconstruction. Seismic refraction studies by *Graindorge* [2004] indicate *P* wave velocities of 6.1 km/s to 7.0 km/s for the Piñón formation, which might be comparable to the *P* wave velocities modeled for the middle and lower crust of the Manihiki Plateau. The Piñón formation is severely faulted and overlain by Cenozoic to recent sediments of the Manabí Basin [*Mamberti et al*., 2004]; therefore, the size of the entire terrane is difficult to constrain (Figure 3b). By comparing the estimated size of the northeastern fragment (east-west 500 km and north-south 1200 km) and the oceanic terranes within the northern Andes, we infer that parts of the plateau, probably those of thinner crustal thickness and most of the lowermost crust, have been subducted in this region. The Piñón formation extends approximately 200 km in east-west direction and 900 km in north-south direction, which infers a subduction of 300 km of overthickened crust along the South American trench. The subduction scenario is comparable to that of the Malaita terranes at the Solomon trench [*Musgrave*, 1990; *Petterson et al*., 1999; *Mann and Taira*, 2004].

To summarize, petrological and geophysical data point to the Piñón formation as a remnant of the Manihiki Plateau within the northern Andes. The Gorgona Plateau might be also related to secondary volcanic activity of the Manihiki Plateau, but the presence of multiple volcanic stages on all LIP fragment is still debatable. The San Juan formation has been formed as part of another oceanic plateau within the same time frame as the main activity of the Ontong Java Nui event.

4.3. Palmer Land Event: Initiated by LIP Subduction?

4.3.1. Cretaceous Subduction at the Gondwana Margin

The Pacific realm of West Antarctica has been described as a mosaic of different terranes—similar to the northern Andes—which were accreted since the establishment of the eastern Gondwana subduction margin (Figure 4b) [e.g., *Ferraccioli et al*., 2006]. The identification of possible remnants of the Manihiki Plateau within the northern Andes can be based on a wide variety of fieldwork and published data. The proposed collision site of the eastern fragment (Figure 4a) of the Manihiki Plateau within the Bellingshausen Sea and Palmer Land of present West Antarctica has been only scarcely investigated and mapped due to difficult accessibility of the ice-covered land region. Rock outcrops are few and extremely difficult to access and sample. The identification of possible LIP fragments is based on the few available rock samples and geophysical data from the area (Figure 4b). Subduction was supposedly stopped by the hard docking of the Hikurangi Plateau, part of former Ontong Java Nui, at the Chatham Rise of Zealandia (conjugate of present Marie Byrd Land sector of Antarctica) at 105–100 Ma (Figure 4a) [*Davy et al*., 2008; *Davy*, 2014].

Unlike the Hikurangi Plateau, the eastern fragment of the Manihiki Plateau, which is also about 25 Ma old at the time of its arrival at the subduction zone, seems to have been subducted almost completely. Small accreted fragments are possibly the source of one of the blocks identified within the magnetic anomaly map, for example, the Charcot Island block (Figure 4b) [*Ferraccioli et al*., 2006]. The subduction at this part of the Gondwana margin did not cease until the end of the CNS. Along with age and the subduction mode, the crustal thickness and buildup is the main factor that determines whether an oceanic plateau is subducted or not [*Cloos*, 1993]. A 23–25 km thick crust has been derived from gravity anomaly modeling for the Hikurangi Plateau [*Davy et al*., 2008]. The eastern Manihiki Plateau fragment must have had a maximum crustal thickness of 20 km at its breakup margin at the Manihiki Scarp with decreasing thickness toward its outer margins at the fringe of Ontong Java Nui. It is, therefore, comparable to the subducted thinner part of the Hikurangi Plateau [*Reyners et al*., 2011]. We suggest that the subduction of the eastern Manihiki Plateau fragment was possible due to its smaller crustal thickness in comparison to that of the Hikurangi Plateau.

4.3.2. Evidence for the Presence of an LIP Within the Antarctic Peninsula

The presence of relatively young buoyant LIP crust could cause flattening of the subducting slab [*Gutscher et al*., 1999, 2000]. A flattened slab leads to the emplacement of adakites within the volcanic arc [*Gutscher et al*., 2000]. In the Palmer Land region, adakitic rocks crop out [*Wareham et al*., 1997] and point to a mixing of different magmatic sources including a young (<25 Ma) oceanic component, which corresponds to the age of the LIP fragment at the time of subduction.

Vaughan et al. [2012] identified two distinct kinematic phases in the Palmer Land region, which they called the Palmer Land Event with phase 1 (about 107 Ma) and phase 2 (about 103 Ma). Both phases fall into the projected collision time of the eastern fragment of the Manihiki Plateau with the margin. Whether the two distinct phases can be associated with soft- and hard-docking events or rotation of the fragment comparable to that of the Hikurangi Plateau [*Davy*, 2014] cannot be distinguished. Structural geological data from rare outcrops and corresponding radiometric ages indicate two different paleostrain axes [*Vaughan et al*., 2012]; a modification of the collision pattern such as a rotation analog to the Hikurangi Plateau [*Davy*, 2014] seems plausible. A possible candidate to host small obducted remnants of the eastern fragment of the Manihiki Plateau may be the yet unexplained, strong Charcot magnetic anomaly, which lies at the continental margin of Palmer Land (Figure 4b). Petrological samples from Alexander Island (Charcot Island block) point to multiple phases of northward moving magmatic activity. This might be related to the subduction of young oceanic crust of the colliding Pacific-Phoenix oceanic spreading ridge [*McCarron and Larter*, 1998] or to the presence of a young oceanic plateau fragment [*Scarrow et al*., 1998]. The presence of an oceanic LIP fragment in the area of Palmer Land and Charcot Island seems plausible, although the exact area and impact of the collision or subduction cannot be better constrained. However, the presence of adakitic rocks indicates a slab flattening possibly induced by the subduction of an oceanic plateau.

4.4. Implications of the Subduction of the Manihiki Plateau Fragments

The reconstruction of LIP fragment paths and their subduction and accretion within the northern Andes and the Antarctic Peninsula contributes to the understanding of these complex accretional margins. The presence of a former oceanic plateau fragment in the subduction regime altered both subduction margins temporarily. The survival and location of such accretionary crust seems to be related to the subduction mechanism, the age, and the buildup of the plateau's crust. The fact that simultaneously emplaced oceanic plateaus shows such a wide range of possible fates while interacting with a subduction zone shows the high complexity and diversity of oceanic LIP crust. For instance, a further assessment of the Piñón formation basalts could give indication on where a possible weak layer within the plateau is situated and if this coincides with either the transition between the initial expansion phase of the Manihiki Plateau or the petrological difference between high and low TiO_2 magmas. This would allow a deeper insight into the crustal structure of LIPs. The presence of a LIP fragment at the Antarctic Peninsula margin helps to decipher the complex structure of this remote and still poorly studied area.

5. Conclusion

Ontong Java Nui emplaced during the Early Cretaceous and disintegrated into multiple fragments within a relative short period of time. Whereas the Manihiki Plateau—at its centerpiece—is located far from any present plate boundaries, all other fragments interacted with the circum-Pacific subduction zones. By blocking the subduction at the Solomon Trench, the Ontong Java Plateau initiated a subduction polarity reversal. The Hikurangi Plateau subducted below the Chatham Rise and the South Island of New Zealand, leading to a cessation of the subduction in this area of the eastern Gondwana margin. We trace the motions of northeastern and eastern fragments of the Manihiki Plateau across the Pacific Ocean to locations where they collided at the subduction margins of the northern Andes and West Antarctica, respectively.

Based on geophysical, petrological, and plate kinematic evidence, we infer that the Piñón formation of the northern Andes represents the remnant of a Manihiki Plateau fragment. The aggregation and partial subduction below the South American craton occurred during the Paleocene. The eastern fragment of the Manihiki Plateau was almost completely subducted in the area of Palmer Land of the southern Antarctic Peninsula in the middle Cretaceous, similar to the subducted part of the Hikurangi Plateau. The connection between the fragments of the Manihiki Plateau and their remnants onshore contributes to the understanding of the highly complex structure of the accretionary margins of the northern Andes and the Antarctic Peninsula.

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