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# Oceanic Isostasy: Seafloor Spreading and Rift Localization

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**Key Points:**

1) Oceanic isostasy may be an important key to the facilitation and localization of seafloor spreading
2) Evidence for and influence of oceanic isostasy on seafloor spreading can be seen across the globe from rifted margins to Iceland to backarc spreading.

**Abstract**

Why Earth is the only known planet with plate tectonics is a long-standing question. The primary driving forces of plate tectonics are widely understood to entail slab-pull, ridge-push, and mantle basal tractions. Mantle tractions are usually understood in terms of active upwellings within the mantle convective system, but such upwelling are fraught with difficulties in sustaining long term continuity with the overriding plates. Recognition that basal tractions can be modified by the isostatic response of thermal subsidence in the presence of an overlying ocean may reconcile the difficulties. In particular, greater thermal subsidence near the ridge axis, requires a systematic outward mass flux beneath the plate to maintain gravitational equilibrium. This flow biases basal tractions from resistance to plate push. This outward directed flow gives the plate a perpetual resistance-free base facilitating self-sustaining plate separation and boundary localization. Localization of the plate boundary can explain why oceanic boundaries tend to narrow in comparison to subaerial boundaries. Oceanic isostasy has likely impacted tectonics from Iceland to backarc spreading to the continental slivers dotting the oceans. That the ocean plays an integral part to perpetuating seafloor spreading may be an important puzzle piece as to why only Earth exhibits plate tectonics.

# 1 Introduction

To some degree, plate tectonics is a surface manifestation of underlying mantle dynamics (Bercovici et al., 2000; Coltice et al., 2017). That plate tectonics is only known to occur on Earth suggests something fundamentally different about the Earth's mantle dynamics than other terrestrial planets (Martin et al., 2008). An important key to the perpetuation of plate tectonics on a planet is the how and why of self-sustaining continuous generation of new seafloor at a spreading axis.

Seafloor spreading poses a bit of a paradox. Seafloor spreading seems a nearly ubiquitous process in the geological record. Seafloor spreading has been active on Earth for perhaps three billion years or more (Cawood et al., 2018). It is responsible for the creation of majority of Earth's surface and an integral part of Earth's current dynamics (Crameri et al., 2019). Earth's supercontinents have come together and ruptured apart multiple times in Earth's history (Nance and Murphy, 2013). Rifted margins of widely varying structure evolve to spreading (Perez-Gussinye et al., 2023; Sapin et al., 2021), sometimes quickly (Umhoefer, 2011; Wang et al., 2019) and with rapid along strike propagation (Taylor et al., 1995).

Simultaneously, seafloor spreading has been fairly difficult to account for. Despite rift structures on virtually all terrestrial planets, no other known planet exhibits seafloor spreading (Martin et al., 2008). The forces required to rupture continental lithosphere do not present a simple pathway to evolve to spreading (Brune et al., 2023; Buck, 2007). Failed rifts, like the Mid-continent rift in North America (Ojakangas et al., 2001; Stein et al., 2018) and the West African Rift System (Ghomsi et al., 2022) are of size and magmatic output comparable or exceeding other rifts that did transition to spreading. Even some rifted margins that did transition, such as the north central Atlantic took remarkably long time to fully rupture after initial rifting (Worthington et al., 2021).

Suggested answers to the question as to why Earth has plate tectonics include large extraterrestrial impacts such as the one that created the moon (O'Neill et al., 2017), plume activity (Gerya et al., 2015), and the presence of water in the mantle (Bercovici, 1998; Dymkova and Gerya, 2013; Regenauer-Lieb et al., 2001; van der Lee et al., 2008). The importance of water in facilitating plate tectonics has been identified as affecting the rheology of the mantle and crust, likely contributing to development of both a weak asthenosphere and enhancing deformation at plate boundaries (e.g., Bercovici, 1998). Additionally, water can enhance faulting by lubrication and increasing pore pressures (Dymkova and Gerya, 2013). While potentially each are important factors, here I look at another direct effect of water - specifically the overlying ocean - in facilitating and localizing seafloor spreading by inducing basal tractions aligned with spreading.

## 2 Plate driving forces

The energy behind plates moving on the planet ultimately derives from heat and gravity. The first viable connection between mobile continents and planetary forces mobilizing them was made in the first half of the 20$^{th}$ century by Arthur Holmes (Holmes, 1931) in what he termed 'substratum convection' - mantle convection as we now know it. His real genius was demonstrating how a planetary heat engine could produce horizontal forces from essentially vertical dynamics. This remains the central condition of all plate tectonic forces where gravitational instabilities result in horizontal forces.

The primary forces driving plate tectonics are some combination of ridge push, slab pull, and mantle tractions (Becker and O'Connell, 2001; Coltice et al., 2019; Forsyth and Uyeda, 1975), again horizontal forces arising from gravitational instabilities. Considerable evidence points to slab pull as being a dominant force (Conrad and Lithgow-Bertelloni, 2002; Forsyth and Uyeda, 1975), although plate motions must already be developed to some degree prior to subduction initiation. The magnitude of ridge push is smaller than slab pull by roughly an order of magnitude (e.g., Fowler et al., 1990). It is a force that arises with the difference in elevation across a plate and varies by plate age as elevation differences increase; Young plates (< 5 Myr) have forces ~$10^{11}$ N/m with values increasing to > $10^{12}$ N/m for plates older than ~25 Myr. Of course, while ridge push increases with plate growth, basal resistance over a stationary mantle also correspondingly increases. While there is broad consensus that mantle tractions are necessary in plate tectonic development (e.g., Bercovici et al., 2000), our understanding of exactly how mantle tractions participate in plate motions still has large deficiencies. For example, it takes only a cursory look at popular explanations of seafloor spreading to find graphics of large convection cells beneath diverging plates with the upwelling cell limbs aligned with the spreading center between. Besides suggesting a likely erroneous Rayleigh number of the mantle (Weeraratne and Manga, 1998), such pictures fail when considering how plates grow, shrink, and migrate. To maintain upwelling limbs beneath a migrating mid-ocean ridge and growing plate, the aspect ratios of the cells will become inconsistent with each other as well as the overlying plates; Subducting ridges would imply superimposed upwelling and downwelling limbs, etc. Still, the prevalence of such figures are likely to remain in the absence of a clear alternative relationship between underlying mantle flow and plate motion.

## 3 Oceanic isostasy

Oceanic isostasy is an only recently explored forcing function on oceanic mantle dynamics and hence mantle tractions (Conder, 2012, 2022a). One reason it has been long-overlooked is that isostasy is typically viewed as a static or transient state and does not immediately lend itself as an obvious contributor to a steady-state process like seafloor spreading. However, under certain geological conditions, isostasy can act in non-transient ways. For example, one of the pioneers in continental isostasy suggested long-term isostatically induced horizontal compression in a system like the North American eastern seaboard; The compression arising from the isostatic return mass flux ‘undertow’ as mass is continually eroded from the nearby mountains and deposited on the continental shelf (Hayford, 1911) (Figure 1). The undertow ceases only with cessation of the erosion-deposition mass redistribution.

Analogous to the process of constant return mass flux in figure 1, differential thermal subsidence of the seafloor involves a continual mass exchange at the surface as water deepens at different rates across the plate. To maintain gravitational equilibrium, this surface water mass exchange must be balanced by a deeper counter mass exchange (Figure 2). Our current geophysical understanding of thermal isostasy in the oceans largely follows the work of Parsons and Sclater (1977). The depth of the seafloor is widely understood to be controlled by Pratt isostasy resulting in the well-known depth-square root of age relationship where the mid-ocean ridge system is topographically high and seafloor subsides as it ages and moves away from the ridge due to cooling. In the decades following this seminal work, the overlying ocean has invariably been treated as a static load correction, necessary to account for the excess mass effect on the shape of the seafloor (e.g., Turcotte and Schubert, 2002 section 4.22). While substantively correct with longstanding predictive power for understanding for depth of the seafloor, the Parsons and Sclater work has perhaps been a little too successful; In particular, the ocean and mantle dynamic interplay required to maintain the isostatic equilibrium that their work depends on remains almost entirely unconsidered.

In essence, to maintain gravitational equilibrium, the flux of ocean mass towards more rapidly thermally subsiding younger seafloor must be balanced by a corresponding flux of asthenosphere mass directed towards older lithosphere. As thermal subsidence is continual and systematic, this isostatic response must likewise be continual and systematic (Figure 2). While isostatic in nature, this dynamically maintained equilibrium does not fit cleanly as either Pratt-type or Airy-type isostasy. Here I term it 'oceanic isostasy'. The 1D analytic solution of the physics of this oceanic isostasy for a constant velocity plate is derived in Conder (2012).

The outward directed asthenosphere flow induced from oceanic isostasy only incrementally leads the overlying lithospheric plate with tractions less than those estimated for ridge push (Conder, 2012). Nonetheless, oceanic isostasy may have profound implications on the plate motions. Firstly, although the forces are small in absolute terms compared to ridge push or slab pull ($10^8$ – $10^{11}$ N/m), asthenosphere leading the lithosphere alters the plate tectonic force balance by giving a systematically positive push to the plate and supplanting what would otherwise be basal resistance. At a minimum, this gives a resistance-free base for plates to glide on allowing forces like ridge push greater efficiency. Further, because these tractions come from the isostatic response to the plate itself, the form of the self-sustaining asthenosphere flow stays aligned with the plate geometry through plate growth and migration. Secondly, the outward tractions can help localize the divergent plate boundary.

While continental rifting precedes seafloor spreading, there are clear fundamental differences between the two processes. Continental rifting typically entails stretching continental crust, frequently, but not always with significant magma emplacement (Bialas et al., 2010). Continental rifts may be wider or narrower (100s of km wide vs 100km or less) (Buck, 2007). Continental rifting involves thicker crust and lithosphere than that near spreading axes. In contrast, seafloor spreading generates entirely new igneous crust (as opposed to stretching) in a more continuous fashion reflected in regular and symmetric gravity and magnetic anomalies.

The ocean has always been treated as incidental to seafloor spreading. However, the above isostatic considerations suggest that it could play an important role in facilitating plate separation during seafloor spreading. As noted above, there are fundamental differences in structure and process between continental rifting and seafloor spreading and to date the tipping point for a system to move from rifting to spreading has been elusive. Arguably, the most fundamental difference is in seafloor spreading being more continuous and self-sustaining in its behavior, reflected in systematic morphology, magnetic anomalies, gravity anomalies, and ultimately the steady-state creation of new oceanic crust. Study of rifts and rifted margins to understand the transition to spreading has led to a wealth of information about crustal behavior and evolution during extension (e.g., GeoPRISMS, 2015). However, the diversity of rifted margins in structure and strain history gives no overarching theory as to the successful tipping points of when a continental rift will transition to seafloor spreading (Brune et al., 2023).

The elimination of basal resistance that arises with thermal subsidence in the presence of an ocean may present this tipping point for the transition of continental rifting to seafloor spreading (Conder, 2022a). It has long been recognized that tractions on the base of the lithosphere are central to plate tectonics; Yet, basal tractions may be either driver or drag to plate motions depending on whether the underlying viscous mantle leads or lags the motion of the overlying plate (e.g., Forsyth and Uyeda, 1975). In mature spreading systems, forces like ridge push and slab pull likely dominate though their effectiveness is greater when basal drag is diminished. As continental rifts must be exerting forces sufficient for lithosphere extension over resistance from the lithosphere itself and from viscous basal tractions, an active rift system experiencing inundation can receive a positive push towards stable extension with the reduction in resistance that comes with the onset of isostatically driven undertow.

## 3.1 Quantitative development of oceanic isostasy

Drawing largely on Conder (2012), a brief sketch of the underlying physics is presented here. A useful starting point for understanding oceanic isostasy is with Pratt isostasy and the square root of age-depth equation for seafloor,

$$d = d_{ridge} + c\sqrt{t}, \tag{1}$$

where $d$ is ocean depth, $d_{ridge}$ is ridge depth and $t$ is seafloor age. The subsidence constant, $c$, depends on several factors, including thermal expansion, mantle temperature, thermal diffusivity, mantle density at reference mantle temperature, $\rho_m$, and ocean density, $\rho_w$. Lumping properties other than densities into a constant, $f$, gives,

$$c = f\,\frac{\rho_m}{\rho_m - \rho_w}, \tag{2}$$

The factor $\rho_m /(\rho_m - \rho_w)$ accounts for the mass due to the overlying substrate. Subaerially, this factor goes to unity. Subaerial cooling is fully Pratt compensated as no mass enters or leaves with cooling and contraction. For mantle density of 3330 kg/m$^3$ and ocean density of 1030 kg/m$^3$, this implies submarine subsidence 45% greater than subaerial subsidence for crust of similar age (Figure 3). As this excess 45% comes from ocean mass being continually redistributed among columns during subsidence, a mechanism other than Pratt isostasy must accommodate nearly 1/3 of the gravitational disequilibrium. Reattaining gravitational equilibrium to account for the water mass redistribution requires a counter redistribution of mass in the asthenosphere.

The amount of mass that ultimately must be redistributed from any column is equal to the excess mass, $\gamma$, increase arising from the overlying ocean, the difference between submarine and subaerial subsidence curves (Figure 3a). Assuming $c_{submarine} \sim 0.35$ km/Myr$^{1/2}$

$$\gamma = \left(0.11 \frac{km}{Myr^{\frac{1}{2}}}\right)\sqrt{t}. \qquad (3)$$

As this is a continually ongoing process with previous disequilibria already adjusted for, the relevant curve to consider is the rate of mass increase (Figure 3b). Differentiating equation 3 with respect to time gives the rate of mass accumulation in each column,

$$\dot{\gamma} = 0.055 \frac{km}{Myr^{\frac{1}{2}}} / \sqrt{t}. \qquad (4)$$

Because younger seafloor subsides more rapidly than older seafloor, excess is added more rapidly to young seafloor requiring a counter asthenospheric flow away from young seafloor towards older seafloor.

The change in mass in a column due to lateral asthenosphere counter flow is the mass flux into the column minus the mass flux out of the column. In 1D, the governing equation becomes,

$$\dot{\gamma}^* - \frac{dU}{dx} = 0, \quad (5)$$

where,

$$\dot{\gamma}^* = \dot{\gamma} - \dot{\gamma}_{mean}, \quad (6)$$

and $U$ is the lateral asthenosphere mass flux (in units of $km^2$/Myr). The flux as a function of position, $U(x)$, can be found by

$$U(x) = \int_0^X \dot{\gamma}^* dx, \quad (7)$$

As shown in the supplementary material of Conder (2022a), associated basal tractions, $\sigma$, may be estimated by assuming $U$ is maintained through Poiselle flow in the asthenosphere. Given a channel width, $H$, and viscosity, $\eta$,

$$\sigma = 1.5\eta U/H^2. \quad (8)$$

To maintain consistency with the previous work, calculations of tractions in this manuscript assume $H$ to be 100 km and $\eta$ to be $10^{20}$ Pa·s. A numerical solution is given in Conder (2022a) and maintained as a MatLab function at: https://opensiuc.lib.siu.edu/geol_comp/.

## 4 Localization of spreading centers

A clear difference of mid-ocean spreading centers relative to continental rifts is the narrow, stable, well-defined axis of extension giving rise to localized seismicity and a regular pattern of magnetic anomalies. While the weakest part of the plate will preferentially rupture under extension, the systematic undertow induced from oceanic isostasy may directly factor in the tighter orderliness of spreading centers. The outward mass flux centers on the locus of greatest submarine subsidence (most rapid ocean deepening) (Figure 4). For most of the ridge system this is the spreading axis, so alignment of outward flow from the axis reinforces mantle tractions tending towards a single locus of extension. A quantification of localization forces can be found by integrating outward tractions as a function of location (Figure 4d). Positive values promote extension and negative values promote compression.

Even in a messy thermal state, potentially like that of a continental rift, a preferential spreading locus may develop with submergence. Figure 5 shows localization preference for an instance with two loci of rifting. Even with two loci, the basal mantle tractions give a preference for localizing at one over the other. Going further, figure 6 shows two examples of randomly distributed thermal ages (a), their associated mass fluxes arising with submergence (b), and the strength of localization across the system (c). Even with randomness in the thermal system, once submergence occurs, there are preferential loci for extension where subsequent new crust is likely to be emplaced, leading to a self-reinforcing feedback loop of raised temperatures and further localization in that place.

## 5 Geological examples

There are a number of areas across the globe worth examining their geophysical character and/or geological history in the context presented here. For instance, considerations could be made at rifted margins, including complex rifts like the Galicia Bank (Druet et al., 2018; Grevemeyer et al., 2022), the Gulf of Mexico (Kneller et al., 2012), and/or backarc spreading centers.

### 5.1 Subaerial vs. submarine spreading

With its subaerial spreading, Iceland presents a unique natural test and illustration of the process at work. Extensional tectonics on the island are known to differ in substantial ways from typical seafloor spreading; Extension takes place over a broader area, dikes and earthquake swarms are common, and ridge-parallel strike-slip faulting is documented (Einarsson, 2008; Karson, 2017; Sigmundsson et al., 2020). Thicker crust (Einarsson, 2008) and/or plume processes (Karson, 2017) likely contribute to these differences, but its subaerial nature may also be a contributing factor. To investigate the relative importance of the subaerial nature of extension on the island relative to thicker crust and/or hotspot processes, one can look to the portion of the spreading system that is both on-platform (thick crust) and submarine; Contrasting this section of the spreading system to adjacent submarine off-platform and subaerial on-platform sections offers a valuable examination at the relative importance of an overlying ocean on spreading center dynamics versus thicker crust and/or hotspot.

As a first order observation, ridge morphology suggests that the coastline may be at least as important of a factor in the spreading process as crustal thickness. Figure 7 shows a digital elevation model where the Reykjanes Ridge crosses the Icelandic platform and onto the island. The transition in morphology of the Reykjanes Ridge from a narrow band of en echelon spreading segments (Saemundsson et al., 2020) to the wider and more distributed volcanic zones of Iceland proper occurs across the Reykjanes Peninsula; In contrast, there is minimal response observed in the crossing from off-platform to on-platform. If the difference in morphology were primarily controlled by the structure of the platform, such as its thicker crust, the morphological differences should track with the edge of the platform rather than the coast as observed.

Likewise, it would be a surprising happenstance for the change to correspond tightly with the coastline if the morphology change were strictly plume-controlled.

More than just ridge morphology, changes in character of earthquake distribution, gravity and magnetic anomalies, and even the eastward shift of the plate boundary are all more tightly associated with the coastline than with the edge of the Icelandic Platform (Conder, 2022a). Together these further the case that the presence of the ocean is an important factor in the observed differences. Figure 8 shows the isostatic gravity anomaly on and around the platform (Bonvalot et al., 2012). Figure 9 shows seismicity and slip vectors of extensional earthquakes from the CMT catalog (*globalcmt.org*) on and around the platform. Events are determined to be extensional, compressional, or strike-slip by the most vertical stress axis of the moment tensor: P, T, or B. Earthquakes on the submarine platform occur along a localized spreading axis and are primarily associated with normal faulting more akin to earthquakes off-platform than on the island. Across the Reykjanes peninsula events are dominated by strike-slip bookshelf faulting rather than normal faulting (Einarsson et al., 2020). Further into the island many events are associated with volcano-tectonic processes rather than simple plate separation (Einarsson, 2008), but nowhere on the island can the same coherency of extensional slip be found as on the submarine platform. Similarly, isostatic gravity anomalies have a greater change in character across the coastline than across the platform edge with anomalies on the island being less spatially coherent and showing minimal definition with the local rift axes. In virtually every geophysical dataset observed, the coastline demarcates a greater change in geophysical character than delineations expected from associations with thick crust or plume processes. The tight correlation of tectonic character changes with the coastline is a strong indicator that the ocean is not passive but an active participant in the seafloor spreading process.

As Iceland is bounded by submarine regions to each side, its tectonic behavior may be subtly different than for a simple submerged/non-submerged dichotomy. Specifically, while the localization of the ridge axis for Iceland may be lacking relative to the totally submerged case in Figure 4, the mantle tractions arising around the island may be important to consider. Figure 10 shows the isostatic flow response and localization for a subaerial axis with submerged flanks. Importantly, the subaerial region does not gain additional mass with thermal subsidence. The location most rapidly accumulating mass via subsidence is the submarine region just beyond the shoreline (Figure 10a). As the outward mass flux aiding seafloor spreading is not present beneath the subaerial region, the flux beneath the plateau may be directed inward (Figure 10b). If the island participates fully in the gravitational equilibrium process with the surrounding oceanic regions, localization is preferred just outside the island's edges (Figure 10d). However, if the island does not efficiently participate in the gravitational equilibration process - as the isostatic anomalies suggest (Figure 8) - localization will preferentially occur within the subaerial region, but no strong preference beyond that (figure 10e). This lack of strong localization may be why the on-island axes develop in multiple zones (figure 7) and routinely shift around the platform (Foulger et al., 2020); Similarly, informing why extensional earthquake slip vectors lose consistency on the island while maintaining consistency on the submarine platform (figure 9).

Progressing back in time, this effect of biasing extensional axes to within subaerial regions flanked by ocean may be seen in the Geological history of the Greenland-Iceland-Faroe ridge.

Past rift axes associated with Iceland were not restricted to the current island; Multiple former rift axes jumped around nearly the entire length of the Greenland-Faroe ridge (Foulger et al., 2020). Importantly, these rift axes were mostly subaerial at the time of their existence as the Greenland-Iceland-Faroe ridge was a land bridge until 10-15 Myr ago (Denk et al., 2011; Ellis and Stoker, 2014). As submergence of the ridge progressed, localization as seen in figure 10e likely continually constrained rift axis jumps to remaining subaerial regions – culminating in the configuration they are now.

## 5.2 Seaward dipping reflectors

Rifted margins are often described in terms of 'magma-rich' and 'magma-poor' end members (Franke, 2013; Peron-Pinvidic et al., 2019; Tugend et al., 2020). Magma-rich margins are typically characterized as having a package of seaward-dipping-reflectors (SDRs), frequently interpreted as thick packages of lava flows occurring during an increase of magma production associated with creation of the first oceanic crust (e.g., Chauvet et al., 2021). It may be the case that seaward dipping reflectors (SDRs) in the distal portion of the rift system are indicators of subaerial spreading (Collier et al., 2017; Jackson et al., 2000; Karner et al., 2020; Mutter et al., 1982) with analogs today in Iceland (Karner et al., 2020) and the Afar subaerial section of the Red Sea Rift (Bastow and Keir, 2011; Corti et al., 2015). Given the prevalence of magma-rich margins (e.g., Geoffroy, 2005), a high frequency of possible subaerial spreading in the geological record appears to point away from the ocean having an important role in the development of seafloor spreading. On close inspection, though, these may be exceptions that prove the rule. Despite their prevalence along the Atlantic and other margins, Iceland and Afar are the only analogs we see today; Iceland and Afar are clearly unusual in the current spreading system in that 1) subaerial spreading accounts for < 1% of the current spreading system, and 2) as noted earlier, their behavior contrasts in fundamental ways with the submarine spreading system.

If distal SDRs are indicators of subaerial spreading, the intrinsic differences to seafloor spreading are important. It seems probable that without the facilitating mantle tractions, extension in Iceland and Afar are largely a kinematic response to actively facilitated spreading in the adjacent submerged plate boundary. Magma-rich in this case may indicate development of a steady state magma chamber with lithospheric thinning but without the localizing effects that come with submergence. Conspicuously, even a magma-rich region like Iceland with long-lived magma chambers appears to be largely driven by far-field tectonics as opposed to magma-driven opening (Kolzenburg et al., 2022). That magma-rich and magma-poor margins can vary over short spatial scales (Gouiza and Paton, 2017; Shillington et al., 2009) is consistent with a model of spreading tracking with inundation but with instances of Icelandic-type extension leading inundation; Spreading leading inundation may occur with adjacent inundated self-sustaining spreading centers forcing the subaerial neighbors in the system. Presumably most of these instances would evolve to stable and submerged seafloor spreading.

### 5.3 Rift to drift

While the number of locations where the mid-ocean ridge system crosses the coastline are limited, additional geological evidence for the ocean actively facilitating seafloor spreading can be found with the rift-to-drift transition. Places like the Afar triple junction with three extensional boundaries at different stages of the rift-to-drift transition as well as Atlantic rifted margins show the rift-to-drift transition tracking more closely with inundation than other factors such as obliquity, rifting style, or even rifting intensity (Conder, 2022a). The list of locations in considering the relationship between rupture and inundation is extensive including rifted margins (Sapin et al., 2021; Unternehr et al., 2010), backarc spreading centers, and rifts on other planets.

One example to look to is rifting in the Gulf of California (GoC), a complex rifting and drifting system (Umhoefer et al., 2020) exhibiting along strike variability in rifting structures (Lizarralde et al., 2007; van Wijk et al., 2017). Extension is primarily accommodated along multiple long transform faults connecting several short spreading segments (Figure 11). Although young, seafloor spreading and extension in the GoC has a complex evolution culminating in fully developed spreading in the south and pull-apart basin extension in the north (van Wijk et al., 2017). Prior to opening, subduction of the Farallon plate occurred off the west coast of Baja California. Rifting began in Baja after the subducting plate was consumed (~12 Ma); Subsequently, the GoC captured the primary plate boundary along the previously active continental arc moving it inland (Michaud et al., 2006; Umhoefer, 2011). Rupture along the ~1500 km long system occurred rapidly, co-opting the plate boundary within 6-10 Myr after cessation of subduction (Umhoefer, 2011). While the rapid coalescence of the boundary must have had much to do with the weak arc lithosphere (Umhoefer, 2011), observed variations in along-strike extension style suggest other superposed tectonic effects at work as well, such as effects of sedimentation (Bialas and Buck, 2009), mantle fertility (Lizarralde et al., 2007), or structural styles (van Wijk et al., 2017). Of particular importance here, marine incursion in the GoC also appears to have a close relationship with the plate boundary coalescence and subsequent evolution and development of seafloor spreading (Umhoefer et al., 2018).

To first order, it is easily observed that seafloor spreading is occurring in the southern gulf which had more ready access to the sea contrasting with earlier stages of rifting occurring in the northern section. Of course, the details are more nuanced. Marine incursion occurred in the southernmost Gulf by 8-9 Ma corresponding with spreading propagation of the East Pacific Rise towards the arc into the southernmost gulf (Umhoefer et al., 2018). Marine incursion reached all the way to the Salton Trough around 6.5-6.3 Ma (Bennett et al., 2015; Umhoefer et al., 2018) at which time, the majority of the plate boundary localized in the GoC and the Guaymas basin in the central gulf fully transitioned to spreading (Umhoefer et al., 2018). Although the Salton Trough is no longer inundated, marine incursion into the trough may have occurred as late as the Pleistocene (Ross and and Jefferson, 2020). Free-air gravity anomalies suggest an isostatically coherent rifted domain bounded by subaerial flanks extending from the southern gulf to the Salton Trough (figure 11). That the gravity coherency crosses the oceanic-continental crust boundary within the gulf, but not across the coastline to either side (Figure 11b) suggests an ocean link to a similar isostatic behavior along the full GoC spreading system. Because of the narrowness of the GoC, it should be underscored that the viability of oceanic isostasy as a tectonic driver depends on load (incursion) width and strength of the lithosphere to support that load (Conder, 2022a); High-strength lithosphere and narrow loads are resistant to developing outward mantle tractions. Still, the gravity signal suggests that the entire rift system participates in the system's isostatic mass exchange.

While the Salton Trough remains a pull-apart basin, it is not wholly clear whether currently submerged northern gulf basins have recently ruptured fully to seafloor spreading or remain in pull-apart mode (Martin-Barajas et al., 2013; van Wijk et al., 2019) as thick sediments obscure straightforward interpretation. In addition to obscuring the northern basins, the thick sediments of the northern gulf have been suggested to affect GoC tectonics. Thick sediments can reduce the buoyancy forces across the rift and tend to narrow rifting and aiding the rift-to-drift transition (Bialas and Buck, 2009). However, the rift-to-drift correlation with the sedimentary package in the GoC runs counter to that suggesting thermal effects may be more important (Martin-Barajas et al., 2013). While sediments are not efficient as a thermal blanket to weaken the lithosphere (Bialas and Buck, 2009), it may be that sediments will tend to smooth the thermal variations, reducing the degree of differential thermal subsidence, thereby slowing the localization from oceanic isostasy and delaying the rift-to-drift transition.

### 5.4 Rafted continental slivers

Many rafted pieces of continental crust litter the world's ocean basins (Ashwal et al., 2017; Grevemeyer et al., 2022; King et al., 2020; Kumar et al., 2019; Nauret et al., 2019; Polteau et al., 2019; Santos et al., 2019; Scotchman et al., 2010). The fluid nature of the ocean can lead to rapid changes of an ocean's influence on underlying mantle tractions; In particular, rapid changes in the preferential locus of spreading can occur as the ocean recedes or advances in specific locations. Competing rifts may be inundated at different times; If a weaker rift is inundated first, it may begin to transition only to give way when a water path inundates a stronger competing second rift. For example, the Sao Paulo High presents an exception to the otherwise regular northward progression of the rift-to-drift transition in the South Atlantic (Heine et al., 2013; Moulin et al., 2010) (Figure 12). During this stall, the block comprising the rise was transferred from the African plate to the South American plate (Heine et al., 2013; Scotchman et al., 2010). It is plausible that a water path northward crossed an immature rift on the South American side of the block to link to the rift system northward. As spreading matured north and south, a second water path may have inundated a rift on the African side of the block transferring the locus of extension to the other side (Conder, 2022b). It is not difficult to imagine similar scenarios for other stranded blocks of continental material following the rift-to-drift transition. While speculative for any individual location, such a process would be an expected implication of inundation playing an active role in seafloor spreading development.

## 5.5 Backarc spreading

While the GoC is a former backarc evolved to spreading system, many current backarcs also have developed spreading (Martinez et al., 2007). Back-arc spreading presents a curious condition of extension in an overall compressional environment. In general, backarcs in subduction systems can be compressional or extensional with extensional systems potentially developing into spreading systems. Backarc spreading commonly results from rifting of an oceanic volcanic arc that generates and migrates into a backarc basin (Martinez and Taylor, 2003). The difference in systems with backarc extension and compression have long been couched in terms of low-stress (extensional) ‘Marianas’ type and high-stress (compressional) ‘Chilean’ type subduction with high-stress more typical for continental arcs and low-stress more frequent for oceanic arcs (e.g., Uyeda, 1987). The variability in subduction and backarc systems has led to considerations of multiple variables to explain the differences such as slab dip (Bott et al., 1989) or sediment thickness (Cloos and Shreve, 1996); More refined distinctions such as accretionary vs erosional margins (e.g., Harris et al., 2014) or tsunamigenic vs non-tsunamigenic (e.g., Bilek and Lay, 2018) have been further advanced. The mechanisms driving the development of backarc spreading are likewise varied in the literature (Goes et al., 2017; Heuret and Lallemand, 2005; Martinez and Taylor, 2006; Sdrolias and Muller, 2006; Wallace et al., 2005). Still, it is clear that extensional backarc basins dominate the oceans while compressional backarcs are more prevalent in continental regions (Heuret and Lallemand, 2005). While admittedly appearing somewhat tautological, the presence of ocean in oceanic backarcs may be a contributing factor to the development of backarc spreading.

An ocean-ocean subduction zone in particular presents a geometrical layout favorable to active mantle tractions for facilitating spreading. The arc presents a strip of crust with thinned lithosphere and elevated temperatures (Ha et al., 2020) and high heat flow (e.g., Von Herzen et al., 2001) and therefore differential thermal subsidence. For largely submerged arc-backarc systems such as the Marianas (Figure 13a), Tonga (Taylor et al., 1996), or South Sandwich (Leat et al., 2016) this differential subsidence will tend to drive flow outward into the backarc away from the arc; Direct loading of volcanic deposits will add to the isostatic response, although likely in transient pulses. The resultant mantle tractions will augment existing tensional stresses in the overlying plate and tend to localize extension beneath the arc. If localization and full rupture does occur, the locus of differential subsidence will be strongest at the spreading axis (Figure 5) potentially leading to sustained spreading and migration into the backarc basin with continued crustal accretion.

Pointedly, the rift-to-drift transition in backarc basins can be fundamentally different than on continental rifted margins (Wang et al., 2019). Inherent differences may arise in that back-arc basins typically begin in a fully submerged state, while continental rifting is frequently associated with advancing inundation. Advancing inundation can invoke a transient additional push for rift opening (Conder, 2022a); Such differences may lead to inherently different opening progressions for backarcs relative to continents. For instance, rather than progressive extension leading to rupture (e.g., Brune et al., 2016; Davis and Lavier, 2017; Lavier and Manatschal, 2006), backarc spreading axes are shown to intrude laterally on diffusely extending regions through propagation or synchronous jumping (Dunn et al., 2013; Taylor et al., 1995; Wang et al., 2019). That an already extensively faulted and extending region preferentially localizes along a crack originating outside the area is peculiar; The implication being that the outside propagating crack has something intrinsic to localization that the already present extensional cracks within do not. That a localized and propagating axis may have outwardly directed mantle tractions that track with its growth in contrast to faults within the area can straightforwardly account for the co-option.

The Okinawa Trough (Figure 13b) is an unusual case of backarc rifting within continental crust (Arai et al., 2017). Extension in the ~1000 km long Okinawa Trough began in the Miocene (Fournier et al., 2001). Current extension rates vary considerably along strike with ~1 cm/yr in the northern trough to ~5 cm/yr in the south (Nishimura et al., 2004). Extension is diffuse in the slower northern trough and localized along narrow axes in the central and southern trough (Arai, 2021; Nishizawa et al., 2019). Such structure must be transient as continued spreading will undoubtedly create oceanic crust within the basin; Still the current literature for the rift-to-drift transition makes little allowance for the possibility of localization within continental crust prior to the transition to spreading. As this is also an unusual case of a submerged continental arc and backarc, this joint occurrence may be a manifestation of the effectiveness of isostatically driven mantle tractions contributing to backarc spreading development and evolution.

## 6 Oceanic vs. continental tectonics

Notwithstanding the remarkable successes of plate tectonics in understanding the history and deformation of the Earth's crust, continental deformation only loosely adheres to the fundamental assumptions of non-deforming plate interiors and narrowly confined boundaries (England and Jackson, 1989; Molnar, 1988; Thatcher, 1995). Many plate boundaries on Earth can be classified as 'diffuse' in that they do not fit the classical definition of deformation being limited to a narrow boundary (Gordon, 1998; Stein and Sella, 2002) (Figure 14). For continents, plate boundaries frequently entail wide topographic and seismically active zones; These being the norm for both compressional (e.g., Tibetan plateau, Hindu Kush) and extensional regions (e.g., Basin and Range, portions of the East African Rift) rather than the exception (Gordon and Stein, 1992). Even 'narrow' single-valley extensional boundaries like the Main Ethiopian Rift (Corti, 2009) and Rio Grande Rift (Hudson and Grauch, 2013) are several tens of kilometers wide and 3:1 being a high length to width aspect ratio for individual basins. Continental strike slip boundaries like the San Andreas can be 100km or more wide (Bennett et al., 2002; Thatcher, 1995). In contrast, oceans tend to more closely adhere to the plate tectonic assumptions of rigid interiors and narrowly confined boundaries. This dichotomy presented a large obstacle for accepting the ideas leading to plate tectonics until after the oceans began to be systematically explored in the 20th century (Sandwell, 2001). The increasing quality and quantity of geodetic data, such as GPS, repeatedly exposes more regional complexity of continental deformation; The complexity frequently requires additional microplates, tectonic blocks, or diffusely deforming regions to adequately explain the data (Hasterok et al., 2022; Stamps et al., 2021; Wang and Shen, 2020). While diffuse boundaries are also recognized in the oceans (Gordon, 1998, 2023; Stein and Sella, 2002), they are of a decidedly different nature than those on continents (Figure 14); Diffuse oceanic plate boundaries are more subtle, being identified primarily by kinematic closure misfits and/or a moderate number of earthquakes (Stein and Sella, 2002) and characterized by smaller strain rates than other boundary zones (Zatman and Richards, 2002). Primary examples are found in the Indian ocean (Conder and Forsyth, 2001; Royer and Gordon, 1997; Wiens et al., 1985) and in the Atlantic between the North and South American plates (DeMets and Merkouriev, 2019).

The dichotomy between continental and oceanic tectonics is typically explained in terms of contrasting crustal rheologies with continental crust being thicker and often with a weak ductile layer in the lower crust (Molnar, 1988; Thatcher, 1995). While the differences in rheological profiles are undoubtedly important controls on deformation, especially for enabling deformation in the third dimension like crustal thickening (Molnar, 1988), it is worth considering whether the presence of an ocean may also play a role in the dichotomy. As discussed above, Iceland's switch from a narrowly confined boundary to wider deformation occurs at the coastline rather than the edge of the submarine platform (figure 7). For mantle tractions to localize spreading centers, there are two requirements: systematic differential subsidence (say from thermal variations) and an overlying fluid layer to redistribute mass with that differential subsidence. As diffuse oceanic boundaries do not tend to reflect major topographical or thermal variability, there is little facilitation of localizing the boundaries until rifting or some other process imposes a systematically varying thermal structure across the region.

## 7 Outstanding issues

The idea of oceanic isostasy as a facilitator of seafloor spreading poses potential answers to many aspects of our understanding of seafloor spreading and plate tectonics. However, it does pose several questions of its own that are currently unanswered and listed here. This list should be considered representative rather than comprehensive.

Possibly most importantly, the generated mantle tractions need to be considered within the broader system of mantle upwellings and convective circulation. The induced flow adds to the overall flow pattern, biasing mantle tractions towards positive push, but does not require the integrated flow at any given location to be in the plate spreading direction. So, other flows in the asthenosphere driven by plumes or other mechanisms may still be dominant mantle tractions when present. There are numerous places across the globe that show evidence for asthenosphere flow in opposition to plate motion directions (Behn et al., 2004; Conder et al., 2002; Conrad et al., 2007; Ghosh et al., 2013; Naliboff et al., 2009). Perhaps biasing the 'average' basal traction to positive is sufficient for the overall spreading system. Or perhaps, the fact that the isostatic response will happen as shallowly as possible, while convective flows may dominate more deeply keeps them from acting strongly in opposition. This will only be answered with a better understanding of the interaction of the various flows acting in the mantle.

The physics in Conder (2012) and used here for discussion assumes an axial high at the ridge axis, which is most appropriate for fast spreading centers. Slow spreading systems like the Mid-Atlantic Ridge exhibit median valleys (Small et al., 1998). Spreading from within a median valley requires dynamic uplift of young crust as it moves away from the axis prior to thermally subsiding (Phipps Morgan et al., 1987), following a different form of subsidence near the ridge axis than that seen in figure 3. This different form will likely reduce the effectiveness of localization of the plate boundary. Neovolcanic zones within median valleys are typically wider than at fast spreading centers, sometimes wandering laterally or broadening to 5-10 km within the median valley (Macdonald, 1982). These variations may be due to the lessened efficiency of localization, but to what degree that derives from having rift shoulders rather than monotonic subsidence remains to be determined.

It is clear than not every inundated or submerged rift transitions to seafloor spreading. In some cases like the Gulf of Suez, it is straightforward as to why; Strong lithosphere can support a narrow overlying load precluding an isostatic response (Conder, 2022a, b). However, locations like the submerged northern Okinawa Trough have wide loads of submergence and likely moderate to low elastic thicknesses and are extending diffusely. It seems possible that the slow rates of opening (Nishimura et al., 2004) may preclude full rupture, although slower spreading systems, such as the Gakkel Ridge in the Arctic (Cochran et al., 2003) have transitioned.

It is fair to say that our understanding of the interplay of oceanic isostasy as a facilitator of horizontal plate motions is in the early stages. There are numerous outstanding questions to be addressed as well as undoubtedly many as yet unidentified research avenues understanding the process’s contributions and limitations to shaping our planet.

## 8 Summary points

Differential thermal subsidence of the crust in the presence of an overlying ocean can induce a flow in asthenosphere to adjust for isostatic equilibration. Roughly 1/3 of oceanic plate gravitational equilibrium must be accommodated in this manner.

This self-sustained, isostatically-driven asthenosphere flow is directed outward from the thermally youngest (and most rapidly subsiding) part of the system, and may be a necessary component to the occurrence of seafloor spreading.

Mantle tractions induced by oceanic isostasy may lead to highly localized extension in contrast to less localized subaerial extension.

Subaerial spreading is different than submarine spreading in that it is not self-perpetuating nor may have a strongly localized boundary; Instead subaerial spreading is likely propelled along by the neighboring spreading system. Although less smooth than seafloor spreading, continual regular extension may be enough to sustain a steady-state magma chamber.

Extensional modes on Iceland contrast more strongly across the coastline than seen across the off-on platform transition or any other closed contour on the island. The coastline acting as the most prominent transition line illustrates a direct impact of the overlying ocean layer on facilitating self-perpetuating, localized seafloor spreading.

In addition to potentially facilitating seafloor spreading, oceanic isostasy may play important roles in development of other seafloor processes such as backarc opening and the existence of stranded continental slivers in the oceans.

The dichotomy of continental vs oceanic tectonics may be viewed at least partially as a dichotomy of subaerial vs submarine tectonics.

Contrary to the long-held view of the ocean as incidental to plate tectonics, the ocean itself may well be a crucial element to the seafloor spreading process and potentially why Earth is the only known planet to exhibit plate tectonics.

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**Availability Statement**

No new data are presented in this manuscript. Software for calculating mass fluxes as isostatic response is housed at the SIU open software site: https://opensiuc.lib.siu.edu/geol_comp/.

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**Figure captions**.

**Figure 1.** Figure from Hayford (1911) showing isostatic response of continual mass exchange at the surface being balanced by a continual counter mass flux at depth. The 'undertow' mass flux will induce stresses on the crust above, in this case resulting in compression in the crust near the base of the mountain belt.

**Figure 2.** Asthenosphere isostatic response to surface mass exchange that occurs with differential thermal subsidence at the seafloor. Cartoon is in a plate-fixed reference frame. Faster subsidence regions (younger ages) gain mass more rapidly as water deepens than regions with slower subsidence (older ages), inducing a counter mass flux that leads the plate in the plate spreading direction. In this plate-fixed reference frame, plate growth is to the left.

**Figure 3.** Subaerial and submarine square root of age (a) and mass gain (b) curves (after Conder, 2012). $\gamma$ corresponds to both the difference in the two curves and the height of asthenosphere column equivalent to the mass of the overlying ocean.

**Figure 4.** Calculations for 1D across-axis isostatic response of differential thermal subsidence at the seafloor. Top panel (a) shows seafloor thermal age to either side of a constant velocity, symmetrically spreading ridge. Second panel (b) shows mass gain that comes with subsidence and a deepening ocean. Dashed line is the mean gain line. To maintain gravitational equilibrium, seafloor with ocean mass gains above the line must transfer asthenosphere mass to seafloor with ocean mass gains below the line which have relative mass deficits. Third panel (c) shows the

resulting asthenosphere mass flux to maintain equilibrium (left axis) and associated basal tractions (right axis). Positive fluxes and tractions are directed to the right; Negative fluxes and tractions are directed to the left. Equilibrium (zero) line is dashed. Directions of flow also denoted by arrows. Bottom panel (d) shows the integrated outward push force at the seafloor as a function of location. Positive values promote extension, negative values promote compression. The push force peaks and tends to localize extension at the axis.

**Figure 5.** Example of two axes of extension. (a) shows thermal age across the system, with two locations having zero age. (b) shows the isostatic mass flux response across the system after submergence (left axis) and associated basal tractions (right axis). Dashed line is equilibrium. (c) shows the outward mantle tractions associated with extension. When submergence occurs, the isostatic response likely has preference for one of the axes to extend and reinforces localization at the expense of the other.

**Figure 6.** Examples of random thermal structure of a rift and isostatic response after submergence. (a) shows (randomly distributed) thermal ages across the system. (b) shows the isostatic mass flux response across the system after submergence. (c) shows the mantle tractions associated with extension. When submergence occurs, the isostatic response will likely result in a preferred locus of extension that will reinforce itself with continued extension.

**Figure 7.** Morphology of the midocean ridge system across the Icelandic platform and coastline. Ridge morphology changes are greatest across the Reykjanes Peninsula as opposed to across the off-platform/on-platform transition. ‘RR’ stands for Reykjanes Ridge. ‘WRZ’, ‘ERZ’, and

'NRZ' stand for Western Rift Zone, Eastern Rift Zone, and Northern Rift Zone, respectively (following *Karson* [2017]). Colors denote depth/elevation. Figure made in GeoMapApp (*geomapapp.org*).

**Figure 8.** Isostatic gravity anomaly of Iceland and surrounding region (Bonvalot et al., 2012). The isostatic anomaly being the difference from expected if the system were in Airy isostatic equilibrium Thin black line denotes coastline. The edge of the platform where intersected by the spreading system is also marked. Labels are same as in figure 6. 'KR' stands for Kolbeinsey Ridge. Figure made in GeoMapApp (*geomapapp.org*).

**Figure 9.** Seismicity and slip vectors for extensional earthquakes on and around Iceland. Blue circles are earthquakes of magnitude >4.5 1960-2022. Thick black line denotes coastlines. Medium thick black line is 400 m contour outlining the Iceland platform and thinner black lines show deeper 200 m contour intervals. Black arrows show predicted slip direction based on MORVEL Euler pole (DeMets et al., 2010). Green, cyan, magenta, and red lines show slip directions for extensional earthquakes from 1990-2023 in the CMT catalog (*globalcmt.org*). Extensional events are absent on the Reykjanes Peninsula as it is dominated by strike-slip bookshelf faulting. Colors show deviations from predicted directions: Green within 15°, Cyan within 30°, Magenta within 45°, and Red greater than 45° from predicted.

**Figure 10**. 1D calculations of mantle tractions for a representative oceanic region with a subaerial portion (gray region) where the thermal age is youngest. Half-spreading rate is set to 10 km/Myr. Top panel (a) shows mass adjustments arising from subsidence with an overlying

ocean. The subaerial region gains zero mass with subsidence. Dashed line is line of equilibration. Middle panels (b&c) show the corresponding mass exchange (left axis) and associated basal tractions (right axis) assuming that either the subaerial region participates in the mass exchange (b) or does not participate (c). Bottom panels (d&e) show corresponding associated outward push from induced mantle traction in b&c respectively. If the subaerial region fully participates in isostatic equilibration, preferred localization is just outside the edges of the island (d); If participation is limited, preferred localization is on the island itself but no preference beyond that (e).

**Figure 11.** Gulf of California topography (a) and free air gravity anomaly (Sandwell v30) (b). Plate boundary marked in red (extension) and black (transform). Boundary follows (Umhoefer et al., 2020) and (Martin-Barajas et al., 2013). EPR = East Pacific Rise. The exact boundary is more speculative in the northern gulf. The area of low free air gravity seen along the Gulf (thick orange line; roughly the 0 mGal contour) extends to and terminates at the Salton trough in the north indicating the extent of the region where isostatic accommodation process is correlated. The continental-oceanic crustal boundary is clear in the gravity across the gulf, but is subdued along the gulf. Figure made in GeoMapApp (*geomapapp.org*).

**Figure 12.** South Atlantic plate reconstructions at 120 and 115 Ma from Heine et al. (2013). Purple arrows show seafloor spreading propagation direction. A continuous northward progression stalled at ~120 Ma at the Sao Paulo High (SPH). Spreading pick up a few million years later to the north, again propagating northward, but also southward back to the SPH. The connection between the two is made on the African side of the high. The orange star denotes the

location of the extinct Abimael propagator (Sandwell et al., 2014) that propagated northward on the South American side of the SPH before succumbing to spreading on the other side of the high.

**Figure 13.** Bathymetry of the Marianas (a) and Okinawa Trough (b) backarc spreading systems. Colorbar is the same for both. The Marianas is a typical ocean-ocean backarc system. The mostly submerged arc system and remnant arc rafted off with spreading can be clearly observed. The Okinawa trough is a ocean-continent backarc system where extension is occurring much like that seen in ocean-ocean systems. Extension has thinned the continental crust to where distinct rift axes can be observed in the southern system. Figures made in GeoMapApp (geomapapp.org).

**Figure 14.** Diffuse plate boundaries from Stein and Sella (2002). Stippling indicates diffuse tectonic boundaries as well as their overall character. Broad seismicity, topography and faulting tends to characterize subaerial plate boundaries. Virtually all subaerial boundaries can be classified as diffuse. A smaller percentage of submarine boundaries can be classified as diffuse. Submarine diffuse plate boundaries tend to be more subtle than subaerial diffuse boundaries and identifiable by plate closure and/or a moderate number of earthquakes. The mid-ocean ridge system tends to be narrow and non-diffuse.

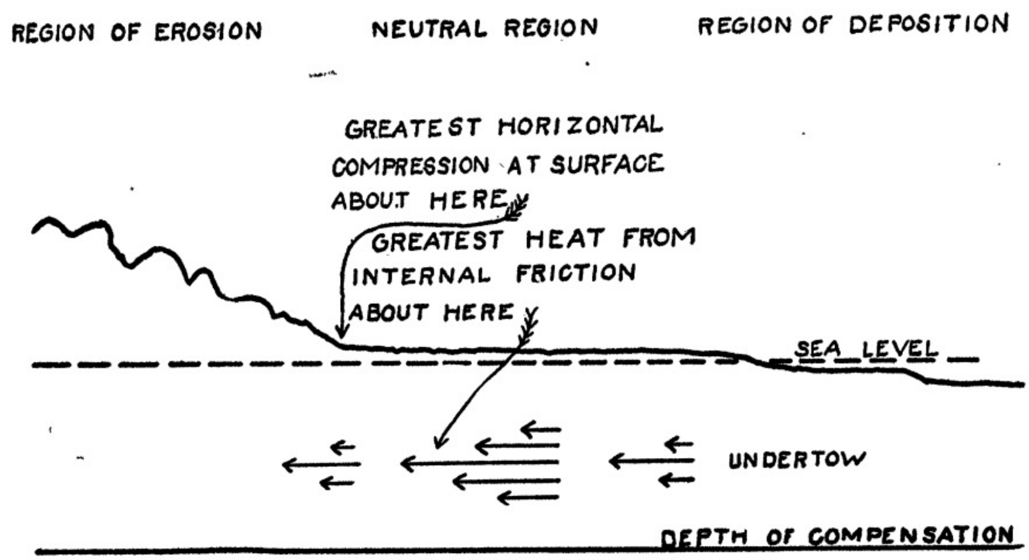

REGION OF EROSION
NEUTRAL REGION
REGION OF DEPOSITION
GREATEST HORIZONTAL COMPRESSION AT SURFACE ABOUT HERE
GREATEST HEAT FROM INTERNAL FRICTION ABOUT HERE
SEA LEVEL
UNDERTOW
DEPTH OF COMPENSATION

Figure 1

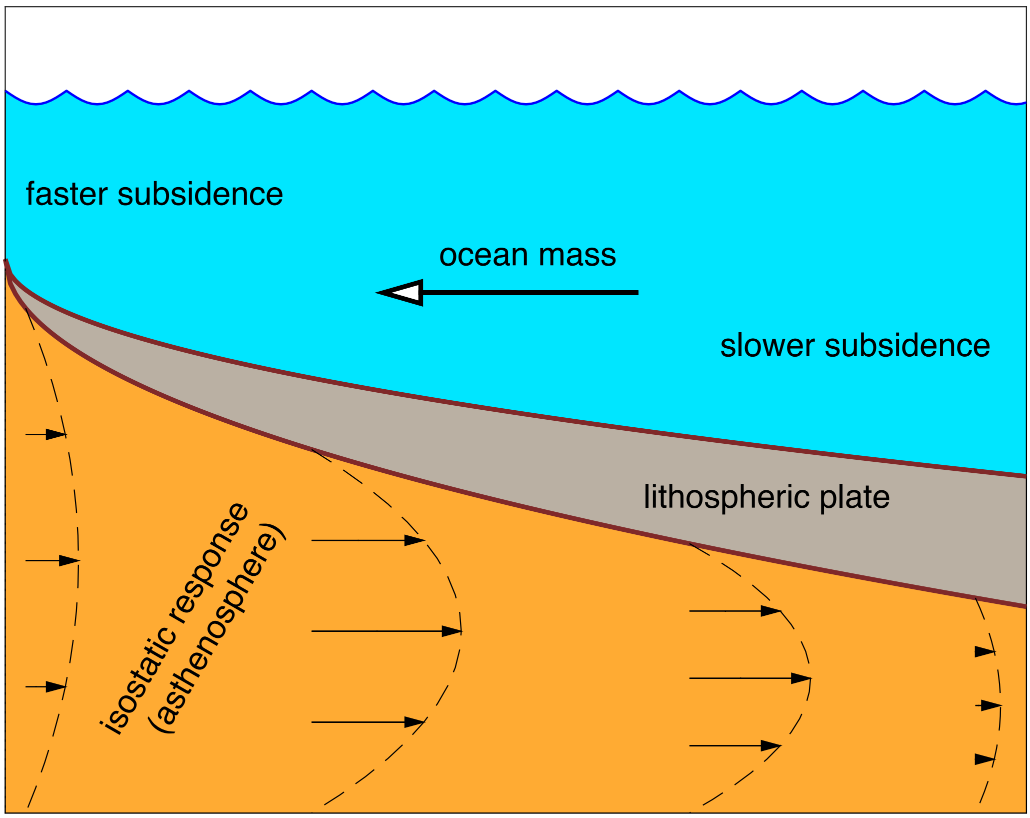

faster subsidence
ocean mass
slower subsidence
lithospheric plate
isostatic response
(asthenosphere)

Figure 2

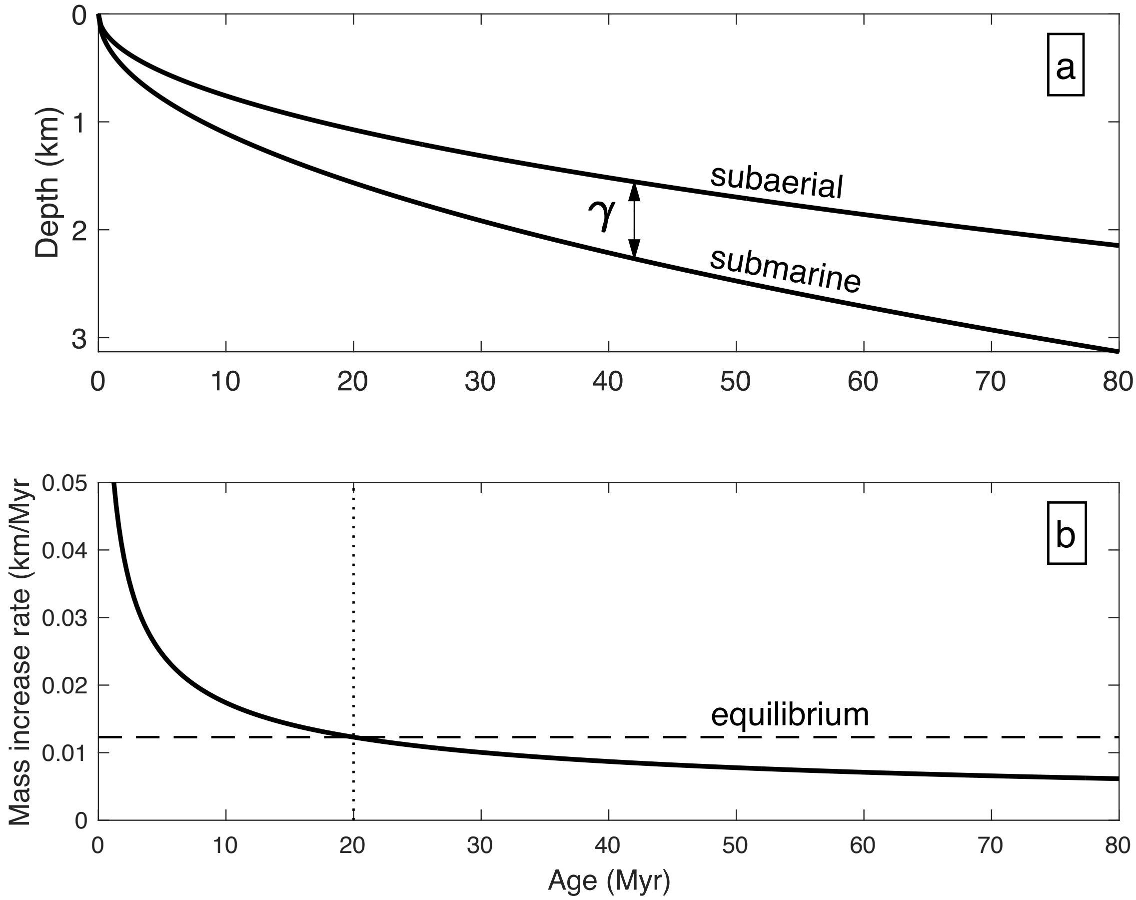
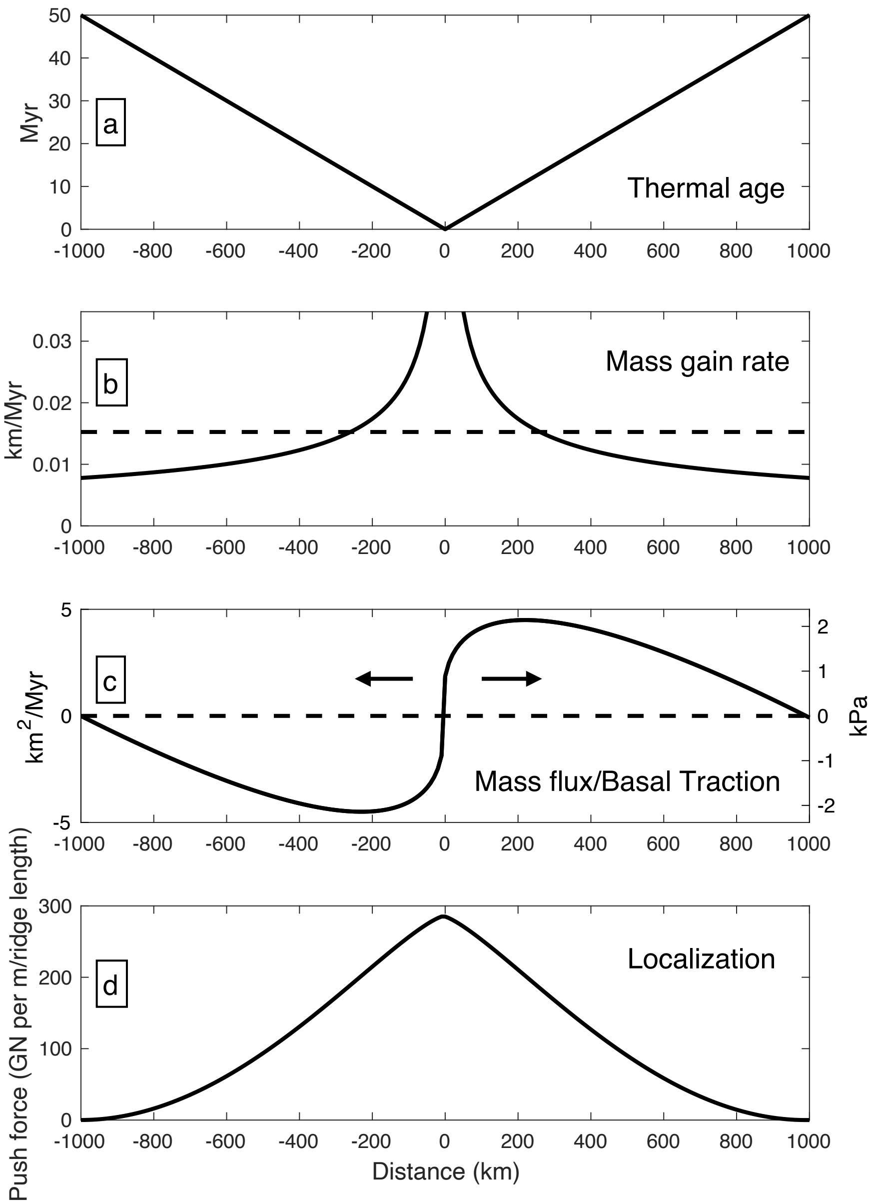

a
Depth (km)
subaerial
submarine
γ
b
Mass increase rate (km/Myr
equilibrium
Age (Myr)

Figure 3



a
Myr
Thermal age
50
40
30
20
10
0
-1000
-800
-600
-400
-200
0
200
400
600
800
1000
b
km/Myr
Mass gain rate
0.03
0.02
0.01
0
c
km$^2$/Myr
kPa
Mass flux/Basal Traction
5
0
-5
2
1
-1
-2
d
Push force (GN per m/ridge length)
Localization
300
200
100
Distance (km)


Figure 4

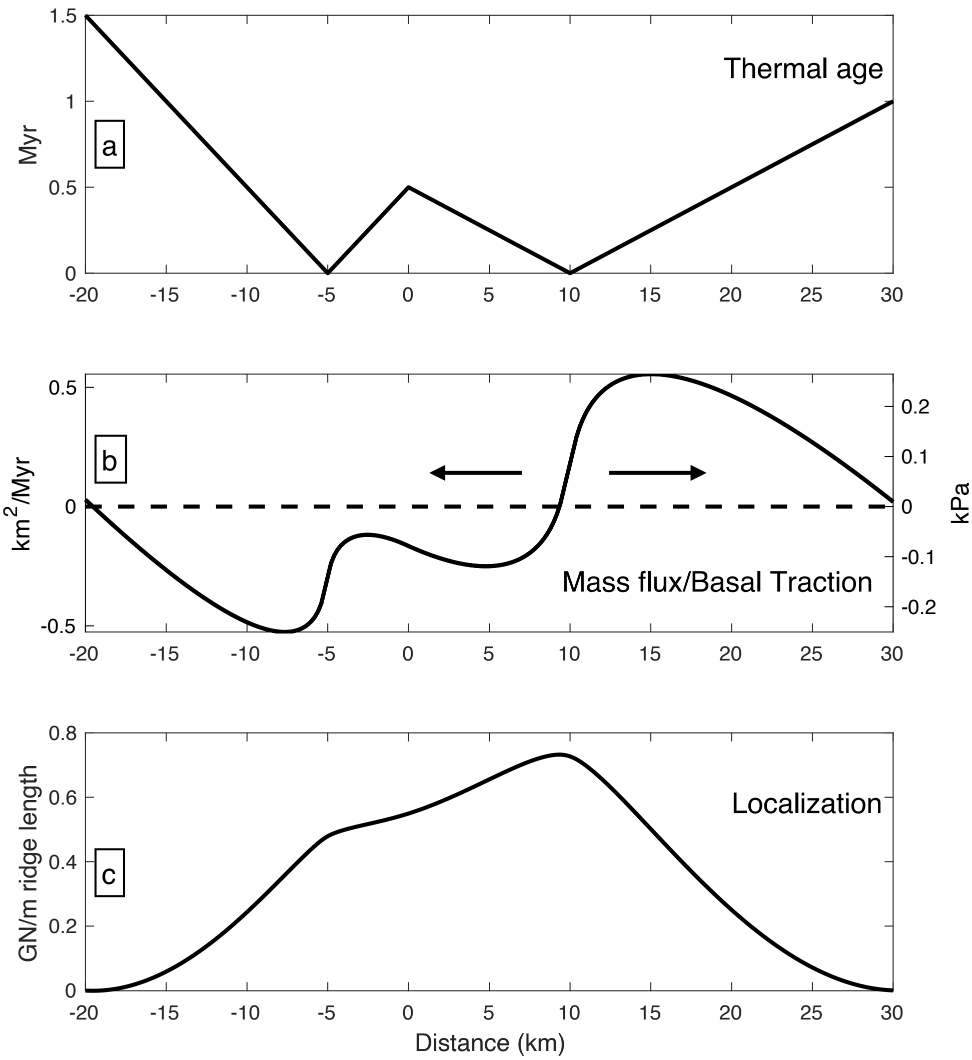

Thermal age
a
Myr
b
km$^2$/Myr
kPa
Mass flux/Basal Traction
c
GN/m ridge length
Localization
Distance (km)

Figure 5

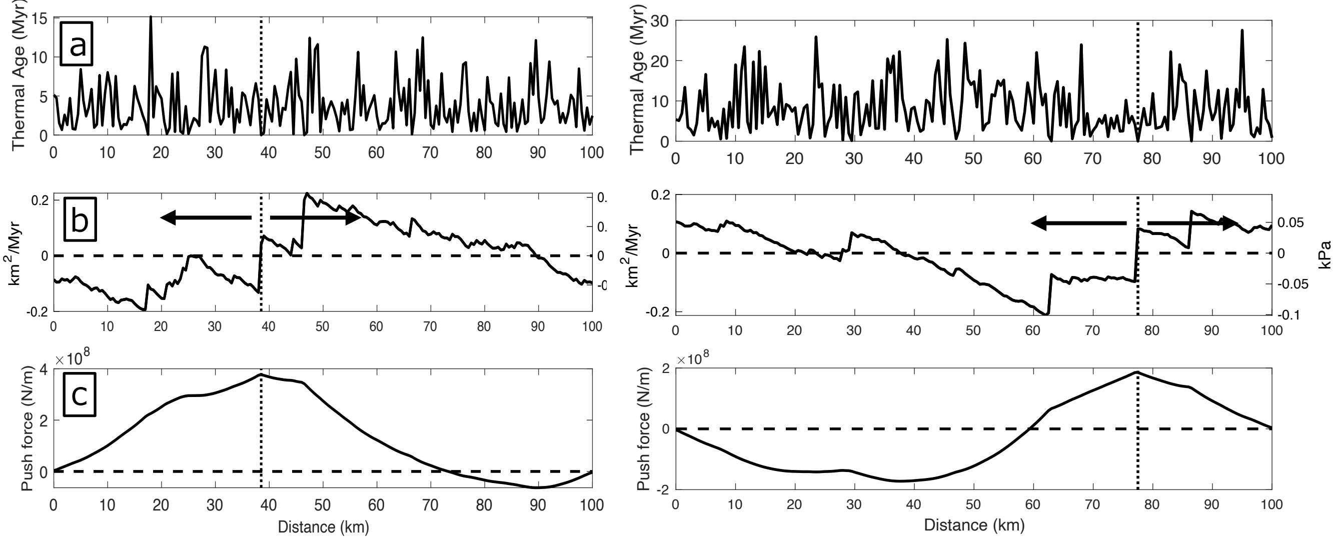



Figure 6

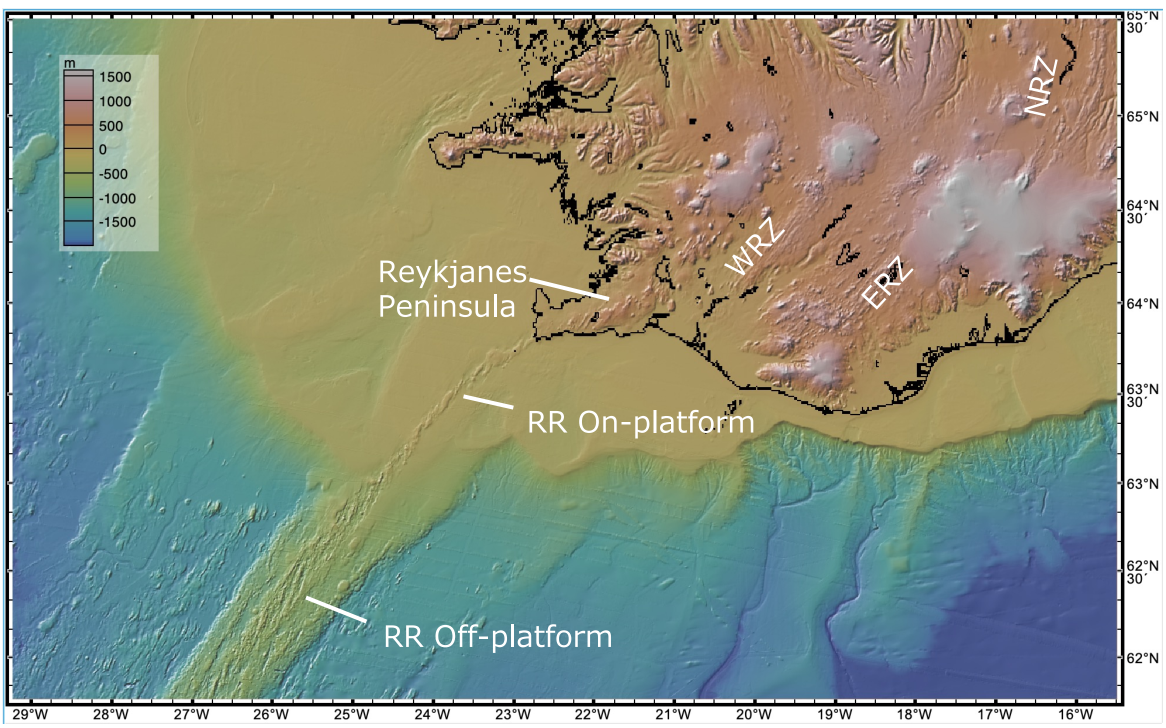
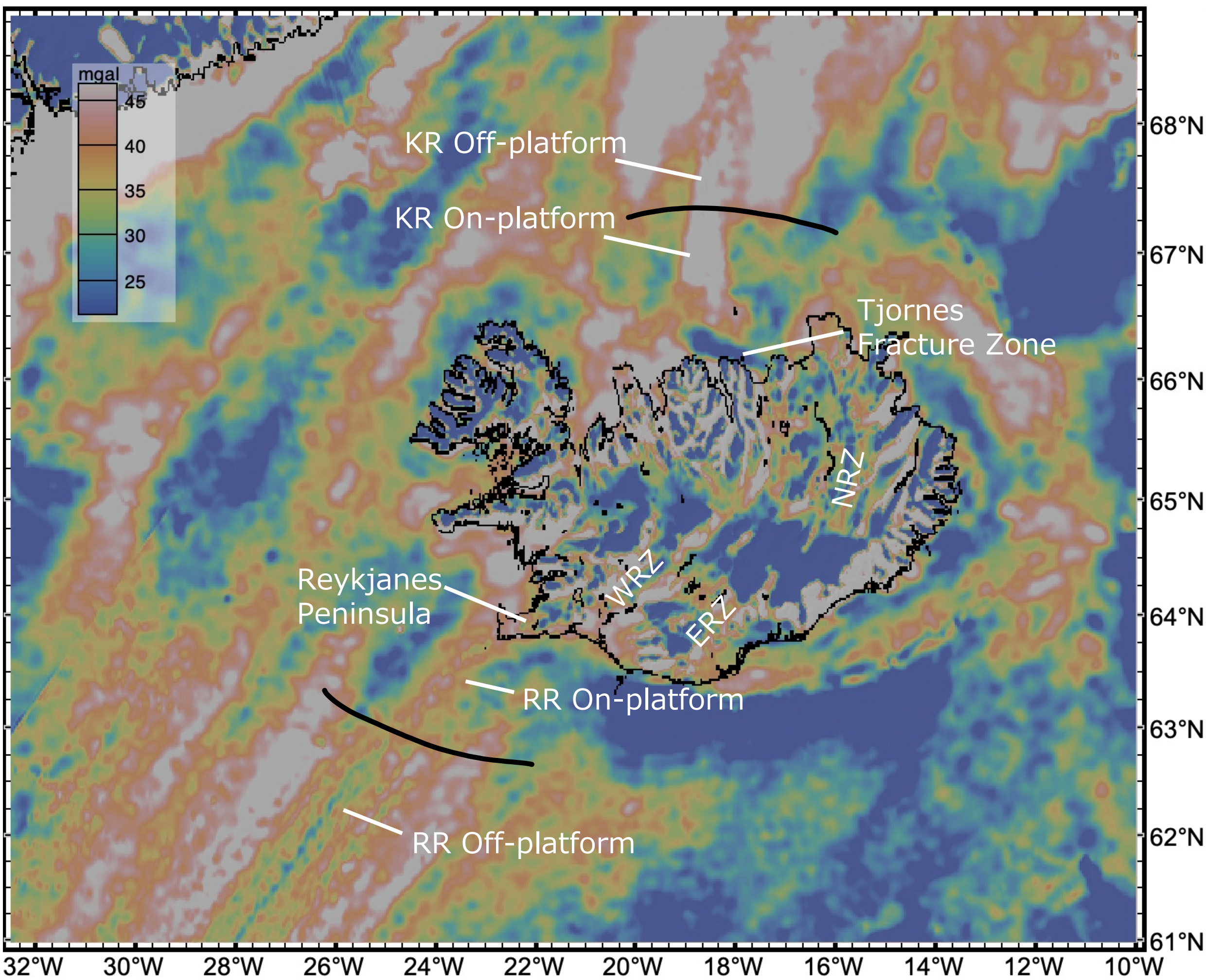

m
1500
1000
500
0
-500
-1000
-1500
Reykjanes
Peninsula
WRZ
ERZ
NRZ
RR On-platform
RR Off-platform
65°N
64°N 30´
64°N
63°N 30´
63°N
62°N 30´
62°N
29°W
28°W
27°W
26°W
25°W
24°W
23°W
22°W
21°W
20°W
19°W
18°W
17°W
16°W

Figure 7



mgal
45
40
35
30
25
KR Off-platform
KR On-platform
Tjornes
Fracture Zone
NRZ
Reykjanes
Peninsula
WRZ
ERZ
RR On-platform
RR Off-platform
68°N
67°N
66°N
65°N
64°N
63°N
62°N
61°N
32°W
30°W
28°W
26°W
24°W
22°W
20°W
18°W
16°W
14°W
12°W
10°W


Figure 8

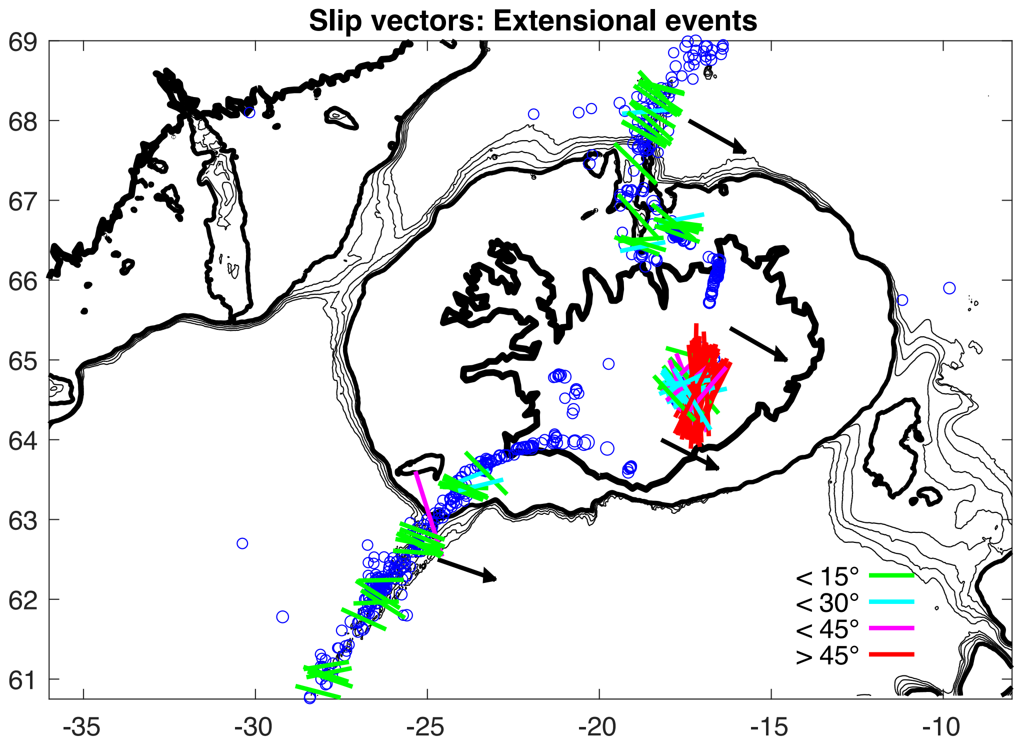

Slip vectors: Extensional events
69
68
67
66
65
64
63
62
61
-35
-30
-25
-20
-15
-10
< 15°
< 30°
< 45°
> 45°

Figure 9

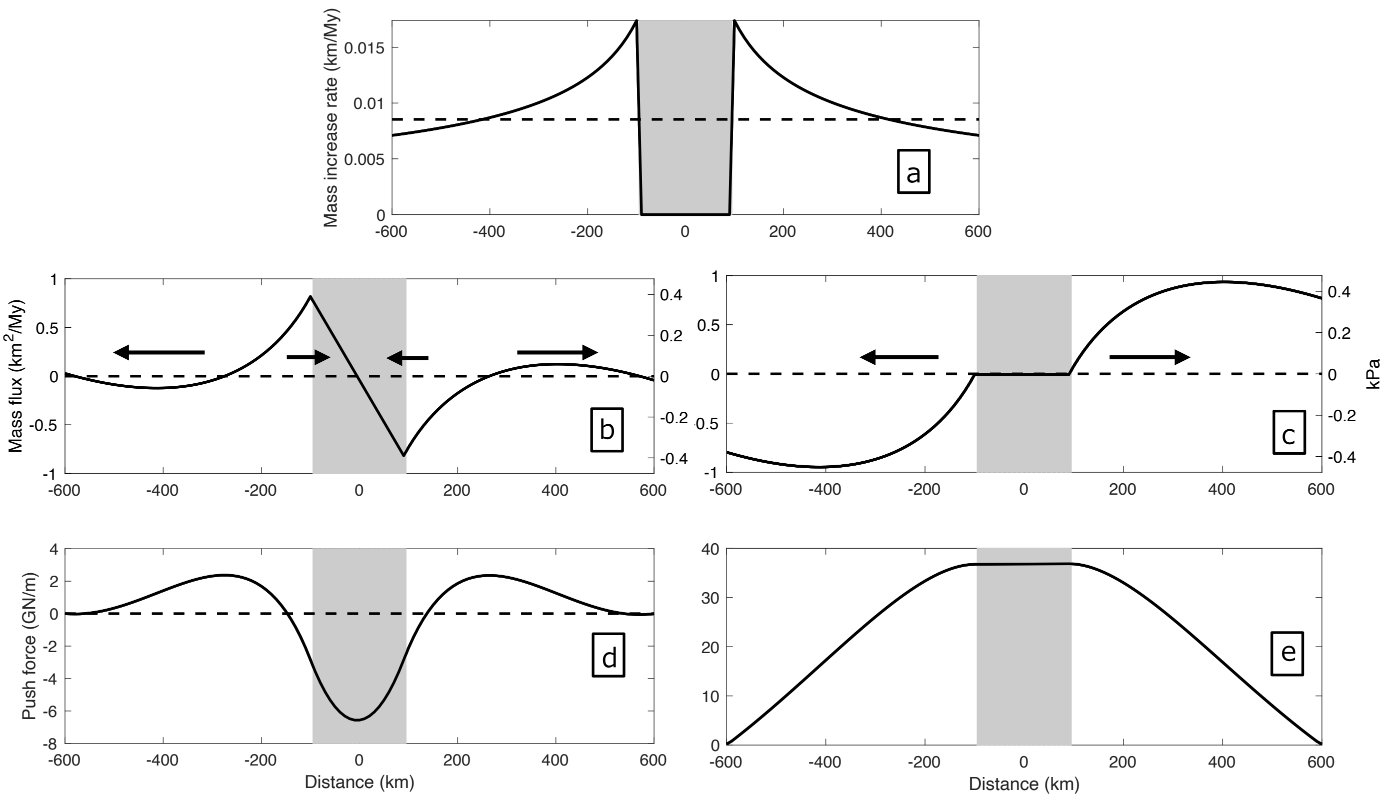

Mass increase rate (km/My)
a
Mass flux (km²/My)
b
c
kPa
Push force (GN/m)
d
e
Distance (km)

Figure 10

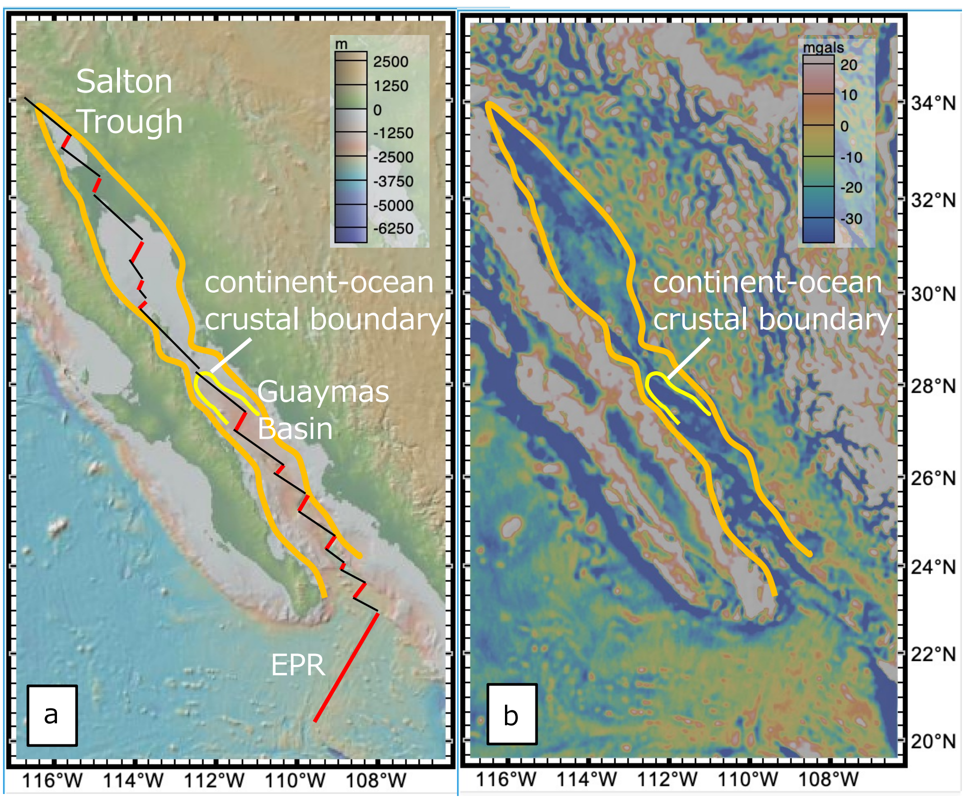
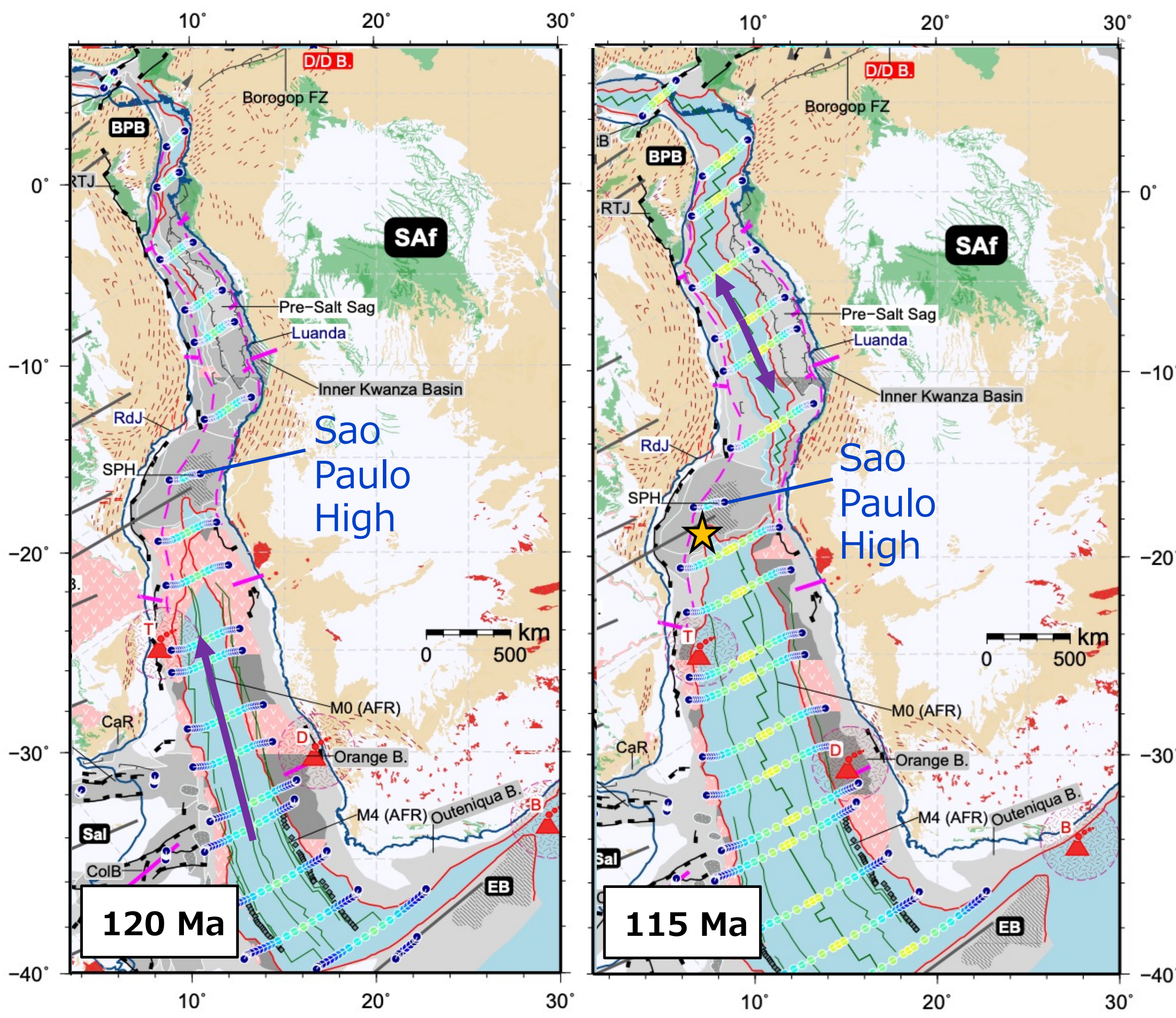

Salton Trough
continent-ocean crustal boundary
Guaymas Basin
EPR
a
b
m
2500
1250
0
-1250
-2500
-3750
-5000
-6250
mgals
20
10
0
-10
-20
-30
34°N
32°N
30°N
28°N
26°N
24°N
22°N
20°N
116°W
114°W
112°W
110°W
108°W

Figure 11



10°
20°
30°
0°
-10°
-20°
-30°
-40°
D/D B.
Borogop FZ
BPB
RTJ
SAf
Pre-Salt Sag
Luanda
Inner Kwanza Basin
RdJ
Sao
Paulo
High
SPH
km
0
500
T
M0 (AFR)
CaR
D
Orange B.
M4 (AFR)
Outeniqua B.
B
Sal
ColB
EB
120 Ma
115 Ma


Figure 12

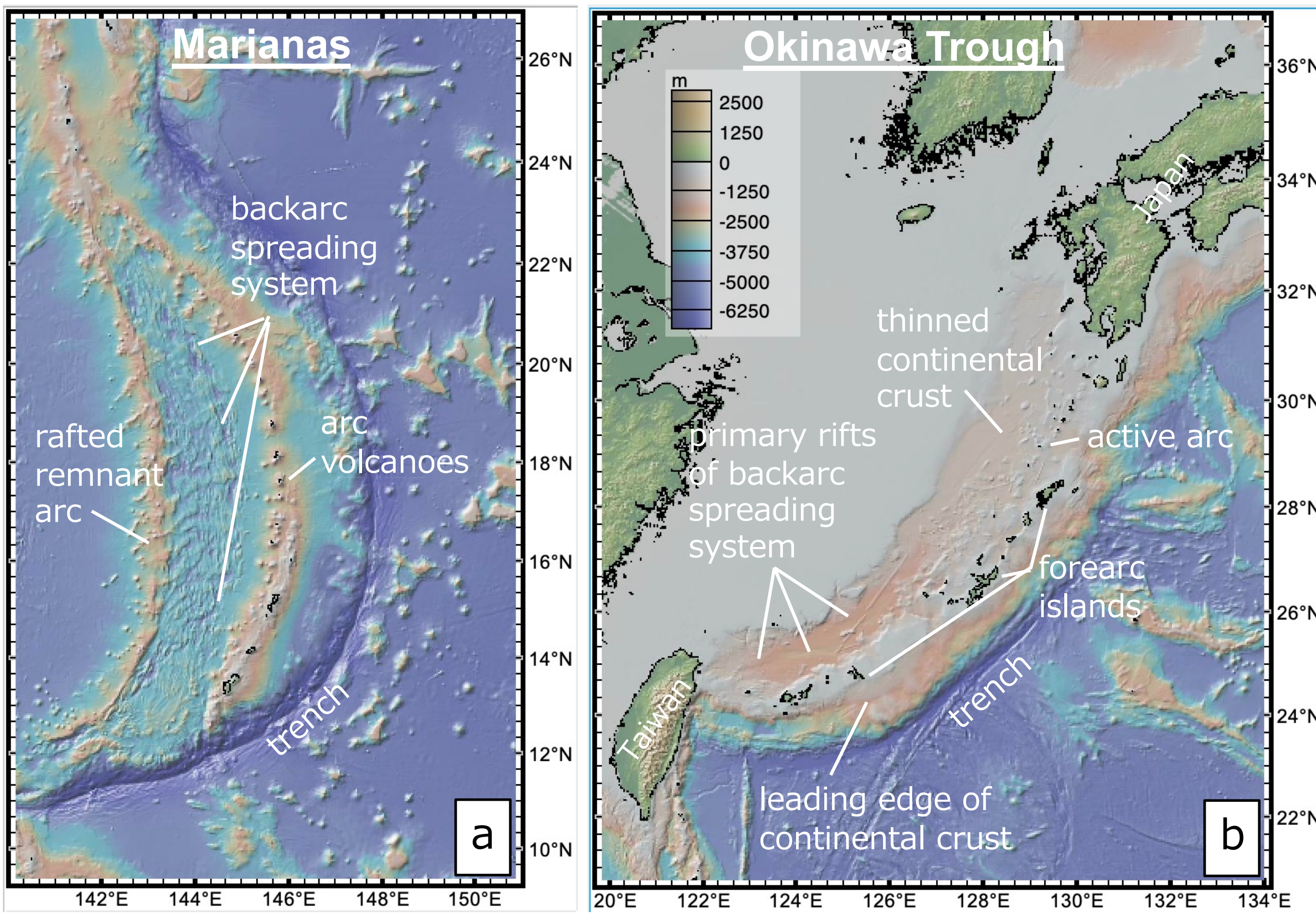
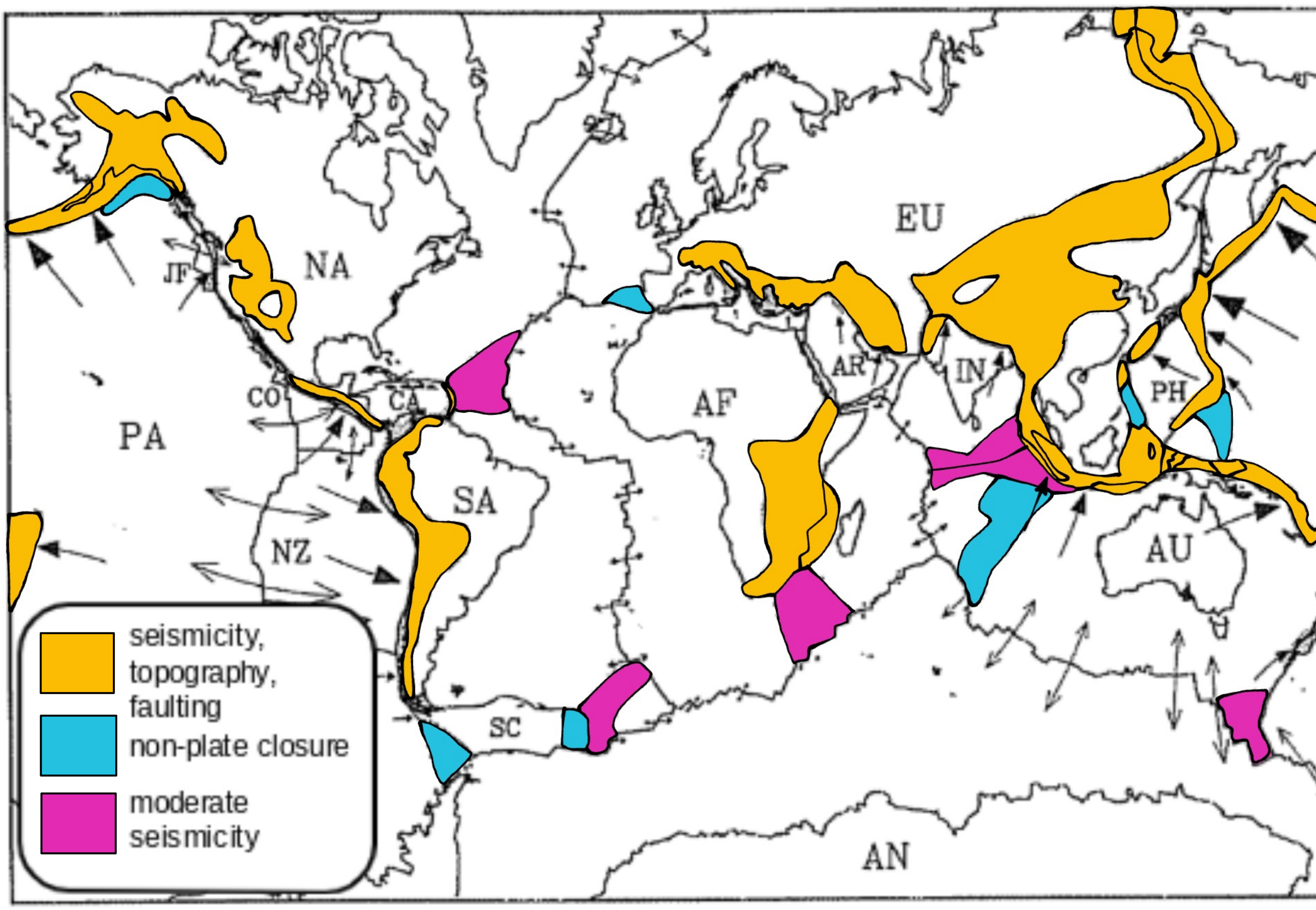

Marianas
backarc spreading system
rafted remnant arc
arc volcanoes
trench
a
26°N
24°N
22°N
20°N
18°N
16°N
14°N
12°N
10°N
142°E
144°E
146°E
148°E
150°E
Okinawa Trough
m
2500
1250
0
-1250
-2500
-3750
-5000
-6250
Japan
thinned continental crust
active arc
primary rifts of backarc spreading system
forearc islands
trench
Taiwan
leading edge of continental crust
b
36°N
34°N
32°N
30°N
28°N
26°N
24°N
22°N
20°E
122°E
124°E
126°E
128°E
130°E
132°E
134°E

Figure 13


EU
NA
JF
CO
CA
PA
AR
IN
PH
AF
SA
NZ
AU
SC
AN
seismicity,
topography,
faulting
non-plate closure
moderate
seismicity