Subduction Initiation by Plume-Plateau Interaction: Insights From Numerical Models

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Abstract It has recently been demonstrated that the interaction of a mantle plume with sufficiently old oceanic lithosphere can initiate subduction. However, the existence of large lithospheric heterogeneities, such as a buoyant plateau, in proximity to a rising plume head may potentially hinder the formation of a new subduction zone. Here, we investigate this scenario by means of 3-D numerical thermomechanical modeling. We explore how plume-lithosphere interaction is affected by lithospheric age, relative location of plume head and plateau border, and the strength of the oceanic crust. Our numerical experiments suggest four different geodynamic regimes: (a) oceanic trench formation, (b) circular oceanic-plateau trench formation, (c) plateau trench formation, and (d) no trench formation. We show that regardless of the age and crustal strength of the oceanic lithosphere, subduction can initiate when the plume head is either below the plateau border or at a distance less than the plume radius from the plateau edge. Crustal heterogeneity facilitates subduction initiation of old oceanic lithosphere. High crustal strength hampers the formation of a new subduction zone when the plume head is located below a young lithosphere containing a thick and strong plateau. We suggest that plume-plateau interaction in the western margin of the Caribbean could have resulted in subduction initiation when the plume head impinged onto the oceanic lithosphere close to the border between plateau and oceanic crust.

1. Introduction

Mantle plumes, which are large volumes of abnormally hot rocks, migrate upward within the mantle due to their positive buoyancy. As a mantle plume reaches the base of the lithosphere, it results in diverse geological processes including the formation of plateaus (Bierlein & Pisarevsky, 2008; Fitton et al., 2004; Hastie & Kerr, 2010; Taylor, 2006), continental breakup (Fromm et al., 2015; Hill, 1991; Pirajno, 2000), triple-junction rifting (Burke & Dewey, 1973; Houseman, 1990; Koptev et al., 2018), and subduction initiation (Baes et al., 2016; Gerya et al., 2015; Stern & Gerya, 2018; Ueda et al., 2008; Whattam & Stern, 2014). Among these tectonic settings, subduction zones are the only ones that represent convergent tectonics. In this aspect, initiation of subduction following the impingement of a hot buoyant mantle plume is a unique feature of plume-lithosphere interaction.

The first recognized natural example of plume-induced subduction initiation (PISI) occurred in the Caribbean region (Whattam & Stern, 2014). The complex tectonic setting of the Caribbean plate (Figure 1) is the matter of several unresolved geoscientific controversies. A particularly controversial issue of Caribbean tectonics is the origin of subduction in the west and southwest of the Caribbean plate. Pindell et al. (2005) proposed that the northeast dipping subduction zone in the west of this plate initiated along an intraoceanic transform fault in the Late Albian (ca. 105–100 Ma). In contrast, Whattam and Stern (2014) used geochemical, geochronological, and isotopic data to suggest that the arrival of a large plume head at about 100–95 Ma, which formed the Caribbean Large Igneous Province (CLIP), induced a new subduction zone in the region. They further argued that the promoting factor for PISI in this region was the compositional and density contrasts between the 140–110 Myr old oceanic plateau and the surrounding oceanic lithosphere. Following this study, Boschman et al. (2019) employed plate kinematic reconstructions to show that subduction in the western Caribbean formed in an intraoceanic environment as a result of plume-lithosphere interaction. Although these studies agree that subduction was induced by a mantle plume, some aspects of how this process evolved remain enigmatic. One of these issues is the formation of a single (instead of multiple) slab in the region. Whattam and Stern (2014) noted that a preexisting subduction zone in the northeast was the key factor in development of a single slab in the Caribbean. Another
example of PISI is the formation of the Cascadia subduction zone in Eocene times (Stern & Dumitru, 2019), where the arrival of the Yellowstone mantle plume head beneath western North America at circa 55 Ma have destroyed the existing Cordilleran subduction zone and initiated the Cascadia subduction zone.

A number of numerical modeling studies have previously investigated PISI. The initiation of continental lithosphere subduction following interaction between a plume and continental lithosphere was first numerically explored by Burov and Cloetingh (2010). They demonstrated that at shallow depth, the subduction of continental lithosphere is controlled by rheological lithosphere stratification and free surface. At greater depth (300–500 km) subduction is governed by phase changes and slab interactions with the surrounding mantle. Using 2-D models, Ueda et al. (2008) investigated mantle plume-induced subduction of oceanic lithosphere. They found that a hot and buoyant mantle plume can break the lithosphere apart and by flowing on top of the broken lithosphere can eventually initiate a new subduction zone. Formation of a subduction zone is controlled by a critical local weakening of the lithospheric material above the plume, which depends on the plume’s volume, buoyancy, and the thickness of the lithosphere. Three-dimensional numerical modeling suggests that in the early Earth, plume-lithosphere interaction led to either episodic lithospheric drips when a plume interacted with a young lithosphere or initiation of a subduction zone when interacting with an older lithosphere (Gerya et al., 2015). The key factors for PISI are the strength and negative buoyancy of the lithosphere, magmatic weakening above the plume head, and lubrication of the slab interface by hydrated crust. In present-day Earth, interaction of a buoyant plume with a lithosphere can result in four deformation responses (Baes et al., 2016): (a) self-sustaining subduction initiation, (b) frozen subduction initiation, (c) slab break-off, and (d) plume underplating. The parameters controlling the development of these deformation regimes are brittle/plastic strength and age of the oceanic lithosphere, the presence/absence of lithospheric heterogeneity, and buoyancy of the plume, which depends on the size, composition, and temperature of the plume.

Baes et al. (2020) conducted a numerical modeling study, in which they investigated the factors controlling the shape of plume-induced subduction zones. They show that the development of single-slab or multi-slab subduction depends on several parameters such as age of oceanic lithosphere, thickness of the crust, mantle temperature, and large-scale lithospheric extension. Their results indicate that single-slab subduction induced by the interaction of a mantle plume with a thick plate (with thickness of 20 km) forms only if the lithosphere has a moderate age of 40 Myr. Large-scale lithospheric extension facilitates subduction initiation especially when the plume hits a lithosphere containing a thick plateau. Based on these results, Baes et al. (2020) suggest that single-slab subduction in the western margin of the Caribbean was formed due to either plume-plateau interaction in an extensional regime or interaction of a mantle plume with a lithosphere with typical crustal thickness of 8 km.

Figure 1. Tectonic setting of the Caribbean region.
Both natural observations (e.g., Whattam & Stern, 2014) and numerical models (Baes et al., 2020) agree that large-scale heterogeneities, which commonly exist within the oceanic lithosphere, can impede PISI. In the 3-D numerical models of Baes et al. (2020), the mantle plume hits a homogenous plateau, which may not be representative of the tectonic setting of the Caribbean region at circa 100 Ma. In the present study we aim to setup 3-D numerical models with heterogeneous crustal thickness, which allow us to explore the effect of relative distance of the plume head and plateau edge on subduction initiation. We also investigate the influence of lithospheric age and strength of the lower crust on PISI.

2. Model Setup

We use the code I3ELVIS, which is based on a staggered finite difference scheme combined with a marker-in-cell technique (Gerya, 2010). We obtain stresses and velocities by solving the momentum, continuity, and energy equations:

\[ \nabla \cdot \mathbf{v} = 0 \]

\[ \frac{\partial \sigma_{ij}}{\partial x_j} - \frac{\partial P}{\partial x_i} + \rho g_i = 0 \]

\[ \rho C_p \left( \frac{D T}{D t} \right) = -\nabla \cdot \mathbf{q} + H_r + H_a + H_s + H_l, \]

where \( D \frac{D}{D t} \), \( P \), \( v \), \( \sigma_{ij} \), \( g_0 \), \( C_p \), and \( q \) are the Lagrangian time derivative, pressure, density, velocity, deviatoric stress tensor, gravity acceleration, heat capacity, and heat flux, respectively. \( H_r \), \( H_a \), \( H_s \), and \( H_l \) denote the radioactive, adiabatic, shear, and latent heat productions, respectively.

The viscosity in our models depends on stress, temperature, and pressure. The effective viscosity is expressed as

\[ \eta_{\text{effective}} = \frac{1}{\eta_{\text{diffusion}}} + \frac{1}{\eta_{\text{dislocation}}}, \]

where \( \eta_{\text{diffusion}} \) and \( \eta_{\text{dislocation}} \) are viscosities for diffusion and dislocation creep, \( P \) is pressure, \( T \) is temperature, \( \dot{\varepsilon}_{II} = \sqrt{\frac{1}{2} \dot{\varepsilon}_{ij} \dot{\varepsilon}_{ij}} \) is the second invariant of the strain rate tensor, \( \sigma_{cr} \) is the diffusion-dislocation creep transition stress, and \( A_D, E, V, \) and \( n \) are the strain rate pre-exponential factor, activation energy, activation volume, and stress exponent, respectively. We limit the viscosity between \( 1 \times 10^{18} \) Pas and \( 1 \times 10^{26} \) Pas.

Plastic deformation is defined by the following modified version of the Drucker-Prager yield criterion (Byerlee, 1978; Ranalli, 1995):

\[ \tau = C + \phi \lambda_{\text{melt}} P, \]

where \( C, \lambda_{\text{melt}}, \) and \( \phi \) are cohesion (rock strength at \( P = 0 \)), melt-induced weakening factor (Gerya et al., 2015), and internal friction coefficient for the confined fractures, respectively. The product of \( \phi \lambda_{\text{melt}} \) is called the effective friction coefficient. We note that melt-induced weakening is implemented locally within the lithosphere above areas of melt extraction. For all other materials, no magmatic weakening is assumed (\( \lambda_{\text{melt}} = 1 \)).

Density, which depends on pressure and temperature, is expressed as

\[ \rho = \rho_0 [1 - \alpha (T - T_0)] [1 + \beta (P - P_0)], \]

where \( \rho_0 \) and \( T_0 \) are the density and temperature under the conditions of \( P_0 = 0.1 \) MPa and \( T_0 = 298 \) K.
(pressure and temperature at Earth’s surface) and $\alpha$ and $\beta$ are the thermal expansion coefficient and the compressibility coefficient, respectively.

Melt extraction and percolation are defined in a simplified manner as described in Gerya (2013) and Gerya et al. (2015). Lagrangian markers are used to track melt extraction. The total amount of melt for every marker is calculated as

$$M = M_0 - \sum_{m} M_{\text{ext}};$$  \hspace{1cm} (5)

where $M_0$ and $\sum_{m} M_{\text{ext}}$ are the standard volumetric degree of mantle melting and total melt fraction extracted during the previous melt extraction episodes, respectively. The extracted melt moves vertically and is added as a volcanic rock layer to the top of the shallowest partially molten mantle above the plume. Melt percolation is considered to be instantaneous (Connolly et al., 2009). Slab dehydration and mantle hydration are based on the water markers approach (Gerya & Meilick, 2011) in which the equilibrium mineralogical water content for the crust and the mantle is defined as a function of pressure and temperature from thermodynamic data by free energy minimization.

The model domain is a cuboid with the dimensions of $1,212 \times 296 \times 1,212$ km$^3$, which is resolved by $404 \times 148 \times 404$ grid points (finer grids in the vertical direction). The uppermost model domain consists of a 20 km-thick layer of sticky air simulating an internal free surface. Below this layer is an oceanic lithosphere, a spherical plume with a radius of 100 km, and an asthenosphere until a depth of 296 km (Figure 2). The oceanic lithosphere consists of three layers: upper and lower crust as well as oceanic lithospheric mantle. The oceanic lithosphere thickness in Model M1 is 50 km, corresponding to a lithospheric age of 20 Myr. The thickness of the crust is not uniform. In our reference model (Model M1 in Table 2), the crust on the right half of the model has a thickness of 8 km, while the left half has a crustal thickness of 20 km (representing a plateau). In this model, the plume head is located directly below the plateau border (the boundary between oceanic crust and plateau). Subsequently, we vary the distance between plume head and plateau edge in order to investigate its effect on the model results (Table 2). We note that our models can

![Figure 2. Reference model setup (Model M1 in Table 2). The upper panel illustrates the initial temperature field of the model. The lower panel represents the compositional field of a cross-section cutting through the middle of the model (color code is at the bottom of the figure).](image-url)
simulate the interaction of a mantle plume with large volcanic plateaus such as the Caribbean and Ontong Java plateaus. The present-day Caribbean plateau is among the largest plateaus on Earth covering an area of ~6 × 10^5 km^2. Burke (1988), however, argued that the plateau may originally have been more than twice of this size and that part of the plateau has been already subducted beneath the North Andes plate.

We employ non-Newtonian viscoplastic rheologies (Ranalli, 1995) for the different layers of our model (Table 1). The upper crust, lower crust, plume, and mantle are considered to be wet quartzite, plagioclase An75, wet olivine, and dry olivine, respectively. To lubricate the plate interface when the lithosphere subducts, the friction coefficient of the upper crust, which is a 2 km-thick layer, is set to zero. This resembles weakening of the uppermost crust due to water percolation and fracturing. The friction coefficient of the other layers, except the lower crust, is 0.2. For the lower crust, generally, a coefficient of 0.1 is used. However, to investigate the effects of the strength of the lower crust, some models use a higher lower crustal friction coefficient of 0.2 (Models M33–M66 in Table 2).

In the reference model, the temperature field of the oceanic lithosphere is based on cooling half-space model for an oceanic lithospheric age of 20 Myr (Turcotte & Schubert, 1982). To investigate the effect of oceanic lithospheric age, we consider additional models with ages of 40 and 70 Myr (Table 2). The temperature of the asthenosphere is determined using an adiabatic gradient of ~0.5 K km^{-1}, and the temperature of the plume is 1,850 K. The thermal boundary conditions are 273 K at the upper boundary including the sticky air layer, while zero horizontal heat flux conditions are applied across the vertical boundaries. An infinity-like external temperature condition (Gerya, 2010) is imposed on the lower boundary. According to this method, an external constant temperature condition is satisfied at a distance of ΔL away from the bottom model boundary. The temperature gradient over this distance is considered to be constant (∂T/∂x = (T - T_{external})/ΔL) (e.g., Gerya et al., 2015). We employed a 200 K higher temperature at the bottom boundary to simulate the plume tail. The mechanical boundary conditions are free slip for all boundaries, except the bottom of the model, which is an open permeable boundary.

### 3. Model Results

We conducted a series of numerical experiments to investigate plume-plateau interaction and the effect of the following model parameters (Table 2): (i) the location of the plateau edge with respect to the plume head, (ii) age of the oceanic lithosphere, and (iii) strength of the lower crust. Model results show four distinctly different lithospheric deformation patterns: (a) formation of an oceanic trench (i.e., a trench that develops within oceanic lithosphere), (b) a circular oceanic-plateau trench formation (a circular shaped trench developing partly within the oceanic lithosphere and partly within the plateau), (c) formation of a plateau trench (a trench forming within the plateau), and (d) no trench formation. We first present the results of the reference model with oceanic trench formation. Then, we explore the results of some models representing the three other deformation regimes.

### Table 1

| Material          | Flow law | ρ (kg m⁻³) | C (MPa) | φ (-) | A (Pa s) | E (KJ mol⁻¹) | V (m³ mol⁻¹) | n   | Kb (W m⁻¹ K⁻¹) | Hf (µW m⁻³) |
|-------------------|----------|------------|---------|-------|----------|-------------|-------------|------|----------------|-------------|
| Sticky air        | Air      | 1          | 0       | 0     | 1.0 × 10¹⁸| 0           | 0           | 1.0  | 2              | 0.0         |
| Upper crust       | Wet quartz| 3,000      | 1       | 0     | 1.0 × 10¹⁸| 0           | 0           | 2.3  | 1.18 + 474/(T + 77) | 0.25 |
| Lower crust       | Plagioclase An75 | 3,000      | 1       | 0.1–0.2 | 4.8 × 10²²| 238         | 0           | 3.2  | 1.18 + 474/(T + 77) | 0.25 |
| Lithospheric mantle| Dry olivine| 3,300      | 1       | 0.2   | 3.9 × 10¹⁶| 532         | 0.8 × 10⁻⁵  | 3.5  | 0.73 + 1,293/(T + 77) | 0.022 |
| Asthenosphere     | Dry olivine| 3,300      | 1       | 0.2   | 3.9 × 10¹⁶| 532         | 0.8 × 10⁻⁵  | 3.5  | 0.73 + 1,293/(T + 77) | 0.022 |
| Plume             | Wet olivine| 3,000      | 1       | 0.2   | 5.01 × 10²⁰| 470         | 0.8 × 10⁻⁵  | 4.0  | 0.73 + 1,293/(T + 77) | 0.022 |

Note: See text for further explanation.

*Flow law for all materials are based on Ranalli (1995). bThermal conductivity (K = [-q/∂T/∂x] = K₀ + a/(T + 77), where K₀ and a are the first constant and coefficient of conductivity, respectively). aHeat production (H = Hr + Hd + Ha + Hf).
| Table 2 | List of Experiments |
|---------|---------------------|
| Age: thickness of oceanic plate (Myr·km) | Plume head-plateau edge distance (km$^2$) | Friction coefficient of the lower crust | Extra conditions | State of deformation |
| M1 20:50 | 0 | 0.1 | — | Oceanic trench formation |
| M2 20:50 | 50 | 0.1 | — | Oceanic trench formation |
| M3 20:50 | 100 | 0.1 | — | Circular oceanic-plateau trench formation |
| M4 20:50 | 150 | 0.1 | — | Circular oceanic-plateau trench formation |
| M5 20:50 | 250 | 0.1 | — | Circular oceanic-plateau trench formation |
| M6 20:50 | 350 | 0.1 | — | Circular oceanic-plateau trench formation |
| M7 20:50 | −50 | 0.1 | — | Oceanic trench formation |
| M8 20:50 | −100 | 0.1 | — | Oceanic trench formation |
| M9 20:50 | −150 | 0.1 | — | Oceanic trench formation |
| M10 20:50 | −250 | 0.1 | — | Circular oceanic-plateau trench formation |
| M11 20:50 | −350 | 0.1 | — | Circular oceanic-plateau trench formation |
| M12 40:70 | 0 | 0.1 | — | Oceanic trench formation |
| M13 40:70 | 50 | 0.1 | — | Oceanic trench formation |
| M14 40:70 | 100 | 0.1 | — | Oceanic trench formation |
| M15 40:70 | 150 | 0.1 | — | Oceanic trench formation |
| M16 40:70 | 250 | 0.1 | — | Oceanic trench formation |
| M17 40:70 | 350 | 0.1 | — | Oceanic trench formation |
| M18 40:70 | −50 | 0.1 | — | Circular oceanic-plateau trench formation |
| M19 40:70 | −100 | 0.1 | — | Circular oceanic-plateau trench formation |
| M20 40:70 | −150 | 0.1 | — | Circular oceanic-plateau trench formation |
| M21 40:70 | −250 | 0.1 | — | Circular oceanic-plateau trench formation |
| M22 40:70 | −350 | 0.1 | — | Circular oceanic-plateau trench formation |
| M23 70:92 | 0 | 0.1 | — | Oceanic trench formation |
| M24 70:92 | 50 | 0.1 | — | Oceanic trench formation |
| M25 70:92 | 100 | 0.1 | — | Oceanic trench formation |
| M26 70:92 | 150 | 0.1 | — | Oceanic trench formation |
| M27 70:92 | 250 | 0.1 | — | Oceanic trench formation |
| M28 70:92 | 350 | 0.1 | — | Oceanic trench formation |
| M29 70:92 | −50 | 0.1 | — | Oceanic trench formation |
| M30 70:92 | −100 | 0.1 | — | Oceanic trench formation |
| M31 70:92 | −150 | 0.1 | — | Oceanic trench formation |
| M32 70:92 | −250 | 0.1 | — | Oceanic trench formation |
| M33 70:92 | −350 | 0.1 | — | Oceanic trench formation |
| M34 20:50 | 0 | 0.2 | — | Oceanic trench formation |
| M35 20:50 | 50 | 0.2 | — | Oceanic trench formation |
| M36 20:50 | 100 | 0.2 | — | Oceanic trench formation |
| M37 20:50 | 150 | 0.2 | — | Oceanic trench formation |
| M38 20:50 | 250 | 0.2 | — | Oceanic trench formation |
| M39 20:50 | 350 | 0.2 | — | Oceanic trench formation |
| M40 20:50 | −50 | 0.2 | — | Oceanic trench formation |
| M41 20:50 | −100 | 0.2 | — | Oceanic trench formation |
| M42 20:50 | −150 | 0.2 | — | Oceanic trench formation |
| M43 20:50 | −250 | 0.2 | — | Oceanic trench formation |
| M44 20:50 | −350 | 0.2 | — | Oceanic trench formation |
| M45 40:70 | 0 | 0.2 | — | Oceanic trench formation |
| M46 40:70 | 50 | 0.2 | — | Oceanic trench formation |
| M47 40:70 | 100 | 0.2 | — | Oceanic trench formation |
| M48 40:70 | 150 | 0.2 | — | Oceanic trench formation |
| M49 40:70 | 250 | 0.2 | — | Oceanic trench formation |
| M50 40:70 | 350 | 0.2 | — | Oceanic trench formation |
| M51 40:70 | −50 | 0.2 | — | Oceanic trench formation |
| M52 40:70 | −100 | 0.2 | — | Oceanic trench formation |
| M53 40:70 | −150 | 0.2 | — | Oceanic trench formation |
| M54 40:70 | −250 | 0.2 | — | Oceanic trench formation |
| M55 40:70 | −350 | 0.2 | — | Oceanic trench formation |
| M56 70:92 | 0 | 0.2 | — | Oceanic trench formation |
| M57 70:92 | 50 | 0.2 | — | Oceanic trench formation |
| M58 70:92 | 100 | 0.2 | — | Oceanic trench formation |
3.1. Oceanic Trench Formation (Reference Model)

Figure 3 displays the temporal evolution of viscosity in the lithospheric layer (upper panels), the compositional field of a cross-section cutting through the center of the model (middle panels), and surface topography (lower panels). When the plume head reaches the bottom of the plate, it penetrates the lithosphere as a result of lithospheric magmatic weakening (Figure 3a). This leads to the formation of a new oceanic plateau above the plume head, due to decompression melting of the plume material. Since the crustal thickness is not homogenous above the plume head, the plume rocks move asymmetrically over the broken segment of the lithosphere on the oceanic side of the model and push the oceanic plate downward into the mantle (Figure 3b). Consequently, a surface depression—representing the future trench location—develops along the border of the newly formed plateau on the oceanic side of the plume head (lower panel of Figure 3b). The negative buoyancy of the oceanic lithosphere along with extra gravitational force coming from the overriding plume material on the oceanic lithosphere provides the necessary driving force for subduction initiation, which results in development of a curved oceanic trench (Figure 3c).

3.2. Circular Oceanic-Plateau Trench Formation (Model M4)

Model M4 is similar to the reference model, except that in this model the plume head is located entirely below the lithosphere with a typical oceanic crustal thickness of 8 km. The distance between the plume head and the plateau border is 150 km (Table 2). Results show that the plume breaks the lithosphere and an oceanic plateau forms within the circular broken area (Figure 4a). As the entire plume body is located underneath the homogeneous part of the oceanic lithosphere (below the 8 km-thick oceanic crust), it spreads symmetrically within the circular area created by plume penetration and below the broken segments of the lithosphere. Movement of plume rocks over the lithosphere pushes the fragmented lithosphere downward. This initiates sinking of the oceanic lithosphere into the mantle. At some point the circumferential stress exceeds the yield stress. As a result, ring confinement, which acts as a resistive force, is overcome by tearing of the slab (Figure 4b). Surface topography shows development of a circular bathymetric depression, which indicates the location of the future trench (lower panels of Figure 4). Subduction continues from all sides of the circular area. Continued sinking of the oceanic lithosphere into the mantle leads to formation of a circular oceanic-plateau trench (Figure 4c).

| Table 2: Continued |
|---------------------|
| **Age: thickness of oceanic plate (Myr:km)** | **Plume head-plateau edge distance (km)** | **Friction coefficient of the lower crust** | **Extra conditions** | **State of deformation** |
| M59 | 70:92 | 150 | 0.2 | — | Oceanic trench formation |
| M60 | 70:92 | 250 | 0.2 | — | Oceanic trench formation |
| M61 | 70:92 | 350 | 0.2 | — | Oceanic trench formation |
| M62 | 70:92 | −50 | 0.2 | — | No trench formation |
| M63 | 70:92 | −100 | 0.2 | — | No trench formation |
| M64 | 70:92 | −150 | 0.2 | — | No trench formation |
| M65 | 70:92 | −250 | 0.2 | — | No trench formation |
| M66 | 70:92 | −350 | 0.2 | — | No trench formation |
| M67 | 20:50 | — | 0.1 | Uniform crustal thickness of 8 km | Circular oceanic trench formation |
| M68 | 20:50 | — | 0.1 | Uniform crustal thickness of 20 km (plateau) | Circular plateau trench formation |
| M69 | 20:50 | — | 0.2 | Uniform crustal thickness of 8 km | Circular oceanic trench formation |
| M70 | 20:50 | — | 0.2 | Uniform crustal thickness of 20 km (plateau) | No trench formation |
| M71 | 40:70 | — | 0.1 | Uniform crustal thickness of 8 km | Oceanic trench formation |
| M72 | 40:70 | — | 0.1 | Uniform crustal thickness of 20 km (plateau) | Plateau trench formation |
| M73 | 40:70 | — | 0.2 | Uniform crustal thickness of 8 km | Oceanic trench formation |
| M74 | 40:70 | — | 0.2 | Uniform crustal thickness of 20 km (plateau) | Plateau trench formation |
| M75 | 70:90 | — | 0.1 | Uniform crustal thickness of 8 km | No trench formation |
| M76 | 70:90 | — | 0.1 | Uniform crustal thickness of 20 km (plateau) | No trench formation |
| M77 | 70:90 | — | 0.2 | Uniform crustal thickness of 8 km | No trench formation |
| M78 | 70:90 | — | 0.1 | Uniform crustal thickness of 20 km (plateau) | No trench formation |

*The positive distances stand for the cases where the plume head is below the oceanic crust, and the negative distances indicate the situations where the plume head is below the plateau and zero represents the cases where plume head is exactly below the plateau border.*
Figure 3. Results of Model M1—which has a 20 Myr old oceanic lithosphere and plume head is located exactly below the plateau border—illustrating temporal evolution of an oceanic trench formation at times (a) 0.025 Myr, (b) 0.084 Myr, and (c) 1.914 Myr. The upper panels show the base 10 logarithm of viscosity within the lithosphere. The middle and lower panels illustrate compositional field of a 2-D cross-section cutting through center of model and surface topography (top view), respectively. The color bars of viscosity and surface topography are shown at the top of the figure, and color code of compositional field is at the right-bottom of the figure.
Figure 4. Results of Model M4—which is similar to the reference model except that plume head is located beneath the lithosphere with oceanic crust of 8 km at 150 km distance from plateau border—which show development of a circular oceanic-plateau trench formation at times (a) 0.062 Myr, (b) 1.251 Myr, and (c) 1.921 Myr. For explanation of panels meanings, see Figure 2 caption.
3.3. Plateau Trench Formation (Model M20)

Model M20 is similar to the reference model except that the age of the lithosphere is increased to 40 Myr and the plume head is instead located below the plateau, 150 km away from the plateau border (Figure 5). The plume rises and breaks the lithosphere. Coeval decompression melting of the plume rocks results in the formation of an oceanic plateau in the central part of the model (Figure 5a). Plume materials move over the fragmented segments of the lithosphere, initiating sinking of the lithosphere into the mantle, which causes the formation of a short cylindrical slab and development of a circular surface depression. Ring confinement is overcome by breaking of the lithosphere from the surface (Figure 5b). The slab breaks off from the side where the lithosphere is not homogeneous (containing normal oceanic crust and plateau) and subduction of the lithosphere on the homogeneous side (plateau side) continues, leading to a plateau trench formation (Figure 5c).

3.4. No Trench Formation (Model M30)

Figure 6 shows the results of Model M30, which features a 70 Myr old oceanic lithosphere and a plume head impinging beneath the plateau, located 100 km away from the plateau border. During plume-lithosphere interaction, the plume penetrates the lithosphere and forms an oceanic plateau inside the circular region created in the middle of the model (Figure 6a). Plume rocks rise over the broken lithosphere, attempting to bend it. This results in the development of a shallow depression on the surface around the new plateau (lower panel of Figure 6b). However, due to the resistance of the old and strong oceanic lithosphere to subduction, the lithosphere does not flex enough to initiate sinking of the slab (Figure 6b). The deformation regime in this case is formation of a new oceanic plateau without subduction initiation.

4. Discussion

Figure 7 summarizes how the four described geodynamic regimes (i.e., different lithospheric responses to plume-plateau interaction) depend on the initial plume placement, age of the lithosphere, and crustal strength. We note that previous studies have examined the effect of parameters such as plume buoyancy, which depends on volume, temperature, and composition of the plume, strength of the lithosphere, and lithospheric heterogeneities (i.e., the existence of a weak zone within the lithosphere or large extensional/compressional regimes) on the lithospheric response to plume impingement (Baes et al., 2016, 2020; Gerya et al., 2015). Our experiments show that regardless of the crustal strength or age of the lithosphere, oceanic subduction initiates if (1) the plume head is located directly below the plateau border or (2) it impinges the lithosphere at a distance from the plateau edge of less than the plume radius (100 km in our experiments; Figure 7a). In this case, when the plume breaks the lithosphere, one side of the circular area created by the plume penetration contains an oceanic plateau while the other side exhibits normal oceanic crust. Due to this high density contrast and the tendency of the lithosphere with normal oceanic crust to sink into the mantle, only the lithosphere on the oceanic crust side will subduct.

If the plume head is located entirely below the young oceanic lithosphere (20 Myr old) containing a weak normal oceanic crust (with a friction coefficient of 0.1), at a distance greater than the radius of the plume from the plateau edge, a circular oceanic-plateau trench forms. In this situation, the plume first breaks the homogeneous oceanic lithosphere, and a circular short slab forms. Slab tearing at depth, which facilitates subduction initiation, is eased due to the low strength of the young lithosphere. This result is consistent with models with uniform oceanic crust (Baes et al., 2020, and Models M67 and M69 in Table 2 and Figure 8a). When the plume head is instead located below the young lithosphere containing a weak plateau (with a friction coefficient of 0.1), at a distance of 150 km or less from the plateau border, an oceanic trench forms. This is due to the resistance of the buoyant plateau to subduction, which is higher for younger lithosphere, and the proximity of normal lithosphere to the broken part of the plate. When the distance becomes large (more than 150 km in our experiments) and the plume head is entirely located under the plateau, a circular oceanic-plateau trench develops, which agrees with the results of models that have a uniform plateau thickness (Baes et al., 2020, and Model M68 in Table 2 and Figure 8b). When the oceanic crust is strong (with a friction coefficient of 0.2), the response of young lithosphere to plume-lithosphere interaction is similar to that of weak oceanic crust. The only difference is that subduction does not initiate if the plume head is located below the strong plateau (Figure 7b). This is due to the fact that high positive buoyancy of the plateau combined with its high strength creates a resistive force, which is much bigger than the driving forces needed to initiate subduction.
Figure 5. Results of Model M20—which is similar to the reference model except that the age of the lithosphere is 40 Myr and the plume head is located below the plateau, 150 km away from plateau border—indicating temporal evolution of a plateau trench formation at times (a) 0.074 Myr, (b) 2.975 Myr, and (c) 3.643 Myr. For explanation of panels meanings, see Figure 2 caption.
For oceanic lithospheres with intermediate age (40 Myr), oceanic subduction initiates if the plume head is located below the oceanic crust. Plateau subduction initiates when the plume head reaches the lithosphere containing a plateau at a distance greater than the plume radius. These results coincide with the outcome of models with uniform oceanic crust and plateau thicknesses (Baes et al., 2020, and Models M71–M74 in Table 2 and Figure 8). The one-sided subduction of the oceanic crust or plateau is eased as a result of the short circular slab breaking from the surface (Figure 5b). Since tearing of the intermediate-aged...
lithosphere at depth is difficult—due to lithospheric strength increasing with depth (Equation 3)—ring confinement is overcome by breaking of the slab from the surface on one side. A circular oceanic-plateau trench forms if the plume head is located below the plateau at a distance less than 150 km from the trench.

![Diagram of oceanic trench formation models](image)

**Figure 7.** Summary of results of models with heterogeneous crust (Models M1–M66 in Table 2). (a) and (b) illustrate the results of models in which the plume head is located beneath the oceanic crust and plateau, respectively.

- **Oceanic trench formation**
- **Circular oceanic-plateau trench formation**
- **Plateau trench formation**
- **No trench formation**
plateau edge. In these models, heterogeneity of the crust helps initiate the tearing of the short cylindrical slab, which facilitates circular trench formation. Figure 7 shows that the strength of the oceanic crust does not affect the results significantly when the plume interacts with an intermediate age oceanic lithosphere.

When the plume reaches beneath a 70 Myr old oceanic lithosphere, oceanic subduction initiates if it is located entirely below the oceanic crust (Figure 7a). Comparing these results with those of homogeneous oceanic lithosphere (Figure 8a), it is clear that heterogeneity in the thickness of the crust facilitates subduction initiation. We speculate that subduction will not initiate in models where the plume head is located at greater distances from the plateau border than those investigated here. In agreement with previous studies where a uniform plateau thickness is assumed, we find that when the plume head is located below the plateau subduction cannot initiate (Baes et al., 2020, and Models M76 and M78 in Table 2 and Figure 7b).

Melt-induced weakening is essential factor in PISI. It substantially reduces the long-term brittle-plastic strength of the rocks that are subjected to melt percolation. Gerya et al. (2015) show that the weakening of the lithosphere critically depends on the availability of melt to create frequent episodes of dike propagation through the lithosphere. Hence, interaction of a hot mantle plume with the lithosphere—which implies large melt production rates and dike injection frequency—is potentially capable of weakening of the lithosphere above the plume head. After penetration of plume head into the lithosphere the main cause of subduction initiation is the sudden uprising of the considerable amount of mantle plume material on top of the lithosphere, which pushes the broken segments of the lithosphere downward into the mantle.

Subduction in the southwestern margin of the Caribbean and NW South America around 100 Ma is suggested as the first example of PISI (Whattam & Stern, 2014). The authors claim that the formation of a curved trench (instead of a cylindrical one) was due to preexisting subduction in the northeast margin, which

Figure 8. Summary of results of models with homogeneous crust (Models M67–M78 in Table 2). (a) and (b) show the results of models in which the plume head is located beneath the oceanic crust and plateau, respectively.
Baes et al. (2020) suggest that plume head arrival beneath either a plateau under extension or a lithosphere with a typical crustal thickness of 8 km might have led to the development of a curved trench in the Caribbean region. According to our model results and considering that the oceanic lithosphere at the time of plume-lithosphere interaction in the Caribbean (at 100 Ma) was older than 40 Myr (Gerya et al., 2015, and references therein), we propose that plume-lithosphere interaction could have resulted in subduction initiation if the plume head arrived beneath the oceanic lithosphere and not below the plateau. As suggested by Whattam and Stern (2014), we believe that two episodes of plume-lithosphere interaction happened in the region. The first one occurred at ~140–110 Ma. During this first event, the mantle plume could not break the lithosphere and initiate subduction due to the high strength of the lithosphere and/or a deficit in the positive buoyancy of the plume (Figure 9a). This led to the formation of a plateau as a result of decompression melting of the plume. The plateau moved northeast following the motion of the Farallon plate (Whattam & Stern, 2014). At ~100 Ma, the second episode occurred as another plume head reached the bottom of the lithosphere (at the same location as the previous one). The old plateau that formed from the first event was located close to the eastern side of where the new plume head arrived (Figure 9b). At this time, the plume penetrated the lithosphere, and the heterogeneity of the crustal thickness helped initiate the subduction of the old (and hence strong) Farallon plate (Figure 9c). We do not, however, exclude other factors that were suggested by previous studies such as interaction of a plume with a lithosphere with normal crustal thickness (i.e., ~8 km) and plume-plateau interaction under extensional regimes (Baes et al., 2020) or the existence of subduction in the northeast margin (Whattam & Stern, 2014) as possible causes of the formation of a curved trench in the western margin of Caribbean.

5. Conclusions

In this study, we explored the lithospheric response to plume-plateau interaction using 3-D thermomechanical models. Our modeling results show that depending on the age of the oceanic lithosphere, strength of the crust, and distance between the plume head and plateau border four different lithospheric responses can result from plume-plateau interaction. The resultant deformation regimes are (a) oceanic trench

![Figure 9. Cartoon illustrating (a) construction of the early 140–110 Ma oceanic plateau (b) followed by another episode of plume-lithosphere interaction at 100 Ma (c), which led to subduction initiation of Farallon plate in the western Caribbean region.](http://example.com/figure9.png)
formation, (b) circular plateau-oceanic trench formation, (c) plateau trench formation, and (d) no trench formation. Our results show that oceanic subduction initiates if the plume head impinges exactly beneath the plateau border or if it is located at a distance less than the plume radius from the plateau edge. Subduction cannot initiate if the plume head impinges onto young oceanic lithosphere (≤20 Myr) containing a strong plateau, or if it is below old oceanic lithosphere (≥70 Myr) involving a plateau. The strength of the crust, which acts as a resistive force in subduction initiation, is mainly significant when a mantle plume interacts with a young lithosphere with a thick crustal component. Despite the results of models with uniformly thick old oceanic lithosphere, which show no trench formation regime, by including heterogeneity in the crustal thickness, we find that old oceanic lithosphere can sink into the mantle, developing a new subduction zone. Our results suggest that subduction along the western margin of the Caribbean could be the consequence of plume-plateau interaction if the plume head impinged the oceanic lithosphere at a distance close to the plateau border.

Data Availability Statement

The data regarding experiments in this study have been provided in GFZ data services (http://doi.org/10.5880/GFZ.2.5.2020.001).

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