Evidence and models for lower crustal flow beneath the Galápagos platform

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Abstract The volcanic Galápagos Islands are constructed upon a broad platform, with their active westernmost islands marking the current position of the hotspot. Built upon young oceanic lithosphere (<15 Ma), this platform exhibits unique morphologic features including a system of stepped terraces on the southwestern escarpment with 3 km relief, contrasting with gentle slopes off the eastern platform toward the Carnegie Ridge. Considering horizontal lithostatic pressure differences associated with this relief, along with high temperatures within the young, hotspot-affected lithosphere, it is likely that lower crustal flow contributes significantly to crustal deformation within the Galápagos platform. Using a 2-D, isostatic, thin-sheet approximation for the Stokes flow equation with (Newtonian) space-time-dependent viscosity, we suggest that the bathymetric rim along the eastern platform region (where gravimetry indicates Airy isostasy) near Española Island may be the expression of a mature lower crustal flow front developed over the last ~3 Myr; horizontal mass displacements (~50 km) associated with this crustal flow episode may have advected mantle plume geochemical signatures toward the southeast, and in directions not necessarily parallel to the hotspot track. Also, the stepped terraces along the southwestern platform may be explained by lower crustal flow-associated backward tilting of the bathymetric surface that, although resulting in small angular changes (~0.1°), effectively hinders the horizontal flow of lava sheets. This backward-tilting process was likely restricted to the last ~1 Ma or less, and may be a unique event involving extrusion of lavas from within the southwestward-marching lower-crustal flow front.

1. Introduction

The Galápagos Islands exhibit many unusual characteristics relative to other ocean-island provinces, features made famous of course by Darwin’s extraordinary geological and biological discoveries there in the 19th Century. From the physiographic and geologic standpoint, this island archipelago and the surrounding seafloor display features that remain largely unexplained to this day, including the early-noticed “Darwinian trends” of structural and volcano alignment, the occurrence of volcanism along a broad elevated platform, the diversity of lava and volcano types, and the complicated spatial and temporal patterns of geochemical signatures that must reflect the interaction between the Galápagos hotspot (mantle plume) and the adjacent Galápagos Spreading Center (GSC), located just to the north of the Galápagos archipelago (illustration on Figure 1a). The thickened crust of the paired Cocos and Carnegie Ridge systems has resulted from the location of the Galápagos hotspot approximately beneath the GSC for the past ~20 million years [Werner et al., 2003].

An unusual characteristic of the Galápagos Archipelago is the broad distribution of volcanism, which extends well outside of the present mantle plume location beneath the lithosphere [Villagomez et al., 2014]. Another related aspect is the somewhat non-time-transgressive eruptive history along the E-W direction of Nazca plate motion relative to the hotspot reference frame [Geist et al., 1988]. The plume is currently located beneath the large, active shield volcanoes that have formed Fernandina and Isabela Islands (Figure 1b), yet there has also been recent volcanism on some of the easternmost islands such as San Cristobal, as well as older eruptions in the central platform, e.g., Santa Fe Island [Geist et al., 1988]. Adding to this complicated scenario is that the trace element and isotopic signatures of the Galápagos volcanoes do not fit the expected pattern—although the plume signature along the GSC becomes stronger as the Galápagos hotspot is approached from both east and west [Detrick et al., 2002; Christie et al., 2005], the most depleted island and seamount signatures are found in the center of the archipelago, while the most hotspot-like (enriched) signatures are found around the margins of the platform [Geist et al., 1988; Hoernle et al., 2000].
This pattern may result from thermal entrainment effects within the plume [Richards and Griffiths, 1989], inherent heterogeneous “streaks” within the plume [Farretani and Hoffman, 2012], lithospheric thickness variations [Gibson and Geist, 2010], or perhaps unknown aspects of plume-ridge interaction [Feighner and Richards, 1995; Ito et al., 1997; Ito and Bianco, 2014; Gibson et al., 2015].

These fascinating features and many others, have been intensively studied since Darwin’s early explorations, with most of the emphasis being on the volcanology and geochemistry of the islands, and more recently of the surrounding seamounts [White et al., 1993; Harpp and White, 2001], as well as the tectonic evolution of the region [Hey et al., 1972; Hey and Vogt, 1977]. Figure 2 illustrates the recent evolution of the Galápagos Archipelago in stages, starting approximately 3 million years ago, during which most of the present-day platform and volcanic islands have developed [Geist et al., 2008]. The sequence of events illustrated in Figure 2 suggests a question that seems to have hardly been addressed in previous studies of the Galápagos system: To what extent has the broad, relatively flat Galápagos platform, much of it lying within just several hundred meters of sea level, formed as the result of volcanic (extrusive) construction that can be associated with the main island-forming volcanoes and surrounding seamounts, and to what extent has it formed by intrusion of plume-derived magma into the lower crust with subsequent horizontal (gravity-current) flow of this material? In this paper, we explore the possibility that after crustal intrusion, horizontal spreading may exercise a first-order control on the construction of the modern Galápagos platform (last ~6 Ma), as well in relation with the analogous Cocos and Carnegie Ridges dating back to 20 Ma.

A number of lines of evidence suggest that lower crustal flow is important in the formation and evolution of the Galápagos platform. First, the breadth and flatness of the platform itself is otherwise difficult to explain solely in terms of volcanic processes. Indeed, the situation may be somewhat analogous to the on-ridge Iceland hotspot, where lower crustal intrusion and horizontal flow are thought to play major roles in shaping the Iceland Plateau as a whole [Jones and Maclellan, 2005]. Second, the morphology and underlying seismic structure of the Cocos and Carnegie Ridges suggest that most of the thickened crust found there is intruded gabbro plus perhaps other more mafic rock [Hooft et al., 2010; Villagomez et al., 2011], with some evidence that lower crustal flow along the Carnegie Ridge at about 1° 10’S latitude likely continued after and extended southward beneath surface volcanic construction along the ridge [Richards et al., 2013; see also Sailhares et al., 2003, 2005].

A third line of evidence, and a major focus of this paper, comes from some of the more recently formed structural features of the modern Galápagos platform. The Great Southwest Escarpment (GSE) is one of the most enigmatic, unexplained features found at any hotspot province on Earth (Figure 3). Running along the SW margin of the present platform, the seafloor here drops about 3 km to abyssal depths over a horizontal distance of only about 30 km, or an average slope of about 10% (or 5.7°). It does not appear that this escarpment is caused by a master normal fault [Feighner and Richards, 1994], but rather that it is made up of huge, back-stepping or terraced lava flows accumulated in a systematic way [Geist et al., 2008]. In this paper, we suggest that the GSE may be explained by a lower crustal flow front capped by lava flows which in turn may originate in neighboring platform volcanoes as well as local breakouts in the flow front itself as it moves southwestward.

Lavas erupting on the platform surface and near the escarpment flow over the subhorizontal portions defined by this surface, which is itself made up of older basaltic lava flows that, under normal conditions, would exhibit slope-angles of less than 1° (S. Self, personal communication, 2014) [Deschamps et al., 2014; Umino, 2012]. Any process altering these surface slope-angles is likely to alter the style (dynamics) of future lava flows, as the driving force is the downslope gravity component with lava viscous braking controlled by cooling.

We hypothesize that southwestward-directed regional mass transport in the deep crustal regions, occurring on time scales of order ~10^5 years, is likely to tilt the (local) bathymetric surface of the platform rendering it more horizontal in time, and thus producing an exceptional environment for lava flows that is less favorable for long distance travel. In this way, new lava flows will likely travel shorter distances, creating a receded, backward-stepping set of terraces. Finally for this area, we will discuss and show that the possibility of additional geomorphic shaping mechanisms, like the ones given by paleosealevels effects, is unlikely.

Figure 1a indicates several other elongated bathymetric escarpments bounding the platform toward the east that, although less dramatic than the GSE, we suggest may also have been shaped by regional deformation of the lower crust. Unfortunately, no high-resolution bathymetric imaging has been done on these latter features, and so it is presently difficult to know to what extent they may be analogous to the
present-day GSE. However, it is reasonable to suppose that the processes occurring at present along the GSE may be similar to the processes that formed the bounding escarpments along the southern slopes of the Carnegie Ridge and the northern slopes of the Cocos Ridge (with some subsequent geomorphic reshaping).

Figure 1. (a) Geographic context and bathymetry of the Galápagos archipelago. The Galápagos islands span equatorial latitudes, and are the subaerial expression of a broad and thick volcanic platform constructed on young (<15 Myr) crust and lithosphere. The Galápagos hotspot is located under the westernmost area of the archipelago, where there is also a higher concentration of active volcanic centers. (b) Geologic setting of the Galápagos: young crust is created at the GSC (reference isochrons in red), and is offset by the right-lateral 91°W transform fault zone (TFZ), creating a ~5 Myr age discontinuity mainly in the northern Galápagos [Mittelstaedt et al., 2012]. The active volcanoes are the surface expression of extensive hotspot-associated magmatism, processes that have constructed the crustal platform and thermally reset the lithospheric age. The orange-dashed ellipse marks the inferred present-day location of the hotspot [Villagomez et al., 2014]. Thick white arrows show the plates’ velocities with respect to the deep-hotspot reference frame [Werner et al., 2003; Cuffaro and Doglioni, 2007; Morgan and Morgan, 2007]. On the eastern platform region, remarkable bathymetric features with elongated rim morphology are the surface expression of Proposed (lower crustal) Flow Fronts (PFF1 and PFF2). Green straight lines are the modeled transects: (SWT) South-Western transect and (EPT) Eastern platform transect. WDL is the Wolf-Darwin lineament.
In this way, a fourth line of evidence and also a major focus of this work is the observation that in the eastern regions of the Galápagos archipelago, the platform decreases in elevation with a striking morphology: a pronounced, gently curved bathymetric escarpment or rim that extends over approximately 180 km, east from Española and San Cristobal islands and representing the platform natural boundary to the east (PFF1 in Figure 1b, where “PFF” refers to “proposed flow front”). This rim-like bathymetric feature, with roughly 1.5 km of platform relief and showing remarkable horizontal continuity, does not seem to be created purely by volcanic construction and regional crustal intrusion. Following several lines of argument, we suggest that this bathymetric feature may have been influenced by lower crustal flow occurring on a hotspot-affected weak crust, and that after roughly ~3 Myr of evolution it may now be in a mature dynamical stage, probably evolving very slowly or perhaps even “frozen.” For this area, we will additionally discuss how paleo-sealevel effects might have played a role in geomorphic shaping, specifically in the upper parts of this escarpment.

East of the PPF1 escarpment, elevations decrease smoothly while otherwise showing mainly one gentle rim (PFF2 in Figure 1b) that also exhibits remarkable continuity, until it finally merges into the Carnegie ridge, the distinctive and broad bathymetric high that smoothly extends toward the Nazca-South America subduction zone, and which is believed to be part of the Galápagos hotspot track [Richards and Griffiths, 1989, Werner et al., 2003] in response to the relative Nazca plate velocity.
Last, the phenomena of lower crustal intrusion (inflation by feeding) and lower crustal flow (relaxation) appear ubiquitous in ocean island settings from both the theoretical standpoint and from numerous high-quality seismic refraction studies along other hotspot tracks [Richards et al., 2013]. Furthermore, at least several hotspot tracks formed either at spreading ridges or on very young oceanic lithosphere (e.g., the Ninety-east Ridge between latitudes 5°N and 27°S on the Indian plate, and the Nazca Ridge offshore of Peru on the Nazca plate) appear to be remarkably smooth, that is, lacking in major volcanic edifices, suggesting most of the volume of excess crust is emplaced by lower crustal intrusion and flow. We hypothesize that in many of these ocean island systems (Galápagos, Iceland, Ninety-east, Nazca, Easter Island, and likely others) the dominant form of crustal thickening and aseismic ridge construction is likely intrusion and lower crustal gravity flow while the mantle plume is located at or near a spreading ridge on young oceanic lithosphere.

Here we develop models for lower crustal flow beneath the Galápagos platform in order to address these observations, and in order to suggest further tests of the lower crustal flow hypothesis. A particular focus is the possibility that the unusual back-stepping lava terraces along the GSE may be shoaled due to backward-tilting as a lower crustal flow front progresses southwestward from the center of the active platform. If such flow is indeed occurring beneath the Galápagos, it may help explain why the platform is so broad, and why it is bordered by conspicuous escarpments. The lateral spread of intrusive magmas may also be related to the curious patterns of geochemical signatures noted above [Happ and White, 2001; Hoernle et al., 2000], as well as the long-noted extraordinary variety of volcanic products and forms occurring on the islands [Christie et al., 1992].

2. Methods

Our modeling approach follows from the work of Jones and Maclellan [2005], who described the crust beneath the Iceland Plateau as a Stokes fluid with variable viscosity. In turn, their study follows from the work of Huppert [1982] and McKenzie et al. [2000]. Our approach goes beyond that of Jones and Maclellan [2005] in several ways that allow better application to the Galápagos. A complete mathematical description is found in Appendix A, notes on the numerical implementation are in Appendix B, and benchmarking and sensitivity analyses are given in Appendix C.

2.1. Physical Formulation

We model lower crustal flow in two Cartesian dimensions, so that 3-D out-of-plane flow is not included. This simplifies the analysis and computation, and is justified because both of the suggested “flow front”
morphologies we focus upon are linear features. Figure 1b shows the two transects: a Southwestern platform transect (SWT) and an Eastern platform transect (EPT). The crust and mantle are assumed to be in isostatic equilibrium, i.e., vertical movements are rapid compared to horizontal flow response times. Mass transport is computed using a 2-D thin-sheet approximation of the Stokes equation, where the crust is described as an incompressible Newtonian fluid whose viscosity varies in time and space. In the thin-sheet approximation, the system is, at all times, of horizontally extended geometry, and all variables are assumed to have much longer characteristic length scales in $x$ than in $z$, so that

$$v_x \gg v_z \quad \text{and} \quad \frac{\partial (\cdot)}{\partial x} \ll \frac{\partial (\cdot)}{\partial z} \quad (1 \text{ and } 2)$$

where the velocities $v_x, v_z$ are with respect to a reference frame attached to the moving Nazca plate.

2.1.1. Boundary Conditions

The following boundary conditions are applied in our models (illustration in Figure 4):

1. We impose zero mass flux across the vertical boundaries at both ends of the model box. Zero horizontal velocity on the platform highland lateral boundary (located over central platform points), taken as $x = 0$ at the left model box boundary, is justified by the fact that the central platform is spreading (by deformation) in opposite directions, thus horizontal velocities are presumably small in the central area. The zero horizontal velocity on the distal boundary (at $x = L$, where $L$ is the transect length) recognizes that normal (background) oceanic crust located far away from the platform and volcanic centers is not affected by lower crustal flow within the platform.

2. Horizontal velocities are zero on the lower boundary (Moho), assuming that upper mantle rocks have higher creep strength than crustal rocks [McKenzie et al., 2000]. This is not an entirely satisfactory assumption within the central platform, where plume-derived magmas are heating both the mantle and crust. Vertical motion at the Moho is determined by isostasy.

3. We impose zero horizontal velocity at the upper boundary (bathymetry), where the cold uppermost crust behaves rigidly with respect to the softer lower crust below. We do not consider brittle failure of the upper crust, although such behavior could be used post facto to infer stress patterns. Vertical motion here is found by solving the governing equations.

The above boundary conditions are formulated mathematically in Appendix A.

2.1.2. Initial Condition

Due to the nature of the partial differential equations we are solving, the initial condition for the velocity field is determined by the initial conditions of the topography (bathymetry) and the initial temperature field, which in turn determines the initial viscosity structure. Let’s consider $x$ as the horizontal coordinate, $t$ the time, and $h(x,t)$ the bathymetric height function.

We adopt an arbitrary but convenient form for the initial bathymetry function:

$$h(x, 0) = h_{\text{crust}} + h_{\text{platform}}(x) \quad (3)$$

where $h_{\text{crust}} = \frac{6}{(1+f)} \ln(\text{km})$.

Figure 4. Model sketch: the 2.8 g/cm$^3$ crust is in isostatic equilibrium with a 3.3 g/cm$^3$ mantle, therefore providing our reference level for the elevation. Mass transport is computed using a 2-D thin-sheet approximation of the Stokes flow equation, where the crust is described as a Newtonian fluid with space and time-dependent viscosity. The imposed boundary conditions are summarized as zero mass flux at the lateral boundaries, and horizontally-rigid upper and lower boundaries.
therefore $h_{\text{crust}}$ is the constant elevation of the top of a 6 km thick background crust over the isostatic equilibrium reference level (determined by the combination crust versus mantle densities). This equilibrium level is taken here as the reference for the vertical coordinate, so the crustal lower boundary is defined simply by

$$h(x,t) = -f \cdot h(x,t).$$

(5)

is the platform thickness function, with $N_{\text{steep}}$ an even number and $H_{\text{plat}}$ the platform’s initial maximum thickness. This initial platform function mathematically corresponds to the Butterworth filter equation.

The platform’s horizontal length scale is $\sim 35\%$ of the model box’s horizontal length, $x_0 = 0.35 \cdot L$. The effect of the value of the exponent $N_{\text{steep}}$ is investigated in the sensitivity analysis of Appendix C. The effect of the functional form for the initial bathymetric profile is also addressed in Appendix C.

### 2.1.3. Variable Viscosity

Like Jones and Maclennan [2005], we assume that viscosity is solely a function of temperature:

$$\eta = \eta_{\text{solidus}} e^{[(\alpha/\eta_{\text{ave}}) - 1]}$$

(6)

upon which we impose the restriction $\eta_{\text{min}} \leq \eta \leq \eta_{\text{max}}$ in accordance with physical criteria (Appendix A3).

This is a reasonable approximation for the range of depths and thermodynamic conditions present in the crust, where pressure-dependence of viscosity is of secondary importance. The thermal field consists of two components: (1) Thermal aging of the crust, described by a half-space cooling model and adapted for subhorizontal topography [Jones and Maclennan, 2005]. Additionally, a constant horizontal age gradient is considered due to the lithospheric age increase moving southward from the Galápagos spreading center. (2) A hot thermal perturbation of variable strength (decreasing away from the platform area) is added to account for the continuous injection of magma into the lower crust due to the melting mantle plume beneath the platform, which, of course, is the cause of the widespread volcanism. This component has the form of a 2-D Gaussian perturbation centered at the platform side of the model box and at Moho depths. For simplicity, its amplitude remains constant in time.

A complete description of the Temperature-Viscosity parameterization is given in Appendix A3 section.

### 2.2. Modeling Calculations

The model equations (Appendix A) are obtained from the thin-sheet approximation of Stokes flow with space and time-dependent viscosity, and are solved numerically using finite differences in space and time. The computer program was written in C language and is available from the first author upon request. Post calculation output graphics, plots, and animations were done in Matlab. The finite difference grid adapts to the time-changing geometry of the crust. Integrals over the vertical coordinate were carried out using Simpson’s rule for quadrature. Time integration of the controlling equation was done using an explicit second-order Runge-Kutta scheme, also known as the “improved Euler method.” This scheme yields second-order accuracy. Numerical solution was intrinsically carried out in nondimensional form with scaling parameters based on $L$ (system’s horizontal length), $c$ (aspect ratio), $\rho_c, \rho_m$ (crust and mantle mass densities), and $\eta_{\text{ave}}$ (average crustal viscosity). Our code was benchmarked successfully with isoviscous solutions from Huppert [1982] and McKenzie et al. [2000] (see Appendix B).

### 3. Results

We present two models here in detail. The first is intended to apply to the Southwestern Escarpment Transect (SWT) and the second to the Eastern Platform Transect (EPT). These two models were optimized to match approximately the bathymetric and age constraints, and the reader is referred to the model sensitivity analysis in Appendix C for a detailed explanation of model parameter trade-offs.
3.1. Southwestern Escarpment Transect

This transect (Figure 1b, “SWT”) is 350 km long and spans from the central platform across the Great Southwestern Escarpment and onto apparently normal seafloor of age \( \sim 20 \) Ma. The central platform lithospheric reference age at the northeastern end of the transect (and west from the 91\(^\circ\)W transform fault) is 10 Ma. These are just reference ages, considering that in our model a hot-thermal perturbation will be superimposed in association with the plume and active volcanism (implying thermal rejuvenation). The reference viscosity values are \( \text{visc}_{\text{min}} = 4 \times 10^{17} \), \( \text{visc}_{\text{solidus}} = 4 \times 10^{18} \), and \( \text{visc}_{\text{max}} = 4 \times (5.5 \times 10^{22}) \), all in [Pa·s], the factor 4 finally chosen to yield an adequate process duration; all in reasonable agreement with Jones and Maclellan [2005], Hirth and Kohlstedt [1996], and Mackwell et al. [1998].

The somewhat arbitrary initial condition includes a 3 km high platform with a slope exponent of \( N_{\text{steep}} = 10 \), approximately matching the present-day heights over the background seafloor. We assumed a 6 km thick background crust, in agreement with Ito et al., [1997], and with seismic studies for the Nazca plate [Contreras-Reyes and Carrizo, 2011]. The initial background thermal field is referenced to 1 Myr ago, consistent with the approximate age of this part of the platform [Harpp and White, 2001; Geist et al., 2008] so that the reference ages of the thermal field were 9 and 19 Ma at the northeastern and southwestern ends of the transect, respectively. The effect of continued heating associated with magmatic intrusion into the platform is simulated by a hot thermal perturbation with length-scale \( L_T = 0.3 \) (see Appendix A), (Table 1 summarizes the model parameter values for the transect). As will be seen, within less than 1 Myr the model quickly relaxes toward a natural quasi steady state profile in bathymetry and particles velocity.

Figure 5 shows snapshots of the evolution of this model over a 1 Myr period from the initial condition in terms of the viscosity and horizontal velocity fields. The viscosity field evolves only slightly (solely by crustal aging or cooling, while dominated by the hot thermal perturbation), while the horizontal velocities decrease by more than an order of magnitude as the model “relaxes” from its initial bathymetric profile shape to a more stable flow front shape. As expected, the flow is mainly confined to a lower-crustal region where there is a constructive combination of low viscosity (hot and weak material) and a nonzero driving force (bathymetric gradient above).
As suggested in the introduction, for this transect we are particularly interested in the evolution of the bathymetric gradient, with the hypothesis that backward tilting of the behind-the-escarpment region may be the cause of the unusual observed backward-stepping lava terraces. With regards to typical submarine volcanic morphologies, mere fractions of a degree of backward tilting would suffice to limit the extent of lava flows spreading away from the platform toward the SW.

Figure 6 shows the bathymetry, its slope, and the time-derivative of the slope (bathymetric tilting rate or angular velocity) at the logarithmic times {9.0, 9.01, 9.1, 10} Ma. As time progresses, there is a slight decrease (total of \( \frac{40}{C_24} \) m) in platform height while an acute escarpment-hinge with an increased slope region just below develops, all accompanied by horizontal advancement. The tilting rate exhibits two opposite-sign regions: a forward tilting (steepening) region at the foot of the escarpment and a backward tilting (becoming more horizontal) region at the top and behind it. Figures 7–9 show essentially the same information, displayed as two-dimensional plots over a continuum time domain (9–10 Ma). The color-coded plot of the tilting-rate shows the backward-tilting region in hot colors. The rates decrease in time (asymptotic behavior) as they translate horizontally with the flow front.

Considering a bathymetric surface segment (between fixed points in the horizontal) and measuring its slope-angle versus time gives an account of the angular conditions that would control the spreading of lava flows over this bathymetric surface segment. For instance, we have isolated a fixed 10 km interval in the horizontal: \((81,91)\) km (grid-dependent), and we have measured the spatial average of the bathymetric slope-angle over it, as a function of time (double checked by integrating the space-average tilting-rate over time), then plotted this in

![Bathymetry functions](image)

**Figure 6.** Topography functions: \(h(x,t)\) and spatiotemporal derivatives (slope angle and slope-angle time rate). Time snapshots are represented by color code. Legend at center-right is valid for the three plots; color goes as green-red-blue-black as time increases: numbers indicate millions of years since reference initial time (9 Myr).

### Table 1. Model Parameter Values

<table>
<thead>
<tr>
<th>Parameter</th>
<th>SW Transect</th>
<th>EP Transect</th>
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<tbody>
<tr>
<td>(L)</td>
<td>350 km</td>
<td>350 km</td>
</tr>
<tr>
<td>(H_{\text{plat}})</td>
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<tr>
<td>(x_c)</td>
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<td>(n_{\text{steep}})</td>
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<td>6</td>
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<tr>
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<td>(\eta_{\text{max}})</td>
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<tr>
<td>(\eta_{\text{initial}})</td>
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<td>1 \times 10^{22} + 10^{23}/2 Pa s</td>
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<tr>
<td>(\rho_{\text{mantle}})</td>
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</table>

\(^a\)The parameters \(H_{\text{plat}}, x_c, n_{\text{steep}}\) apply solely as initial conditions.

\(^b\)The parameters \(x_c, n_{\text{steep}}, LTx, LTz\) are non-dimensional.

\(^c\)The age variation “\(\text{Age}_{\text{transect}}\)” along the EP transect could be larger (by \(\sim 5\) Myr), depending upon the history of motion of the 91°W transform, following Mittelstaedt et al. [2012].
Figure 10. The result is that an initial slope-angle of $\sim0.28^\circ$ is progressively reduced in time, rendering the surface more horizontal, down to a final slope-angle of $\sim0.01^\circ$ after 1 Myr. The process is rapid at the beginning: roughly 90% of the angular reduction occurs during the first quarter million years. This behavior can also be appreciated in the bathymetry-derivatives profiles of Figure 6.

Similar results are obtained for other averaging intervals in the proximity of the platform edge and the escarpment, and in general for other sets of parameter values. The initial average slope over the chosen interval, the process duration, and the final slope over the interval were found after a parameter-space search, and are geologically relevant in a submarine volcanic environment (S. Self, personal communication, 2014). For instance, the viscosity values were chosen in order to adjust the duration of the process, and are specific for this study transect. Owing to the viscosity values and the length-scale of the hot thermal perturbation, the horizontal particle displacements (mass transport) in the lower crust during the 1 Myr time span are about 8 km: the flow is braked by the high viscosities beneath the escarpment and the off-platform crust. In this way, we tried to mimic the flexural conditions prevailing there as suggested by the gravimetry-inferred lithospheric compensation mechanisms [Feighner and Richards, 1994].

Mass transport away from the platform-center due to lower crustal flow causes the subhorizontal portion of the platform bathymetry to become more horizontal as the flow progresses. While this process occurs generally for these types of dynamical settings, in our model a highly viscous (stiff) crust suffering minimal mass transport produces the required backward-tilting (fractions of a degree) of the subhorizontal platform region behind the escarpment.

### 3.2. Eastern Platform Transect

This transect (Figure 1b, "EPT") is 350 km long and extends from present-day background crust reference thermal ages of 7.5 Myr under the central platform (and east from the 91°W transform fault), to 16 Myr (or possibly as much as 20–22 Myr following Mittelstaedt et al. [2012]) off the platform toward the southeast.

The reference viscosity values are $\nu_{\text{visc,min}} = 10^{17}$, $\nu_{\text{visc,liquidus}} = 10^{18}$, $\nu_{\text{visc,max}} = 5.5 \times 10^{22}$; all in (Pa s). The value of $\nu_{\text{visc,max}}$ was chosen as the arithmetic mean between $10^{22}$ and $10^{23}$ (the system is not very sensitive to these parameters, and the values chosen are
reasonable for the shallow crust). The other two viscosity values were chosen to obtain reasonable time scales and a significant degree of flow-front advance in the bathymetry (a likely effect owing to the weak thermomechanical conditions) (see sensitivity analysis in Appendix C, section on time scales and front advancement versus viscosities). These values are in reasonable agreement with Jones and Maclennan [2005], Hirth and Kohlstedt [1996] and Mackwell et al. [1998].

For the initial condition on bathymetry, a 2 km thick platform was chosen. This platform thickness value is justified because: (i) present-day local platform relief is between 1 and 2 km (approximately); and (ii) owing to the mechanically weak conditions on the eastern platform region, the construction of a thick, high elevation volcanic platform is precluded by relaxation of the lower crust. The initial bathymetry slope exponent is $N_{steep} = 6$. This value was chosen after exploration of its effect on the final state of the bathymetry after 3 Myr of evolution (see Appendix C). We assumed a 6 km thick background oceanic crust, in reasonable agreement with Ito et al. [1997] and with Contreras-Reyes and Carrizo [2011]. Table 1 summarizes the model parameter values for the transect.

The initial condition on the thermal field is referenced to 3 Myr ago, so the reference thermal ages at the transect extremes are 4.5 Myr at the platform-center end and 13 Myr (or perhaps up to 18 Myr) at the off-platform end toward the southeast (this uncertainty does not have a big effect on the system behavior due to locally high viscosities). Nevertheless, the duration of this tectonic episode, even after parameter exploration, is considered a weakly constrained quantity. Certainly, construction and morphological shaping episodes within this area of the platform have occurred during the last ~5 million years [Geist et al., 2008], but the timing uncertainties are large (see discussion section). The hot thermal perturbation’s horizontal length scale was, after parameter space investigation, chosen to be $L_x = 0.45$. This rather large value is consistent with the weak central platform and hotspot track type of environment [Feighner and Richards, 1994], and with the transect location and orientation. Figure 11 shows snapshots of the model evolution over the 3 Ma time-span. Owing to the asymptotic behavior exhibited by the system, snapshot time-spacings are chosen accordingly, and the snapshot times represent the reference ages of the background crust at the platform-center end of the transect (from 4.5 to 7.5 Myr).
The viscosity field evolves only slightly (solely by thermal aging and barely resolved by the color scale) and is dominated by the hot thermal perturbation. The horizontal velocities decrease dramatically in time: the region of high velocities, initially with values of about 50 cm/yr, becomes progressively confined to a narrow and deep crustal region with final values of order 0.1 cm/yr causing the “relaxation” of the system from its initial bathymetric profile shape to a more stable flow front. As expected from the theoretical standpoint, flow is mainly confined to a lower-crustal region where there is a constructive combination of low viscosity (hot and weak material) and a nonzero driving force (bathymetric gradient above). The initial platform thickness is 2 km (full bathymetric height 2.909 km above the isostatic reference level), and after 3 Myr of evolution the final thickness is 1.79 km (2.699 km over the reference) (see Figures 12 and 13), close to the observed bathymetric height differences across the eastern platform escarpment of about 1.5 km (over horizontal distances of around 20–30 km, measured along the transect). This height reduction is a consequence of mass conservation: the platform “deflates” by evacuation of the material from its lower regions. The system develops a well-defined bathymetric front of increasing slope. Figure 14 (top) shows the locus of maximum slope (green) versus time, and the loci of slope 20% of that maximum (red). Within these limits, the average slope is computed (region of at least 20% the maximum) and it is found to increase in time up to a

Figure 11. Eastern platform transect system evolution. Note the changing scale of the horizontal velocity component. The system evolves rapidly during the early stages, hence the uneven spacing of the snapshot times. The material portion affected by lower crustal flow driving forces moves away from the hot and weak region. After a transient period lasting less than 1 Myr, the system has developed a bathymetric flow front that advances without changing its shape significantly. The flow front evolves asymptotically toward a steady state. Original model box’s right end has been cropped for illustration purposes. We have estimated the position of three particles versus time to show accumulated displacement: drawn as white markers leaving a tail (shown only on the left plots), with their final position as black dots. Vertical motion has been neglected.

Figure 12. Topography functions: h(t), and spatiotemporal derivatives (slope angle and slope angle-time rate). Time snapshots are represented by color code. Legend at center-right is valid for the three plots; numbers indicate millions of years since reference initial time (4.5 Myr).
final value of ~5.4% (Figure 14, bottom). This value is approximately the observed present-day slope value of the eastern platform escarpment in the vicinity of Isla Española (Figure 1a).

The system experiences surface flow front displacement of order 30 km during the 3 Myr time-span (Figures 11–14), but we note that given the boundary condition the horizontal velocity is zero on the bathymetric surface. Therefore this front merely follows the isostatic response to the redistribution of mass underneath. Total maximum horizontal particle displacements within the deforming lower crust are of order ~50 km (shown in Figure 11). This displacement is away from the central platform and along the direction of the transect. This result has geochemical implications that are treated in the discussion section.

In summary, the bathymetric contrasts (lithostatic pressure gradients) on this young hotspot-affected lithosphere suggest that lower crustal flow is likely, and our simulations account reasonably for some basic observed bathymetric features; hence, our label “Proposed Flow Front One” (PFF1) in Figure 1b.

4. Discussion

4.1. Southwestern Escarpment Transect

Feighner and Richards [1994] showed that the Great Southwestern Escarpment (GSE) and nearby off-platform crust are located over flexurally supported lithosphere with elastic thickness ~12 km, but that its upper edge is located over a narrow transition to near-Airy isostatic compensation toward the northeast [see Feighner and Richards, 1994; Figures 12a and 13a]. Therefore, the thermally weakened regions close to the escarpment may undergo lower crustal flow away from the platform toward the southwest, with flow blocked by the stronger off-platform lithosphere to the southwest that remains undisturbed by Galápagos-related volcanism.

A key focus of this study is to try to understand what processes control the emplacement of the spectacular lava flow terraces off the GSE as mapped by Geist et al. [2008]. A working hypothesis is that the seemingly unique backward-stepping nature of these terraced flows indicates regional (“backward”) tilting toward the central platform during emplacement. Surface slope-angles of massive lava flows cooled in oceanic environments, averaged over distance, are typically less than ~1°. These are the slopes that control the subsequent lava runoff of future extrusive events (S. Self, personal communication, 2014) [Deschamps et al., 2014; Umino, 2012]. Given that the driving force of any lava flow is the along-the-slope gravity component, a mere fraction-of-a-degree modification of the slope-angle on these surfaces would alter the runoff distance of future lava flows.

For the GSE, we have demonstrated that even very small amounts of lower crustal flow (less than 10 km of horizontal mass displacement) may suffice to tilt the bathymetric subhorizontal upper parts of the platform (behind the escarpment’s edge) in a backward sense, with total angular decreases of more than 90% of the initial slope-angles during a time span of 1 Ma. The platform surface, becoming more horizontal with the ongoing mass-transport in the lower crust, experiences a decreasing along-the-slope gravity component with time. Thus, lavas flowing over this backward-tilted surface would experience a decreased driving-force, which eventually may cause the full braking of the lava flow (by viscous dissipation), on the time-scale of thermal diffusion (cooling of the lava). Lava flows will then cool down and stop before the flow edge of older events (past eruptive episodes), creating a receded step. This mechanism may operate successively, thus creating a system of stepped terraces, significantly differing from typical basalt flow-fields in which
lava flows cool down over horizontal distances not necessarily ordered with respect to previous flows. In this way, the mechanism outlined here and shown to operate for appropriate model conditions applying to the southwestern escarpment region, may explain the unique morphology of the GSE. A key aspect on the applicability of our idealized isostatic model to the real Earth scenario in which there is a spatial transition from isostasy (central platform) to flexure (off-platform) in the vicinity of the platform-edge, is given by the high viscosities on the under-the-escarpment and off-platform crustal regions chosen for the model, which are all above $10^{21}$ Pa·s, primordially owing to the short thermal perturbation length-scale ($L_T = 0.3$, see snapshots on Figure 5), with minimal mass transport. We believe this “highly damped” crustal flow scenario is appropriate for a region of the Galápagos crust bounded by a flow-resistant flexural lithosphere.

Aside from the volcanic sources on the platform, magmas might be extruded directly from the escarpment surface, stemming from a dike-and-sill complex that likely exists there beneath the platform surface as part of the magmatic system. The locations of the vents for the terrace-building lava flows are unknown, but the

Figure 14. Bathymetric Slope Angle function (above) with markers for the maximum slope location (green) and the points where the slope is 20% of the maximum (red). Using these slope-value defined bounds, the spatial average flow-front slope is computed as the mean value inside that region, for all times. These values are then plotted as slope and slope angle (below).
high-resolution imaging of Geist et al. [2008] suggests that lavas may be extruded directly from the GSE itself as massive “break outs,” producing a system of basaltic lobes.

Along the GSE numerous steps/terraces have been imaged [Geist et al., 2008], exhibiting widely different horizontal extents and vertical relief. A plausible scenario is that each step is the accumulation of several shoaled lava flows stemming from a single eruptive episode [Geist et al., 2008]. The differences between the individual heights and volumes of different steps could then be related to the spatiotemporal eruptive behavior, modulated by bathymetric tilting associated with lower crustal flow. According to our models (Figures 5–10), eruptive events occurring during episodes with total durations of $<10^3$ years would experience a quasistatic bathymetric profile (owing to system response timescales $>10^4$ years), and progressive lava flows should produce the natural slope-angles of the bathymetric surface after cooling. For subsequent eruption episodes occurring $\sim 10^4 - 10^5$ years after a quiescence period (with few or no eruptions), owing to the accumulated tilting given the system’s response, lava flows will encounter a more subhorizontal surface. Figure 15 illustrates how long-term tilting between lava flows could result in backward-stepping lava terraces. Variations in the durations and volumes of eruptive episodes, as well as in the platform tilting-rates, would cause variations in the thicknesses and volumes of the terrace steps. The aforementioned timings, in accordance with our modeling results, are considered reasonable for Galápagos volcanism [Naumann et al., 2002; Geist et al., 2014].

Finally, we recognize that in some situations volcanic platforms in oceanic settings may be “planed off” by sea level erosional processes, as suggested for certain features along the Hawaiian ridge [e.g., Mark and Moore, 1987; Moore, 1987]. However, Geist et al. [2008] discuss at length why these mechanisms cannot explain the
deeper (>1 km depth) terraces along the GSE, and show that these terraces consist of a vast collection of rugged and irregular lobes, or “tongues” of basaltic lava flows. Some degree of continuity and smoothness (more familiar of coastlines erosional processes) are found only in the most shallow slopes of the GSE, where sea level effects might plausibly have come into play over the last few million years during which the GSE must have formed. However, the paleoeseas hores suggested in the paleogeographic study of Geist et al. [2014] do not suggest that the GSE volcanic terrace features identified by Geist et al. [2008] were formed at or above sea level, instead, they suggest that such region of the platform grew totally under submarine conditions.

4.2. Eastern Platform Transect

The eastern platform exhibits less prominent bathymetric contrasts than the GSE, but with a distinctive morphological feature: a rim-like, asymmetric ridge, or escarpment. This structure is about 180 km long, with Isla Española forming the most elevated part, and with mild horizontal curvature along its northern parts, just east of Isla San Cristobal. When measured along the study transect and neighboring areas, the associated elevation contrasts are ~1.5 km over horizontal distances of ~20–30 km, corresponding to slopes of ~5–6% (~3.2°). The overall shape of this bathymetric structure suggests a front associated with lower crustal flow. Due to the eastward migration of the plate with respect to the Galápagos hotspot [Hey, 2010], the eastern platform was not only younger but considerably hotter and weaker in the recent past, likely undergoing intense deformation. Based on these considerations, we hypothesize that this bathymetric rim is the surface expression of lower crustal flow away from the platform and towards the southeast, occurring over the last ~3 Ma, and now having reached a mature, or perhaps “stagnant” stage.

Feighner and Richards [1994] inferred a transition from near Airy isostasy to elastic flexure southeast from the present-day location of PFF1, close to Española island. Nevertheless, considering that 3 Ma ago this region was hotter and weaker due to the presence of the Galápagos plume [Geist et al., 2008; Werner et al., 2003], we propose that this area may have been relatively weak compared to the present-day GSE. In our modeling, this weakness is reflected in the younger effective plate ages, smaller viscosity values (factor of ~3), and larger horizontal length scale of the hot thermal perturbation ($L_{x_{e}}=0.45$). Crustal age uncertainties in the southeastern Galápagos (shown e.g., in Figure 1b) would imply bigger age gradients along the transect, but ultimately, they have only a small effect on the results presented here.

Our simulations used initial platform thicknesses of 2 km, as explained in the Results section. In our preferred simulation, after 3 Myr of evolution the model exhibits a final platform thickness of 1.79 km and a final average slope across the escarpment of ~5.4% (although this value is dependent on how the average is taken), consistent with observations. There is an initial transient period that depends on our arbitrary initial conditions, and lasts for about 0.5 Myr during which high strain-rates and rapid front advancement occur. As seen in Figures 13 and 14, the position of the surface front advances monotonically at a speed that decreases with time. The front’s horizontal position behaves asymptotically in time (all variables showing decreasing time-rates).

In these simulations, the lower crust undergoes high shear-strains and suffers subhorizontal mass displacement, represented by channelized flow whose velocity distribution is shown in Figure 11. Buoyant mantle-plume melts ascending through this deforming crust would not follow purely vertical trajectories, as they would be partially advected with the flow (away from their original mantle upwelling regions in the central-west platform) toward the southeast. Calculating typical values of total horizontal displacements due to flow of the lower crust in the 3 Myr time-span of our simulation, we find values of ~50 km (with more than 80% of it occurring during the first 1 Myr). (In relation with the uncertainties of these estimates, the off-platform crustal ages uncertainty for this transect may be as large as 5 Myr (following the study of Mittelstaedt et al. [2012]), but nonetheless, their effect on the mobility of the lower crust is unimportant, as the viscosities in the off-platform region would be above $10^{21}$ Pa·s, and following our sensitivity analysis, an error of 10 km of total displacement would be too high). This 50 km displacement (shown in Figure 16), whose magnitude is roughly the distance between Santa Fe and San Cristobal islands, is a significant fraction of the platform’s horizontal extent. The horizontal displacement of these mantle-plume melts is proportional to the time they take in traversing the crust, which is in turn dependent upon whether they are creating a new flow channel (at the forefront of the mantle-plume tail) or exploiting a preexisting one (at some distance down-stream the hotspot track). Keller et al. [2013] show that the ascent speeds for melts that are creating a path through the crust, is of order 10–50 km/Myr; therefore, they would require a time of order ~1 Myr to ascend through the ~20 km thick Galápagos crust. Accordingly, 3 or 4 Myr ago, the mantle-plume-derived melts may have
ascended slowly enough to have been advected horizontally by bulk transport within the lower crust, presumably carrying the plume isotopic signatures, for distances of order 40–50 km. If sufficiently buoyant to erupt at the surface, these melts might have given rise to volcanic centers extended toward the southeast of their mantle origins. Although speculative at this point, we mainly just wish to make the case that this advection process may need to be considered in future interpretations of the pattern of geochemical anomalies in the Galápagos.

This transport might therefore influence the horizontal spatial patterns of plume-like geochemical signatures, which have been the subject of considerable debate in the Galápagos [Geist et al., 1988; White et al., 1993; Hoernle et al., 2000; Harpp and White, 2001]. Consider for instance the observation that the Galápagos platform exhibits more plume-like signatures on its margins [Geist et al., 1988]. We note in particular the work of Harpp and White [2001], where complex patterns can be observed. See for example Figure 16 where the ratio \(^{87}\text{Sr}/^{86}\text{Sr}\) shows an elongated maximum towards the southeast, perhaps consistent with directed lower crustal flow as along our eastern platform transect study, where the material flows away from the center towards the southeast. Harpp and White [2001]; attribute southeasterly elongated geochemical anomalies, such as the one exhibited by the high \(^{87}\text{Sr}/^{86}\text{Sr}\) of the FLO component (one of the isotopic components present in the Galápagos) to lithospheric dragging in the direction of plate motion with respect to the hotspot. Our models suggest a possible additional influence on the plume geochemical signature transport due to the deforming lower crust. Mass transport due to lower crustal flow may operate in a direction not parallel to plate motion.

The southernmost edge of the platform between Floreana and Española islands, in conjunction with the PFF1, exhibits considerable geometrical (horizontal) extension away from the hotspot track. Southeast-directed lower crustal flow occurring during the last ~3 Ma may have played a role in spreading and extending the crust locally (S. Gibson, personal communication, 2014), therefore increasing, perhaps even controlling, the areal extent of the platform.

Another interesting feature of the eastern platform is that the present-day PFF1 is not entirely a one-sided escarpment (dipping east), as it also exhibits an elevation drop on its western side toward the central platform. Thus, there is a two-sided, or “symmetric-ridge” aspect to its morphology. The steeper eastern flank dominates the morphology, and we have modeled it as a one-sided surface flow-front. The western flank is less pronounced (lower relief and slope), and cannot be explained by the gravity flow model for lower crustal flow advancing from the central platform. We hypothesize instead that the small western slope is explained by cooling of the interior of the platform, representing a more mature stage of thermal evolution than for the GSE.

Regarding paleosealevel effects, the proto eastern platform began forming roughly ~6 Myr ago, when the hotspot located fully beneath the Nazca plate. At that time, the growing volcanic edifice was being...

Figure 16. Map of the \(^{87}\text{Sr}/^{86}\text{Sr}\) anomaly over the Galápagos archipelago, adapted from Figure 7a of Harpp and White [2001]. Dashed black line marks the location of the Eastern platform escarpment, which we infer to be the surface expression of a lower crustal flow front (PFF1). Green arrows indicate the direction and magnitude of the lower crustal mass displacement, with a map-scaled length of 50 km. The elongated geometry of the Strontium isotopic anomaly might be partially explained by plate-hotspot relative movement, but here we suggest an extra component: lower crustal flow may play a role in transporting plume-like signatures with respect to the moving plate, slightly away from the hotspot track.
constructed upon young plume-affected lithosphere, likely starting from fairly shallow water depths of order ~1 km. With local sea level responding to climate shifts and plume-related dynamic topography, it is plausible that the upper parts of this protoplatform (perhaps at depths of ~0.5 km) might have formed paleoshorelines. Accordingly, Figure 8 of Geist et al.’s [2014] paleogeographic reconstructions shows paleoshorelines approximately along the present-day eastern platform escarpment ~0.5 km water-depth contour. However, a more prominent delineation of the present-day escarpment lies at the 1 km depth contour, markedly lower than the paleoshorelines proposed by Geist et al. [2014]. But we also note that our eastern transect model suggests ~200 m of subsidence over a 3 Myr time span, better placing the paleoshorelines as a cause for the present-day morphology. In any case, we suggest that it is likely that both lower crustal flow perpendicular to PFF1 and possibly paleoshoreline effects may have contributed to the morphology of PFF1, and sorting out these effects would require high-resolution bathymetric imaging of this feature.

4.3. Other Features Possibly Related to Lower Crustal Flow

A number of other bathymetric features of the Galápagos platform (and the Carnegie Ridge to the east) may have formed as a result of episodes of lower crustal flow (Figures 1a and 1b). For example, a notable ridge to the east of the platform and forming part of the Carnegie ridge, labeled PFF2 (proposed flow front 2), shows gentle but distinctive elevation contrasts with a horizontal extent of ~250 km. The geometry of this feature is hypothesized to result from lower crustal flow starting approximately 9 Myr ago, when the hotspot was located beneath the lithosphere there. Flow would have been directed toward the SSE. Owing to local thermal ages of ~8–11 Myr and noting the lack of obvious major centers of volcanic construction, we infer relatively high lower crustal viscosities. Consequently, this flow front may be close to stagnation. We note that the reflection/refraction study of the Carnegie Ridge by Sallares et al. [2005] indicates a remarkable feature consistent with the above inferences, where the thickened lower crust along the southern margin of the eastern Carnegie Ridge appears to extend horizontally beyond the bathymetric expression of volcanic construction toward the south. In other words, it seems possible that lateral spreading of warm lower crust may continue even after major volcanism has ceased. For this area, a role for paleoasealevels in shaping the morphology is plausible, and indeed the paleoaseashores (Geist et al. [2014]) between 5 and 3 Myr bear some resemblance with the PFF2 escarpment. However, the local paleoaseashore is rather composed of paleoislands, perhaps overlying a more continuous and connected feature like the one observed under the ocean today, so that sealevel erosion likely shaped only the most elevated parts of the present-day escarpment. We therefore suspect that this bathymetric feature, after been formed by magmatic intrusion, was mainly controlled by lower crustal flow and deformation.

Also shown in Figure 1b are localized bathymetric escarpments along the southeastern and eastern regions of Isabela island, and the northern regions of Santa Cruz island. These escarpments are important at the island length-scales and may have been created by recent volcanic construction. However, we suggest that they may be at least partially the result of recent lower crustal flow beneath this region of the platform, which is the hottest and weakest. Additional processes may have also shaped them, such as carbonate deposition on submarine slopes responding to eustatic changes.

Except for localized islands and seamounts, the central platform undergoes a smooth transition to regional background oceanic floor toward the north: no significant escarpments are observed [Mittelstaedt et al., 2012]. The northern platform region represents a different dynamical regime, due to younger crustal ages and time-dependent plume-ridge interaction associated with the northward migration GSC relative to the hotspot [Hey et al., 1972; Ito et al., 1997].

4.4. Remarks on Platform Evolution

The Galápagos platform may be considerably more voluminous now than in the past [Geist et al. 2008] (Figure 2). Repeated and episodic crustal intrusion events are not only plausible, but also consistent with recent seismic studies. Hooft et al. [2010] presented seismic refraction results for the Galápagos platform, finding thickened crust and platform edges characterized by low P-velocity consistent with high porosity, whereas for the interior of the platform magmatic intrusion accounts for most of the crustal section, as evidenced by seismic velocities, where the gabbroic lower crust constitutes roughly 2/3 of the total crustal thickness. These results are consistent with the modeling results of Feighner and Richards [1994], where a platform crustal thickness of slightly more than 16 km is inferred, with the lower ~6–8 km likely being gabbroic intrusion added to the background crust [see also Richards et al., 2013]. We also note that the westernmost islands of Isabela and
Fernandina are constructed upon older lithosphere than the central platform [Feighner and Richards, 1994], and thus the process of lower crustal intrusion and flow there may be less prevalent.

4.5. Future Studies
The results presented above suggest future studies that might lead to a better understanding of the evolution of the Galápagos Archipelago and GSC system. The greatest need is for combined seismic reflection and wide-angle refraction studies to resolve the structure of the crust, perhaps along transects such as those studied here. Seismic anisotropy studies could help confirm lower crustal flow, for instance on the eastern platform, due to a potential strain-driven orientation of minerals in the lower crust. In addition, the modeling work should be extended to include 3-D effects and to include predictions for stresses in the shallow crust associated with lower crustal flow, which in turn could be constrained by seismotectonic studies and topographic and bathymetric studies of faulting structures on the platform and islands. High-resolution gravimetric studies could help in developing a more robust model of crustal and elastic lithosphere thicknesses. Further consideration of the effects of 3-D (sub-horizontal) mass transport in the lower crust on the pattern of geochemical signatures might also modify interpretations of the latter. Last, more investigations are needed to clarify the nature of the extraordinary lava terraces along the SW escarpment, their sources, and related tectonic structures.

4.6. On the Asymptotic Behavior of the Flow Fronts
In our models, the velocities of both the lower crust particles and the advancing bathymetric front become progressively slower with time. Different histories are found for the two study transects (owing to different initial conditions, parameters, and temperature fields), but both evolve toward asymptotic states.

The asymptotic behavior of flow fronts arising in gravity currents is mathematically and experimentally understood for isoviscous systems, for example, Huppert [1982]. After long times, our simulations reach monotonic asymptotic states. For our case with space and time-dependent viscosity, the crustal velocities are locally proportional to the product \( \eta^{-1}h^2 \frac{\partial h}{\partial x} \), where \( x \) is the horizontal coordinate of some point, \( \eta \) the viscosity, and \( h \) the bathymetric height above it.

The reasons for these systems to acquire asymptotic trends are several:

1. There is a shrinking of the driving forces-affected region, due to horizontal narrowing of the escarpment, therefore a decrease in the volume of the mobile region.
2. The platform becomes more horizontal in time, therefore the deep regions under it (located at the rear of the mobile region) become less mobile (even though their viscosity is almost invariable due to the locally dominant hot perturbation).
3. The off-platform crust viscosities, the highest in the system at all times, increase due to the half-space cooling component there dominant. (Owing to the continuity equation, conditions (i, ii, and iii) already imply a progressive stagnation of the mobile region.)
4. The platform height \( h \) decreases in time (at all \( x \)), causing a reduction in the driving forces (also below the escarpment).
5. Both the escarpment and the mass below it move away from the hot and weak regions, thus their viscosities increase with time.
6. In the vast majority of our simulations, the escarpment slope increases in time, but finally acquires a constant value.

Mathematically, systems whose velocities trend toward zero can tend to different asymptotes, depending upon how each of these velocities behave asymptotically. They can asymptotically approach a final, bounded displacement, or a monotonically growing, unbounded one. Unfortunately, from our simulations, we cannot distinguish between these two cases. Finally, the question is geologically irrelevant: the Galápagos platform will be subducted beneath South-America in a finite amount of time.

5. Conclusions
The Galápagos platform, formed perhaps with a larger component of intrusion than extrusion, exhibits several characteristics that suggest flow of its lower crust. We have presented numerical models that include...
viscous relaxation of the lower crust and exhibit flow-fronts whose geometries resemble the observed present-day bathymetric escarpments in the Galápagos platform.

The conspicuous escarpment along the southwestern margin is formed by a system of stepped terraces that might be explained by lower crustal flow, which would produce backward tilting of the subhorizontal platform surface thereby hindering the long-distance flow of lavas and potentially producing the stepped morphology. Both the observed present-day morphology and the paleogeographic reconstructions of Geist et al. [2014] rule out the possibility of paleosealevels as having an effect here.

The extended escarpment on the eastern platform region may, according to our models, be satisfactorily explained by flow of the lower crust during the last ~3 Myr, producing a well-defined bathymetric front, with the whole system evolving asymptotically toward the present-day configuration. This lower crustal flow episode may be associated with horizontal mass transport: mantle plume geochemical signatures could potentially be advected 50 km toward the southeast, which might be related to the geographically elongated isotopic anomalies found by Harpp and White [2001]; and increase the platform’s horizontal extension. Both the paleobathymetric configuration and the paleogeographic reconstructions of Geist et al. [2014] seem to indicate that paleosealevels may also had an effect in the upper parts of the eastern platform.

Similar dynamical regimes may have occurred in the past to the east of the platform, such as the southern escarpment on the Carnegie ridge, and perhaps currently in other smaller, more recent escarpments on the central/western platform.

The phenomenon of lower crustal flow as a mechanism for tectonic deformation, already noted for Iceland by Jones and MacIennan [2005] and here inferred to operate in the Galápagos, must be common in hotspot provinces emplaced upon young oceanic lithosphere [Richards et al., 2013], such as the Azores and Easter islands as well as in aseismic ridges like the Ninety-east and Nazca ridges.

Appendix A: Physicomathematical Description

A1. Conservation Laws

We consider a 2-D Stokes fluid with velocity \( \mathbf{v}(x, z, t) \) field, with \( i = 1, 2 \) representing \( x \) and \( z \) components and \( t \) the time; subjected to a homogeneous gravitational field \( \mathbf{g} = -g \hat{z} \) (which implicitly defines the orientation of the vertical axis). The relevant conservation laws are:

\[ \rho \mathbf{v}_{,i} = 0 \quad \text{Conservation of mass for incompressible fluid (continuity)} \quad (A1) \]

\[ \sigma_{ij} + \rho \mathbf{g}_i = 0 \quad \text{Conservation of linear momentum for Stokes flow} \quad (A2) \]

and

\[ \sigma_{ij} = -p \delta_{ij} + 2\eta \left( \frac{1}{2} \mathbf{v}_j + \mathbf{v}_i \right) - \frac{1}{3} \delta_{ij} \mathbf{v}_k \mathbf{v}_k + \psi v_k \delta_{ij} \]

\[ (A3) \]

is the general constitutive equation for a Newtonian isotropic fluid with symmetric stress tensor (with shear viscosity \( \eta \) and volume viscosity \( \psi \)).

which, using (A1), simplifies to:

\[ \sigma_{ij} = -p \delta_{ij} + \eta \left( \mathbf{v}_j + \mathbf{v}_i \right) \]

\[ (A4) \]

In this formulation, we will allow \( \eta = \eta(x, z, t) \).

Given the continuity and constitutive equations, the horizontal component of the momentum equation can be written as:

\[ 0 = -p_\times + 2 \left[ \eta v_x \right]_x + \left[ \eta \left( v_x + v_z \right) \right]_z \]

\[ (A5) \]

where we have assumed a 2-D system, so that all the partial derivatives with respect to the other horizontal component \( (y) \) vanish.

We use a thin-sheet approximation defined by the two following conditions:

\[ v_x \gg v_z \quad \text{subhorizontal mass flow} \]

and
\[ \frac{\partial a}{\partial x} \ll a_{z} \] where \( a \) is any dependent variable of the system, which means that the system has, at all times, horizontally elongated features.

Keeping only dominant terms of the momentum equation (in relation with the thin-sheet approximation), and regarding the pressure gradient as an unknown quantity, we get:

\[ 0 = -p_{x} + (\eta v_{x})_{z} \] (A6)

which is an approximate version of the horizontal component of the momentum equation.

We assume a lithostatic pressure field:

\[ \rho(x, z, t) = \rho gh(x, t) - z \] (A7)

with \( h(x, t) \) being the height of the bathymetry over the isostatic reference level, and \( z \) the vertical coordinate, positive upward.

Thus, \( p_{x} = \rho gh_{x} \) (A8)

and consequently, the momentum equation can be recast as:

\[ (\eta v_{x})_{z} = \rho gh_{x} \] (A9)

Integration over the vertical coordinate \( z \) gives:

\[ \eta(x, z, t) v_{x}(x, z, t) - \eta(x, z_{0}, t) v_{x}(x, z_{0}, t) = \int_{z_{0}}^{z} \rho g h_{x} dz \] (A10)

which, after trivially integrating its right side, we will write in a more generic form:

\[ v_{x}(x, z, t) = \frac{1}{\eta(x, z, t)} \left[ c(x, t) + \rho gh_{x} z \right] \] (A11)

with \( c(x, t) = \eta(x, z_{0}, t) v_{x}(x, z_{0}, t) - \rho gh_{x} z_{0} \)

being the “integration constant” for \( z \), which is a function of \( x \) and \( t \) here. We do not know the vertical gradient of the horizontal velocity at this point, so it is not possible to evaluate this function for now. Instead, we will readily obtain a computable expression for it when evaluating the boundary conditions.

Another integration over the \( z \) coordinate gives:

\[ v_{x}(x, z, t) = v_{x}(x, z_{0}, t) + \int_{z_{0}}^{z} \frac{1}{\eta(x, z', t)} \left[ c(x, t) + \rho gh_{x} z' \right] dz' \] (A12)

A2. Boundary Conditions

Upper and lower boundaries are considered horizontally rigid:

\[ v_{x}(x, h, t) = 0 \quad \text{and} \quad v_{x}(x, -h, t) = 0 \] (A13 and A14)

where \( f = \frac{h}{h_{t}} \) is a “crust-mantle” isostatic equilibrium factor, and always with \( h = h(x, t) \).

Taking \( z_{0} = -fh \) (isostatic Moho), the condition at the lower boundary yields:

\[ v_{x}(x, z, t) = \int_{-fh}^{z} \frac{1}{\eta(x, z', t)} \left[ c(x, t) + \rho gh_{x} z' \right] dz' \] (A15)

and then, the condition at the upper boundary yields:

\[ c(x, t) = -\int_{fh}^{h} \frac{1}{\eta(x, z', t)} dz' \left. \int_{-fh}^{z} \rho gh_{x} z' \eta(x, z', t) \right|_{-fh}^{h} \] (A16)

So, factoring out, finally we can write:
\[ v_x(x, z, t) = \rho g h, x \left[ \int_{-h}^{h} \frac{z'}{\eta(x, z', t)} \, dz' - s(x, t) \int_{-h}^{h} \frac{1}{\eta(x, z', t)} \, dz' \right] \]  \hspace{1cm} (A17)

with \( s(x, t) = \int_{-h}^{h} \frac{1}{\eta(x, z', t)} \, dz'' \)

(with \( z', z'' \) being auxiliary integration variables with the same scaling and dimensions as \( z \)).

We assume zero mass flux at the vertical walls \((x = 0 \text{ and } x = L)\) of the model box, more specifically:

\[ v_x(x=0, z, t) = 0 \quad \text{and} \quad v_x(x=L, z, t) = 0 \]  \hspace{1cm} (A18 and A19)

for the whole range \(-h \ll z \ll h\).

### A3. Temperature and Viscosity Structure

The thermal structure consists of two ingredients: (1) A half-space cooling model (with locally defined depth coordinate, i.e., \( \text{depth} = \text{depth}(x, z, t) \)) and thermal age being horizontally dependent upon the age gradient \( \frac{\partial \text{Age}}{\partial x} \)), and (2) A hot thermal perturbation with Gaussian form in \( x \) and \( z \) as described in the main text.

1. Starting with a definition of homologous temperature \((T_{\text{om}} = \frac{T}{T_{\text{solidus}}})\) that depends on the solidus thermodynamic conditions, the analytic solution of the temperature versus depth and time given by the half-space cooling model (with initial temperature \( T_{\text{solidus}} \) and upper boundary temperature \( T_0 \)) is:

\[ T_{\text{om}} = \text{erf} \left( \frac{\text{depth}}{2 \sqrt{(\lambda t)}} \right) \]  \hspace{1cm} (A20)

where the time \( t = 0 \) corresponds to when the half space was at the solidus temperature, zero thermal age; therefore \( t \) can be approximately equated with the age of the background crust (related to its distance to the spreading ridge). If we adapt this model to a thin-sheet system, in which the bathymetric surface is subhorizontal, we can take

\[ \text{depth}(x, z, t) = h(x, t) - z \]  \hspace{1cm} (A21)

and follow up with the aforementioned solution, with negligible error regarding the flatness of the bathymetry. This is exactly the same description adopted by Jones and Maclennan [2005]. If in addition, in this study, we consider the crust as having a differential age progression along the horizontal, in such a way that we can define a gradient in crustal age, \( \text{Age}_{\text{grad}} = \frac{d\text{Age}}{dx} \), provided that we know the location of the study transect. Here, this gradient is assumed constant and applied along the full transect length.

With these ingredients, a modified and approximate version of the half-space cooling temperature solution can be adopted:

\[ T_{\text{om}}^{\text{SC}}(x, z, t) = \text{erf} \left( \frac{h(x, t) - z}{2 \sqrt{(\lambda (t + \text{Age}_{\text{grad}}x))}} \right) \]  \hspace{1cm} (A22)

2. In addition to the previous description, given the evidence of diapirism and volcanism throughout the Galápagos platform and with it being sustained in geologic time, it is likely that not only a high vertical geothermal gradient is present, but also that a high-temperature region extends from the Moho and up to shallow depths in the central platform area under the volcanoes, which creates a horizontal thermal gradient component relative to the cold off-platform background crust. Consequently, we have imposed a hot thermal perturbation in trying to mimic this geologic scenario in a simple way. The perturbation is taken to be a bidimensional, unit-amplitude Gaussian function (with horizontal and vertical length scales \( L_x \) and \( L_z \), respectively). A Gaussian function was preferred because it represents the kernel of the diffusion equation.
\[
\tau_{\text{om}}(x, z, t) = \exp \left[ -\left( \frac{x}{L_{T_x}} \right)^2 - \frac{[f h(0, t) + z]^2}{L_{T_z} f h(0, t)} \right]
\]

(A23)

Which is a Gaussian function centered on the platform side of the model box \((x = 0)\) and at the time-dependent Moho depth \((z(t) = -f h(0, t))\). This definition implies the vertical advection of the perturbation.

The vertical length-scale value \(L_{T_z}\) is chosen to produce acceptable surface values of the viscosity (considered volumetric averages over length scales given by the grid size). On the other hand, the value of \(L_{T_x}\) determines how horizontally extended the weak area under the central platform is, and is considered a model parameter to be investigated.

This hot thermal perturbation, taken to be homologous, is added to the temperature function previously defined.

\[
T_{\text{om}} = T_{\text{om}}^{\text{SC}} + T_{\text{om}}^{\text{hot}}
\]

(A24)

Homologous temperatures greater than one are obtained in the vicinity of the hot thermal perturbation's center, which implies that the system has, on volumetric local averages (over grid-point-centered domains with grid spacing side-lengths), a fraction of partial melting.

Using a thermal diffusivity of \(\kappa = 10^{-6} \text{ m}^2/\text{s}\), and after transforming temperature units (Celsius to Kelvin), the viscosity is computed as a function of the homologous temperature:

\[
\eta = \eta_{\text{solidus}} \exp \left[ (\sigma/T_{\text{om}}) - \alpha \right]
\]

(A25)

with \(T_{\text{om}} = T_{\text{om}}(x, z, t)\) and where a nominal value \(\alpha = 40\) was adopted from Jones and MacIennan [2005].

Finally, in consideration of homologous temperatures greater than unity and with the overall behavior of the exponential equation for viscosity, some restrictions are applied: a lower bound on the viscosity value is imposed following geological constraints that render the exponential formula inaccurate when the viscosity is well below the solidus viscosity due to partial melting. For example, viscosities lower than one tenth of the solidus viscosity (here assumed close to \(10^{18} \text{ Pa}\cdot\text{s}\)) are not allowed. Similarly, a viscosity upper bound is imposed with regard to constraints on tectonic processes, from which it is understood that the real macroscopic viscosity values are not well represented by the expression for viscosity we are using. For instance, maximum viscosities greater than \(10^{23} \text{ Pa}\cdot\text{s}\) are not considered. This approach is in agreement with Jones and MacIennan [2005].

\[\eta_{\text{min}} \leq \eta \leq \eta_{\text{max}}\]

with these bounds being among the parameters investigated in this study.

### A4. Governing Equations

As we see from the equation for velocity, the controlling variable is the bathy/topographic height \(h(x,t)\), so we need an equation for it:

The mass conservation equation (A1) allows us to write:

\[
\frac{\partial v_{z,x}}{\partial z} = -\frac{\partial v_{x,x}}{\partial z}
\]

so that

\[
\int_{z_1}^{z_2} \frac{\partial v_{z,x}}{\partial z} \, dz = -\int_{z_1}^{z_2} \frac{\partial v_{x,x}}{\partial z} \, dz
\]

(A26)

Now, choosing \(z_1 = -f h(x,t)\) (isostatic Moho), and \(z_2 = h(x,t)\) (bathymetry)

\[
\frac{\partial v_{z,x}}{\partial z} = -\frac{\partial v_{x,x}}{\partial z}
\]

(A27 and A28)

(\text{where the upper dot represents the partial derivative with respect to time}).
Thus, equation (A26) can be written as:

$$v_z(x, h, t) - v_z(x, -f(t), t) = - \int_{-h}^{h} v_z(x, z, t) dz$$  \hspace{1cm} (A29)

The boundary conditions (A13) and (A14) allow the partial derivative with respect to $x$ inside the integral in (A29) to be taken outside of it, and then considering (A27) and (A28), equation (A29) yields:

$$\dot{h}(x, t) + f(x, t) = - \int_{-h}^{h} v_z(x, z, t) dz$$

which can be finally written in a more extended and explicit form:

$$\frac{\partial h(x, t)}{\partial t} = \frac{-1}{1 + f} \int_{-h}^{h} v_z(x, z, t) dz$$  \hspace{1cm} (A30)

This equation means that any material column in the crust changes its total vertical length ((1 + $f$) $\cdot h(x, t)$) in time according to the net volume flux across its boundaries. This is a direct consequence of mass conservation.

The numerical solution of the two governing equations ((A17) and (A30)) is carried out in a nondimensional form with the following reference scales:

Length: $L$ (the horizontal length of the model box, here being equal to 350 km for both study transects); defining a scale for $x$.

System aspect ratio (vertical over horizontal): $c = [(6 \text{ km})/(1 + f)] + H_{\text{plat}}/L$; defining a scale for $z = cLz'$ with $[6 \text{ km})/(1 + f)]$ being the elevation of the upper surface of a 6 km thick (oceanic) crust over the isostatic reference level in compensation with the mantle, and $H_{\text{plat}}$ being the initial platform maximum thickness.

Viscosity: a mean value $\eta_{\text{ave}}$ is used to scale with the overall system behavior. To estimate the right value using a simple algorithm before run, it is difficult due to the strong spatial variability of the viscosity. We have instead, estimated it by trial and error, with the objective of fitting the model timescales, which are mainly controlled by the low-viscosity mobile regions.

For example, a useful expression is: $\eta_{\text{ave}} = (\eta_{\text{min}}\eta_{\text{ave}}\eta_{\text{max}})^{1/(a+b+c)}$ (which is a power-weighted geometric mean of the system reference viscosities) where $a > b > c$ (an example being $a=8, b=5, c=1$).

Mass-density and gravity are referenced to $\rho_x = 2.8 \text{ gr/cm}^3$ and $g = 9.8 \text{ m/s}^2$, respectively.

With these quantities, a natural scale for velocity can be obtained from the final equation for the horizontal velocity component (equation (A17)):

$$V_0 = \rho_x gc^2L^2/\eta_{\text{ave}}$$

and with it, a possible time scale can be found: $\tau = L/V_0$ so that $\tau = \frac{\eta_{\text{ave}}}{\rho_x gc L}$. In the code, a further constraint is applied on $t$ to obtain desired times scales.

In these last two expressions, we note their dependence on the driving force, the aspect ratio, the absolute size of the system, and the average viscosity.

With these characteristic scales, the governing equations (in nondimensional form) to be considered are:

$$\frac{\partial h'(x', t')}{\partial t'} = - \frac{V_0\tau}{(1 + f) L} \frac{\partial}{\partial x'} \left( \int_{-b(x', t')}^{b(x', t')} v_z'(x', z', t') dz' \right)$$  \hspace{1cm} (A31)

$$v_z'(x', z', t') = \rho' g' h' x' \left[ \int_{-h}^{z} \frac{\dot{z}}{\eta'(x', z', t')} dz' - s(x', t') \int_{-h}^{z} \frac{1}{\eta'(x', z', t')} dz' \right]$$

with $s(x', t') = \int_{-h}^{z} \frac{1}{\eta'(x', z', t')} dz'$  \hspace{1cm} (A32)

in which $x', h', z', \dot{z}, t', v_z', \eta', \rho', g'$ are dimensionless quantities referenced to the aforementioned scales.
It is worth mentioning that these two governing equations are coupled partial differential equations that depend on integrals over the vertical coordinate. The only horizontal derivatives are for vertically integrated quantities. This particular feature allows for a rather simple grid definition and a rather easy numerical solution.

Appendix B: Numerical Treatment

B1. Numerical Solution: Finite Differences in Space and Time

The integrals over the vertical coordinate, both in equation (A32) when calculating \( v_x \) and in equation (A31) when calculating the integral of \( v_x \) over depth (volume flux), are computed using Simpson’s rule for quadrature. This method uses quadratic polynomials every three points to approximate the unknown function and integrate it. It is a good compromise between accuracy and simplicity. At the initial point of quadrature (first point above the Moho), trapezoidal integration with one tenth of grid spacing was used (avoiding lower order method contamination).

For the partial derivative of the volume flux with respect to \( x \), a classical two-point, centered (regarding the equation symmetry) derivative is used.

For time integration of equation (A31), we have used the explicit Runge-Kutta 2, also known as “improved Euler method.” This scheme is, theoretically, second-order accuracy.

Therefore when solving equation (A31), updating the height \( h(x,t) \) implies updating the isostatic Moho as well; consequently, the space grid adapts to the time-changing geometry of the crust at each time step. The horizontal coordinates of the grid points are kept fixed for all times and they consist of an equally spaced set of points. For instance, in the nondimensional space, the grid spacing is \( dx = 1/(nx-1) \) at all times, with \( nx \) being the total number of points on the horizontal (a constant in the program). The vertical coordinates of the grid points change with time and they also depend on \( x \). They are defined as an equally spaced set of points over each column length vertical interval \((1+f)h(x,t)\) in such a way that we define a matrix for the \( z \) coordinates as:

\[
z_i(t) = -fh(x_i,t) + i \cdot dz
\]  
(B1)

where \( i \) is the index that sweeps over the vertical, \( j \) over the horizontal, and \( dz \) is the vertical increment.

This grid, being extremely simple, can handle the calculation of the velocity field (equation (A32)) at each grid point at each time, because even though it is inhomogeneous and time-dependent, it constitutes a well-defined Eulerian frame. The upper and lower mesh boundaries are the actual position of the bathymetric surface and the Moho, respectively.

B2. Numerical Instabilities’ Treatment

When needed, a symmetric, weighted, three-point, moving average stability filter was selectively applied to \( h(x,t) \) over \( x \), with the aim of damping spurious oscillations arising at some time steps, in some particular simulations. The filter is applied only at some time steps, under the criterion of an “instability threshold trigger” given by slope contrasts every three spatially adjacent samples of the bathymetry. The filter code’s name is “smoother.3w,” and its equation is:

\[
sh_j = 0.5(1-w)h_{j-1} + wh_j + 0.5(1-w)h_{j+1}
\]  
(B2)

where the index \( j \) refers to the \( x \) coordinate, \( h \) being the input height function, and \( sh \) the output smoothed height function.

The filter was investigated exhaustively, and the desired performance was given by a low threshold trigger (adjacent segments’ slope changes of 50%–80%), but acting with a tiny amount of damping: typical values of \( w = 0.99995 \) or higher were employed. The filter’s effect is subtle, and only intended to suppress small, unphysical oscillations and not the desired features of the bathymetry and velocity field. Keeping the bathymetric flow front displacement virtually undamped was among the controlling criteria in choosing filter parameters.

The filter was needed and applied occasionally in the preparation of animated plots on those simulations with a rather extended hot thermal perturbation (large \( L_T \)) in which, due to the overall mechanical weakness of the system, large velocities, and deformations are produced. For instance, the filter was not needed when extracting topographic functions (slopes, tilting rates, etc.) for analysis.
B3. Benchmarking

Huppert [1982] provides an analytic solution for a gravity current over a horizontal surface, but at the same time sets the essential principles from which McKenzie et al. [2000] derive a more general partial differential equation for mass redistribution in a constant-viscosity thin fluid layer of variable-thickness (based on the Stokes equation). We have considered this equation and implemented its solution numerically, for the case of horizontally rigid upper and lower boundaries. This equation is much simpler because it is derived from an analytic expression for the velocity (a parabolic profile), and it reads:

$$h = \frac{(1+f)^2 \rho c g}{12\eta_c} (h^3 h_x)_x$$

(upper dot being time derivative, Einstein notation in space)

The next plots compare numerical solutions of this relatively simple equation, with outputs from our main code (using numerical solution of full integrals applied on the space and time-dependent viscosity formulation) but now simply set up for a constant viscosity. Two different cases are shown in Figure B1.

Appendix C: Sensitivity Analysis

The purpose of this section is to illustrate some features of the model’s behavior, more specifically, to show the effect of the controlling parameters. The system’s behavior is determined by the initial condition of the bathymetry, the boundary conditions, and the viscosity structure parameters. The boundary conditions were described previously, and their effects will not be investigated further here.

C1. Initial Condition of Bathymetry

The initial bathymetric function controls the system evolution to some degree. As was explained in the methodology section, the initial bathymetry is described by a Butterworth filter equation. We prescribed the horizontal length scale as $x_0 = 0.35 - L$ a priori as a reasonable value for our purposes. We have chosen the initial platform thickness $H_{\text{platform}}$ with regards to geological criteria particular to each study transect.

In the following, model sensitivity will be shown in the vicinity of one of our model simulations, the one associated with the Eastern platform. The value of initial platform thickness is set to $H_{\text{platform}} = 2 \text{ km}$ (overlying a background 6 km thick oceanic crust).

---

**Figure B1.** Comparison of the reference solution given by equations in McKenzie et al. [2000], with the solution given by our code when set up for constant viscosity.

Case 1: $\text{Visc} = 10^{19} \text{ Pa s, } H_{\text{plus}} = 4 \text{ km, } x_0 = 0.35, N_{\text{step}} = 8, L = 350 \text{ km}$

Case 2: $\text{Visc} = 10^{20} \text{ Pa s, } H_{\text{plus}} = 5 \text{ km, } x_0 = 0.25, N_{\text{step}} = 12, L = 350 \text{ km}$

Note: The “platform maximum thickness” is the space-maximum value of the platform thickness (the platform considered as built over the background crust), always found at $x = 0$, and evolving in time. This quantity is slightly different from the bathymetry elevation (function of $x$ and $t$), which is simply the platform thickness (function of $x$ and $t$) plus the isostatically defined constant fraction of the background crust, as explained in the methods section. This note applies to all the subsequent figures.
An important parameter in this function is the exponent $N_{\text{steep}}$ that controls how steep and wide the initial bathymetry ramp is.

C2. Definition of the Viscosity Structure
Following the methodology section, the viscosity structure is defined by the following geometrical, thermal, and rheological parameters:

1. Horizontal Length scale of the Gaussian hot thermal perturbation (provided that the vertical length scale was fixed a priori), defining how extended the weak region is inside the system.
2. Initial age of the background crust, here taken as the reference initial age at $x = 0$ (central platform area), from which the horizontal age gradient defines the initial ages over the system for $0 < x \leq L$.

For example, variations in the values of the maximum viscosity have little effect on the system’s behavior, therefore will not be shown here.

As can be seen in Figure C1, the index $N_{\text{steep}}$ determines the initial steepness of the bathymetric ramp and also the horizontal extension of it. For the combination of parameters defining this model (the location in the parameter space), the effect of $N_{\text{steep}}$ is not important over the platform thickness over time.

Figure C1. (top) Different initial bathymetry profiles versus the slope exponent $N_{\text{steep}}$ (determining the escarpment steepness and width). (bottom) Model sensitivity to the exponent $N_{\text{steep}}$, here shown for two model output variables: the platform maximum thickness versus time, and the escarpment’s maximum slope versus time.
Nevertheless, its effect over the spatial maximum platform slope is significant. As can be seen, the ordering relation of the slope values is conserved, higher initial slopes remaining higher over time. The slope increase is due to the development of a flow front. The time scale is related to the relevant geologic time for the Eastern platform evolution (last 3 Myr). The time is referred to the reference thermal age of the background oceanic crust under the central platform which (not considering thermal rejuvenation due to volcanism in the platform) nowadays is roughly 7.5 Myr (between the 5 and 10 Myr isochrons). This time definition will apply to all of the items of this section.

Figure C2 shows the effect of different values of the horizontal length scale of the hot-thermal perturbation have on the spatial maximum slope (left) and the horizontal position of the points where that maximum slope is reached (recall that the horizontal position of the material points over the bathymetric surface is fixed for all times due to the applied boundary condition). Larger values of the thermal length scale \( L_{\text{Tx}} \) are expressed as larger extensions of the mechanically weak zone, and are therefore associated with larger deformation volumes and enhanced mobility, producing greater front slopes and larger front displacements.

![Figure C2](image)

**Figure C2.** Model sensitivity to the hot thermal perturbation's horizontal length scale \( L_{\text{Tx}} \), here shown for two model output variables: the escarpment's spatial maximum slope versus time, and the escarpment's maximum slope locations versus time.

![Figure C3](image)

**Figure C3.** Model sensitivity to the background crust's initial reference age (including the horizontal gradient in age), here shown for two model output variables: the platform maximum thickness versus time, and the escarpment's maximum slope locations versus time.
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From the inspection of the sensitivity analysis and of all the simulations performed, we learn that high-viscosity systems (e.g., a = 4, 10) not only evolve slower in time relatively to low-viscosity systems (e.g., a = 0.25, 0.5), but also that high-viscosity systems finally do not reach the same states as low-viscosity systems. The systems considered in this study show asymptotic behavior, yet at long-time scales, high-viscosity systems evolve to different asymptotes and tend to “freeze” in different final states than low viscosity systems.

In Figure C3, model sensitivity to the initial reference age of the background oceanic crust is shown for the maximum platform-thickness versus time (left) and the position of the bathymetric flow-front (left). The effect of this parameter is not so important on this region of the parameter space, due to: (i) relatively old initial ages >4.0 Myr, and (ii) the presence of a hot-thermal perturbation that controls the system’s evolution to a greater degree. Nevertheless, the different systems evolve to different final asymptotes.

Figure C4 illustrates the effect of the simultaneous amplification of the three viscosity reference values (minimum, solidus, maximum) by a constant factor (0.25,0.5,1,2,4,10), observed over the spatial maximum platform-thickness and the horizontal position of the maximum slope, versus time. At earlier times (roughly 4.5 to 5.0 Myr), the systems’ evolution time scales seem to be proportional to their viscosity reference values (as predicted by linear theory for time-independent viscosity Stokes fluids). At later times, the proportionality is explicitly broken: higher viscosity systems (a = 4, 10) evolve more slowly in time, and they take longer (than proportional) to reach the same values of “maximum thickness h” than lower viscosity systems (a = 0.25, 0.5). This behavior is due to the time-dependent nature of the viscosity of the systems here considered: the thermal cooling due to aging imposes a natural and independent clock that produces higher viscosities at later times, therefore increasing the space-average viscosities in time, say, for corresponding time intervals with lengths proportional to the reference viscosity values (minimum, solidus, maximum).

Figure C4. Model sensitivity to the simultaneous scaling of the three viscosity reference values (minimum, solidus, maximum), here shown for two model output variables. Here, “a” is the scale factor for the viscosities. Note that the scale for the maximum thickness “h” starts at 1.75 km, so the relative difference among the curves is small.


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