Analysing two potential outlets of a shrinking Pacific mantle reservoir: The Scotia Sea and the Caribbean

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Content

Summary

This thesis deals with new plate reconstruction models of the Scotia Sea and the Caribbean - regions that share very complicated plate tectonic histories. Each model and their implications are discussed, e.g. it is shown that the Central Scotia Sea is most likely not a Mesozoic plate fragment as suggested by some authors. On the other hand, the presented reconstruction model of the Caribbean confirms the link between the Caribbean Plateau and the Galapagos hotspot, i.e. that the rising plume head of the paleo-Galapagos hotspot was the source of the rocks that compose the present-day Caribbean Plateau. This is consistent with abundant geochemical evidence but contrary to previous reconstruction models in an Indo-Atlantic hotspot reference frame. It is outlined in detail, which assumptions are necessary to establish this link.

The reconstruction models were further employed to derive regional age-grids. These in turn were used to calculate the present-day residual/dynamic topography of each region to test the hypothesis, whether the Scotia Sea and the Caribbean are asthenosphere outlets of a shrinking Pacific into a growing Atlantic mantle reservoir. The underlying idea is that due to the shrinkage of the Pacific a pressure gradient drives the asthenosphere through the proposed outlets, such that global mass balance can be achieved by lateral shallow mantle flow alone. If true, these regions should be characterized by a gradual decrease in dynamic topography from west to east. It is shown that the Scotia Sea does not serve as the suggested outlet, while some mantle inflow in the Caribbean realm occurs. However, because the amount of shallow mantle flow from the Pacific into the Atlantic realm seems to be rather small, it is concluded that deep mantle processes underneath Africa are the far more relevant source to achieve global mass balance.

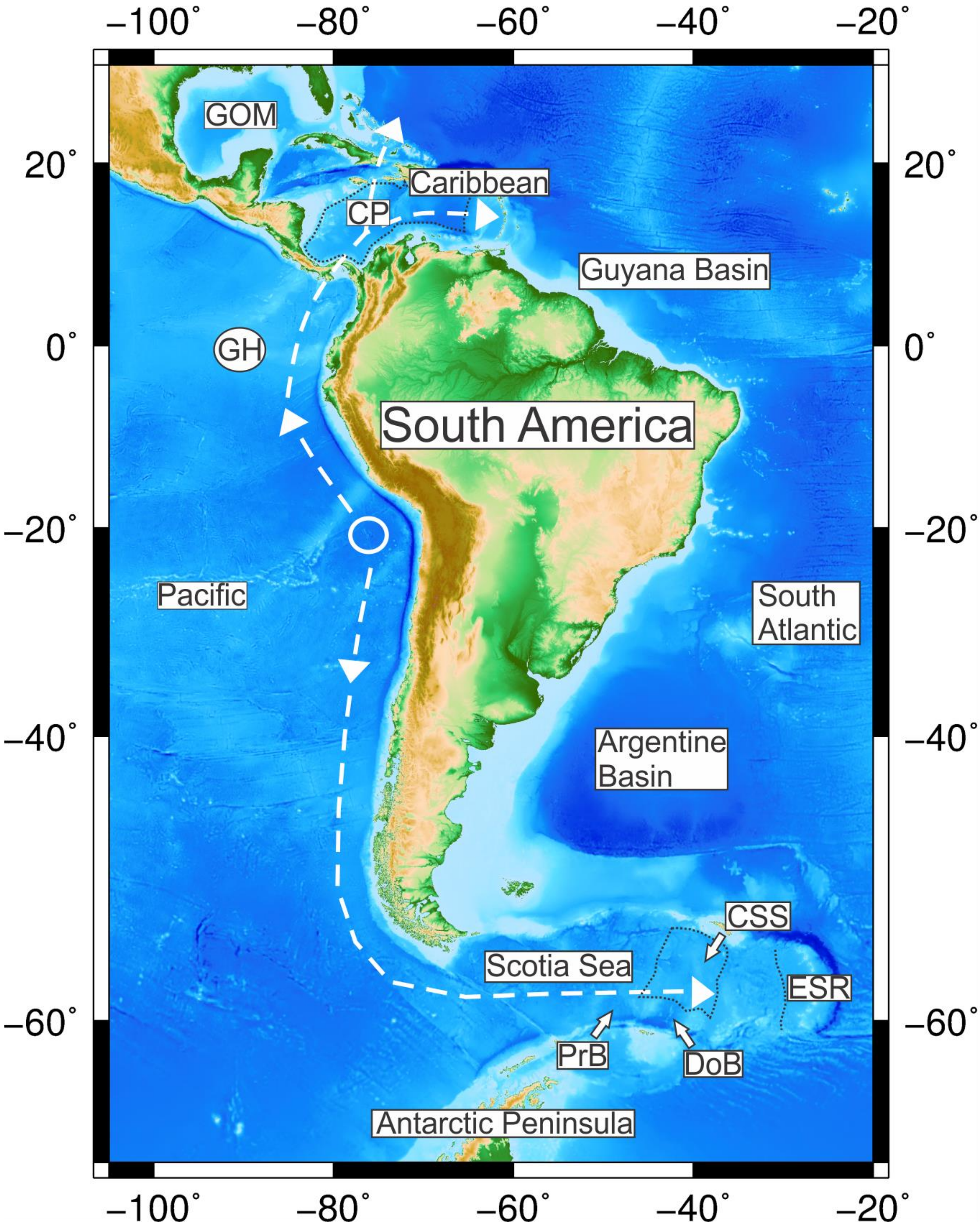
Note on this manuscript

This final version of my thesis has been updated relative to the original version submitted to the examination board. Chapter 2 is now *published* (status in original thesis version: *submitted*) and the follow up paper/chapter 3 *submitted* (status in original thesis version: *in preparation for submission*). This has led to small changes in the manuscript relative to the original thesis version, but none of the calculations or overall conclusions have been modified.

Introduction

Based on the pioneering continental drift theory of Alfred Wegener (1880-1930), Arthur Holmes' (1898-1965) ideas regarding mantle convection, and Harry Hess' (1906-1969) seafloor spreading hypothesis, the theory of plate tectonics has been accepted since the 1960s. According to this theory, the Earth is subdivided into several rigid plates that move with a few centimeters per year relative to each other and with respect to the underlying mantle. The theory offered answers to observations such as volcanic outbursts, earthquakes, ocean basin as well as mountain belt formation; phenomena that were otherwise difficult to explain. However, despite the general consensus that lithospheric plates glide on top of a comparatively weak asthenosphere, Earth's plate tectonic history is far from being comprehensively understood. While the relative motions of large-scale plates, such as the South American plate relative to the African plate, are rather well constrained, regional details are often much less understood and remain debated.

Two of these debated regions are the Scotia Sea and the Caribbean (Fig. 1), which share complicated plate tectonic histories. Arguably the most controversial aspect with respect to the Scotia Sea is the origin of the Central Scotia Sea, which has been interpreted as a Miocene back-arc basin or as a Mesozoic plate fragment (Barker, 2001; Eagles, 2010), respectively. Also, the spreading progression of Dove and Protector Basin are not well understood, because the observed magnetic anomaly sequences are very short and ambiguous (Eagles et al., 2006).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Guyana Basin
GH
South America
Pacific
South
Atlantic
Argentine
Basin
CSS
Scotia Sea
ESR
PrB
DoB
Antarctic Peninsula

Figure 1: Overview of the study areas, the Scotia Sea and the Caribbean. Shallow mantle flow paths according to Alvarez (1982) are indicated in white. The Central Scotia Sea (CSS) and the Caribbean Plateau (CP) are approximately outlined in black. Abbreviations are as follows: East Scotia Ridge (ESR), Protector Basin (PrB), Dove Basin (DoB), Galapagos hotspot (GH), Gulf of Mexico (GOM).

The Caribbean Sea is even more complicated, as most of the regional ocean floor depicts a large oceanic plateau, prohibiting entirely the detection of magnetic anomalies due to the thickened crust. However, it is widely accepted that the Plateau is underlain by a trapped piece of former Farallon lithosphere that moved in between the diverging North and South American plates. In this regard and due to its present-day proximity, the Galapagos hotspot has been interpreted as the source of the rocks that compose the Caribbean Plateau (Fig. 1). According to this theory, the rising plume head of the paleo-Galapagos hotspot and associated volcanic outbursts some 90 Ma ago created the Plateau on top of wide parts of Farallon lithosphere. This view was supported by geochemical investigations of rock samples, which suggested a hotspot origin (Hauff et al., 1997; Geldmacher et al., 2003; Thompson et al., 2003). Yet, plate reconstruction models failed to show this link and continuously predicted an offset of >1000 km between the Plateau and hotspot locations at the proposed time of formation (Pindell et al., 2006; Seton et al., 2012).

However, not only the complicated and controversial plate tectonic histories make the Scotia Sea and the Caribbean interesting regions to study. In fact, both have been viewed as outlets for shallow asthenospheric mantle flow from the Pacific into the Atlantic mantle domain (Fig. 1). This attributes potential geodynamic key roles to both of them. It is based on the reasoning that contrary to the remaining eastern Pacific plate margin, neither deep continental roots nor subducted slabs, i.e. tectonic features generally understood as barriers to mantle flow, are present on the western margins of these two regions (a slab window is proposed for the western Caribbean, southwest of Costa Rica). Based on this observation, it was Walter Alvarez (1982), who proposed that Pacific mantle must be driven out of the Pacific realm through the Scotia Sea and the Caribbean, in order to achieve mass balance between the closing Pacific and opening Atlantic mantle reservoirs, respectively. The Pacific mantle reservoir is shrinking, because the Pacific basin is surrounded by subduction zone systems in which the Pacific lithosphere is constantly being subducted. On the contrary, the Atlantic is opening as large-scale subduction zone systems on the Atlantic plate margin are absent and continuously new seafloor is created along the mid-Atlantic ridge. Shear-wave splitting and geochemical studies have addressed the question, whether the Scotia Sea and the Caribbean truly serve as the proposed outlets, but they have partly led to ambiguous results (e.g. Pearce et al., 2001; Russo et al., 1996, Helffrich et al., 2002).

In this thesis, the above described open questions with respect to the plate tectonic histories of the Scotia Sea and the Caribbean as well as the debate upon Alvarez' (1982) mantle flow hypothesis are revisited. To this end, the plate tectonic histories for each region were modeled based on a literature review. Thorough tests of suggested tectonic scenarios were performed, whether the suggested ideas comply with general principles of plate tectonics, e.g. plate overlap was avoided. This work was necessary as most of the previously published plate reconstruction models were presented only diagrammatically. Such models are not based on rotation files, which contain the longitudinal and latitudinal positions of Euler poles as well as the time specific rotations about these poles to describe the relative motion of either two plates or of one plate relative to the absolute (mantle) reference frame. Therefore, these "old" models merely represent illustrations of an idea rather than a well constrained, reproducible model. Contemporary reconstruction models should, however, be reproducible by using Euler rotation poles and a specific mantle reference frame. Thus, new plate kinematic models of the Scotia Sea and the Caribbean that fulfill the modern needs of science were necessary.

These kinematic models were also useful to address the question, whether the Scotia Sea and the Caribbean are the suggested mantle outlets or not. By means of a geodynamic approach, the present-day dynamic topography of the two regions was calculated presuming that an asthenospheric flow should be indicated by a gradual decrease in dynamic topography from west to east through the Scotia Sea and the Caribbean, respectively. Thereby, dynamic topography refers to the mantle component of the observed topography, originating from mantle up- or downwellings that lead to positive or negative dynamic topography signals, respectively (Braun, 2010). The mantle flow itself is driven by density contrasts within the mantle. To unravel the dynamic component of the observed topography, a theoretical basement depth is subtracted from the isostatically corrected, observed bathymetry. As the theoretical basement depth is based on the regional lithospheric age distribution, plate kinematic models are essential to calculate the present-day dynamic topography of a region, as they allow for the derivation of an age-grid. This instance connects the newly developed plate reconstruction models mentioned above with the mantle flow hypothesis.

Chapter 1 concentrates on the Scotia Sea. A new plate reconstruction model is introduced and an age-grid was derived. The latter was used to calculate the present-day dynamic topography in the Scotia Sea. The results were published in Tectonophysics in 2012.

Chapter 2 evaluates different theories with respect to the Caribbean plate tectonic history. Particularly the link between the Caribbean Plateau and the Galapagos hotspot is investigated. The results were published in GeoResJ in 2014.

Chapter 3 focuses on the regional dynamic topography in the Caribbean. The required age-grid was derived from the plate reconstruction model described in chapter 2. Chapter 3 is therefore an extension of chapter 2 and has been submitted to GeoResJ.

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Chapter 1

The Scotia Sea gateway: no outlet for Pacific mantle

This chapter has been published in Tectonophysics:

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Abstract

The Scotia Sea in the South Atlantic holds a prominent position in geodynamics, because it has been proposed as a potential outlet of asthenosphere from under the shrinking Pacific into the mantle beneath the opening Atlantic. Shear wave splitting and geochemical studies have previously tested this hypothesis. Here, we take a different approach by calculating present-day dynamic topography of the region in search for a systematic trend in dynamic topography decreasing from west to east in response to a flow-related pressure gradient in the sublithospheric mantle. To this end, we reconstruct the kinematic history of the Scotia Sea, which is characterized by complex back-arc spreading processes active on a range of time scales. Our plate reconstructions allow us to derive an oceanic age-grid and to calculate the associated residual (dynamically maintained) topography of the Scotia Sea by comparing present-day isostatically corrected topography with that predicted from our reconstruction. The results provide no indication for a systematic trend in dynamic topography and we conclude that the material needed to supply the growing subatlantic mantle must be derived from elsewhere.

1.1 Introduction

According to Alvarez (1982) the isolation of Antarctica by opening of the Drake Passage in the Scotia Sea (Fig. 1) had profound consequences for the global mantle circulation system by establishing subsurface mantle flow from under the Pacific into the Atlantic Ocean domain. As he conceived continental roots and subducting slabs as effective barriers to lateral mantle flow from under the Pacific into the Atlantic hemisphere, he proposed the newly formed seaway as an outlet for flow within the asthenosphere and a potential mechanism to establish mass balance between the shrinking Pacific and the growing Atlantic mantle reservoirs.

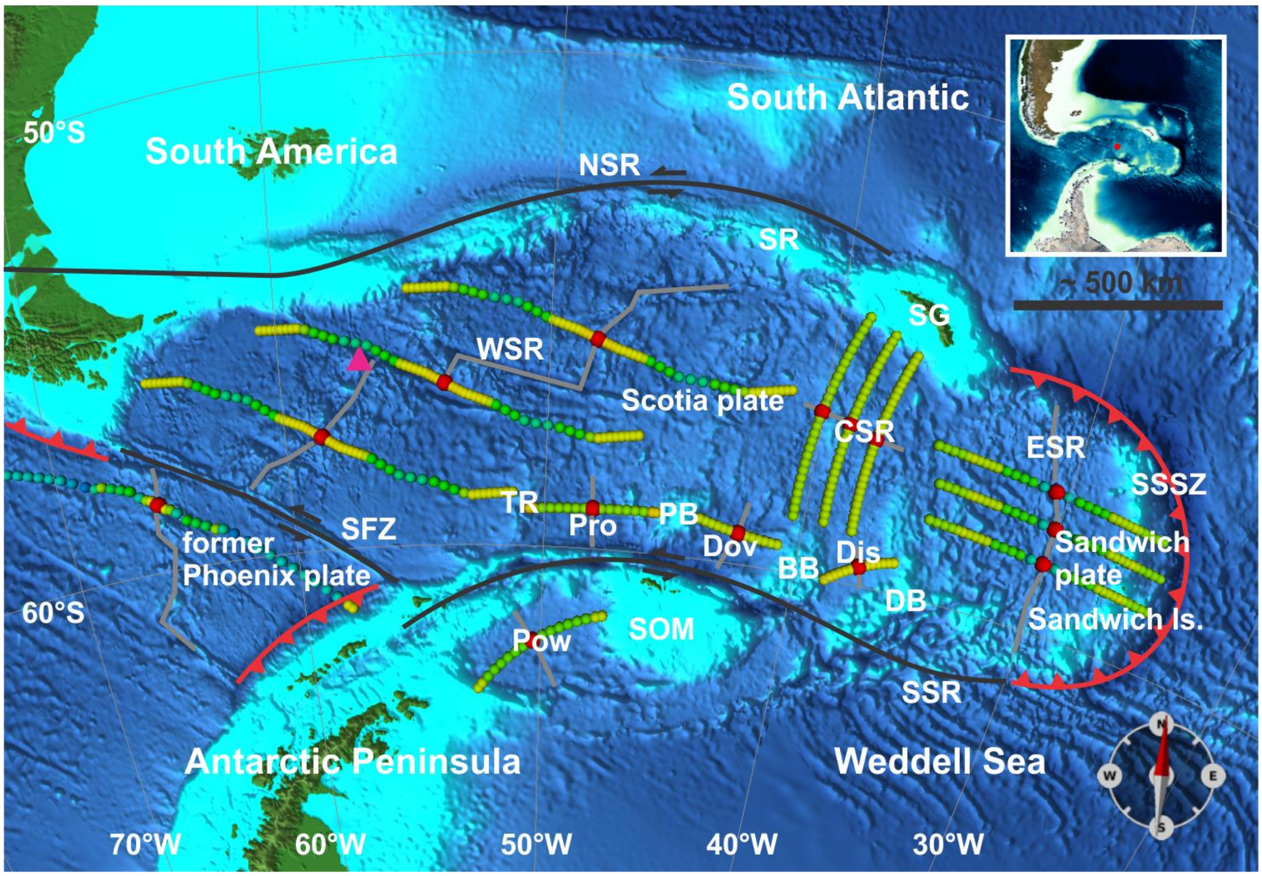

Figure 1: View of Scotia Sea consisting of the Scotia and Sandwich plates, located in between the Antarctic Peninsula and South America (see also insert map, where the red dot indicates the center of the topview map). The region is framed by transform boundaries in the north (North Scotia Ridge (NSR)), south (South Scotia Ridge (SSR)), west (Shackelton Fracture Zone (SFZ)), and the South Sandwich subduction zone

(SSSZ) in the east. Other features are Shag Rocks (SR), South Georgia (SG), and the South Orkney Microcontinent (SOM). Flowlines displaying motion paths of different continental fragments are shown in green. Active spreading (East Scotia Ridge (ESR)) exists in the East Scotia Sea. Extinct spreading ridges are found in the West Scotia Sea (West Scotia Ridge (WSR)), Central Scotia Sea (Central Scotia Ridge (CSR) which remains controversial; see subsection 2.2) and in the Protector (Pro), Dove (Dov), and Discovery Basins (Dis), respectively, which are bounded by presumably - as some discussion on their origin persists - South American continental fragments (Terror Rise (TR), Pirie Bank (PB), Discovery Bank (DB)). Extinct ridges are also found on the boundary between the Antarctic plate and former Phoenix plate as well as in the Powell Basin (Pow). Location of the dredge sample with Pacific mantle type signature (Pearce et al., 2001) [see discussion] is marked by a triangle.

The notion of an asthenosphere has a long history in the geophysics, dating from 19th century investigations of isostatic support of mountain belts (see Watts 2001 for a review). Early on in the 20th century, geodynamicists conjectured it provides a zone of weakness over which plates glide easily (Chase, 1979). Modern geophysical evidence for an asthenosphere comes in the form of geoid and postglacial rebound studies (Hager and Richards, 1989; Mitrovica, 1996), supported by investigations into global azimuthal seismic anisotropy (Debayle et al., 2005), and seismic (Grand and Helmberger, 1984) and mineralogical (Stixrude and Lithgow-Bertelloni, 2005) work on the properties of the low seismic velocity zone found at a depth of between 100-400 km beneath oceanic and tectonically active regions. 3-D spherical mantle convection models are consistent with the view of a weak upper mantle. The models show that a low-viscosity asthenosphere has a profound effect on convection by promoting a long-wavelength convective planform with only a few large elongated mantle convection cells (Bunge et al., 1996), comparable to the long wavelength pattern that characterizes global mantle flow. The dominant influence of the asthenosphere on the convective planform has also been inferred from analytic fluid dynamic considerations (Busse et al., 2006), and it has been proposed that the existence of an asthenosphere is essential in stabilizing the unusual plate tectonic style of convection that prevails on Earth (Richards et al., 2001).

Complementary to the mobility of the asthenosphere are mechanically stable keels, termed tectosphere, which may exist beneath old portions of the continental lithosphere (Jordan, 1978). Such keels are presumed to pierce through the entire asthenosphere, promoting coupling between continents and the deeper mantle, and restricting the asthenosphere to the subsurface beneath oceanic realms (Conrad and Lithgow-Bertelloni, 2006).

Here we revisit the hypothesis of Pacific-to-Atlantic asthenosphere flux around the tip of South America. The Scotia Sea provides an ideal location, as noted by Alvarez (1982), to explore the pattern of upper mantle flow, as no continental roots or an active subduction zone provide a barrier to asthenospheric flow. By reconstructing the plate kinematic history of the Scotia Sea we present a geodynamic approach that allows us to detect a dynamically maintained component of topography associated with viscous stresses created by upper mantle flow via Drake Passage through the Scotia Sea, rather than by thermal subsidence related to cooling within the oceanic lithosphere (Braun, 2010).

Our paper is organized as follows: we begin with a general estimate on the pressure gradient required to generate sufficient Pacific mantle flow through the Scotia Sea to achieve mass balance in the Atlantic. Thereafter, we describe the applied dynamic topography deconvolution method, which is followed by a results section and a discussion.

1.2 Methodology

1.2.1 Asthenosphere flow estimates

We begin this section with a simple scaling argument, under the assumption that the mass balance required to accommodate Atlantic opening is purely achieved through Pacific mantle flow through the Scotia Sea. We estimate the volume growth rate (Q) of the South Atlantic to be ~ 50 km^3/yr, for which we assumed a length of 10.000 km and a full-spreading rate of 2.5 cm/yr of the South Atlantic mid-ocean ridge system, respectively, as well as a 200 km thick asthenosphere directly under the ridge.

Based on this rate, the pressure difference (Δp) between the western and eastern ends of the Scotia Sea required to compensate the Atlantic growth by Pacific-to-Atlantic mantle flow can be estimated from a simple Poiseuille flow, viewing the Scotia Sea as a channel with thickness h = 200 km (i.e. asthenosphere thickness), width Δy = 1000 km, and length $\Delta x = 3000$ km. The pressure difference is given by:

$$\Delta p = \frac{Q*12\mu*\Delta x}{\Delta y*h^3} \tag{1}$$

Assuming an asthenospheric viscosity (μ) of 10^{19} Pa*s, we arrive at $\Delta p \sim 600$ bar, so that the expected dynamic topography in the West Scotia Sea is about 1800 m, large enough to be detectable, although we note that major uncertainties exist regarding the thickness and viscosity of the asthenosphere.

1.2.2 Dynamic topography deconvolution method

The dynamic component of the actual bathymetry can be unraveled by calculating the difference between the expected bathymetry based on standard cooling models and the isostatically corrected, observed ocean depth (Kido and Seno, 1994). A precise age-grid is therefore crucial, which can be derived from plate reconstruction models which are typically based on magnetic anomaly interpretations. To this end we developed a reconstruction model of the Scotia Sea by using 4DPlates (Clark et al., 2012), a new software package developed by Simula Research Laboratory and Statoil. Deriving an age-grid for the Scotia Sea directly from magnetic lineations is difficult. Indeed various magnetic lineations have been discovered on the ocean floor of the Scotia Sea, but most of them can be correlated to multiple sections of the magnetic reversal scale (Eagles et al., 2006). For this paper, we did not re-interpret magnetic data. Instead we developed a reconstruction model in the moving Atlantic-Indian hotspot reference frame (O'Neill et al., 2005) by testing various age suggestions for each basin within the Scotia Sea directly for plate tectonic consistency (i.e. a "best-fit" reconstruction by avoiding plate overlap).

Our model is shown in figure 2. It displays that the kinematic history of the Scotia Sea is characterized by back-arc spreading processes causing the initial opening of Protector and then Dove basin. We note that both basins may also have opened purely as a result of divergence between South America and Antarctica (Livermore et al., 2005). The early opening phase was followed by further eastward subduction retreat and spreading along the West, Central, and East Scotia Ridges (Table 1).

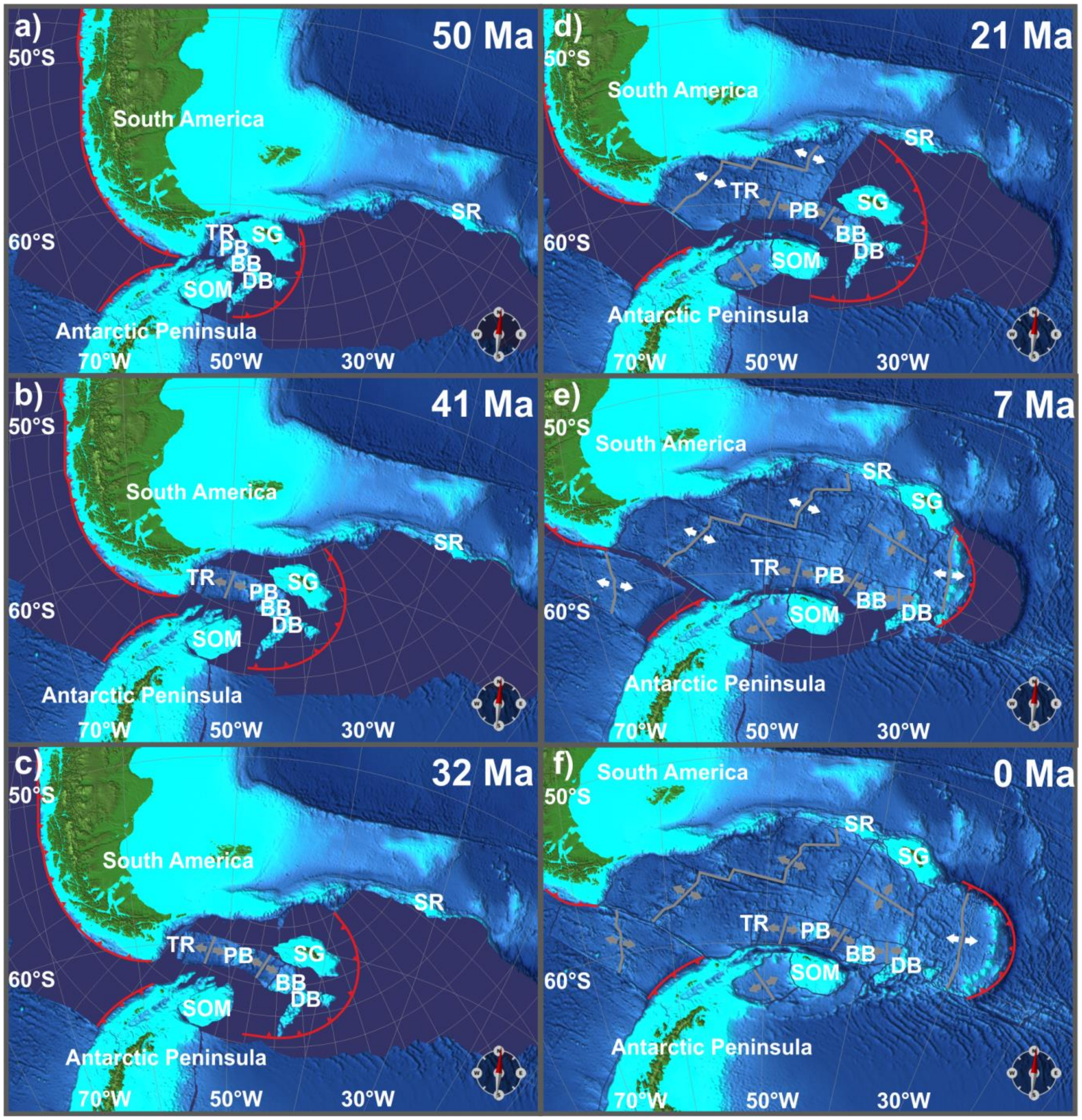

Figure 2: Tectonic reconstructions corresponding to 50 (a), 41 (b), 32 (c), 21 (d), 7 (e) and 0 Ma (f); labels as in Fig. 1; present-day bathymetry is shown (i.e. no age-masking). White/grey arrows indicate active/inactive spreading. 50 Ma ago, South America and Antarctica are connected by a coherent bridge of continental fragments. At 41 Ma, separation between South America and Antarctica and subduction behind Discovery Bank and South Georgia has started, as suggested by Barker (2001). Protector Basin has opened. 32 Ma ago, active spreading occurs along the West Scotia Ridge but has ceased in Dove Basin. At 21 Ma, further subduction behind Discovery Bank leads to opening of Discovery Basin. Moreover, subduction behind South Georgia is about to cause back-arc spreading in the Central Scotia Sea. Powell Basin is already fully open,

placing the South Orkney Microcontinent to its present position relative to Antarctica. Seafloor spreading lasts along the West Scotia Ridge. 7 Ma ago spreading occurs along the West and East Scotia Ridges and ceases along the Central Scotia Ridge.

The age of the Central Scotia Sea is controversial. We treated it as a Miocene back-arc basin, based on for example Barker (2001) (Fig. 2). Yet, heat flow measurements, lack of a visible fossil ridge, fracture zones, and triple-junction traces, as well as the flexural strength of the lithosphere, for example, have led Eagles (2010) and De Wit (1977) to propose that the Central Scotia Sea might be a Mesozoic plate fragment. We could not develop a kinematically consistent reconstruction model that avoids large plate overlap based on this idea, but we acknowledge Eagles' (2010) reconstruction model. However, because he did not provide rotation poles along with his paper and showed only very large time steps, we could not reappraise his model.

Table 1

Ages, rotation poles, and rotation angles as implemented in the reconstruction model (Fig. 2). All other rotations are based on the global rotation file[1] provided by the EarthByte group.

Plate pairs	age	rotation pole (long)	rotation pole (lat)	rotation angle (degree)
Sandwich Plate East -	0.0	0.0	0.0	0.0
Sandwich Plate West	1.7	80.0	-16.57	-1.50
(East Scotia Ridge)	6.0	80.0	-16.57	-4.11
	10.0	80.0	-16.57	-5.76
	11.5	80.0	-16.57	-6.30
	15.0	80.0	-16.57	-8.04
South Georgia -	0.0	0.0	0.0	0.0
Bruce Bank	7.0	0.0	0.0	0.0
(Central Scotia Ridge)	21.0	-57.38	-21.44	24.56
Drake Passage North -	0.0	0.0	0.0	0.0
Drake Passage South	6.25	0.0	0.0	0.0
(West Scotia Ridge)[2]	10.95	6.21	-27.89	-0.97
	16.73	37.71	-23.37	-2.16
	20.13	27.89	-23.64	-3.60
	23.07	31.76	-27.49	-5.00
	26.55	17.21	-32.12	-6.64
	34.0	20.96	-36.46	-8.20
Former Phoenix Plate -	0.0	0.0	0.0	0.0
Antarctica[3]	3.3	0.0	0.0	0.0
	5.23	68.07	91.96	-1.83
	6.57	69.21	97.32	-3.07
	8.07	68.98	90.41	-5.85
	9.31	69.77	92.17	-7.94
	10.95	69.98	94.12	-11.23
	12.40	70.23	96.68	-13.38
	14.61	70.02	94.97	-18.94
	23.41	70.02	94.97	-38.94
Pirie Bank -	0.0	0.0	0.0	0.0
Terror Rise	41.0	0.0	0.0	0.0
(Protector Basin)	48.5	80.82	135.66	-6.84
Bruce Bank -	0.0	0.0	0.0	0.0
Pirie Bank	34.7	0.0	0.0	0.0
(Dove Basin)	41.0	-50.04	-39.80	-10.0
Discovery Bank -	0.0	0.0	0.0	0.0
Bruce Bank	14.4	0.0	0.0	0.0
(Discovery Basin)	21.0	-62.56	-34.38	25.84
South Orkney Microcontinent -	0.0	0.0	0.0	0.0
Antarctic Peninsula	21.8	0.0	0.0	0.0
(Powell Basin)	29.7	-65.13	-45.38	-43.79

[1]http://www.earthbyte.org/Resources/earthbyte_gplates.html

[2]rotations are based on Eagles et al. (2005)

[3]rotations are based on Eagles (2003)

Table 1 continued: Literature sources for the kinematic reconstruction

Basin name	Reference for	Reference for spreading
onset	spreading cessation	
	(Ma)	(Ma)
West Scotia Ridge	Livermore et al. (2005)	Eagles et al. (2005)
Central Scotia Ridge	Barker (2001)	Barker (2001)
East Scotia Ridge	Larter et al. (2003)	still active
Protector Basin	Eagles et al. (2006)	Eagles et al. (2006)
Dove Basin	Eagles et al. (2006)	Eagles et al. (2006)
Discovery Basin	Lodolo et al. (2010)	Lodolo et al. (2010)
Powell Basin	Eagles and Livermore (2002)	Eagles and Livermore (2002)
Phoenix Plate	Manually extrapolated	Eagles (2003)

Another controversial topic is the opening sequence of Dove (41 - 34.7 Ma) (Eagles et al., 2006) and Protector Basin. The latter has been suggested to be of Miocene (17.4 - 13.8 Ma) (Galindo-Zaldivar et al., 2006), Oligocene (34 – 30 Ma) or Eocene age (~ 48 – 41 Ma) (Eagles et al, 2006), respectively. A Miocene aged Protector Basin in our model results in plate overlap of the southern continental fragments with the South Orkney Microcontinent. Also, the Oligocene age assumption would cause plate overlap in our reconstruction model, which we avoided by adopting the Eocene age suggestion. Different age suggestions exist also for the other basins, e.g., the onset of spreading along the West and East Scotia ridges remains under discussion, but the variations are not as large as for the Central Scotia Sea or Protector Basin.

4DPlates uses flowlines to derive an age-grid from our reconstruction model. They are based on the rotation file (see table 1) and represent the motion of the continental fragments and microcontinents for each time step, respectively. By exporting the flowlines from the program, we created a synthetic age-grid (Fig. 3) containing pairs of longitude and latitude as well as the associated age for each location within the Scotia Sea.

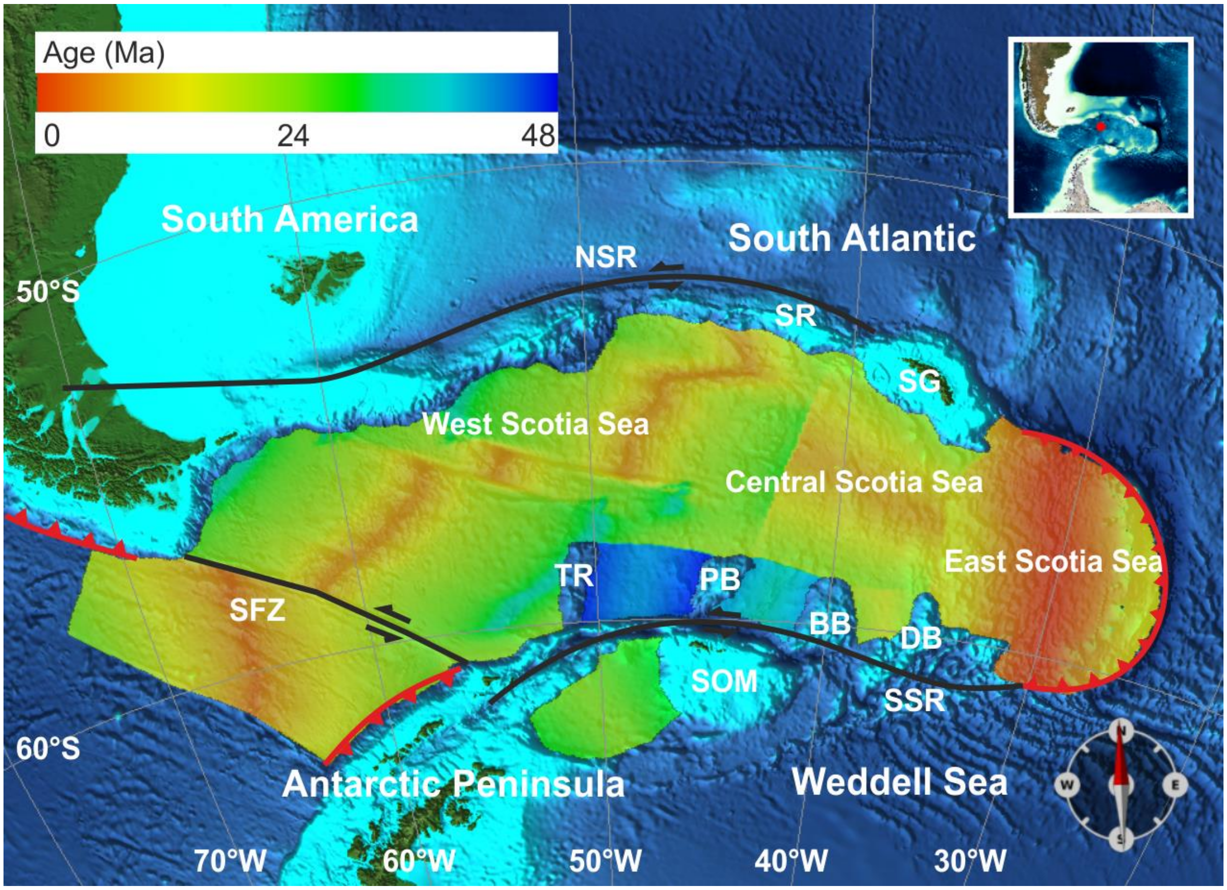

Figure 3: Synthetic age-grid based on the reconstruction model as shown in figure 2.

It is well known that a half-space cooling model predicts deeper basement depths for old ocean floor than so called plate models. For the sake of comparison, we therefore converted our synthetic age-grid (Fig. 3) into expected basement depth based on three standard lithospheric cooling models. The half-space cooling model (Turcotte and Oxburgh, 1967) is uniformly based on the function: depth = 2600 + 345 * √(age). The GDH1 (Stein and Stein, 1992) and PSM (Parsons and Sclater, 1977) plate models deviate from the continuous "square-root of age" assumption in that these models assume exponential functions for ages greater than 20 Ma and 70 Ma, respectively.

The observed bathymetry (Amante and Eakins, 2009) was isostatically corrected for sediments following Sykes (1996) and using the dataset provided by the National Geophysical Data Center (Fig. 4). The difference between the age corrected depth and the isostatically adjusted observed bathymetry was identified as dynamically maintained topography, assuming it is related to mantle flow. Potential sources of error that we will return to in the discussion section are (1) inaccuracies in sediment amounts, (2) uncertainties in crustal thickness as well as (3) in the assumed basin ages (as already pointed out). Since dynamic topography is a long-wavelength feature we low-pass filtered (cut-off wavelength: 500 km) our results.

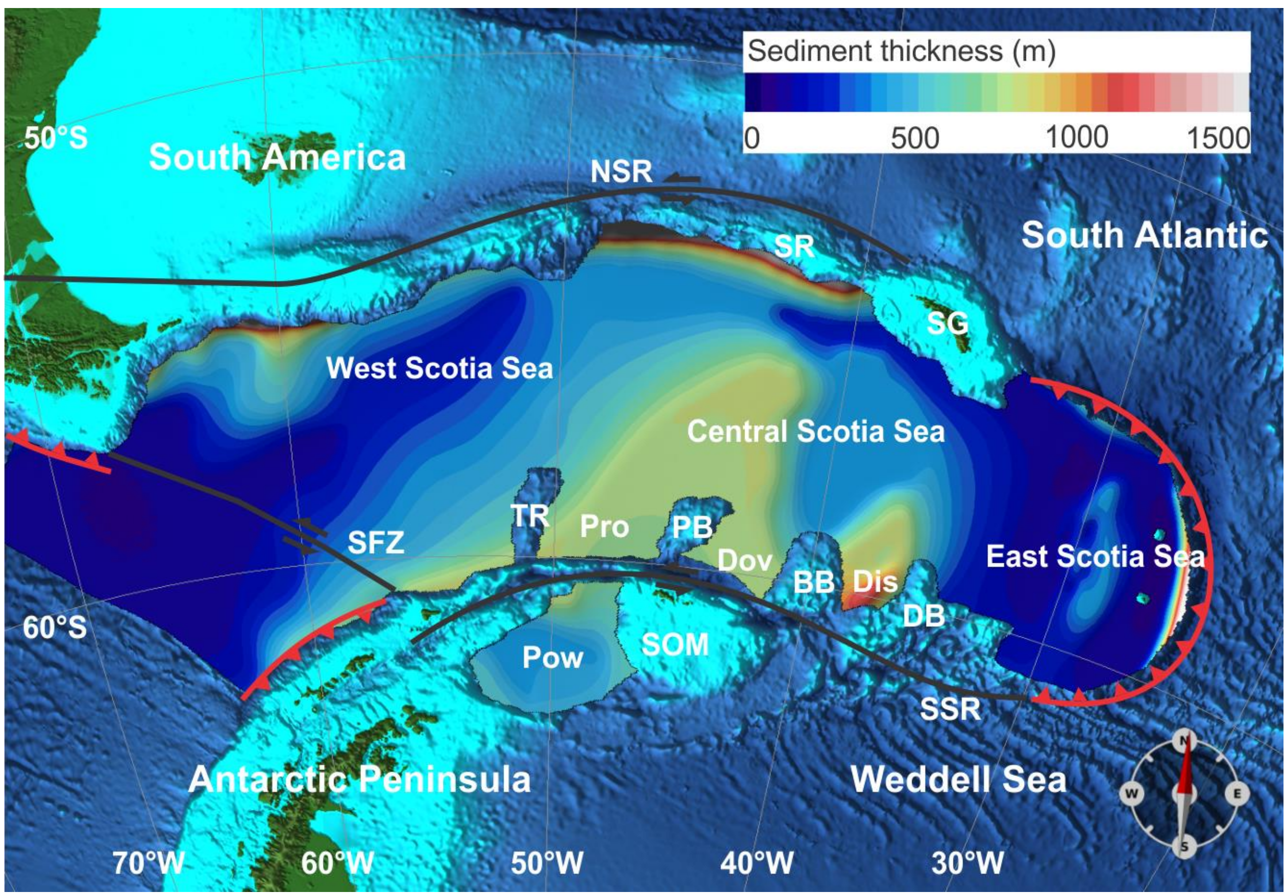

Figure 4: Sediment distribution in the Scotia Sea based on the 5 arc-minute by 5 arc-minute global grid that is available from the National Geophysical Data Center (NGDC). The sediment thickness is generally rather low (< 1000 m). Note that this also applies to the Central Scotia Sea but uncertainties exist (see discussion).

1.3 Results

Figure 5 shows the estimated dynamic topography based on the half-space cooling model (Turcotte and Oxburgh, 1967). The difference in the calculated dynamic topography is negligible between the different cooling models, as we assume a rather young history of the Scotia Sea in our reconstruction model. Thus, we report values based on the half-space cooling model only and refer to figure 6a for the inferred dynamic topography along a profile across the Scotia Sea (Profile A in figure 5) based on the different cooling models.

We see from figure 5 that the calculated present-day dynamic topography of the Scotia Sea is generally low. The western part, including the former Phoenix plate, shows no significant dynamic topography, while the eastern part shows a signal of 90 ± 330 m. A negative signal is observed on the northern flank of the West Scotia Ridge (-100 ± 130 m), while the southern flank shows a positive signal (80 ± 180 m). The East Scotia Ridge shows negligible dynamic topography on the western side (10 ± 270 m) and dynamic topography of 150 ± 350 m on the eastern side (Sandwich plate). The ridge and nearby areas show a negative signal of up to 470 m, while the Central Scotia Sea signal (290 ± 90 m) is on the order of those calculated for the southern basins (i.e. Protector, Dove, and Discovery Basin), which on average show some 330 m of positive dynamic topography. The Powell Basin shows a positive signal of 520 ± 120 m.

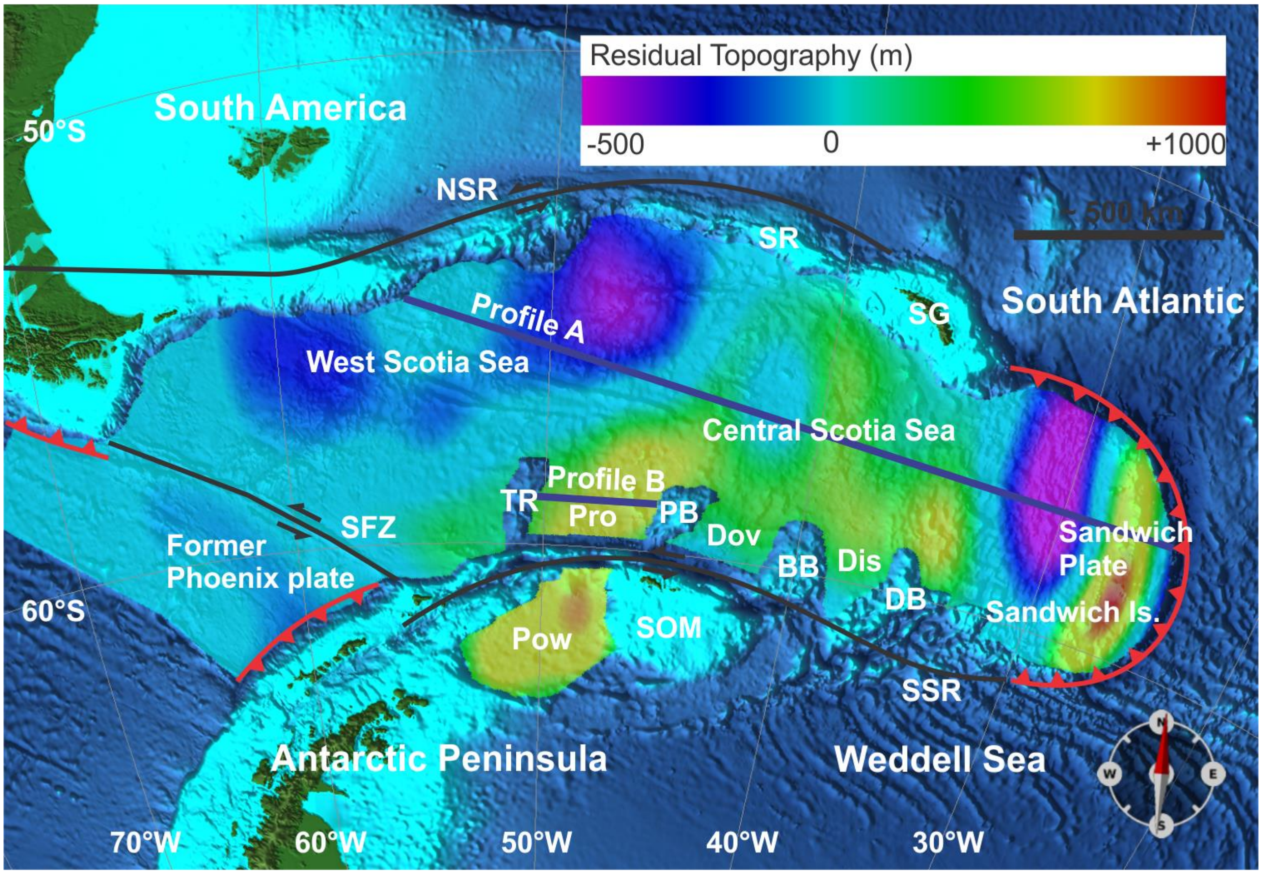

Figure 5: Present-day bathymetry with superimposed residual (dynamic) topography based on a half-space cooling model (Turcotte and Oxburgh, 1967). Note the regional low dynamic topography in the north-west and the high in the east. Note also the strong local dynamic topography signal associated with the subduction of the Scotia arc. The blue lines labeled Profile A and B indicate the location of the profiles illustrated in figure 6 and 7, respectively.

1.4 Discussion

Our results are of great interest as they show no indication for a systematic eastward decrease in dynamic topography (Fig. 5) that one would expect in the presence of substantial asthenosphere flux from the Pacific into the Atlantic around the southern tip of South America (Alvarez, 1982).

This finding agrees well with teleseismic shear-wave splitting studies (Helffrich et al., 2002), which map upper mantle flow directions in the southern Atlantic. The seismically inferred flow directions parallel the absolute plate motion of South America but provide no indication for direct asthenosphere flow through the Drake Passage. Similarly, while there is evidence from geochemical studies for a mantle domain boundary between the Drake Passage and the East Scotia Sea and hence for some Pacifc-to-Atlantic upper-mantle transport through the Drake Passage, the inferred amount is minor and probably related to regional tectonic constraints, rather than global mass-balance requirements (Pearce et al., 2001) (Fig. 1).

In figure 6b we tested the consequences of assuming a Mesozoic plate fragment (assuming an age of 80 Ma) in terms of dynamic topography. Note that approximately 2 km of dynamic topography would be implied in this case, no matter what lithospheric cooling model is used, in order to compensate for the fact that a Mesozoic plate fragment would yield much deeper ocean floor in the Central Scotia Sea than what is actually observed. Such large dynamic topography is difficult to understand on geodynamic grounds, as the Scotia Sea is generally characterized by low dynamic topography < 1 km (see results and Fig. 6a + b). Of course, uncertainties exist regarding the sediment amount and crustal thickness of the Central Scotia Sea, as noted before. With respect to the sediments, we had access to two global sediment grids, i.e. a 1 degree by 1 degree grid (Laske and Masters, 1997) and the eventually used 5 arc-minute by 5 arc-minute grid from the National Geophysical Data Center (Fig. 4). Both grids are mostly based on the same data for the southern oceans provided by Hayes and LaBrecque (1991), so unsurprisingly we found no major deviations between these two grids for the Central Scotia Sea (Table 2). Both indicate that the sediment coverage in the Central Scotia Sea is similar to the surrounding environment and rather low. Nevertheless, in parts of the Central Scotia Sea, the sediment thickness seems to be higher, since Eagles (2010) suggests sediment thicknesses between 800 m and 1500 m based on seismic profile data. In this case, the dynamic topography signal of the Central Scotia Sea would obviously be smaller and potentially make the assumption of an old Mesozoic plate fragment more reasonable. As to the crustal thickness we also cannot entirely rule out thickened crust. Unpublished work presented at the 11th International Symposium on Antarctic Earth Science held in Edinburgh in July, 2011 points in this direction, but overall in the Scotia Sea it is likely to be small because of the presumably young history of the region.

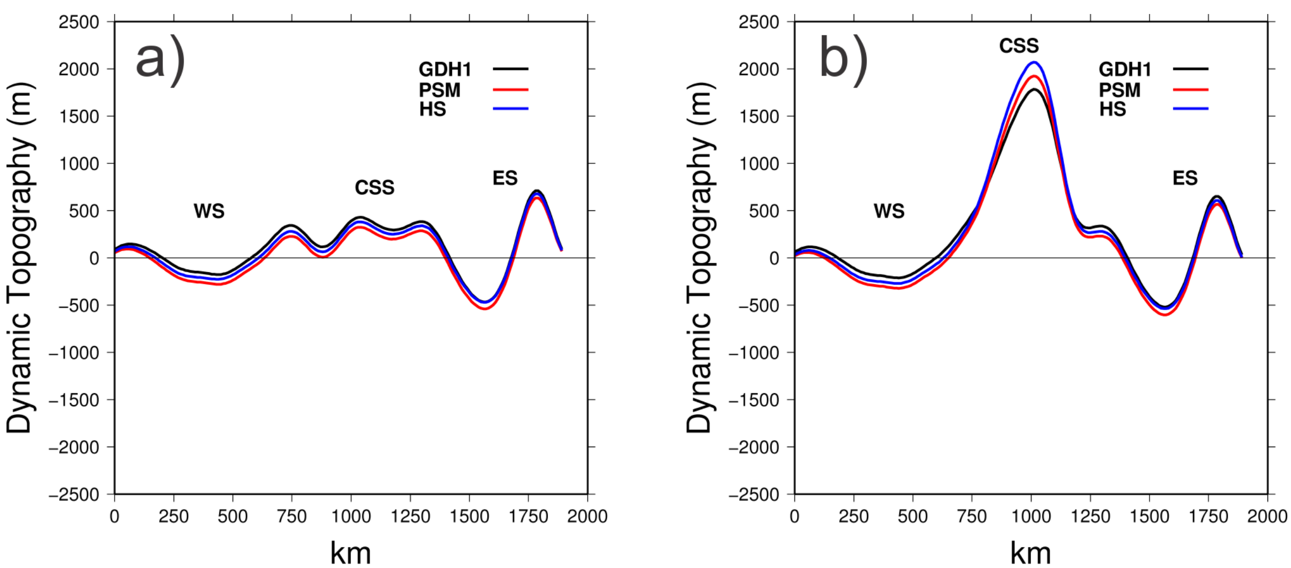

Figure 6: Dynamic topography based on three different lithospheric cooling models along profile A (see figure 5 for location). Labels are: West Scotia Sea (WS), Central Scotia Sea (CSS), East Scotia Sea (ES), and half-space model (HS). a) Dynamic topography based on the reconstruction model as shown in figure 2. b)

Dynamic topography based on an 80 Ma old plate fragment in the Central Scotia Sea (CSS). Note that such an old fragment implies ~ 2 km of dynamic topography. The neighboring regions, i.e. the West Scotia Sea (WS) and East Scotia Sea (ES), show considerably lower dynamic topography.

Table 2 | Comparison of the sediment amounts in the Central Scotia Sea based on the 1 degree by 1 degree sediment grid provided by Laske and Masters (1997) and the 5 arc-minute by 5 arc-minute sediment grid available from the NGDC database.

	Laske and Masters (1997)	NGDC
Minimum (m)	325	189
Maximum (m)	1063	844
Average (m)	566	471
Standard deviation (m)	162	171

We also calculated the dynamic topography of the Protector Basin, based on the various age suggestions that were mentioned in section 2.2. In figure 7a,b,c we illustrate the expected magnitude of dynamic topography along profile B in figure 5. A Miocene aged Protector Basin yields negative dynamic topography of ~ 140 to 240 m (Fig. 7a), depending on which cooling model is considered. This result would be difficult to understand, as the surrounding area shows positive dynamic topography. The suggested Oligocene and Eocene ages instead imply a positive dynamic topography signal (Fig. 7b+c) on the same order of magnitude as the surrounding area. The Oligocene age (Fig. 7b), however, results in plate overlap in our reconstruction so that we implemented the Eocene age (Fig. 7c) in our model as was pointed out above.

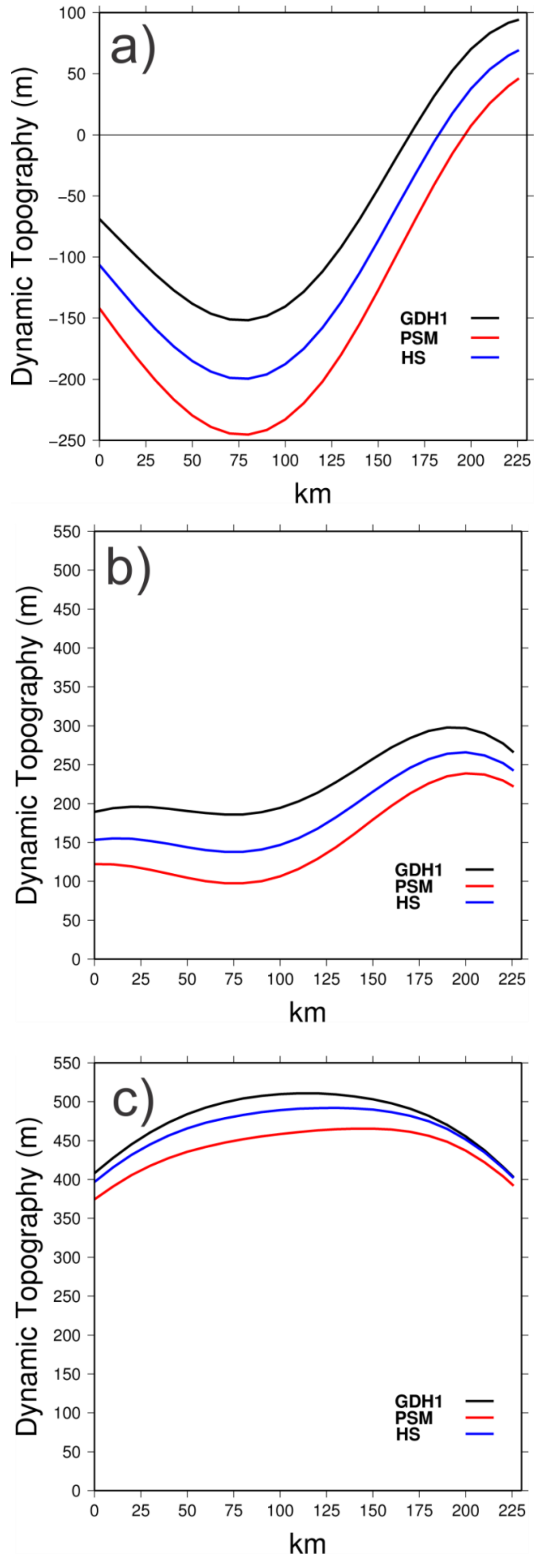
a)
Dynamic Topography (m)
km
GDH1
PSM
HS
b)
Dynamic Topography (m)
km
GDH1
PSM
HS
c)
Dynamic Topography (m)
km
GDH1
PSM
HS

Figure 7: Dynamic topography based on different lithospheric cooling models along profile B located in the Protector Basin (see figure 5). Labels as in figure 6 a) Assuming a rather young age of this basin, as was proposed by Galindo-Zaldivar et al. (2006), results in negative dynamic topography in parts of this basin. Assuming an b) Oligocene or c) Eocene age (the age adopted in our plate reconstruction model), as proposed by Eagles et al. (2006), results in positive dynamic topography on the order of a few hundred meters.

The low dynamic topography in the West Scotia Sea may be related to the prominent dynamic topography low of the Argentine Basin (Fig. 8), while small scale positive dynamic topography signals in the Central Scotia Sea and in the smaller basins in the south have been attributed to mid-ocean ridge subduction (Guillaume et al., 2009). Magnetic anomalies within the northern Weddell Sea (Fig. 1) south of the Scotia Sea, for example, become progressively younger northwards, suggesting ridge-crest subduction (Barker et al., 1984). These collisions are proposed to be youngest to the east and oldest to the west, with an extent probably as far west as the Antarctic Peninsula at 50 degree west (Barker, 2001).

The apparent lack of upper mantle flux from the Pacific into the Atlantic around the southern portion of South America raises questions about the origin of the material needed to supply the growing South Atlantic mantle reservoir. According to Alvarez (1982), the main gateway from the East Pacific into the Atlantic in addition to the Drake Passage is located in the Caribbean Sea. Shear wave splitting studies by Russo et al. (1996) support the notion of Pacific mantle outflow through this region. Likewise, Heintz et al. (2005) deduced along-strike variation at the western margin of South America with lowered upper mantle seismic velocities corresponding to the intersection of the Carnegie and Chile ridges with the continent. They argued for a slab window in this region and a weakness in the subducted plate which might accommodate asthenospheric mantle transfer from the Pacific to the Atlantic region. A similar slab window was proposed by Hole et al. (1994) for the Antarctic Peninsula region.

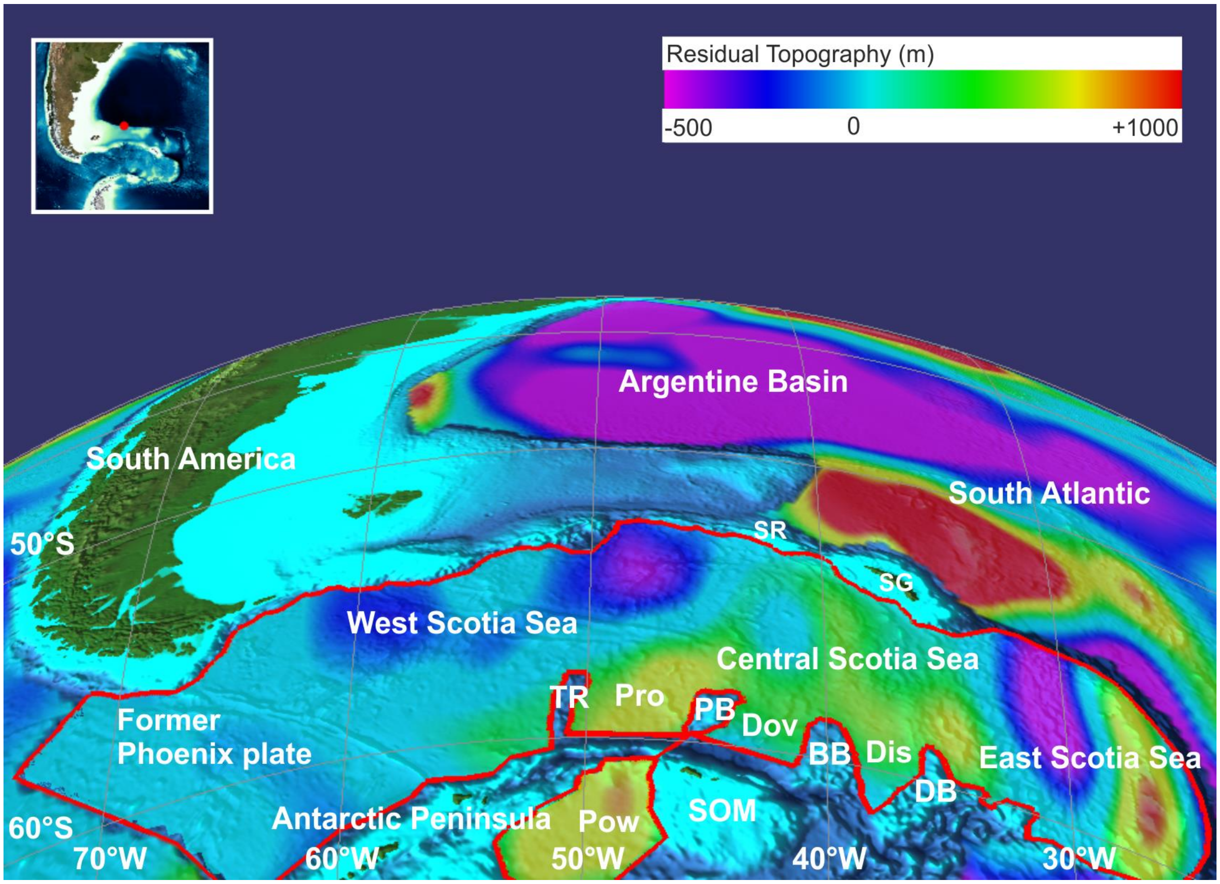

Figure 8: Northeast view from Antarctica. Note the large residual (dynamic) topography low (> 500 m Winterbourne et al., 2009) in the Argentine Basin (based on a half-space cooling model (Turcotte and Oxburgh, 1967).

A far more likely source to supply the growing South Atlantic mantle reservoir, however, is located in the deep mantle beneath Africa. Courtillot et al. (2003) identified a number of primary hotspots in this region and several strong mantle plumes have been imaged in the Sub-Atlantic mantle in recent seismic tomography models that account for finite frequency effects (Montelli et al., 2004). The so-called African superplume, which up-lifts much of the southeastern Atlantic Ocean and parts of the African continent by as much as ~ 1km (Lithgow-Bertelloni and Silver, 1998) in the so-called African superswell (Nyblade and Robinson, 1994), is particularly well known. Geodynamic studies favor a substantial hotspot contribution to the mantle energy budget (Bunge et al., 2001). Also, a significant role of plumes in general mantle circulation would adequately explain the nature of the prominent seismic low velocity anomaly harbored in the deep mantle beneath Africa (Schuberth et al., 2009 a, b). The elevated topography of the African superswell contrasts with significant negative dynamic topography of up to ~ 1 km inferred for the conjugate South American margin in the Argentine Basin (Fig. 8) (e.g., Winterbourne et al., 2009). The remarkable dynamic topography gradient across the South Atlantic is consistent with westward flow emanating from the African superplume, as suggested by Behn et al. (2004) and Husson et al. (2012), rather than eastward flow through the Drake Passage. Equally important is the fact that the magnitude of the basal shear tractions associated with westward fluxing subatlantic asthenosphere may be sufficient to balance the budget of driving and resisting forces acting on the South American plate (Iaffaldano and Bunge, 2009). Thus a variety of evidence suggests that mass balance between the Pacific and Atlantic mantle domains is achieved through deep mantle processes rather than shallow upper mantle flux in the asthenosphere.

We close by noting that our regional dynamic topography results confirm earlier results by Husson (2006), which show a slab related negative dynamic topography signal in the East Scotia Sea, while the South Sandwich island arc accounts for the positive signal in the easternmost Scotia Sea. The larger amplitude inferred by Husson (2006) relative to our results is probably due to three reasons: firstly, our dynamic topography results are band-pass filtered, as noted before; secondly we adopt an earlier spreading onset (15 Ma) based on identified magnetic lineations (Larter et al., 2003) compared to Barker (2001); thirdly, we have used a higher resolution sediment thickness grid.

In summary, our dynamic topography results for the Scotia Sea provide no indication for large-scale asthenosphere flux from the Pacific into the Atlantic through the Drake Passage, consistent with previous findings from seismology and geochemistry, and we suggest that mass balance between the Pacific and Atlantic mantle domains must be achieved through deep mantle related plume processes rather than upper mantle flux.

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Chapter 2

Reconstructing the link between the Galapagos hotspot and the Caribbean Plateau

This chapter has been published in GeoResJ:

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Abstract

Most authors agree that parts of the Caribbean plate are an igneous Plateau underlain by Farallon lithosphere that was trapped in between the North and South American plates. However, the origin of the thickened crust is debated. The theory of oceanic plateaus forming as magmatic outpouring related to a plume arrival became prominent when Large Igneous Provinces could be traced back to hotspots. The present-day proximity of the Galapagos hotspot made it an obvious candidate for associating its plume head arrival with the formation of the Caribbean Plateau. However, it was shown that in a fixed or moving Indian-Atlantic hotspot reference frame, plate reconstructions predicted the Galapagos hotspot a thousand or more kilometres away from the Caribbean plate at the time of Plateau formation (~88−94 Ma). Here, we calculate the goodness of fit for the Pacific hotspot reference frame and the recently developed Global Moving Hotspot Reference Frame. We show that both frames lead to good correlations between the paleo-positions of the Caribbean Plate and the Galapagos hotspot, when a docking time of the Caribbean plate to South America of 54.5 Ma is assumed. As this result is consistent with abundant evidence that lends support for a Galapagos hotspot origin of the rocks that form the Caribbean Plateau, proposed alternative mechanisms to explain the thickened crust of the Caribbean Plateau seem to be unnecessary. Finally, based on our model, we also derived an age distribution of the lithosphere underneath the thickened crust of the Caribbean Plateau.

2.1 Introduction

The Caribbean Sea is a complex tectonic system with numerous components. Because of the lack of sea-floor anomalies, the origin of these components is widely debated. The largest and perhaps most controversial component is the Caribbean Plateau that makes up the bulk of the Caribbean seafloor (Fig. 1). This Plateau is a large igneous province of up to 20 km thick crust (Mauffret et al., 2001) and it is predominantly between 88 and 94 Ma old (Fig. 1) (Kerr et al., 2009; Hastie et al., 2012). The mechanism and place of formation of this Plateau is still the focus of debates. While the Caribbean Plateau is often referred to as the Caribbean Large Igneous Province (CLIP) or the Caribbean-Colombian Plateau, these terms also refer to regions that have been accreted onshore (Mamberti et al., 2003). As such, we will use the term Caribbean Plateau to focus on the submarine section of the Plateau.

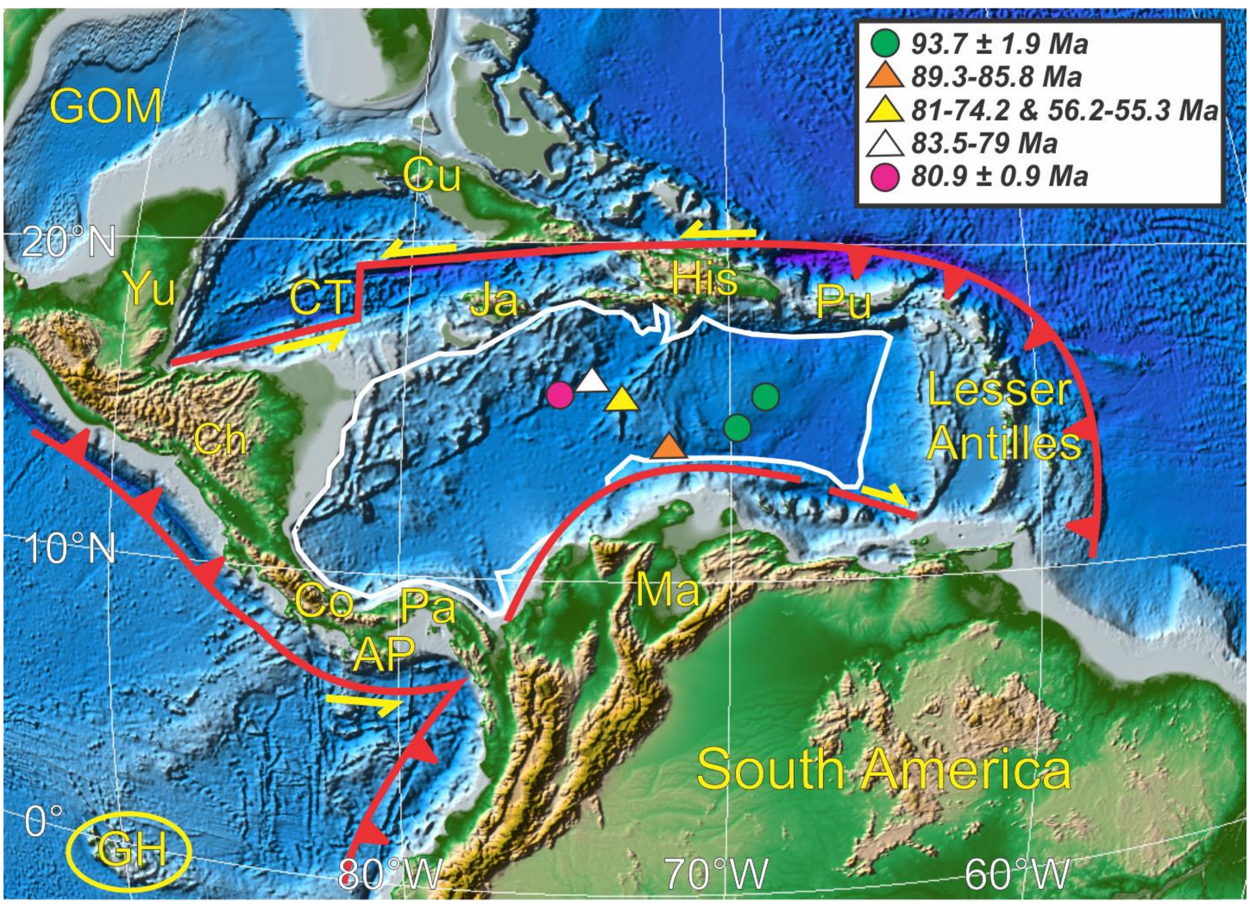

Figure 1: Overview of the Caribbean Sea. Plate boundaries are indicated in red. At present-day the northern plate boundary is characterized by transtension and the southern boundary by a complicated transpressional regime (van der Lelij et al., 20010). Subduction zones along the Lesser Antilles arc and Central America form the plate boundaries on the western and eastern edges of the region, respectively. White outline shows the extent of the Caribbean Plateau (Mauffret et al., 2001). Circles (radiometrically dated) and triangles (not possible to be radiometrically dated) indicate dredge sample locations. Sample ages are given in insert box (Kerr et al., 2009). Abbreviations: Galapagos hotspot (GH), Maracaibo Block (Ma), Panama (Pa), Azuero Peninsula (AP), Costa Rica (Co), Chortis Block (Ch), Yucatan Block (Yu), Jamaica (Ja), Cuba (Cu), Hispaniola (His), Puerto Rico (Pu), Cayman Trough (CT), Gulf of Mexico (GOM).

Agreement exists that the Caribbean Plateau was built on top of a piece of former Farallon lithosphere that was trapped in between North and South America (Duncan and Harggraves, 1984; Hastie et al., 2013). To accommodate the Farallon lithosphere between the Americas, westward subduction initiated in either the Lower (Pindell et al., 2009; Pindell et al., 2012) or Upper Cretaceous (Burke, 1988; Hastie et al., 2013) at a transform boundary (Pindell et al., 2012) or alternatively subsequent to polarity reversal (Hastie et al., 2013; Pindell et al., 2009). Subduction initiated either as a response to the collision of the Caribbean Plateau with the Great Antillean Arc (Burke et al., 1978; Hastie and Kerr, 2010; van der Lelij et al., 2010) or possibly because of a change in the spreading rate of the Mid-Atlantic Ridge (Pindell et bal., 2009; Pindell et al., 2012).

Disagreement remains on the causal mechanism for the thickened crust of the Caribbean Plateau. One theory invokes that the Caribbean Plateau formed above the rising plume head of the paleo-Galapagos hotspot and is due to the associated volcanic outbursts (Burke et al., 1978; Duncan and Hargraves, 1984), analogous to the creation of the Ontong-Java Plateau above the Louisville hotspot (Duncan and Richards, 1991). This is supported by geochemical studies of the

Caribbean Plateau, which suggest an impact of a plume head at the base of the lithosphere Hauff et al., 1997; Geldmacher et al., 2003; Kerr et al., 2003; Thompson et al., 2003). It was also noted that only plume melts have the required temperature to generate the lavas on the Caribbean Plateau (Herzberg and Gazel, 2009; Hastie and Kerr, 2010). Furthermore, the basalts from the Galapagos hotspot have a deep mantle signature combined with other complexities associated with depleted mantle and recycled oceanic crust consistent with the Caribbean Plateau basalts (Sallarès et al. 2005). Finally, the short duration of the bulk formation of the Caribbean Plateau suggests plume-induced melts (Kerr et al., 2009).

Nevertheless, Pindell et al. (2006) have challenged this theory for two reasons: the first being the patchy record of accreted hotspot tracks between 21.2 Ma and 51.9 Ma, probably lost due to subduction under Central and South America with only one sample from the southern Azuero Peninsula (Fig. 1) leading to an age of 32.8 ± 0.5 Ma in this range (Hoernle et al., 2002); and secondly, because of the thousand kilometres between the Galapagos hotspot and the reconstructed position of the Caribbean Plateau at the time of formation predicted by reconstruction models that are based on fixed or moving Indian-Atlantic hotspot reference frames (Pindell et al., 2006; Seton et al., 2012). These difficulties have led Pindell et al. (2006) to propose a model in which the Atlantic asthenosphere upwelled through a remnant slab window caused by westward subduction of the proto-Caribbean spreading centre.

It was shown, however, that the basalts from the Plateau have radiogenic isotopes and trace-element ratios showing similarity to deep mantle sources that are inconsistent with ambient shallow mantle upwellings, such as beneath a slab-window (Kerr et al., 2009). Additionally, Pindell and Kennan (2009) demonstrated that when using a fixed Pacific hotspot reference frame (Wessel and Kroencke, 2008), a good fit between the paleao-Galapagos hotspot and the reconstructed position of the Caribbean Plateau at the appropriate time of Plateau formation (~94 -88 Ma) can be reached.

Thus, as the available geochemical evidence points to the rising plume head of the paleo-Galapagos hotspot as the source of the Caribbean Plateau rocks, we revisit Pindell and Kennan's (2009) observation here and develop a new reconstruction model using two recent reference frames for the Pacific hotspots: Wessel and Kroenke(2008) and Doubrovine et al. (2012). The necessary assumptions to reach a good correlation between the paleao-Galapagos hotspot and the Caribbean Plateau between ~94 and 88 Ma are outlined in section 2.2 and 2.3. In addition, we show how these reconstructions can be used to derive an age-grid for the former Farallon lithosphere underneath the Caribbean Plateau in section 2.4. In section 2.5, we address the regional tomography. Finally, in section 2.6 we discuss our findings, which are summarized again in the conclusions (section 2.7).

2.2 Reconstructing the Caribbean

To calculate the relative positions in our kinematic model, we used 4DPlates (Clark et al., 2012) and assumed that the Caribbean Plateau has a Pacific origin, being fixed to the Farallon plate and at some point 'docked' between the Americas. Here, docked refers to the point in time when the Plateau detached from the Farallon plate, and was wedged between North and South America and as a result became fixed to South America. Prior to docking, the following plate circuit was used: Caribbean Plateau – Farallon Plate – Pacific Plate – Antarctica – Africa, with South and North America moving relative to Africa. For ages older than 83.5 Ma, the Pacific Plate – Antarctica circuit is broken (Seton et al., 2012) and we reconstructed the Americas and associated blocks assuming that their position is given by the moving Atlantic-Indian Hotspot frame (O'Neill et al., 2005). In order to reconnect the circuit, we assumed that there is no relative motion between this frame and Pacific hotspot reference frame of Wessel and Kroenke (2008), the latter giving the relative positions of the Pacific plate to the frame. For times older than 100 Ma, we use the paleo-magnetically-derived true polar wander corrected reference frame of Steinberger and Torsvik (2008). This reference frame is necessary for the age-grid derivation (see section 2.4), but does not affect the correlation between the Caribbean Plateau and the Galapagos hotspot (see section 2.3).

For the various blocks in the Caribbean (Fig. 1), we based our Euler rotations on those provided by Ross and Scotese (1988) with the few exceptions noted below. Firstly, unconformities underlying volcanics on some of the islands surrounding the Caribbean plate (Escuder Viruete et al., 2008; Wright and Wyld, 2011) are indicative for a subduction polarity reversal that supposedly occurred sometime in the Late Cretaceous after the present-day Caribbean Plateau had collided with the Great Antillean Arc (Hastie and Kerr, 2010). Ross and Scotese (1988) model this event to occur at 100 Ma. However, we held the Great Antillean Arc (restricted to Cuba, Hispaniola, Jamaica and Puerto Rico) fixed with respect to North America until 86.5 Ma: implying north-eastward subduction of former Farallon lithosphere along the western boarder of the arc until this time, consistent with recent suggestions of Hastie et al. (2013). The subduction reversal is followed by the consumption of the Proto-Caribbean seafloor by the retreating subduction zone to the north-east as well as transform-fault motion between the edges of the arc and North and South America.

To the west of the Plateau, the reconstruction of the Chortis Block remains debated (Moran-Zenteno et al., 2009). The rotation poles of Ross and Scotese (1988) were used, placing the Block adjacent to North America. However, the Chortis Block could have equally been treated as part of the Great Antillean Arc with a paleo-position further to the west and adjacent to the Great Antillean Arc (Moran-Zenteno et al., 2009). In line with previous suggestions (Leroy et al., 2000 Rojas-Agramonte et al., 2008; Stanek et al., 2009), the collision of Cuba with the Bahamas platform was set to be in the Early Eocene (51 Ma), followed by the onset of seafloor spreading in the Cayman Trough (49 Ma – present-day). Our complete reconstruction of the Caribbean Sea is presented in Fig. 2.

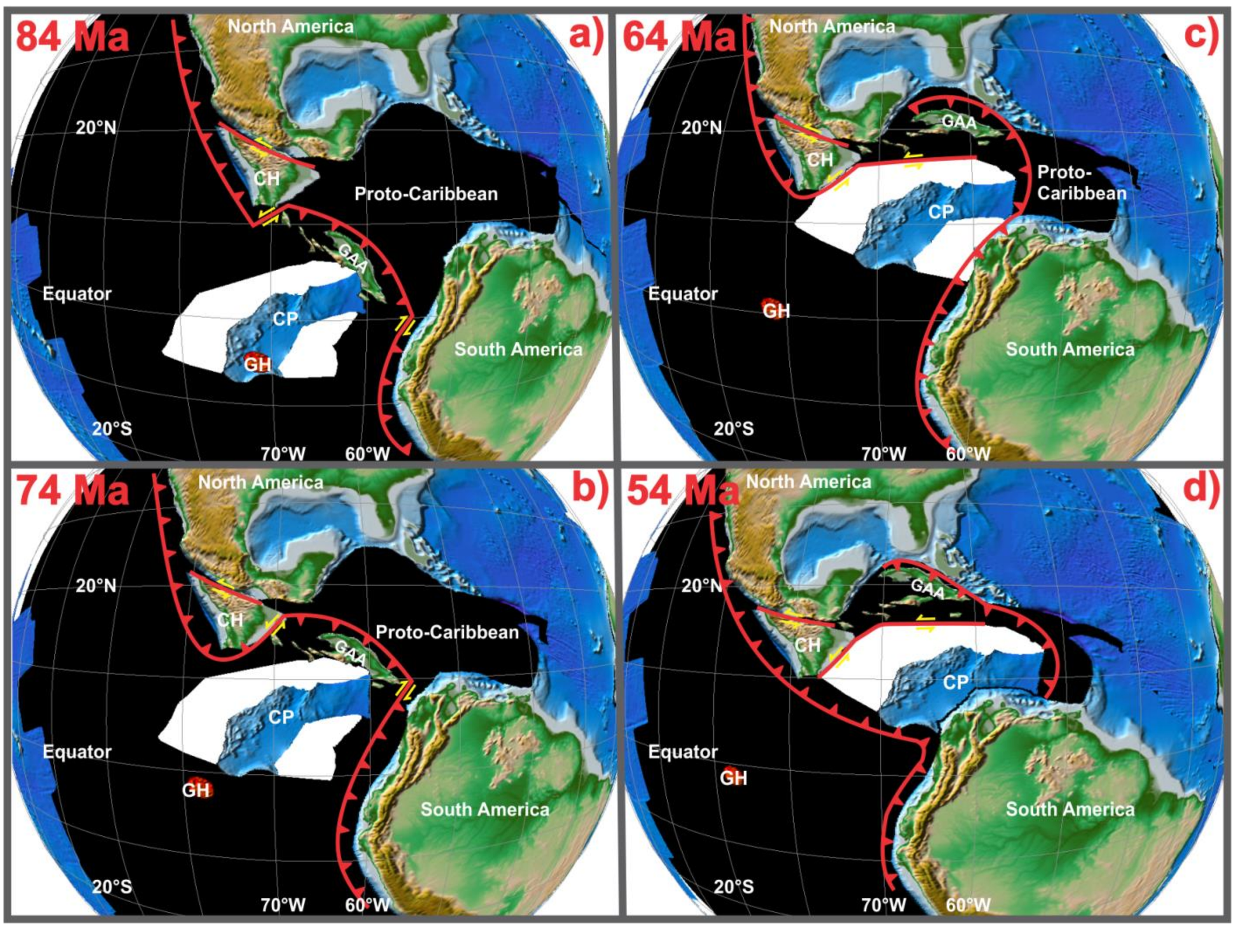

Figure 2: Caribbean plate reconstructions, in which the Caribbean Plate (CP) is composed of the Caribbean Plateau (shaded relief) and associated parts of the Farallon plate (in white) that have either been subducted or obducted. The reference frame of Wessel and Kroencke (2008) is used in this model. Topography is shaded by present-day relief (black indicates seafloor that has since subducted). Abbreviations: Great Antillean Arc (GAA), Chortis Block (CH), Galapagos hotspot (GH). (a) 84 Ma: the Farallon lithosphere crosses the paleo-

Galapagos hotspot at the south-western edge of the present-day Caribbean Plateau. The proto-Caribbean seafloor is being subducted by a north-eastward retreating subduction zone. Farallon/Phoenix lithosphere is subducted beneath western North and South America, respectively, and continues to do so in the following time slices. Transform motion occurs between the Chortis Block and North America (see section 2 for a possible alternative suggested by Moran-Zenteno et al. (2009)). (b) 74 Ma: the Caribbean Plateau resides on top of the Farallon plate and moves away from the paleo-Galapagos hotspot. Further north-eastward retreat of the Great Antillean Arc as well as some northward subduction under the Yucatan block occurs. Transform motion took place along the north-western South American margin. (c) 64 Ma: the Farallon/Caribbean Plate was moving mostly eastward implying transform motion along the south-eastern edge of the plate and northern South America, consistent with unconformities identified on the islands offshore South America (Wright and Wyld, 2011). Further subduction retreat of the Great Antillean Arc. (d) 54 Ma: the Caribbean Plateau was fixed to the bulk of South America and subduction initiated across the northern margin in response to convergence between North and South America. Also, transform motion and compression between north-western South America (Maracaibo Block) likely occurred since then, as the block is characterized by an independently moving complicated strike slip regime. It is suggested to be the source of the present-day transpressional regime (van der Lelij et al., 2010).

2.3 Correlating the positions of the Galapagos hotspot and the Caribbean Plateau

The major uncertainty we want to test is the relative positions of the Caribbean Plateau and Galapagos hotspot at the time of former's formation. In order to do so, we varied the 'docking' of the Caribbean Plateau between North and South America in two reference frames. For the Galapagos hotspot we assumed either a Pacific fixed hotspot reference frame (Wessel and Kroencke, 2008) or the Global Moving Hotspot Reference Frame of Doubrovine et al. (2012). In the former case, as explained in the Caribbean Reconstruction section above, we fixed the moving Indian-Atlantic hotspot frame to the Pacific one for times older than 83.5 Ma. In the latter case, we used the absolute positions for Africa and the Pacific. For the Pacific rotations, the stage rotations of Wessel and Kroenke (2008) were applied from 90 Ma. The extent of the Caribbean Plateau is based on that of Mauffret et al. (2001).

In both reference frames, the Plateau's relative position to the Galapagos hotspot between 94 and 88 Ma is very sensitive to changes in the timing of the docking age as shown in Table 1. Very young docking ages place the Plateau thousands of kilometres away from the Galapagos hotspot, as others have reported (Pindell et al., 2006; Seton et al., 2012). However, for a narrow range of docking ages the Plateau crosses over the position of the Galapagos hotspot at ages closely matching those of the dated dredge samples shown in Fig. 1. For either model, docking ages between 50 Ma and 60 Ma gave reasonable fits for the geochemical data, with the Wessel and Kroenke (2008) model giving distances between 190 and 650 km while Doubrovine et al. (2012) gave distances between 340 and 925 km.

TABLE 1 | DISTANCE BETWEEN THE CARIBBEAN PLATEAU AND GALAPAGOS HOTSPOT AT 89 MA BASED ON DIFFERENT DOCKING TIMES AND REFERENCE FRAMES *

Docking Time (Ma)	Doubrovine et al. (2012) (km)	Wessel and Kroencke (2008) (km)
5.0	4865	5150
10.0	4565	4850
20.0	3600	3900
30.0	2540	2810
40.0	1600	1850
50.0	340	450
54.5	490	190
60.0	925	650
65.0	1280	990
70.0	1770	1500
75.0	2270	2000

*** The distance was measured between the orange triangle (sample dated between 89 and 85.8 Ma) from Figure 1 and the centre of the Galapagos hotspot polygon (red) as shown in Figure 4.**

Within this time block, there was a switch from divergence to convergence between North and South America at 54.5 Ma (Müller et al., 1999). Using 54.5 Ma as the docking age, the Plateau crosses the Galapagos hotspot at the presumed age of formation (88 – 94 Ma) for either reference frame. The virtual hotspot trail for the reference frames of both Wessel and Kroenke (2008) and Doubrovine et al. (2012) using this docking age are shown in Fig. 3 and Fig. 4, respectively, while the rotation parameters are presented in Table 2. Both reference frames lead to an age progression of increasing ages west to east as well as locating the Plateau above the hotspot for the reported sample ages between about 81 Ma and 94 Ma. Note, however, that the massive basaltic flows thought to make up the Caribbean Plateau (Kerr et al., 2009) would not be point-source eruptions (as it appears from the small number of collected samples) but would cover large areas, making a precise correlation difficult.

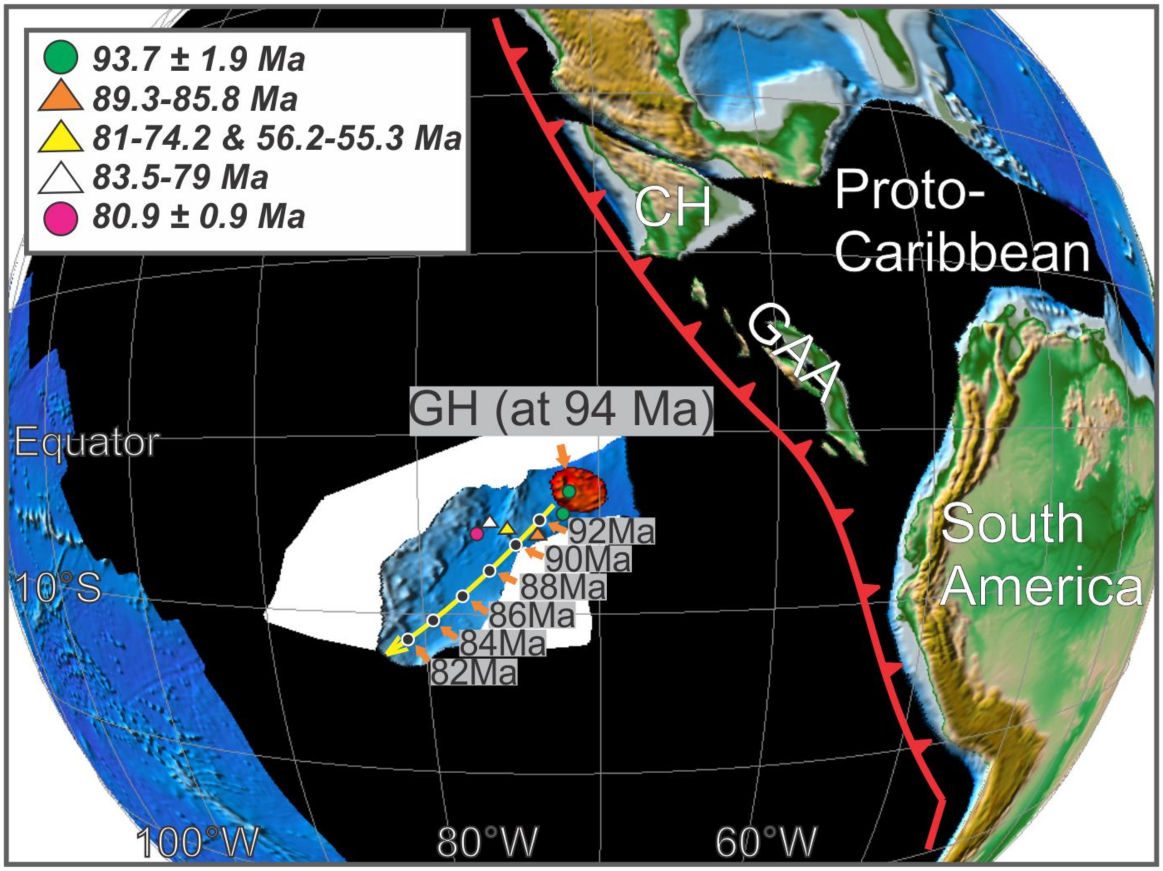

Figure 3: Coincidence of the Galapagos hotspot (red) and the Caribbean Plateau-polygon at 94 Ma, assuming a fixed Pacific mantle reference frame (Wessel and Kroencke, 2008) for the Galapagos hotspot. The white shaded area refers to the portion of Farallon lithosphere that was subducted after it has entered the opening gap between North and South America (see Fig. 2). The yellow arrow and associated points indicate the motion path and timing of the Galapagos hotspot relative to the Caribbean plate (present-day bathymetry is shown). Circles and triangle indicate the same sample datings as in Fig. 1.

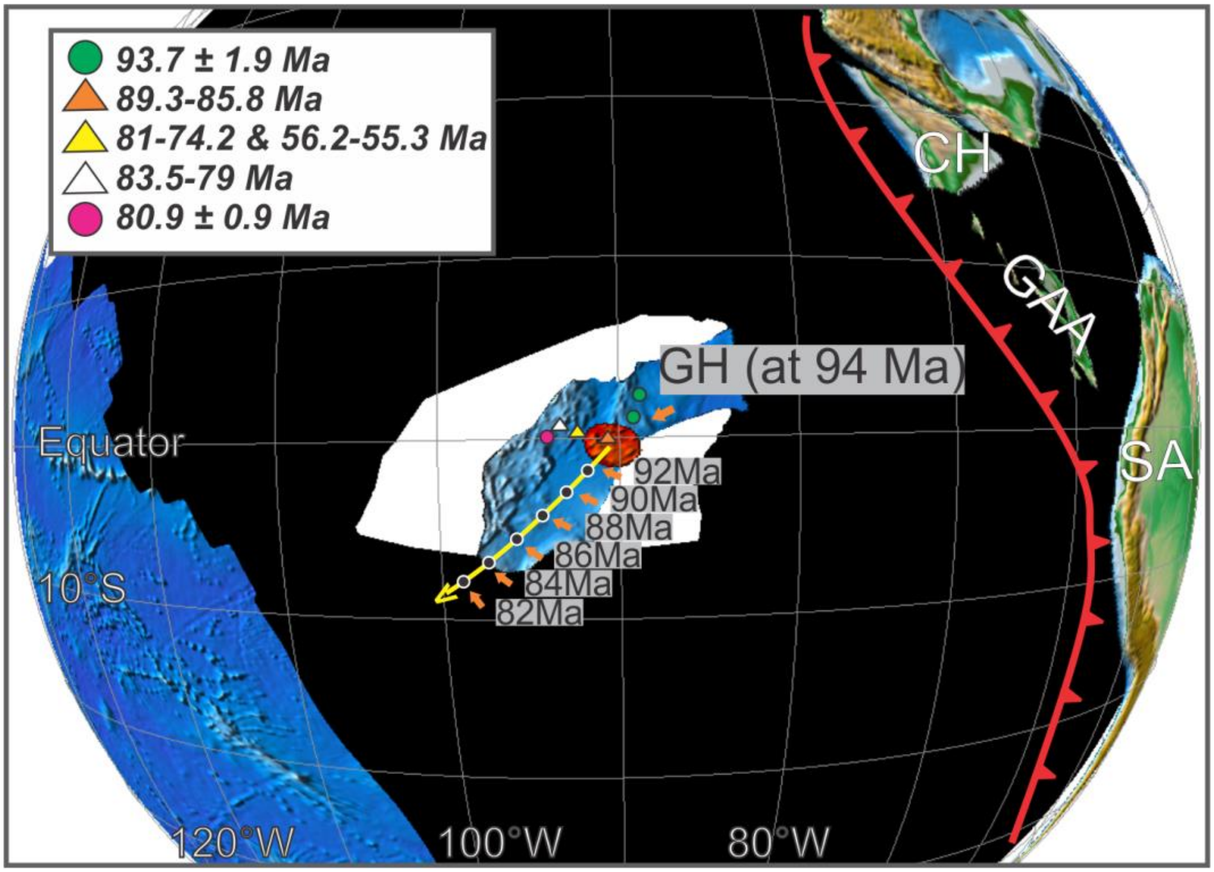

Figure 4: Coincidence of the Galapagos hotspot (red) and the Caribbean Plateau-polygon at 94 Ma, using the global moving hotspot reference frame of Doubrovine et al. (2012). Labels are as in Fig. 2. Note the shift in longitude as a consequence of this reference frame relative to the fixed Pacific hotspot reference frame of Wessel and Kroencke (2008). Note also that the general agreement between the locations of the Caribbean Plateau and the Galapagos hotspot is very similar in both explored reference frames (compare to Fig. 3).

TABLE 2 | ROTATION POLES FOR THE CARIBBEAN PLATEAU-POLYGON IN

Time (Ma)	Finite rotation			Fixed plate (plate ID)
	latitude (°)	longitude (°)	angle (°)	
Based on a fixed Pacific hotspot reference frame (Wessel and Kroencke, 2008)				
0.0	0.0	0.0	0.0	South America (201)
54.5	0.0	0.0	0.0	South America (201)
54.5	67.6642	144.355	50.6645	Farallon Plate (902)
105.0	67.6642	144.355	50.6645	Farallon Plate (902)
Based on a global moving hotspot reference frame (Doubrovine et al., 2012)				
0.0	0.0	0.0	0.0	South America (201)
54.5	0.0	0.0	0.0	South America (201)
54.5	68.75	142.37	52.43	Farallon Plate (902)
105.0	68.75	142.37	52.43	Farallon Plate (902)

2.4 Age-grid

To derive an age-grid based on the assumptions described in the previous section, we rotated the Caribbean Plate/Plateau-polygon to the location of the mid-ocean ridge that once formed the plate boundary between the Pacific and Farallon plates (Seton et al., 2012); and found it touches the ridge at ~105 Ma in either model, although we used the reference frame of Wessel and Kroenke (2008) to be consistent with the reconstruction in Fig. 2. Fig. 5 shows the reconstructed polygon with the paleo-age grids of Müller et al., (2008a). The age distribution covered by the rotated polygon was extracted and then rotated back to the present-day position, adding the intervening 105 million years to the grid (Fig. 6). According to our model, the oldest parts underneath the Caribbean Plateau are located in the east and are approximately 144 Ma old.

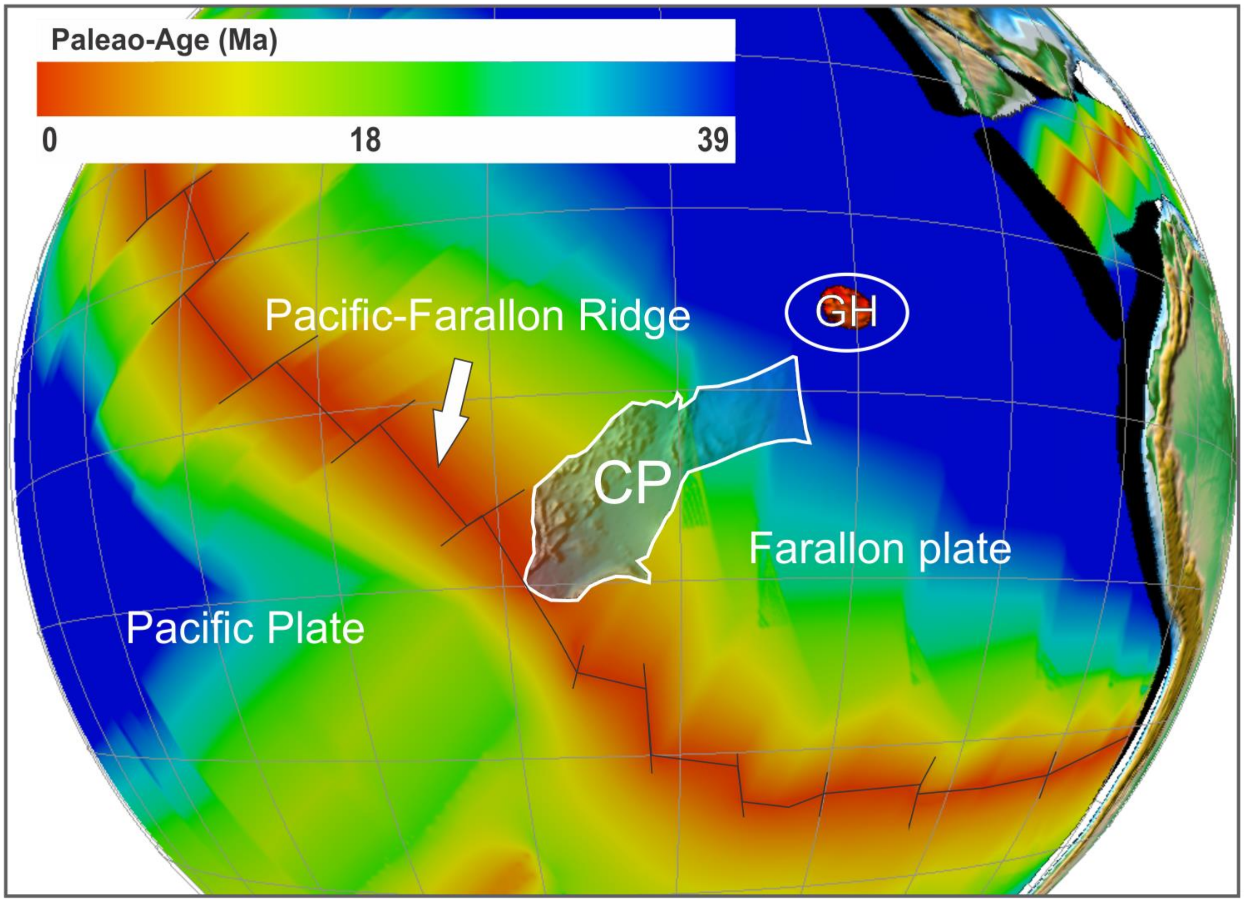

Figure 5: The paleo-age-grids of Müller et al. (2008a) were used to extract the age distribution covered by the rotated Caribbean Plate/Plateau polygon to the former Pacific-Farallon ridge. Based on the reconstruction model presented in Fig. 2, the Caribbean Plate/Plateau polygon touches the ridge 105 Ma ago. Note that the absolute positions are given here by a paleo-magnetically derived true polar wander corrected reference frame (Steinberger and Torsvik, 2008).

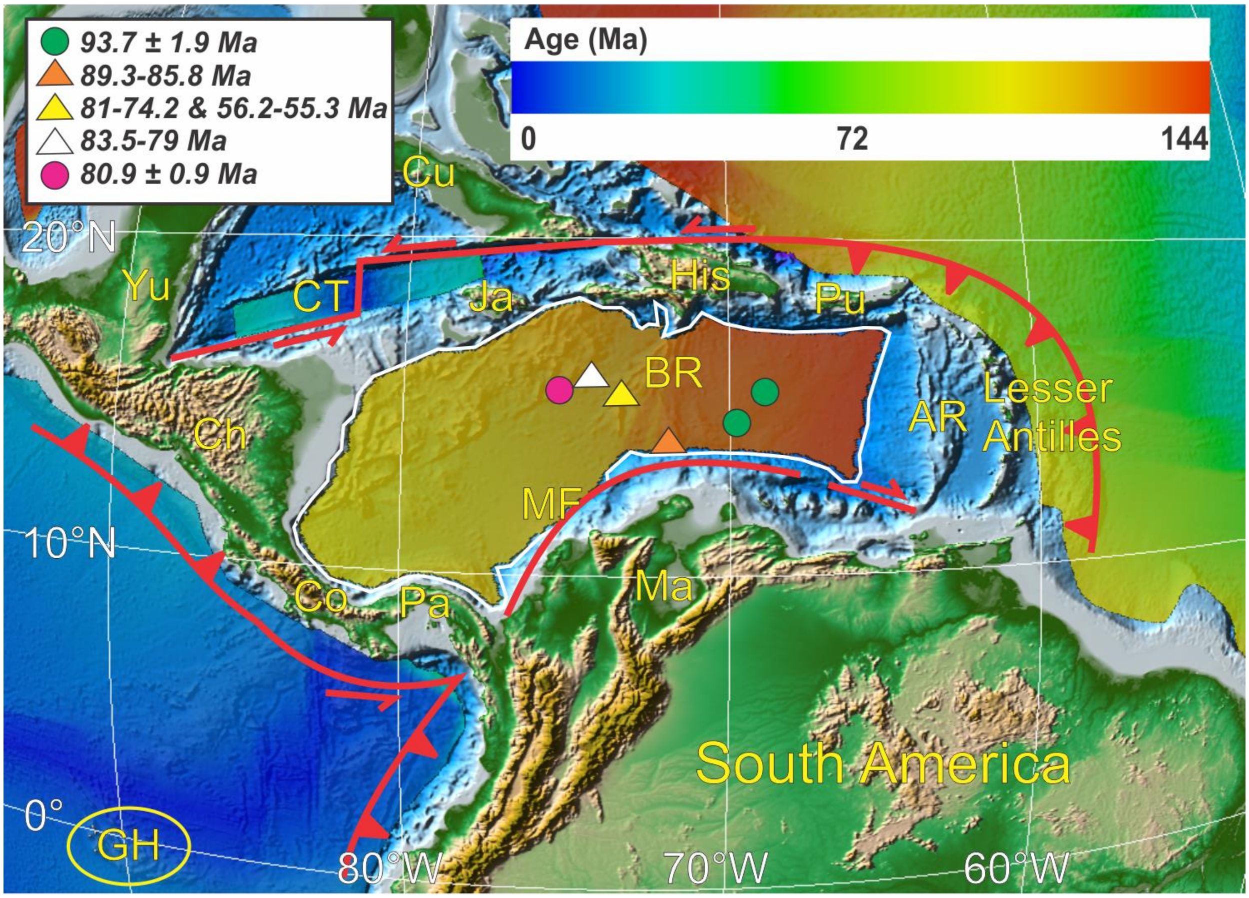

Figure 6: Present-day age distribution of the former Farallon lithosphere underneath the Caribbean Plateau, based on paleo-age-grids of Müller et al. (2008a) (see text for a discussion and Fig. 5). For the Pacific, Cayman Trough, Gulf of Mexico and the Atlantic, the age distribution according to the age-grids of Müller et al. (2008b) is shown. Note that both reference frames tested in this study lead to the same age-distribution, as it is based on the relative motion of the Farallon-Pacific Plates only.

2.5 Seismic Tomography

Duncan and Richards (1991) noted the concentration of hotspots in the area of high geoid residuals. With the advent of high resolution seismic tomography, the underlying anomalies or Large Low Shear Velocity Provinces (LLSVPs) at the base of the mantle could be correlated with these hotspot locations (Throne et al., 2004). In both cases, the Galapagos hotspot lies on the edge of the eastern most extent of the Pacific geoid anomaly (Duncan and Richards, 1991) and the Pacific LLSVP (Fig. 7, insert). With vertical mantle transit times of 150 million years (Bunge et al., 1998), it is likely that the arrival of the Galapagos plume ~88-94 million years ago would still be seen in mantle tomography. In Fig. 7, a large upwelling can be observed tilting from the eastern extent of the Pacific LLSVP to the northeast and continuing upwards to the base of the Galapagos hotspot (purple arrow in Fig. 7).

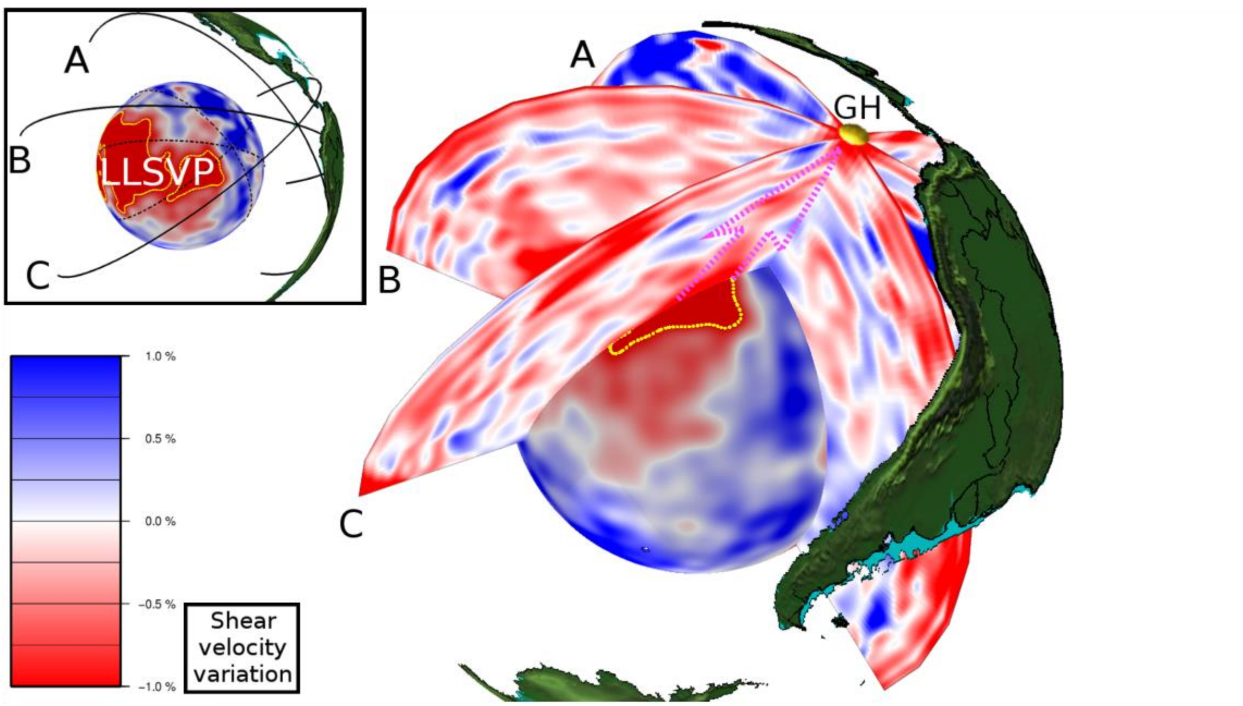

Figure 7: Shear-wave velocity anomalies from the S40RTS model (Ritsema et al., 2010) at 2800 km depth with three cross-sections (A, B and C) showing shear-wave velocity variations. Insert shows the 2800 km anomalies with the traces of the top (solid black line) and bottom (dashed black line) of the cross-sections for reference. The -1.0% anomaly at 2800 km depth is highlighted by a dashed yellow line and indicates the Large Low Shear Velocity Province (LLSVP) underneath the Pacific. The dashed purple line indicates a tilted anomaly from the LLSVP to underneath Galapagos hotspot (GH) located at the yellow dot. National boarders and topography (down to 500 m depth) are shown for reference.

2.6 Discussion

The relative distance between the Galapagos hotspot and the Caribbean Plateau at the time of large igneous province formation is largely dependent on the docking age of the Caribbean Plateau, as shown in Table 1, with 54.5 Ma giving a relatively small offset as well as coinciding with the change in relative motion between the Americas. However, the idea of Pindell and Kennan (2009) that the relative drift between the Indian-Atlantic and Pacific hotspot reference frames might be responsible for the apparent mismatch in locations is also significant. Here, we used the recent Global Moving Hotspot Reference Frame (GMHRF) of Doubrovine et al. (2012) as well as the Pacific hotspot frame (Wessel and Kroencke, 2008). The latter frame gives a better fit of the position of the Galapagos hotspot relative to the Caribbean Plateau for our docking age of 54.5 Ma. While we could tune the docking age to improve the fit of the GMHRF, the derivation of this timing from the initiation of convergence between the Americas gives an external constrain to the reconstruction.

Using two reference frames reduces the uncertainty of our conclusions, although such uncertainties remain because of the uncertainties in Wessel and Kroenke (2008) for ages > 90 Ma and in the GMHRF for ages between 50 and 80 Ma (Doubrovine et al., 2012). In addition, our paper does not prove a causal relationship between the Galapagos hotspot and the Caribbean Plateau, but removes one of the main plate kinematic objections to such a theory.

Finally, recent debate has focused on whether the correlation between the LLSVP edges and hotspot locations mean that deep mantle upwellings are tilted from the centre of the LLSVPs or occur at the edges and proceed straight up in the mantle (Throne et al., 2004; Steinberger and Torsvik, 2012). In the mantle tomography presented here, there appears to be a tilted upwelling from the centre of the eastern extent of the Pacific LLSVP, agreeing with the former suggestion.

2.7 Conclusion

We have shown that a good correlation between the relative positions of the Galapagos hotspot and the Caribbean Plate at the time of Caribbean Plateau formation (main phase: 88 − 94 Ma) can be reached in the Pacific (Wessel and Kroencke, 2008) as well as in the Global Moving Hotspot Reference Frame (Doubrovine et al., 2012). The docking time of the Caribbean Plateau to South America proved to be critical in this regard and best fits were reached when a timing of 54.5 Ma was assumed. Interestingly, this timing is coincident with the switch from divergence to convergence between the Americas, possibly implying some long-wavelength consequences. Our findings are consistent with abundant geochemical evidence. Therefore, alternative mechanisms to explain the formation of the Caribbean Plateau, such as asthenosphere inflow through slab windows, seem to be unnecessary.

Finally, our model enabled us to derive an age-grid of the former Farallon lithosphere underneath the Caribbean Plateau. This grid will be useful for various geodynamic investigations, including the calculation of the regional present-day dynamic topography.

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Chapter 3

An outlet for Pacific mantle: The Caribbean Sea?

This chapter has been submitted to GeoResJ.

Abstract

The Pacific Ocean is surrounded by subduction zone systems leading to a decreasing surface area as well as sub-surface mantle domain. In contrast, the Atlantic realm is characterized by passive margins and growing in size. To maintain global mass balance, the Caribbean and the Scotia Sea have been proposed as Pacific-to-Atlantic transfer channels for sub-lithospheric shallow mantle. We concentrate on the Caribbean here and test this idea by calculating the present-day regional dynamic topography in search for a gradual decrease from west to east that mirrors the pressure gradient due to the shrinkage of the Pacific. To calculate the dynamic topography, we isostatically correct the observed topography for sediments and crustal thickness variations, and compare the result with those predicted by lithospheric cooling models. The required age-grid was derived from our recently published reconstruction model. Our results confirm previous geochemical and shear-wave splitting studies and suggest some lateral asthenosphere flow away from the Galapagos hotspot. However, they also indicate that this flow is blocked in the Central Caribbean. This observation suggests that rather than through large scale Pacific-to-Atlantic shallow mantle flow, the global mass balance is maintained through some other process, possibly related to the deep mantle underneath Africa.

3.1 Introduction

The conception of a relatively mobile, mostly laterally flowing asthenosphere between 100 and 400 km depth, on which tectonic plates glide easily, has been accepted for a long time (Chase, 1979). Several lines of evidence are consistent with this view and comprise isostatic considerations of mountain belts (see Watts, 2001 for a review), geoid and postglacial rebound studies (Hager and Richards, 1989; Mitrovica, 1996), seismic surveys (Grand and Helmberger, 1984; Debayle et al., 2005) as well as mineralogical investigations (Stixrude and Lithgow-Bertelloni, 2005). Also, numerical mantle convection models (Bunge et al., 1996) and analytical fluid dynamic calculations (Busse et al., 2006) are consistent with a relatively weak upper mantle. In fact, it has been proposed that an asthenosphere is essential in sustaining plate tectonics on Earth (Richards et al., 2001).

This understanding of the asthenosphere has great consequences for the general perception of mantle exchange processes between the Pacific, Atlantic and Indian Ocean mantle domains, respectively, which differ from each other in their isotope and trace element chemistry (Schiano et al., 1997). While the Pacific is shrinking, because subduction of Pacific lithosphere in the marginal subduction zones occurs faster than the creation of new lithosphere along the Pacific mid-ocean ridges, the Atlantic and Indian Oceans are growing as more lithosphere is being created than subducted. This poses the question where the underlying mantle reservoirs of the Indian and Atlantic Oceans are fed from, and where the material underneath the shrinking Pacific goes to.

To this end, Walter Alvarez (1982) suggested a concept of asthenosphere exchange between the three mantle domains as a mechanism to achieve global mass balance. He viewed the Caribbean, for which a slab gap in the Costa Rica/Panama region has been suggested

(Johnston and Thorkelson, 1997) (Fig. 1), and the Scotia Seas as gateways for Pacific-to-Atlantic asthenosphere flux. The Australian-Antarctic discordance was suggested for asthenosphere transfer between the Pacific and Indian Ocean mantle domains.

Alvarez (1982) viewed these outlets as necessary, because direct lateral asthenospheric mantle flow from the closing Pacific into the opening Atlantic/Indian mantle domains is thought to be blocked by sinking slabs and deep continental roots, known as the tectosphere (Jordan, 1978), everywhere else along the Pacific Plate margin. The flow is supposedly triggered by the shrinkage of the Pacific realm, which is envisaged to cause a pressure gradient driving the asthenosphere through these proposed outlets.

We have tested Alvarez's (1982) hypothesis for the Scotia Sea already (Nerlich et al. 2013), and focus in this paper on the Caribbean Sea (Fig. 1). We follow the conceptual view of Nerlich et al. (2013) and calculate the present-day dynamic topography assuming that it is indicative of asthenospheric mantle flow. If the asthenosphere flows from the Pacific mantle domain around the northern tip of South America into the Atlantic realm, we expect a gradual decrease in dynamic topography from the Pacific realm throughout the Caribbean Sea.

3.2 Geological Setting

At present day the northern plate boundary is characterised by transtension while the southern plate boundary by a complicated transpressional regime (van der Lelij et al., 2010). Subduction zones along the western Central American margin and Lesser Antilles arc form the plate boundaries on the western and eastern ends of the region, respectively (Fig. 1). In addition, there is a slab gap west of Panama and Costa Rica (Johnston and Thorkelson, 1997). The relatively few sample points from Caribbean basement (Fig. 1) indicate that it was formed between ~ 94 and 80 Ma ago.

Positive free-air gravity anomalies of up to 30 mGal are observed in the northwest of the Caribbean Plateau as well as above the Beata Ridge, while negative anomalies of up to 45 mGal are found at the western and eastern edges of the region (Fig. 1). Relatively thin sediment cover (<2 km) is present in the northeastern part of the region around the Beata Ridge (Fig. 2), whereas the Magdalena Fan (MF, Fig. 2) has up to 6 km of sediment cover. The thickest crustal (>20 km) regions are the western section of the Plateau as well as to the northwest of the Magdalena Fan (Fig. 3). The Magdalena Fan itself and the easternmost part of the Plateau have a lower crustal thickness of up to 12 km. The remaining parts of the Plateau are between 15 and 18 km thick, i.e. more than twice as thick as typical oceanic crust.

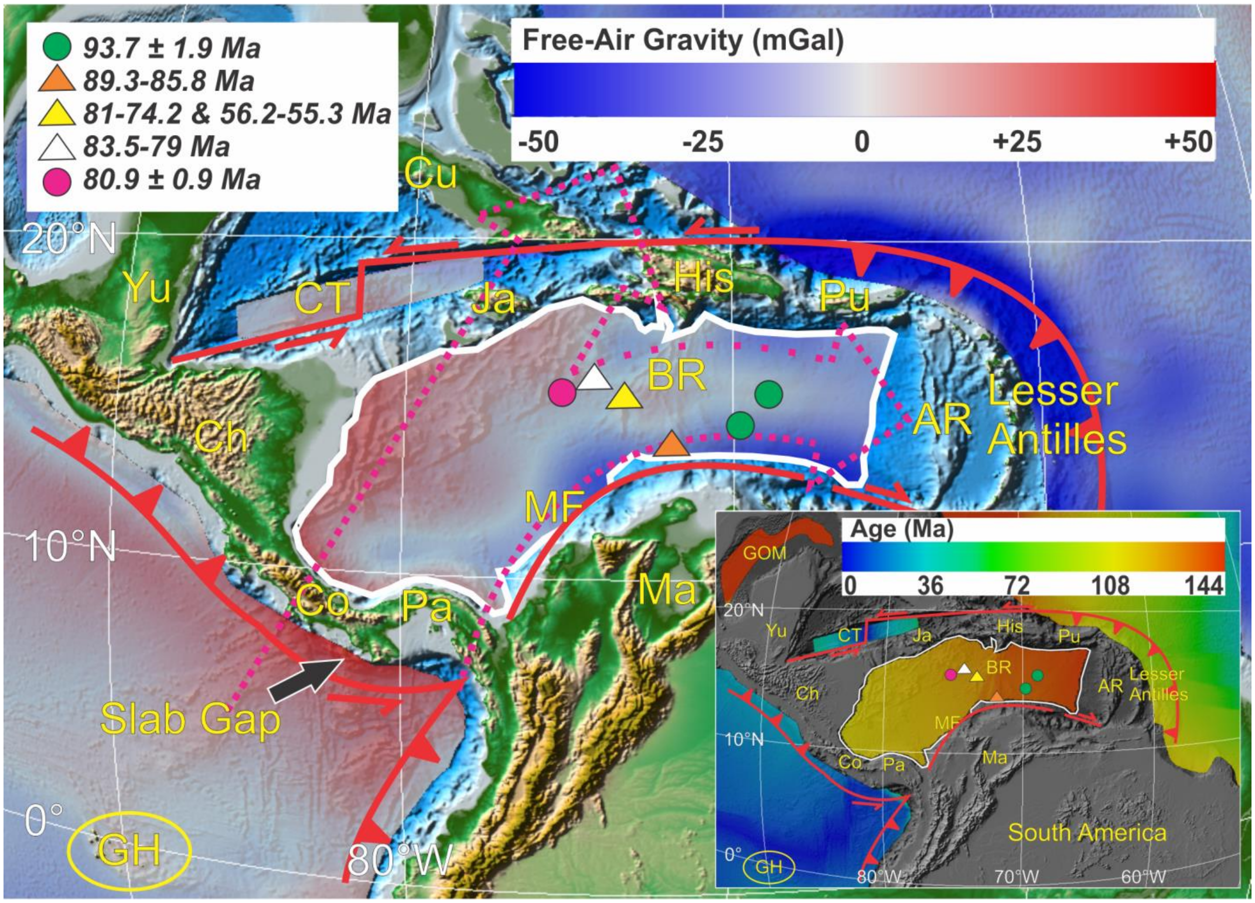

Figure 1: (<u>Main</u>) Overview of the Caribbean Sea with superimposed long-wavelength free-air gravity anomalies (low-pass filtered to 5 degrees) from Tapley et al., 2005 shown in areas with age-grid coverage (topography shown in other areas). (<u>Insert</u>) Regional age-grid from Nerlich et al. (2014), ranging in the Caribbean realm from 105 to 144 Ma. (<u>Both</u>) The white outline shows the region of the Caribbean large igneous plateau (Mauffret et al., 2001). Asthenosphere flow paths according to Alvarez (1982) are dotted in purple. The age of rock samples (Kerr et al., 2009, Hastie et al., 2013) drilled from basement in the region is indicated by dots (Ar/Ar dated) and triangles (dated by inclusions or overlaying strata). Insert on the top left hand side shows sample age ranges. Abbreviations are as follows: Galapagos hotspot (GH), Maracaibo Block (Ma), Panama (Pa), Costa Rica (Co), Chortis Block (Ch), Yucatan Block (Yu), Jamaica (Ja), Cuba (Cu), Hispaniola (His), Puerto Rico (Pu), Gulf of Mexico (GOM), Cayman Trough (CT), Aves Ridge (AR), Beata Ridge (BR), and Magdalena Fan (MF). Note, all figures but Figure 4 and 7 are presented using 4DPlates software (Clark et al., 2012).

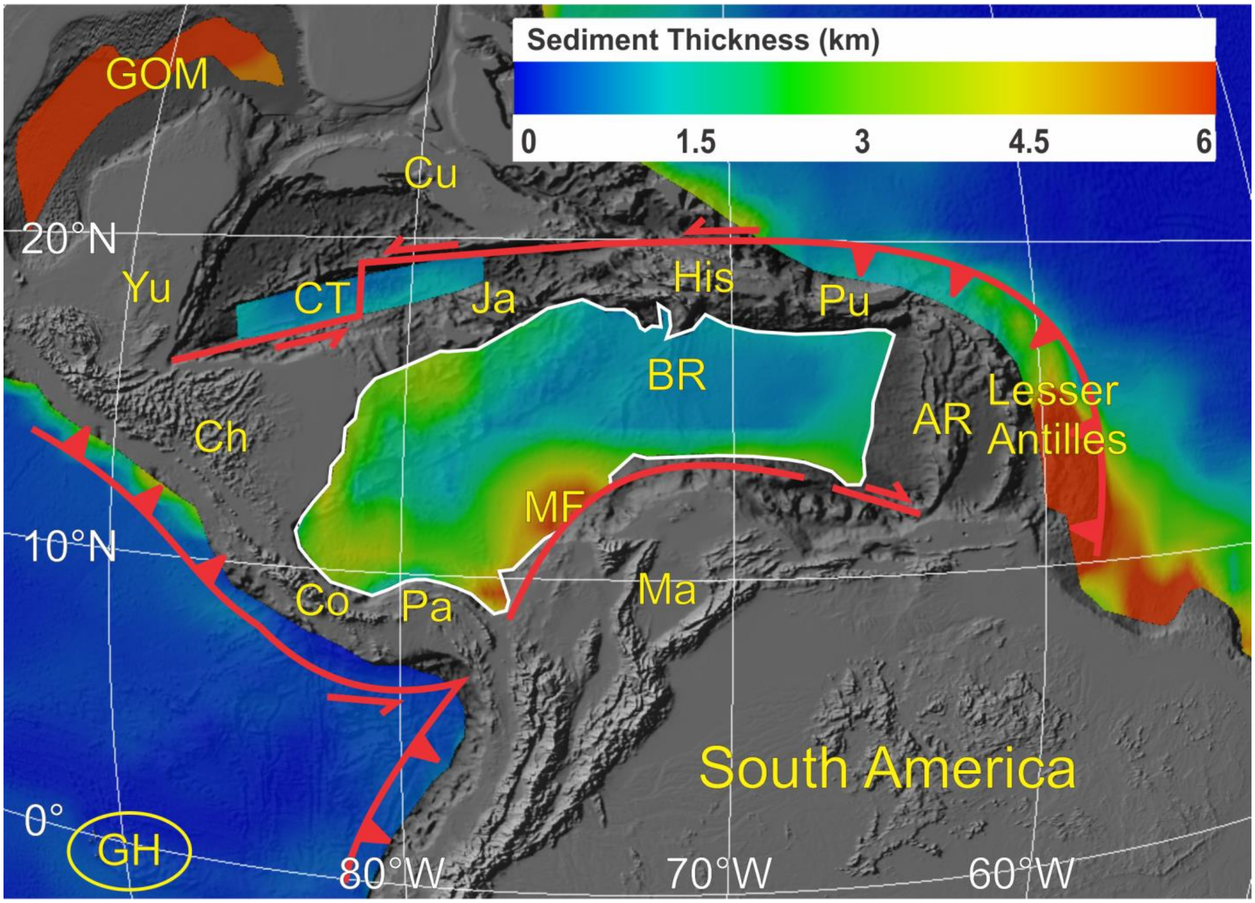

Figure 2: Sediment distribution in the Caribbean realm based on CRUST1 (Laske et al., 2013). Abbreviations are as in Figure 1.

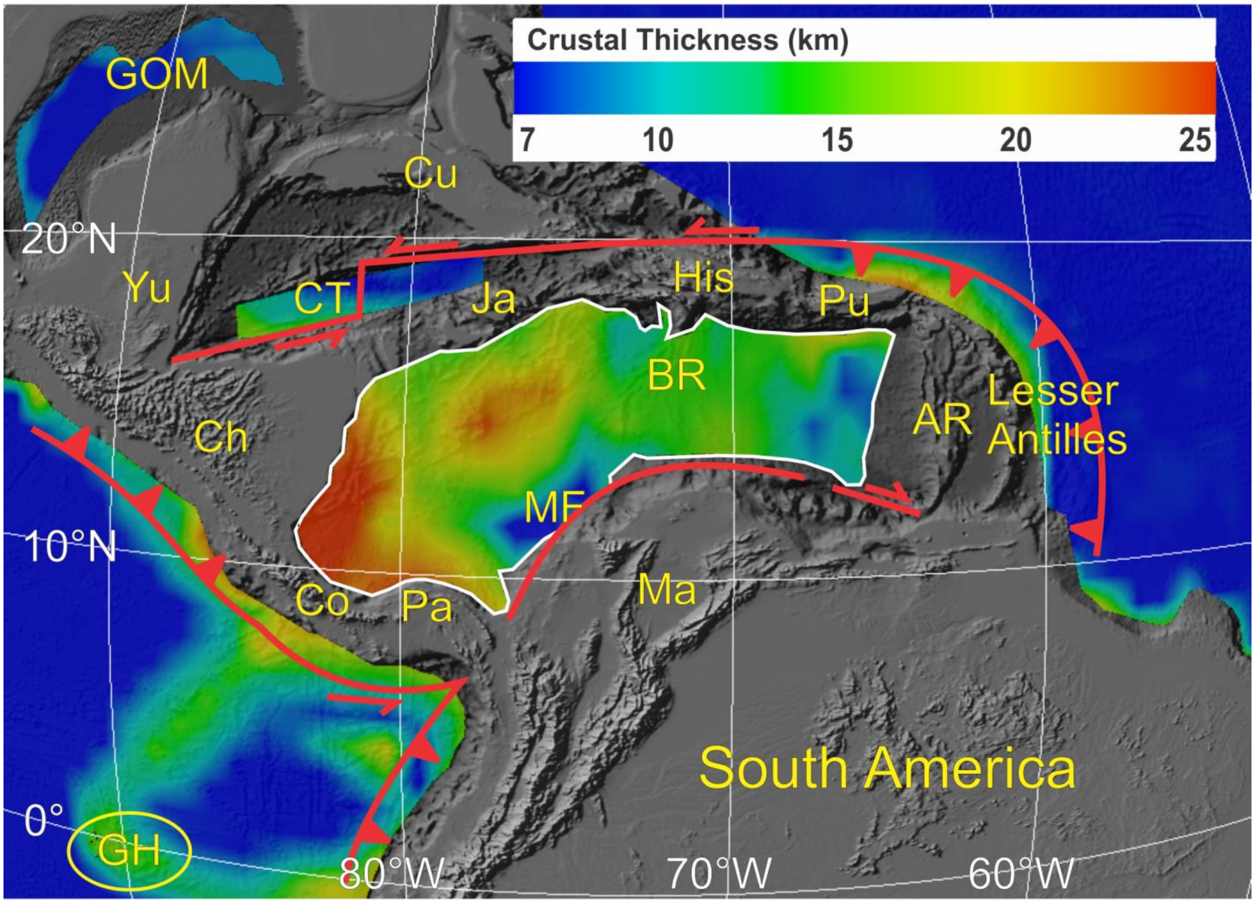

Figure 3: Crustal thickness variations according to CRUST1 (Laske et al., 2013). Abbreviations are as in Figure 1.

3.3 Residual (dynamic) topography deconvolution methodology

Dynamic topography refers to the mantle component of the observed topography, which is caused by mantle up- or downwellings resulting from density variations within Earth's mantle (Braun, 2010). For oceanic regions it can be calculated in three steps: (1) The observed bathymetry is corrected for sediments and crustal thickness variations, assuming local isostacy; (2) Based on standard cooling models (Turcotte and Oxburgh, 1967; Stein and Stein, 1992; Parson and Sclater, 1977) an expected basement depth can be derived, for which a precise lithospheric age distribution is essential; (3) The difference between the expected basement depth from step (2) and the isostatically corrected observed basement depth from step (1) is the 'dynamic' component of the convecting mantle (Kido and Seno, 1994).

For step (1), we made use of the recently released 1x1-degree global crustal and sediment thickness model CRUST1 of Laske et al. (2013), see Figures 2 and 3. The model is divided into three sediment and three crystalline crustal layers and contains density distributions for each one of them. The isostatic correction for sediments (S_c) is calculated by:

$$S_c = \left(\frac{\rho_s - \rho_w}{\rho_a - \rho_w}\right) * S_t \qquad (1)$$

ρ_s, ρ_w, and ρ_a refer to the densities of the sediments, water (1.019 g/cm^3), and asthenosphere (3.2 g/cm^3), respectively. S_t is the total sediment thickness at each respective grid point. The sediment correction was calculated for the sediment layers separately and subtracted from the sediment loaded, observed bathymetry (Amante and Eakins, 2009). We excluded bathymetry that was shallower than the mean water depth minus two times the standard deviation (for the Pacific/Atlantic, we used 4 times the standard deviation in order to reflect the greater variation due to the greater polygon sizes), to avoid the inclusion of continental crust.

The isostatic correction (C_c) for the crystalline crust is given by:

$$C_c = \left(\frac{\rho_a - \rho_c}{\rho_a - \rho_w}\right) * (C_t - C_{avg}) \qquad (2)$$

ρ_c and C_t indicate the density of the crust and the crustal thickness at each grid point, respectively. C_{avg} refers to the average crustal thickness of oceanic crust, which is approximately

7.1 km (Winterbourne et al., 2009). We calculated a weighted mean average density for all three crustal layers. This mean density was used in Equation (2) to derive the isostatic crustal correction for each of the three crystalline crustal layers, which were added to the sediment corrected bathymetry.

For the calculation of a theoretical basement depth, the age distribution is usually derived from magnetic isochron interpretations. However, the thickened crust of the Caribbean Plateau prohibits the direct measurement of magnetic anomalies, such that we derived an age-grid (Fig. 1, insert) from our own plate reconstruction model (Nerlich et al., 2014). The model fits the reconstructed position of the Caribbean Plateau with the paleo-position of the Galapagos hotspot at the main phase of plateau formation between ~94 and 88 Ma (Kerr et al., 2009, Hastie et al., 2013), indicating that the Plateau was built as a consequence of the Galapagos plume head arrival, consistent with geochemical studies (Hauff et al., 1997; Geldmacher et al., 2003; Thompson et al., 2003).

The dynamic topography (*DT*) is given by the following formula:

$$DT = d_{age} - (d + S_t - S_c + C_c), \tag{3}$$

for which depths are positive downwards and d_{age} is the predicted depth. The terms in the brackets in (3) represent the observed bathymerty (d), the sediment thickness (S_t) and the two isostatic corrections from Equations (1) and (2).

3.4 Uncertainty determination

Because of the uncertainty and relative sparse resolution of the sedimentary thickness data from CRUST1.0 with respect to the bathymetric data, we applied two approaches to evaluate the uncertainty of our approach. The first approach consists of a series of quasi-Monte Carlo simulations (Kroese et al., 2011) in which we approximate the uncertainty of the dynamic topography by considering the parameters in Equations (1) and (2) as probabilistic rather than deterministic or discrete values. Table 1 presents the various stochastic parameters that we used in the calculation, where $\mathbf{P}_s \sim$ Normal(2200,200), for example, (following the regular statistical nomenclature) means the stochastic sediment density was distributed with mean 2200 kg/m^3 and standard deviation 200 kg/m^3. For each of the distributions in Table 1, *N* samples were drawn using the Holton Sampling technique (Feinberg and Clark, 2013) for each point in the domain. The *N* results of Equations (1) and (2) were then calculated using these sampled values as parameters and the dynamic topography calculated according to Equation (3). The *N* resulting dynamic topography results were then used to calculate the sample mean and standard deviation at each point.

To determine convergence of the quasi-Monte Carlo method, the mean was calculated analytically. The RMS error (for the domain) between the estimated and analytical mean is shown in Figure 4 as a function of *N*. Linear convergence in the log-log plot is achieved up to N=10,000, while convergence is achieved at about N=100,000 after which no significant improvement in the error is achieved with N and the relative RMS error is smaller than 0.025, corresponding to about 15 m in dynamic topography. The calculated variance of the dynamic topography, based on N=10^7, is presented in the results section below.

Table 1: Stochastic Parameters used in the quasi-Monte Carlo simulation.

Parameter	Symbol	Stochastic Parameter	Stochastic Representation	Units
Sediment Density	ρ_s	$\mathbf{P}_s$	$\mathbf{P}_s \sim$ Normal(2200,200)	kg/m^3
Crustal Density	ρ_c	$\mathbf{P}_c$	$\mathbf{P}_c \sim$ Normal(2800,200)	kg/m^3
Asthenospheric Density	ρ_a	$\mathbf{P}_a$	$\mathbf{P}_a \sim$ Uniform(3100,3300)	kg/m^3
Sediment Thickness	z_s	$\mathbf{Z}_s$	$\mathbf{Z}_s \sim$ Normal(z_s, 500)	m
Crustal Thickness	z_c	$\mathbf{Z}_c$	$\mathbf{Z}_c \sim$ Normal(z_c, 500)	m

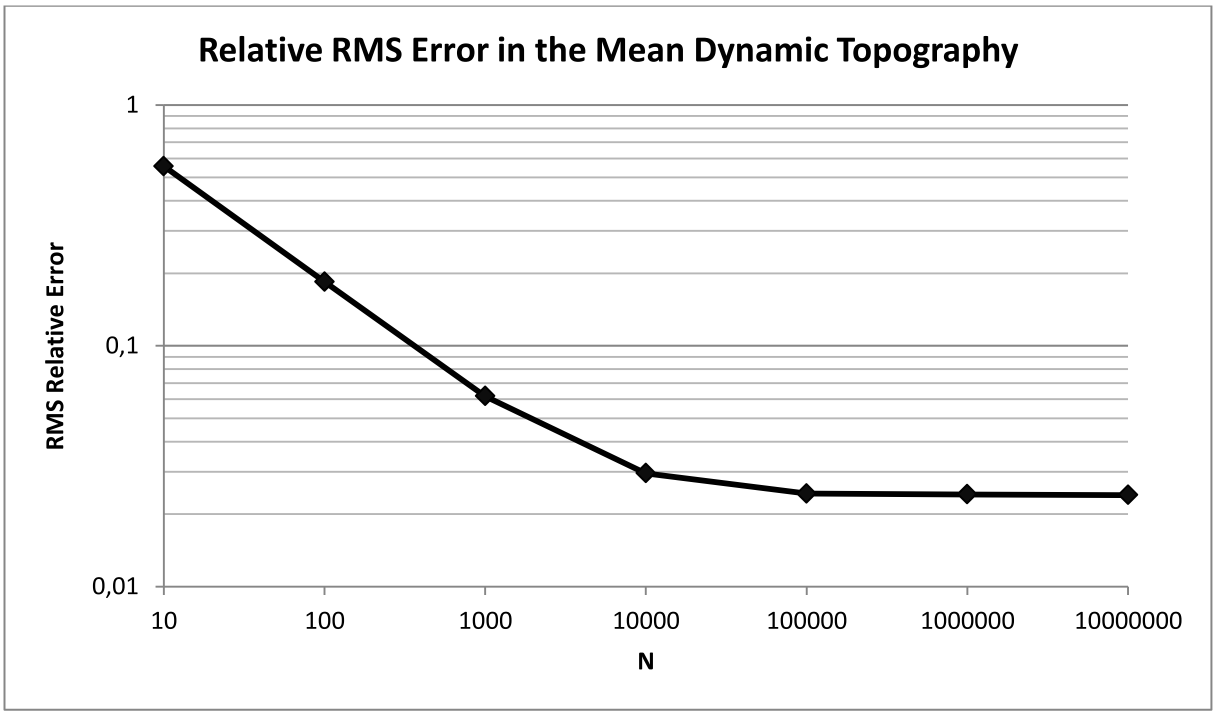

Figure 4. Log-log plot of the RMS error in meters between the analytical and estimated mean of the dynamic topography vs the number of quasi-Monte Carlo simulations, N.

The second method to constrain the uncertainty was to vary the lithospheric cooling model used to determine the theoretical ocean depth, since lithospheric cooling models differ from each other for the oldest ocean floor (between 100 and 150 million years old). We calculated the dynamic topography using three different models. The half-space cooling model (Hs) (Turcotte and Oxburgh, 1967) is uniformly based on the function: basement depth = 2600 + 345 × √(age). The GDH1 (Stein and Stein, 1992) and PSM (Parsons and Sclater, 1977) plate models deviate from the continuous "square-root of age" assumption in that these models assume negative exponential functions for ages greater than 20 Ma and 70 Ma, respectively.

3.5 Results

The standard deviation of the dynamic topography from the quasi-Monte Carlo approach is presented in Figure 5. The regions adjacent to Costa Rica show high standard deviation in the dynamic topography because of the uncertainty in the sediment and crustal densities impacting the combined larger correction for higher mean sediment and crustal thicknesses. While the mean standard deviation is 960 m, the mode which is less sensitive to the outliers mentioned above, is 743 m and is more representative of the standard deviation of dynamic topography in the Magdalena Fan and Beata Ridge (MF and BR, Fig. 6).

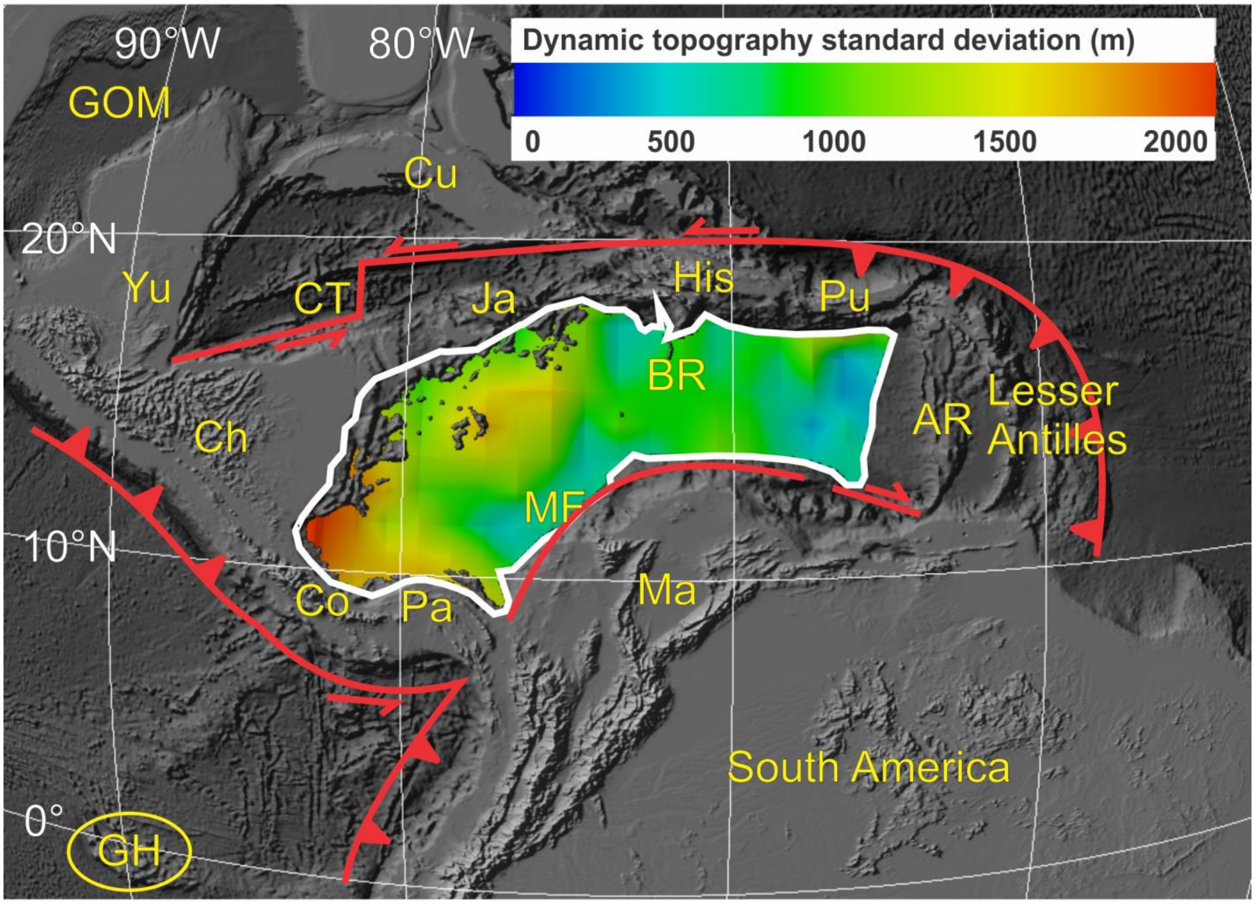

Figure 5: Standard deviation of the dynamic topography derived from the quasi-Monte Carlo simulations, with N=10^7. The mode deviation is 743 m, the median is 862 m and the mean is 960 m.

The calculated dynamic topography signal based on the GDH1 lithospheric cooling model is shown in Figure 6. The figure also includes the locations of profiles visualizing the amplitude of dynamic topography (Fig. 7) based on the three tested cooling models. All profiles have their origin at the Galapagos hotspot (GH) and continue eastward through the proposed slab window and into various directions within the Caribbean realm. The diagrams visualize the discrepancy of dynamic topography predictions between the different lithospheric cooling models: While the predictions in terms of the dynamic topography amplitude for young seafloor (< 60 Ma) such as for the region west of Central America in the Pacific are very similar, they differ substantially in dynamic topography amplitude in the Caribbean realm, where the ocean floor is presumed to be very old (> 100 Ma). However, despite these absolute amplitude differences the patterns are very similar and thus useful to infer mantle flow. Based on the GDH1 model, the calculated minimum/maximum dynamic topography signals of the outlined Caribbean Plateau are -560 /+760 m (half-space model: -183/+1390 m; PSM model: -450/+920 m). The largest positive signal is observed around the Beata Ridge and southeast of Jamaica. The largest negative signal is located north of the Magdalena Fan (GDH1: ~ -750 m). Additionally, two dynamic topography lows are located at the western and eastern ends of the Caribbean Plateau. The Galapagos hotspot in the Pacific realm is characterized by positive dynamic topography of ~ 500 m (according to all cooling models), leading to a relative difference between the dynamic topography low north of the Magdalena Fan and the hotspot of ~ 1250 m (Fig. 7, based on GDH1). Finally, a local maximum is observed just to the east of Panama (Fig. 7).

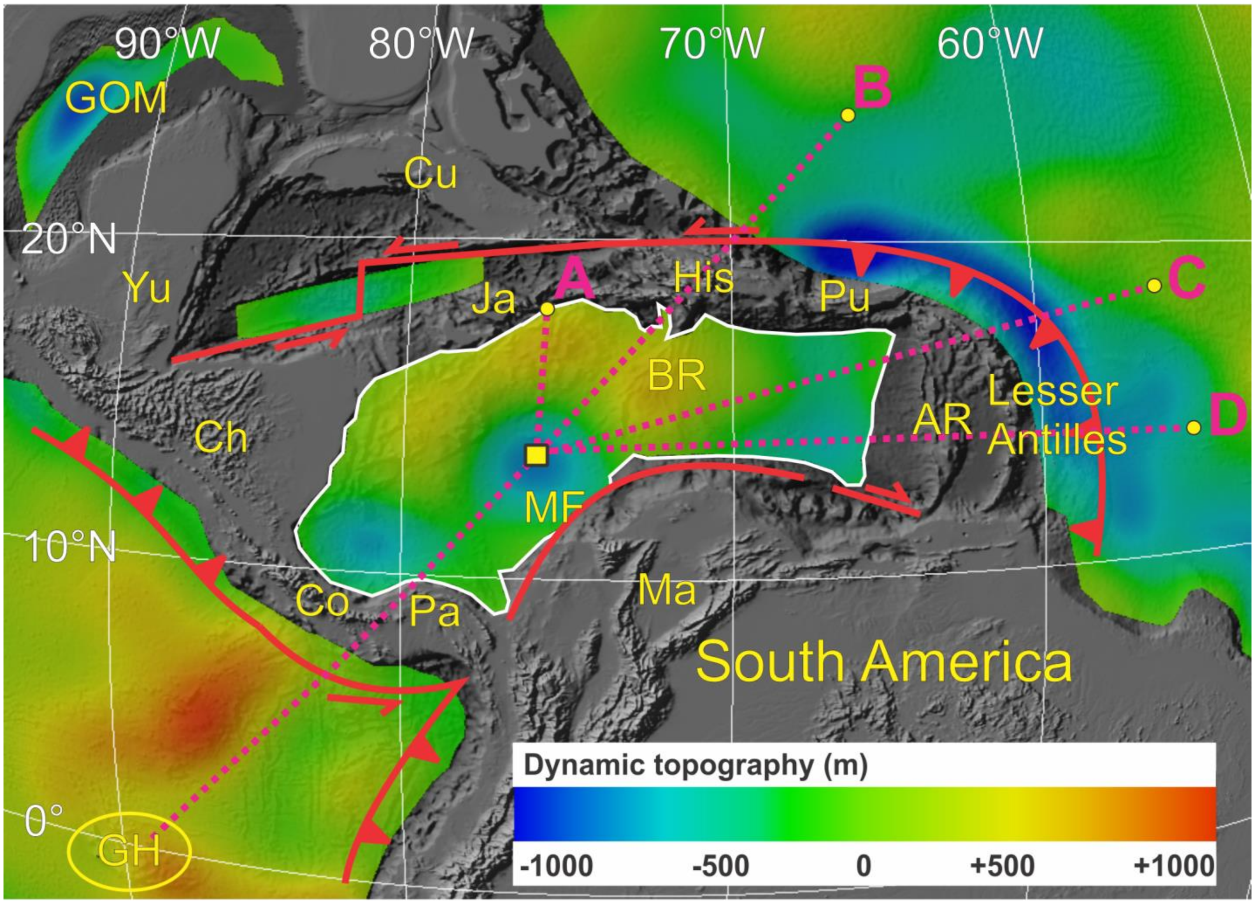

Figure 6: Present-day bathymetry (gray-scale) with superimposed dynamic topography based on the GDH1 lithospheric cooling model (Stein and Stein, 1992). Note, because dynamic topography is a long-wavelength feature, the results presented here were low-pass filtered (cut-off wavelength: 5 degrees, i.e. ~550 km). The central part of the Caribbean Plateau (location is indicated by a yellow square) shows a regional dynamic topography low. The northern part of the region and the Beata Ridge are characterized by positive, the western and eastern ends of the region by negative dynamic topography, respectively. Locations of profiles A-D (Fig. 7) originating at GH are illustrated by the dotted line.

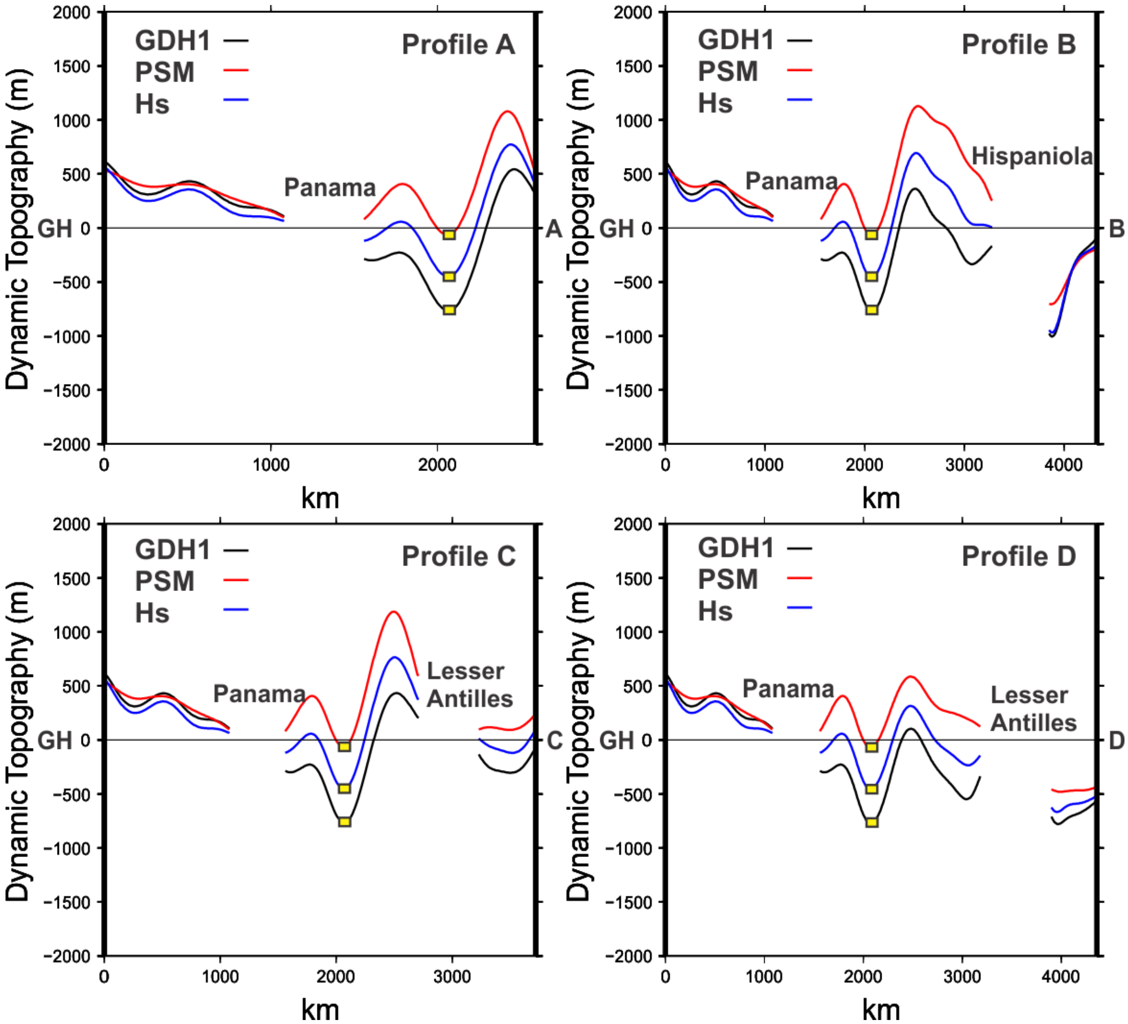

Figure 7: Dynamic topography based on three different lithospheric cooling models along the profiles shown in Figure 6: (1) GDH1 (Stein and Stein (1992) in black, (2) PSM (Turcotte and Oxburgh, 1967) in red, and (3) Hs (Parsons and Sclater, 1977) in blue. The three models lead to similar patterns but differ in the predicted dynamic topography amplitude. The yellow rectangle corresponds to the position of the central dynamic topography low in the Caribbean Sea (see Fig. 6). Note the decrease in dynamic topography from west to east away from the Galapagos hotspot towards the central part of the Caribbean. Note also the dynamic topography increase in all directions past this central Caribbean dynamic topography low.

3.6 Discussion

Regional geochemical variations along the Central American Arc have been known for a long time and been linked to different mantle sources (Carr, 1984; Herrstrom et al., 1995): While rocks from Nicaragua and Guatemala are associated with a slab-metasomatized, mid-ocean ridge basalt (MORB)-source mantle, rocks from Costa Rica and Panama, which are little affected by subduction metasomatism, indicate an ocean-island basalt (OIB)-type enriched mantle origin. Additionally, a striking similarity of the rocks found in Costa Rica and Panama with those from the Galapagos Islands has been noticed (Johnston and Thorkelson, 1997; Abratis and Wörner, 2001). Shallow mantle flow away from the Galapagos hotspot through a seismically quiet zone related to a slab gap between the subducting Cocos and Nazca plates into the Caribbean realm was suggested by these authors as a mechanism to explain the geochemical variations along the Central American Arc. Alternatively, it was argued that variations in the incoming plate's geometry as well as in the composition and fluid content, respectively, could be responsible for the chemical differences (Rüpke et al., 2002).

Support for the mantle flow hypothesis comes from several shear-wave splitting studies, which measure the lattice preferred orientation of olivine crystals in the upper mantle (Russo and Silver, 1994; Russo et al., 1996; Pinero-Feliciangeli and Kendall, 2008; Growdon et al., 2009; Masy et al., 2011). These studies indicated the presence of a regional west-east oriented asthenosphere flow underneath the Caribbean Sea. Yet, these studies were mostly limited to the northeastern margin of South America and the Lesser Antilles arc, rather than Panama and Costa Rica. The only measurement just north of the proposed slab gap (Fig. 1) indicated trench parallel flow (Pinero-Feliciangeli and Kendall, 2008). Also, it should be noted that relative flow directions in general cannot be estimated from shear-wave splitting measurements but merely the orientation.

Compared to geochemical and shear-wave splitting studies, our approach offers a broader sense of the regional asthenosphere flow. The profiles shown in Figure 7 indicate a decreasing dynamic topography signal away from the Galapagos hotspot towards the proposed slab gap and into the Caribbean realm. This observation indicates shallow mantle inflow from the Pacific into the Caribbean realm and confirms the above mentioned studies and Alvarez (1982) mantle flow hypothesis. Undulations along the profiles are observable; foremost the dynamic topography high east of Panama stands in contrast to the general trend. However, the amplitude of this dynamic topography high is strongly model-dependent and most prominent using the PSM model and almost negligible according to the GDH1 model (Fig. 7). Furthermore, the region east of Panama shows rather high variance, whereas the low in the Central Caribbean is a relatively robust feature (Fig. 5). This also applies to the dynamic topography high around the Galapagos hotspot, where the age-grid is derived from magnetic isochrones and the seafloor is young, leading to almost identical predictions of all tested lithospheric cooling models.

Additionally, it is noticeable that the profiles of Figure 7 show a relatively large positive dynamic topography signal in all direction past the dynamic topography low in the central Caribbean (Fig. 6). This suggests that only a tongue of Pacific derived asthenosphere has reached the central Caribbean realm, where the flow is blocked, possibly by the subducting Lesser Antilles slab. If true, an upward reflection of the flow could perhaps cause the observed positive dynamic topography signals further to the east and north of the central dynamic topography low. The negative signal observed on the eastern edge of the Caribbean Plateau most likely reflects the sinking slab of the Lesser Antilles subduction zone. Similarly, the negative signal east of

Costa Rica may be related to the subducting Cocos Plate slab. However, it could also be due to a regional over prediction of the crustal thickness in the CRUST1 model (Laske et al., 2013), which is well above 20 km in this area (Fig. 3).

Finally, it is noteworthy that our dynamic topography amplitudes are all in the commonly accepted range of +/- 1000 m (Braun, 2010) (except the maximum dynamic topography signal predicted by the half-space cooling model). A quantitative comparison of our dynamic topography results (Fig. 6) with the regional long-wavelength free-air gravity signal (Fig. 1) shows a weak positive correlation (r = 0.4). This is in broad agreement with the global results of Winterbourne et al. (2009). Both of these results suggests that the derived age distribution and our dynamic topography results are reasonable.

Despite some potential, most likely Galapagos plume-derived inflow into the Caribbean realm, the absence of a gradual decreasing dynamic topography signal from west to east throughout the entire Caribbean region indicates the absence of a continuous flow into the Atlantic mantle domain. It seems unlikely that the growing Atlantic mantle reservoir is compensated through shallow mantle inflow through the Caribbean realm alone. We arrived at the same conclusion for the Scotia Sea (Nerlich et al., 2013), and suggest that the deep mantle beneath Africa is a far more likely primary source to supply the growing Atlantic mantle reservoir. The elevated topography of the African superswell (Nyblade and Robinson, 1994), which is inferred to be due to a deep mantle upwelling (Lithgow-Bertelloni and Silver, 1998), contrasts with significant negative dynamic topography of up to ~ 1 km at the conjugate South American margin in the Argentine Basin (e.g., Winterbourne et al., 2009; Shepard et al., 2012) as well as with wide areas northeast off the northern margin of Brazil (Guyana Basin). These topography lows are possibly related to the subduction of the former Phoenix and Farallon plates, respectively, that nowadays reside in the lower mantle (Bunge and Grand, 2000). The remarkable dynamic topography gradient across the entire South Atlantic is consistent with westward flow emanating from the African superplume, as suggested by Behn et al. (2004) and Husson et al. (2012). It is likely that this large scale flow has more relevance compared to regional inflow through the Caribbean or Scotia Sea in the process of establishing mass balance between the Pacific and Atlantic mantle reservoirs.

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4. Conclusions

The plate tectonic histories of the Scotia Sea and the Caribbean were investigated in this thesis, resulting in two novel plate reconstruction models for each region. It was shown, for example, that from a geodynamic point of view it is unlikely that the Central Scotia Sea is a Mesozoic plate fragment. It was also demonstrated, which assumptions in a tectonic model are necessary to link the Caribbean Plateau to the paleo-Galapagos hotspot and to show that the latter may well have been the source of the rocks that compose the present-day Plateau.

The established kinematic models were used to derive age-grids and to calculate the present-day dynamic topography. These results were discussed in relation to mantle exchange processes between the shrinking Pacific and growing Atlantic mantle reservoirs. It was presumed that shallow mantle flow from the Pacific realm into the Atlantic mantle domain would be indicated by a gradual decrease from west to east in dynamic topography.

The results for the Scotia Sea do not indicate such a gradient and it was inferred that shallow mantle flow does not occur in this region. In contrast, a decreasing dynamic topography signal away from the Galapagos hotspot into the Caribbean realm has been observed, indicating some regional inflow.

Based on the absence of an asthenosphere flux through the Scotia Sea and despite the potential presence of a minor asthenosphere inflow in the Caribbean realm, it was concluded that deep mantle processes underneath Africa are the more likely source to achieve mass balance in the Atlantic mantle domain.