

RIFTS AND RIFTED MARGINS: A REVIEW OF GEODYNAMIC PROCESSES AND NATURAL HAZARDS

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ABSTRACT

This review provides an introduction to the geodynamic processes that influence tectonic rift evolution and rifted margin architecture. With a strong focus on numerical modeling, I summarize classical and recent insights on rift evolution with differentiation between 2D and 3D concepts and models. One of the key processes during rift evolution is crust-mantle coupling, which controls not only the width of a rift system but also crustal hyper-extension and the degree of final margin asymmetry. Accounting for 3D rift geometries allows investigating along-strike heterogeneities, rift segmentation and rift obliquity. Large amounts of sediments have accumulated at rifted margins, especially at the mouths of large rivers and former glaciers, providing important stratigraphic archives and georesources. Shifting the focus from the geological scale of continental extension to the human time scale, natural hazards are discussed regarding earthquakes and volcanic eruptions during active rifting. Finally, I review natural hazards due to passive margin seismicity as well as slope instabilities at heavily sedimented continental margins that have the potential to generate large landslide tsunamis.

1. INTRODUCTION

Continental rifts result from geodynamic processes that continuously shape the surface of our planet. Present-day Earth provides snapshots of many stages of continental rift evolution (Fig. 1): some extensional zones, such as the Okavango rift arm of the East African rift [Kinabo *et al.*, 2008], embody the early stages of rifting with along-axis linkage of fault segments and the formation of immature border faults. Others, like the Kenya rift [Ebinger *et al.*, 1999], represent an intermediate rift phase featuring large border faults, pronounced axial valleys, and magmatic activity. Rifting in the Afar triple junction, for example, is in the late stage of continental extension [Bastow and Keir, 2011] and is perhaps on the precipice of forming new passive margins. The world's rifted margins like those of the Atlantic and Indian Ocean, are covered with thick sedimentary sequences deposited during the syn-rift and the post-rift phases. However, many rifts do not progress to continental break-up even after long periods of extension: aborted rifts can be found in the interior of continents such as the West African and Central African rift system.

At present, continental rifts comprise a small portion of plate boundaries, and their current extent is only a fraction of the total length of rifts generated from the break-up of Pangea. During the dispersal of this supercontinent, more than 100,000 km of passive margins were formed, rendering them the most common tectonic features on our planet, two times longer than spreading ridges or convergent plate boundaries [Bradley, 2008].

Over Earth's history, several supercontinents were assembled and subsequently dissected by continental rifts. The latest fragmentation of a supercontinent, Pangea, commenced ~250 million years ago and molded most of the present-day passive margins. Pangea break-up has been reconstructed by way of combining regional geological and geophysical observations with global sea-floor spreading histories [e.g. Seton *et al.*, 2012]. The fragmentation began with the Central Atlantic rifting that separated Laurasia from Gondwana in the Triassic (Fig. 2). At around 175 Ma, Gondwana split into an eastern and western part along the East African coast. From 140 to 110 Ma, the largest global rift episode of the Phanerozoic occurred with simultaneous extension between (i) Africa and South America, (ii) Australia and Antarctica, (iii) India and Antarctica and (iv) North America and Eurasia in the North Atlantic. Separation of North America and Greenland from Eurasia involved a protracted rift history, where final rupture took place merely 30 million years ago.

Rifts can be classified according to their tectonic environment as being "Atlantic-type", back-arc rifts, syn-orogenic rifts, and post-orogenic rifts. "Atlantic-type" rifts are those that initiate in continental interiors and often lead to separation of major landmasses by generating a new ocean basin. Back-arc basins form in response to subduction dynamics and are controlled by properties of the down-going slab, mantle flow, and mantle wedge dynamics [Sdrolias and Müller, 2006; Schellart and Moresi, 2013]. Back-arc rifting can lead to crustal break-up and the opening of small oceanic basins like the South China Sea, or the Sea of Japan. However, back-arc basins are not as long-lived as Atlantic-type ocean basins and can be closed if the subduction configuration changes. Syn-orogenic rifts

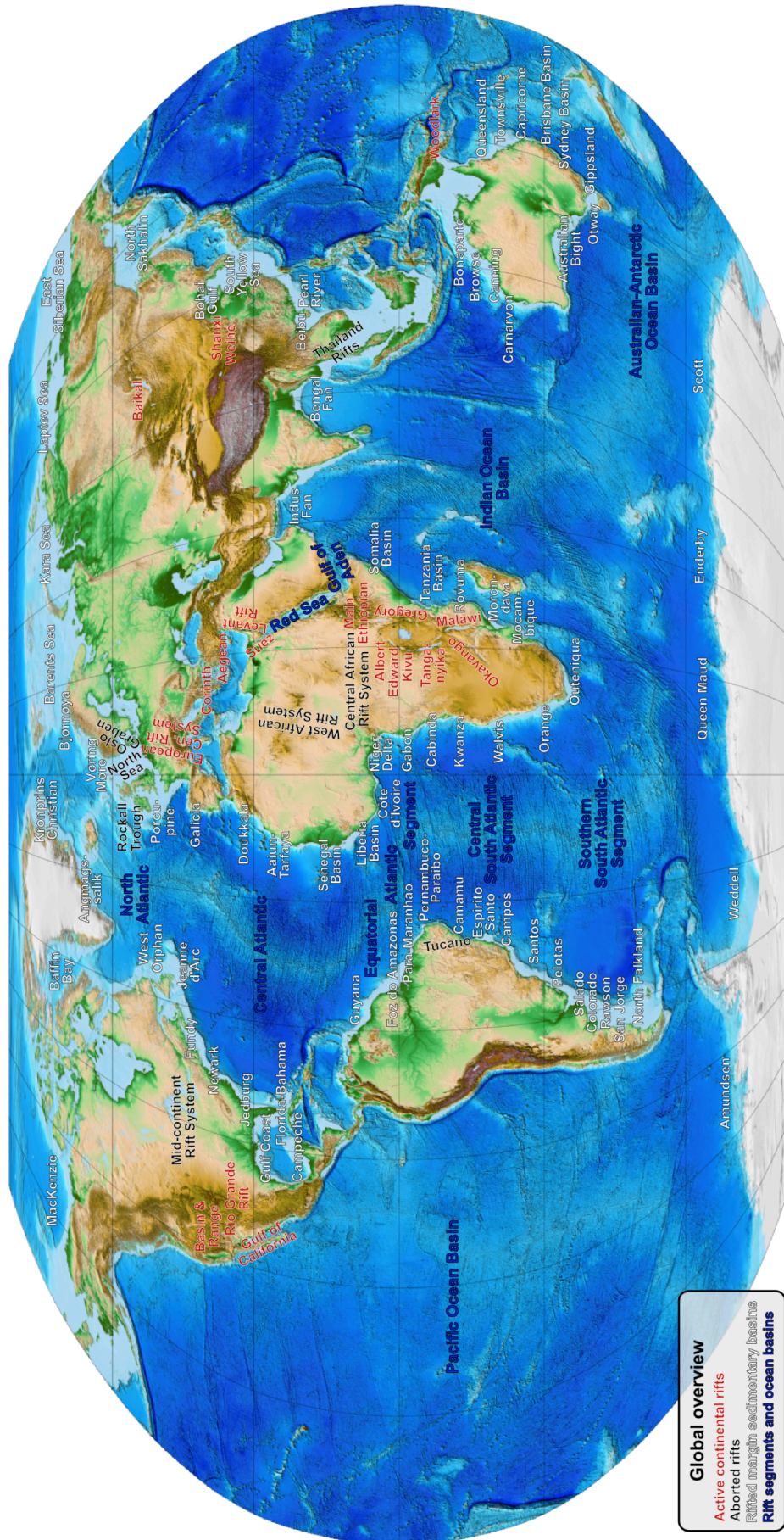


Fig. 1: Global overview map. Map showing a selection of major active rifts, aborted rifts, and rifted margin basins that are important from a scientific and exploration point of view. Basin names are from Meyer et al. [2007], and the background map is based on Etopo1 [Amanite and Eakins, 2009].

develop as a result of stress field changes due to mountain building, which can be caused by collisionally-reactivated inherited weak zones. Two archetypical syn-orogenic rifts are the Upper Rhine Graben portion of the European Cenozoic rift system (ECRIS) and the Baikal rift (Fig. 1). Post-orogenic rifts develop during collapse of young mountain belts that contain overthickened and hot crust, such as the Basin and Range province of North America [e.g. *Malavieille, 1993; Tirel et al., 2008*]. Here, high strain rate extension localized in an array of listric normal faults culminating in the formation of metamorphic core complexes [*Wernicke, 1985*]. The present-day geometry of the Basin and Range province illustrates that continental lithosphere can be severely extended without continental break-up and formation of a new ocean basin.

Below, rift strength and tectonic forces are discussed. I then summarize classical and recent insights on rift evolution gathered from both 2D (Section 2) and 3D (Section 3) concepts and models. Section 4 shifts the focus from the geological scale of plate boundary deformation to the human time scale. Hazards related

to earthquakes and volcanic eruptions in active rift zones are discussed along with submarine slope failures and landslide-generated tsunamis on rifted margins.

1.1 Rift strength and tectonic forces

A variety of studies indicate that the available plate tectonic forces seem to be too small to rupture normal continental lithosphere [e.g. *Buck, 2004*]. However, even plates that are not attached to the strong pull exerted by subducting slabs exhibit continental extension, such as the East African rift system. A combination of inherited weakness and dynamic weakening mechanisms have been proposed as controlling factors of rift initiation: weak suture zones originate from the amalgamation of tectonic plates and are preferentially reactivated in accordance with the Wilson Cycle theory [*Wilson, 1966*]. The impingement of mantle plumes on active rift zones also reduces lithospheric strength by heating and thermal erosion. This process can trigger the final continental break-up [*Buiter and Torsvik, 2014*]. However, it cannot be the cause for continental rifting since extension commences prior to plume arrival in

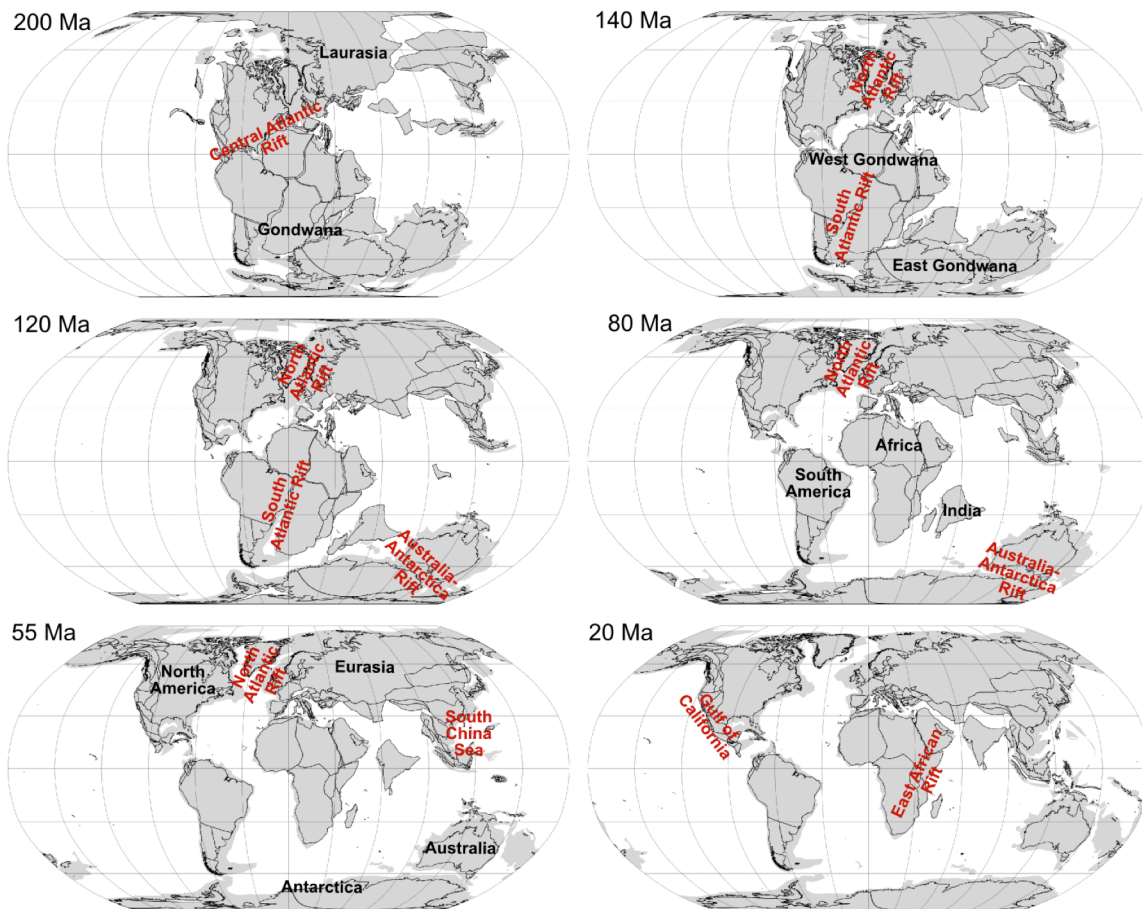


Fig. 2: Past rift episodes during Pangea dispersal.

Gray filled polygons depict the extent of continents including stretched crust. Black contours show coastlines and intra-continental boundaries of terranes and smaller plates. Names of major continental landmasses are denoted in black. Rift systems during time of activity are shown in red. All polygons and rotations are from [*Seton et al., 2012*] reconstructed using GPlates (www.gplates.org).

most cases. Partial melting of the asthenosphere can lead to the emplacement of dikes in the lithosphere, causing weakening due to efficient heating and mechanical strength reduction if magma intrusion rates are high [Bialas *et al.*, 2010; Daniels *et al.*, 2014]. If intrusion rates are low, the frozen mafic intrusions may strengthen the lithosphere, inhibiting or deflecting the rift.

While the mechanisms mentioned above reduce the strength of the rift, another process may locally enhance the extensional force: stress focusing occurs when break-up does not take place simultaneously in a rift zone, but propagates along strike, as was the case of the northward opening of the South Atlantic [Torsvik *et al.*, 2009; Moulin *et al.*, 2010]. Here, the resistive strength of the rift systems was reduced due to the transition from continental rift to weak mid-ocean ridge in the south, while the extensional force imposed by trench suction, mantle drag, and gravitational potential is assumed to have remained constant. Hence, the remaining rift axis in the northern part of the South Atlantic must have experienced a net increase of the extensional line force.

1.2 Modeling approaches

The number of processes posited to affect rift evolution are wide and varied, echoing the diversity of rifts and rifted margins themselves. Conceptual models struggle to quantitatively connect geological and geophysical observations with rock deformation and thermal evolution of the rift. Hence, analog, analytical, and numerical models are often used to conduct rifting experiments and to investigate controlling parameters and associated processes. Even though these models cannot reproduce the natural complexity of these systems, they are nonetheless very useful in isolating the impact of several separate processes that work collaboratively to mold rifts and rifted margins. Here, I shortly summarize the three main modeling approaches.

In *analog models*, physical experiments are conducted using scalable materials like sand or silicon putty to reproduce brittle or ductile deformation, respectively. In nature, tectonic processes evolve on temporal scales of thousand to million years and spatial scales of tens to hundreds of kilometers. Analog models work to scale down these processes by using materials that exhibit dynamically similar behavior on spatio-temporal scales that are favored for laboratory settings (i.e. temporal scales of minutes to days and spatial dimensions of ~0.01 to 1 m [Hubbert, 1937]). One class of analog rift models investigates continental extension within a single brittle layer reminiscent of the brittle upper crust, and such models have provided essential insight into the processes controlling the evolution of fault patterns [Withjack and Jamison, 1986; Tron and Brun, 1991; McClay and White, 1995; Mart and Dauteuil, 2000; Corti *et al.*, 2003; Sokoutis

et al., 2007; Philippon *et al.*, 2015]. More recently, developments in lithospheric-scale models have allowed for realistic modeling of processes such as crustal thinning, where a model asthenosphere accounts for full isostatic balancing [Corti, 2008; Agostini *et al.*, 2009; Autin *et al.*, 2010, 2013; Cappelletti *et al.*, 2013; Corti *et al.*, 2013a; Nestola *et al.*, 2013, 2015].

Analytical modeling of rifts and sedimentary basins first came about in the late 70s and early 80s and captured the first-order thermal evolution of a rift basin. The power of analytical solutions lies not only in the transparency of the method but also in their ability to be evaluated quasi-instantaneously on today's computers, and thus, they can be easily incorporated in more complex modeling approaches. Obviously, this speed comes at a price: only very simple kinematic problems can be solved analytically. The first significant pure shear model of continental extension was presented by McKenzie [1978]. An alternative model was proposed by [Royden *et al.*, 1980], where extension was assumed to result from dike injection. Both models assumed instantaneous thinning. Soon after, a time-dependent component was added to McKenzie's uniform thinning model [Jarvis and McKenzie, 1980] and depth-dependent thinning within a two-layer lithosphere was accounted for [Hellinger and Sclater, 1983]. More recently, an analytical model for time-dependent rifting was suggested by Karner *et al.* [1997], where lithospheric extension is modeled as a series of discrete instantaneous rifting events, each followed by a cooling phase.

Numerical modeling techniques emerged in parallel with analytical models, and several pioneering studies addressed tectonic deformation in the 80s [e.g. Beaumont *et al.*, 1982; England and McKenzie, 1982; Houseman and England, 1986; Zuber and Parmentier, 1986; Bassi and Bonnin, 1988]. These numerical approaches allowed studying rifting in a self-consistent dynamic framework and opened the way to incorporate model components that were not accessible to analytical experiments such as radiogenic heating, non-linear rheologies, and complex geometries. During the last decade, a variety of numerical codes have been developed and 2D numerical modeling became a standard tool to investigate rift processes [e.g. Nagel and Buck, 2004; Lavier and Manatschal, 2006; Pérez-Gussinyé *et al.*, 2006; Buiter *et al.*, 2008; Gueydan *et al.*, 2008; van Wijk *et al.*, 2008; Jammes *et al.*, 2010; Rosenbaum *et al.*, 2010; Wallner and Schmeling, 2010; Huet *et al.*, 2011; Huismans and Beaumont, 2011; Rey *et al.*, 2011; Armitage *et al.*, 2012; Beaumont and Ings, 2012; Choi and Buck, 2012; Chen *et al.*, 2013; Chenin and Beaumont, 2013; Gueydan and Précigout, 2013; Watremez *et al.*, 2013; Brune *et al.*, 2014; Liao and Gerya, 2014; Clift *et al.*, 2015; Petersen *et al.*, 2015;

Sharples et al., 2015]. The major limiting factor when conducting 3D numerical models is the model resolution. To work around this issue, some investigate only the crustal deformation of a 3D rift system, assuming simplified rheologies as well as no crust-mantle coupling [Katzman et al., 1995; Allken et al., 2011, 2012]. However, such approach must be applied with caution, as mechanical coupling of crust and mantle can only be neglected in wide rift systems and only prior to significant lithospheric necking. Nonetheless, recent advances in numerical model development allow for the use of sufficiently high resolution even for 3D model setups with crust and mantle layers and realistic rheologies [Dunbar and Sawyer, 1996; Van Wijk and Blackman, 2005; van Wijk, 2005; Gac and Geoffroy, 2009; Gerya, 2010; Brune et al., 2012; Le Pourhiet et al., 2012; Gerya, 2013; Heine and Brune, 2014; Burov and Gerya, 2014; Le Pourhiet et al., 2014; Koopmann et al., 2014a; Brune et al., 2013; Brune, 2014; Liao and Gerya, 2014; Koptev et al., 2015; May et al., 2015]. Key insights of 2D and 3D modeling studies are discussed in the following sections.

2. RIFTING IN TWO DIMENSIONS

Rifts and rifted margins worldwide show a large structural variety as a result of the interaction of rock rheology, tectonics, magmatism, deformation rates, inherited lithospheric architecture, basement grain, obliquity, climate and sediment supply. The relative importance of these controlling parameters and processes differs between individual rift systems, which impedes the derivation of universal rules for rift evolution. One important way of investigating this complex topic is to focus on simple 2D settings and isolate processes that are relevant during specific rift phases or for certain end-member cases. Following this approach, the research community was able to distill some of the main mechanisms of rift evolution.

2.1. Crust-mantle coupling

Variations in crustal structure are thought to be of primary significance for rift dynamics as they feedback into the thermal structure, stress field, fault evolution, isostatic adjustment, erosion, and sedimentation. During rift initiation, when decompression melting and surface processes exert

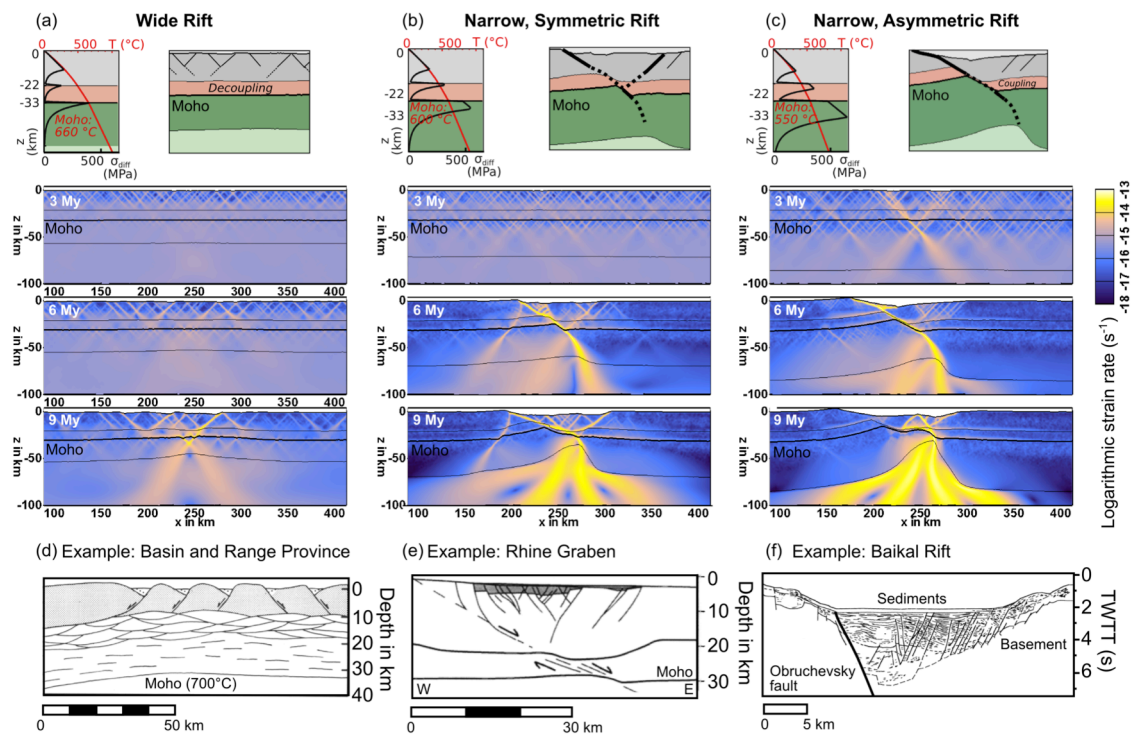


Fig. 3: Symmetry of rifting and crust-mantle coupling.

The three models illustrate the impact of the geothermal profile on rift localization in a laterally homogenous lithosphere. **(a)** Hot and weak lower crust decouples brittle crust and mantle generating a wide rift. **(b)** A colder Moho leads to a stronger lower crust, less decoupling and a narrow, predominantly symmetric rift valley. **(c)** Coupled deformation of crust and mantle in a cold lithosphere leads to narrow, asymmetric rifting with a single master fault. **(d)** A sketch of wide rift tectonics for the Basin and Range province, based on geological and geophysical observations [after Hamilton, 1987]. **(e)** Interpreted structure of the southern Rhine Graben [after Michon and Merle, 2000], based on DEKORP-ECORS seismic data [Brun et al., 1992]. **(f)** Cross-section through the central Baikal rift. Note that the vertical scale is much smaller than in (d) and (e), which is why the Moho is not visible and the boundary fault on the left appears much steeper than in reality [modified after van der Beek, 1997]. Except for the initial geothermal profile, all 3 models (a-c) apply the same parameter set of 8 millimeters per year full extension velocity and strong frictional softening [see Brune et al., 2014 for more details].

only a minor influence, the vertical geometry of a rift system is dominated by tectonic processes. Indeed, it was shown that the style of rifting, the width of a rift system, and its symmetry are controlled by the degree of coupling between upper crust and mantle [Buck, 1991; Huismans and Beaumont, 2003; Gueydan et al., 2008]. Crust-mantle decoupling generates wide rifts like the Basin and Range province or the South China Sea, where lower crustal flow compensates fault activity and suppresses the formation of a necking instability for a long time (Fig. 3a). Such decoupling requires a low-viscosity crustal layer that might result from elevated lower crustal temperatures (like in Fig. 3a), an extraordinary weak crustal rheology, or a thickened crust, such as in the orogenic plateaus. Contrarily, if crust and mantle are coupled, extension produces few large faults that form a narrow rift (Fig. 3b,c). Classic examples include the Baikal rift, the Rhine Graben, the Rio Grande rift, the Gulf of Suez and the East African rift.

In early-stage rifts, both symmetric and asymmetric fault patterns can be observed in narrow rift settings. The process of fault formation can be described as competition between brittle and ductile deformation of individual lithospheric layers [Huismans and Beaumont, 2003; Buiter et al., 2008]. Asymmetric fault formation (Fig. 3c) dominates if the lithosphere involves a strong brittle layer and if a high amount of frictional strain softening is prescribed. A ductile crustal layer, on the other hand, favors more symmetric extension (Fig. 3b).

2.2. Magma-poor and magma rich end-members

Two end-member scenarios of non-volcanic rifts (e.g. Baikal) and volcanic rifts (e.g. Afar) are often used as analogs for the early rift history of magma-poor (e.g. Iberia) and magma-rich (e.g. Southern South Atlantic) rifted margins. There is no clear definition of how much magma constitutes a “magma-poor” margin, as even the Baikal rift, for example, was shown to feature magmatism that is capable of compensating crustal thinning [Thybo and Nielsen, 2009]. Nevertheless, there are several fundamental differences between magma-poor and volcanic rifted margins.

Magma-poor rifted margins feature small melt fractions during crustal thinning, wide transitions between stretched continental and steady-state oceanic crust, and mantle exhumation prior to oceanic spreading [Whitmarsh et al., 2001]. They also feature wide areas of highly thinned, so-called hyper-extended crust with a thickness of less than 10 km, where crust and mantle deformation appears to be tightly coupled [Brune et al., 2014; Mohn et al., 2014]. Well-studied examples of magma-poor margins comprise the Iberia-Newfoundland conjugates [Hopper et al., 2004; Reston, 2007; Ranero and Perez-Gussinye, 2010; Sutra and Manatschal, 2012],

the Central South Atlantic segment [Contrucci et al., 2004; Aslanian et al., 2009; Mohriak and Leroy, 2012], the Australian North West Shelf [Karner and Driscoll, 1999], the eastern Gulf of Aden [Leroy et al., 2012] and the South China Sea [Zhou and Yao, 2009; Franke, 2013].

Volcanic rifted margins, in contrast, involve large amounts of volcanic flows and unusually thick (>10 km) oceanic crust [Mutter et al., 1982; Eldholm et al., 2000]. Volcanic rifted margins like the southern South Atlantic segment and the north-east Atlantic are thought to be commonly associated with mantle plumes and the generation of large igneous provinces [Coffin and Eldholm, 1994; Menzies et al., 2002], though this has recently been disputed [Franke, 2013]. Additional complexities in distinguishing magma-poor and magma-rich end members arise due to deviations from these general patterns; an early rift phase may show magma-poor characteristics, whereas a later rift stage can involve large amounts of volcanism (e.g. the North Atlantic [Lundin and Doré, 2011]).

2.3. Rift Migration

Hyper-extended crust is often distributed between conjugate margins with pronounced asymmetry. Offshore Iberia for example, the width of the highly thinned crustal region amounts to about 70 km [Whitmarsh et al., 2001] while the conjugate margin of Newfoundland exhibits only about 20 km of hyper-extended crust [Hopper et al., 2004]. In the central South Atlantic, the asymmetry is even larger: up to 200 km offshore Angola [Contrucci et al., 2004], and about 30 km at the conjugate Brazilian margin [Unternehm et al., 2010]. Seismic observations as well as kinematic reconstructions of the Iberia/Newfoundland conjugates suggest that an array of ocean-ward younging sequential faults bears responsibility for the typical asymmetric architecture [Ranero and Perez-Gussinye, 2010]. Until recently, there was no thermo-mechanically verified model that explains the existence of highly thinned crust, margin asymmetry, and sequential fault activity simultaneously.

A recent study addresses this problem [Brune et al., 2014], proposing rift migration as a key process during magma-poor margin formation (Fig. 4). The model starts with a dominant single fault and a few antithetic faults, which is typical for narrow, brittle-dominated rifts such as the Baikal or East African rift. During continued extension, elevated temperature and viscous strain softening generate a pocket of weak lower crust at the tip of the major fault. In analogy to the subduction channel at zones of plate convergence, this localized zone of deformation has been dubbed an “exhumation channel” (Fig. 4d). Two processes – i) cooling and strengthening at the footwall of the exhumation channel and ii) softening within the adjacent crust – generate a horizontal strength gradient

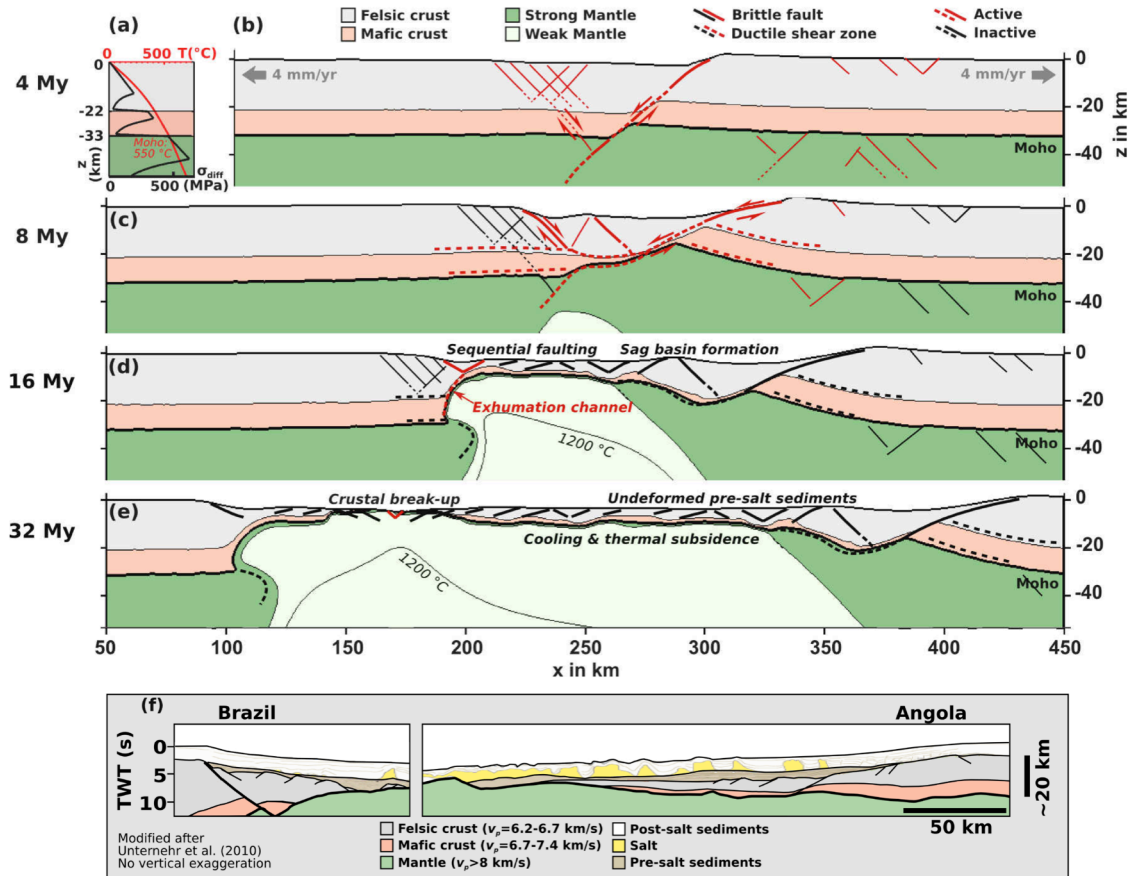


Fig. 4: Numerical model of rift migration.

(a-e) The rift center moves laterally by more than 200 km during 20 Ma creating a wide margin (right) and a narrow margin (left). The wide margin is formed through sequentially active faulting towards the future ocean. Hence, thick undisturbed pre-salt sediments pre-dating breakup are predicted by the model to be deposited in the landward part of the margin. Active faults are shown in red, inactive faults in black. Brittle faults are indicated with solid lines, ductile shear zones with dashed lines. The setup is identical to Fig. 3c. See Brune et al. [2014] for more information. (f) Asymmetry of the South Atlantic continental margins. The image shows an interpreted seismic cross section of the conjugate continental margins of the Campos basin (Brazil) and the Kwanza basin (Angola). Basin locations see Fig. 1. Original seismic interpretations from Unternehr et al. [2010] and Contrucci et al. [2004].

that leads to the migration of the rift in a steady-state manner. Once migration starts, new faults emerge that are sequentially active. Weak crust flows towards to the down-dip end of these faults, which counteracts the faults' tendency to uplift the Moho. This mechanism generates a wide margin on one rift side, where crust is thinned to less than 10 km. The degree of rift migration is controlled by the lower crustal viscosity adjacent to the moving rift, whereas the viscosity of the lower crust is a function of lower crustal composition, initial thermal structure, intensity of strain softening, and most importantly, extension velocity. The effect of lower crustal viscosity was likely most relevant in shaping the Iberia-Newfoundland margins [Brune et al., 2014]. Their structure is best described by an initially decoupled, somewhat asymmetric fault phase followed by crust-mantle coupling, sequential faulting, and moderate degree of rift migration [Sutra et al., 2013].

Steady-state rift migration may be also relevant for other highly thinned and asymmetric margin pairs, such as the southern South Atlantic segment and the East Australia/Lord Howe Rise conjugate. The evolution of the north-east Atlantic margins has been suggested to involve a more discrete form of rift migration [van Wijk and Cloetingh, 2002]. Here, the hyper-extended crust is confined to Early Cretaceous rift basins that have since become inactive [Lundin and Doré, 2011; Rüpke et al., 2013]. Instead, subsequent deformation localized in a region further west, which created several elongated ribbon-like continental fragments [Péron-Pinvidic and Manatschal, 2010]. A very recent study highlights the role of brittle strength during discrete rift episodes that are separated by a phase of tectonic quiescence [Naliboff and Buiter, 2015]. This study illustrates that a rift jump away from the former rift axis is controlled by the relative integrated brittle strength between the initial rift and surrounding regions as opposed to the total integrated strength.

3. RIFTING IN THREE DIMENSIONS

The 2D assumption is a convenient approach as it reduces the complexity of the system while simultaneously enabling higher resolution in the 2D domain. In reality, continental rifting involves several important factors that generate along-strike variability: inherited structures, segmentation, plume-lithosphere interaction, and oblique extension. All of these processes have the potential to overprint 2D properties of the rift system.

3.1. Segmentation

Tectonic and magmatic segmentation occurs in active rifts [e.g. *Keranen et al.*, 2004; *Keir et al.*, 2015] and at passive margins [*Fournier et al.*, 2004; *Franke et al.*, 2007]. Segmentation is thought to occur during the initial rift phases and is characterized by segmented en-échelon border faults. These faults become inactive during subsequent basin-ward localization, as new segments emerge. If this happens in volcanic rifts, where extension is accommodated by dike intrusion, a new type of magmatic segmentation emerges [*Ebinger and Casey*, 2001]. Strong along-strike variations of crustal asymmetry and rift geometry are also found in the highly oblique Gulf of California. These variations have been linked to changes in sedimentation, magmatism, and mantle properties [*Lizarralde et al.*, 2007; *Bialas and Buck*, 2009; *Wang et al.*, 2009]. The South Atlantic rifted margins are segmented by extrusive volcanism, and the segment boundary between volcanic and magma-poor segments often coincides with a fracture zone and can be surprisingly sharp [10s of km, see *Shillington et al.*, 2009; *Koopmann et al.*, 2014b]. Such an observation cannot be explained by gradual, along-strike changes in mantle properties.

3.2 Oblique Rifting

Oblique extension takes place when the relative extension direction of two diverging plates is at an angle to the zone of deformation. Rift obliquity is thought to be one of the major causes of rift segmentation and governs the geometry of currently active rifts [*Díaz-Azpiroz et al.*, 2016], such as the Main Ethiopian rift [*Corti*, 2008], the Levant rift system including the Dead Sea rift [*Mart et al.*, 2005] the Gulf of California rift [*Atwater and Stock*, 1998; *Fletcher et al.*, 2007] and the Upper Rhine Graben [*Bertrand et al.*, 2005]. Many past oblique rifts, such as the Gulf of Aden [*Leroy et al.*, 2012], the Equatorial Atlantic rift [*Heine et al.*, 2013], Africa/Antarctica rift [*Eagles and König*, 2008] and the Antarctica/Australia rift [*Whittaker et al.*, 2013] succeeded in forming an ocean basin. Moreover, patterns of oblique extension have been studied at mid-ocean ridges such as the Reykjanes and Mohns Ridge [*Dauteuil and Brun*, 1993] and the South West Indian Ridge [*Dick et al.*, 2003; *Montési et al.*, 2011].

A major obstacle in assessing the impact of oblique extension on continental rifting stems from the superposition of processes, such as the reactivation of pre-rift structures, sedimentation, and dike dynamics. Recent analog and numerical modeling of oblique extension have proven very useful in decomposing the role of rift obliquity [*Agostini et al.*, 2009; *Autin et al.*, 2010; *Brune and Autin*, 2013]. The structures that arise from a laterally homogenous model under oblique extension have been summarized by Brune [2014]. In these 3D rift experiments, a constant extension velocity is applied to a continental plate until the lithosphere was broken. These 3D numerical models reproduce all possible rift obliquities (i.e. rift-orthogonal extension, low obliquity, high obliquity and strike-slip deformation). Surface stress information is extracted from the model and interpreted in terms of stress regime (extensional, strike-slip, compressional) and optimal small-scale fault orientation [*Brune and Autin*, 2013]. Along-strike model borders use a periodic boundary condition that mimics an infinitely long rift zone. The evolution of a highly oblique rift system that is extended at a constant velocity of 10 millimeters per year (full rate) is illustrated in Fig. 5. Initial small-scale faults strike with an azimuth that precisely bisects the rift-parallel and the extension orthogonal direction. This agrees with established analogue modeling results and theoretical considerations [*Withjack and Jamison*, 1986]. The small-scale faults coalesce into an en-échelon system at 6 My. The phase of en-échelon deformation is accompanied with smoothly varying fault orientations that exceed the rift-parallel azimuth of 0° (Fig. 5e,f). Following the phase of en-échelon deformation at 17 My, pronounced rift-parallel faults emerge adjacent to the rift center, and strong lithospheric necking occurs below the central rift domain. Simultaneously, a thin, strike-slip dominated area emerges in the rift center, indicating strain-partitioning (Fig. 5d). Critically, fault directions are often used to infer paleo-plate movements, however, several independent studies show that local changes in crustal stress field and fault orientation may arise intrinsically during rift maturation, and thus may not require plate motion changes [*Morley*, 2010; *Corti et al.*, 2013b; *Brune*, 2014; *Philippon et al.*, 2015]. Instead, oblique rifts follow a characteristic sequential fault pattern that depends on the maturity of the rift system [*Agostini et al.*, 2009; *Brune*, 2014].

3.3 Rift strength

The mechanical strength of a rift system is controlled by rheological properties of the extending lithosphere [*Burov*, 2007], the presence of weak zones inherited from past sutures [*Buiter and Torsvik*, 2014], melt generation and diking [*Buck*, 2007], and localization feedbacks [*Regenauer-Lieb et al.*, 2006]. Rift strength

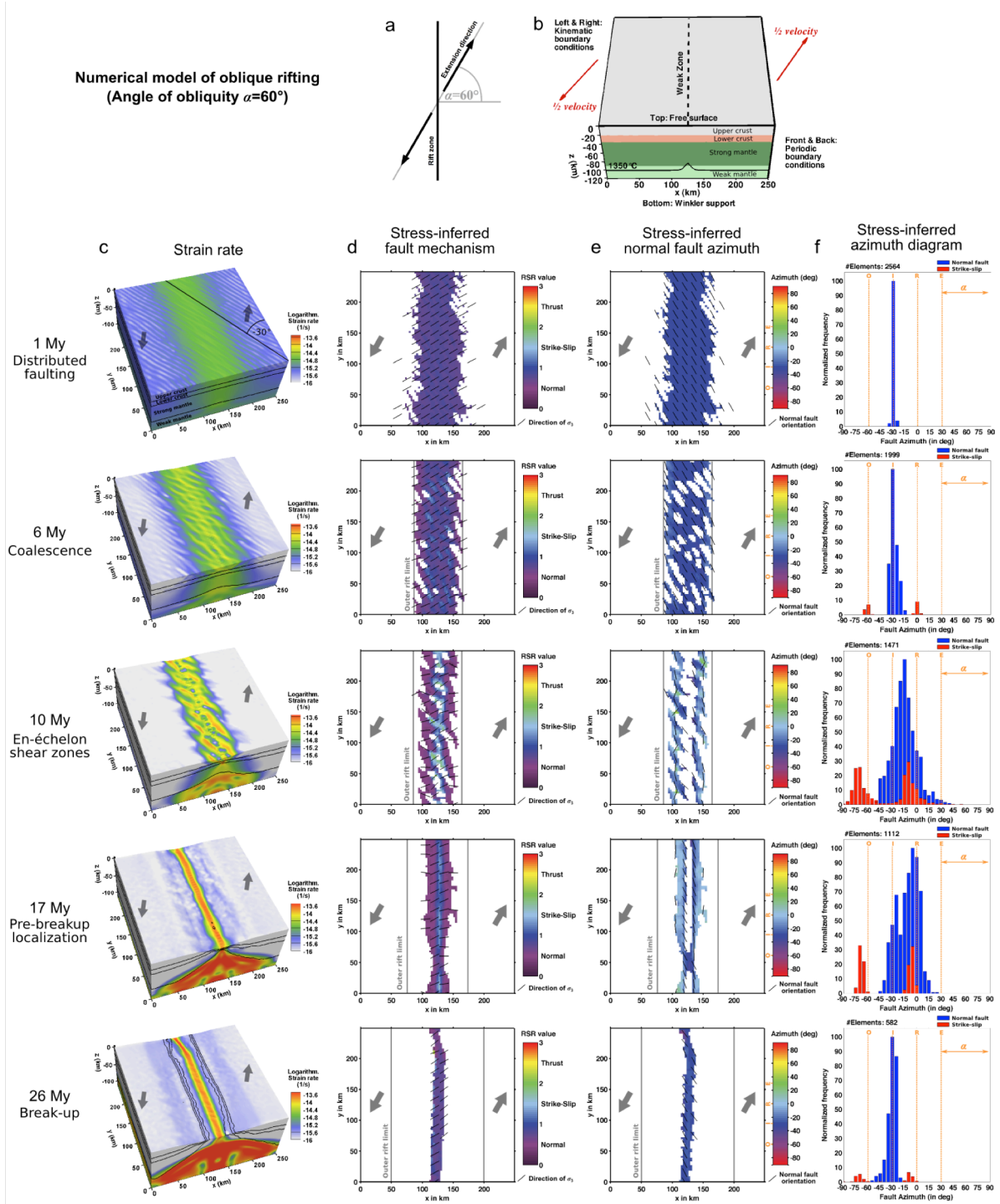


Fig. 5: Stress and fault evolution in oblique rifts.

(a) Geometry of oblique extension. The angle of obliquity $\alpha=60^\circ$ is defined as the angular difference between extension velocity and rift normal. (b) The numerical model setup involves four layers: felsic crust (grey), mafic crust (red), strong mantle (olive), and weak mantle (light green). Extensional velocities are prescribed at the boundaries in x -direction. Boundaries in y -direction are connected via periodic boundary conditions. (c) The surface strain rate at 1 My shows small-scale shear zones that correspond to the theoretically derived value of 30° [Withjack and Jamison, 1986]. Further evolution involves fault coalescence, rotating sigmoidal shear zones and basin-ward localization until crustal break-up is achieved. (d) Visualization of surface stress in terms of regime stress ratio (RSR). An initially transtensional stress regime gives way to a strike-slip zone in the rift center (light blue colors) and a normal fault domain adjacent to the rift center (purple). (e,f) Stress-inferred normal fault azimuths rotate from intermediate to rift-parallel while lithospheric necking takes place as seen in (c). Rift-parallel faulting ends abruptly during basinward localization, followed by intermediate fault orientations. Abbreviated directions indicate the rift-parallel azimuth (R), intermediate fault orientation (I), extension-orthogonal direction (O), and the direction of extension (E). See Brune [2014] for more information and for models of the entire spectrum of extensional obliquity.

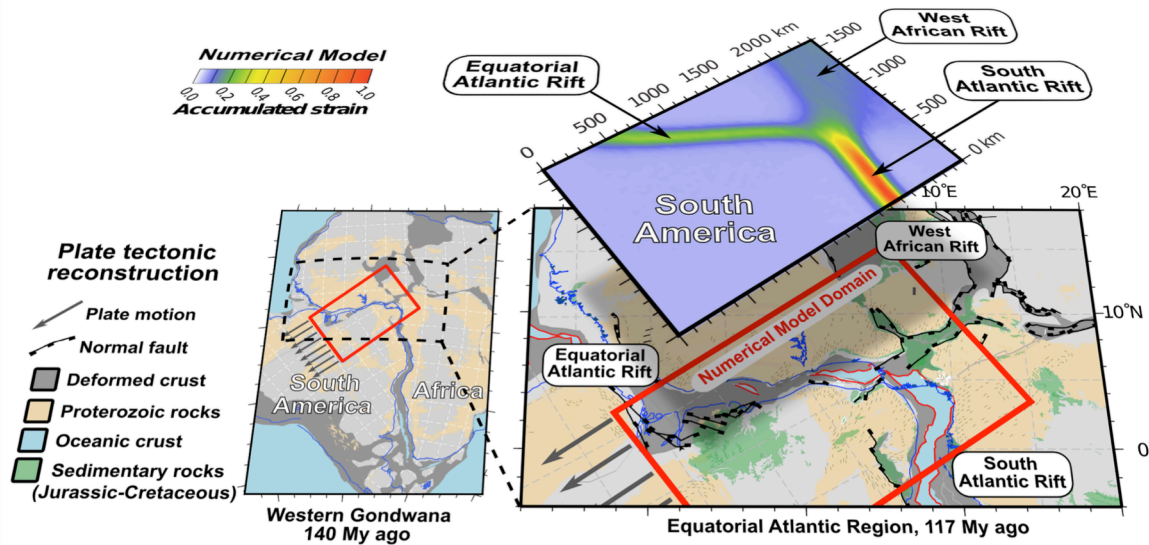


Fig. 6: Rift competition during Equatorial Atlantic rifting.

The plate kinematic reconstruction (bottom) is based on detailed geological, sedimentological and geophysical data sets. The 3D numerical model (top) reproduces the rift zone evolution at Earth's surface. The south-west directed movements of South America activates the Equatorial Atlantic rift with highly oblique extension. Oblique rift systems are energetically favored over orthogonal ones [Brune *et al.*, 2012], which is why the West African rift system with less obliquity offers distinctly more mechanical resistance. After 20 Ma of rift competition between both systems, the West African rift failed while the Equatorial Atlantic ocean basin opened [Heine and Brune, 2014].

however also depends on the obliquity of the rift system; the force required to maintain a given rift velocity can be computed from simple analytical and more realistic numerical models alike, and both modeling approaches demonstrate that less force is required to perpetuate oblique extension [Brune *et al.*, 2012]. The reason is that plastic yielding requires a smaller plate boundary force when extension is oblique to the rift trend. Comparing strike-slip and pure extension end-member scenarios, it was shown that about 50% less force is required to deform the lithosphere under strike-slip. This result implies that plate motions involving significant rift obliquity are mechanically preferred.

This behavior is exemplified by the Early Cretaceous separation of South America from Africa (Fig. 6), where the Equatorial segment of the South Atlantic rift competed with the West-African rift zone [Heine and Brune, 2014]. Upon combining plate kinematic modeling and forward 3D numerical experiments, this study demonstrated that after 20 to 30 My of coexistence, strain localization along the Equatorial Atlantic rift continued as the rift progressed into seafloor-spreading mode. Meanwhile, the West-African rift failed soon after South America broke away from Africa. It appears that the success of the Equatorial Atlantic rift was bolstered by its higher obliquity. Conversely, the less oblique extensional domains within the African plate became inactive. Moreover, an originally orthogonal extensional rift system can weaken over time if the extension direction changes. The Australian plate, for instance,

started to rift away from Antarctica in a nearly rift-orthogonal direction at an initially slow rate. At about 100 Ma, the rift direction became skewed, hence the system became weaker and the rift velocity increased [Ball *et al.*, 2013; Whittaker *et al.*, 2013]. The same process has been invoked for opening of the Gulf of California [Bennett and Oskin, 2014].

Rift strength is clearly a time-dependent parameter, as the lithosphere is substantially thinned (and thus weakened) during advanced stages of continental extension. This has strong implications for the kinematic evolution of mature rifts [Brune *et al.*, 2016]. Assuming that large-scale plate driving forces (e.g. slab pull, trench suction, basal drag, ridge push) vary slowly over time, numerical models with a constant force-boundary condition can investigate extension velocity as an independent parameter [Christensen, 1992; Hopper and Buck, 1993; Brune *et al.*, 2012, 2013]. The constant force boundary condition is applicable to major rifts where the integrated strength of the entire rift system is comparable to the plate driving forces, while this boundary condition cannot be implemented for minor rift basins that are too small to provide dynamic feedback on large-scale plate motions. Fig. 7 depicts the velocity evolution for the aforementioned numerical Equatorial Atlantic rift model [Heine and Brune, 2014]. Upon inception, when rift strength is large, the extension rate is slow. With continued extension, lithospheric necking and strain softening processes cause the rift system's strength to deteriorate. This generates a dynamic rift-weakening

feedback; weakening accelerates rifting, which in turn results in further weakening. This feedback mechanism induces a rapid increase in rift velocity, a phenomenon which is concordant with plate kinematic reconstructions [Torsvik *et al.*, 2009; Heine *et al.*, 2013; Brune *et al.*, 2016].

4. SEDIMENTATION

Surface processes redistribute large volumes of material into the rift basin. Deposited sediment sequences can provide comprehensive records of past kinematic evolution and climatic changes. However, they can also directly exert control on the tectonic evolution of a rift basin [e.g. Burov and Poliakov, 2003], where rift-driven topographic changes can have substantial reciprocal influence on regional climatic regimes, namely rainfall intensity and river drainage area.

4.1 Syn-rift sedimentation

Sediment accumulation strongly affects the thermal rift structure in a rift basin by generating a low-conductive layer at the rift's surface. This thermal blanketing effect increases temperatures in the basement and facilitates partial melting [Lizarralde *et al.*, 2007]. At the same time, a thick sedimentary layer inhibits hydrothermal circulation, leading to even higher thermal blanketing. In the northern Gulf of California, large amounts of sediments are deposited by the Colorado River, which are rapidly buried and heated to ultimately form a metamorphosed crustal

layer [Dorsey, 2010]. The suggestion that high sedimentation rates induce enhanced melting, which promotes localization and narrow rift geometry [Lizarralde *et al.*, 2007], however, has been amended through an alternative mechanism [Bialas and Buck, 2009]. These authors propose that the weight of such sediments counter-balances thinning-related changes in crustal buoyancy forces, thus facilitating narrow rifting.

4.2. Post-rift sedimentation

The total amount of sediment deposited in continental margins is estimated at 16 and 20 Gigatons per year [Milliman and Syvitski, 1992; Ludwig and Probst, 1998]. While convergent margins with rugged topography feature a large number of small streams, passive margins host the Earth's largest rivers. Sedimentary cover at passive margins is especially thick at high-discharge river mouths (Fig. 8). At high latitudes, sediment accumulates due to elevated erosion rates that accompany Arctic and Antarctic glaciation cycles.

Moreover, the surface of the sedimentary wedge is constantly reshaped by contour currents, submarine mass movements (e.g. landslides or turbidity currents), and the continued deposition of new sediments from continental run-off. During and after deposition, the sedimentary package undergoes compaction, causing decreases in porosity and thus providing a favorable environment for elevated pore pressures and focused fluid migration [Berndt, 2005;

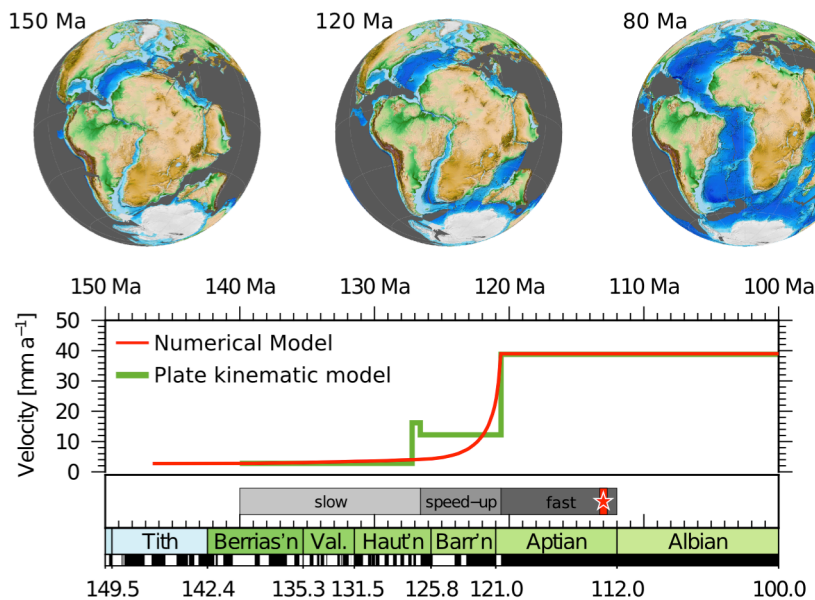


Fig. 7: Evolution of rift velocity during South Atlantic opening.

Rifting between South America and Africa initiated during the early Cretaceous about 145 Ma ago. During the following 30 Ma, the South Atlantic opened like a zipper from south to north (top). During this process, the relative velocity between South America and West Africa changed dramatically (bottom). The numerical and the plate-kinematic models exhibit a slow initial rift phase prior to 127 Ma ago, followed by a ten-fold increase within 6 Ma [Heine and Brune, 2014]. During this period, the last continental bridge between Africa and South America weakened successively due to rift propagation. Continental break-up along the passive margin segments occurred in this region about 113 Ma ago (designated by the red star).

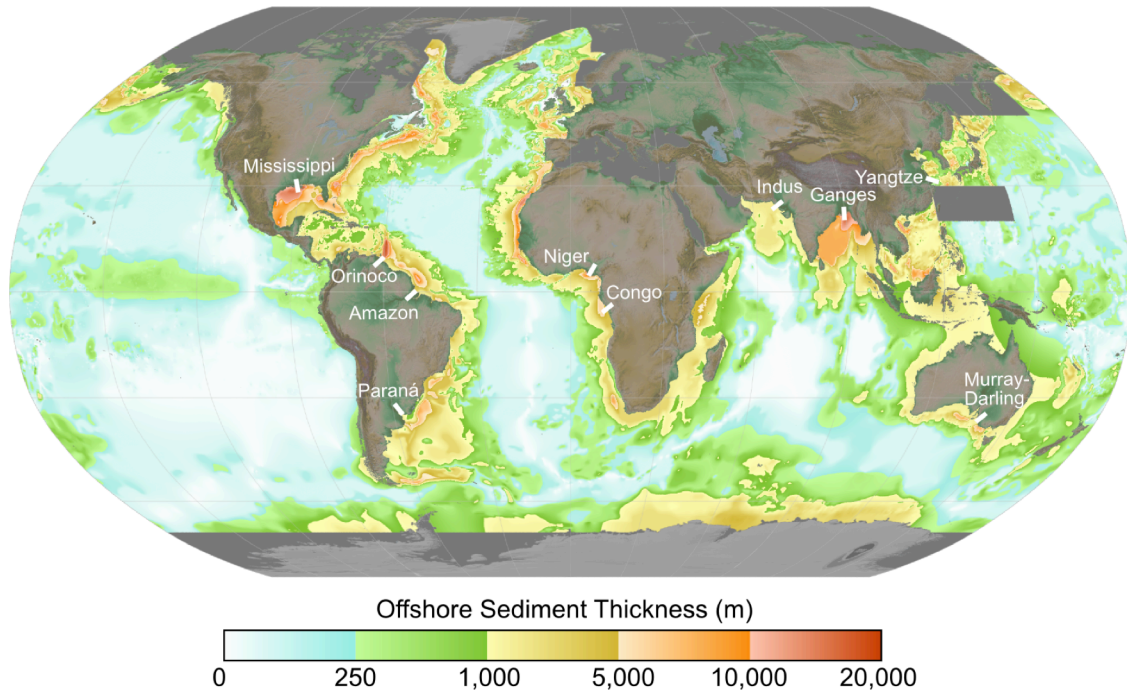


Fig. 8: Global sediment thickness map.

Deep sedimentary basins straddle the rifted continental margins, especially at high-discharge river mouths (shown in white). Sediment thickness depicts data from [Divins, 2003; Whittaker et al., 2013]. Shaded continental topography is based on Etopo1 [Amante and Eakins, 2009]. The image was generated using GeoMapApp (www.geomapapp.org) and GPlates (www.gplates.org). Please see Fig. 1 for names of individual basins.

[Gradmann and Beaumont, 2012; Gradmann et al., 2012]. The thermal evolution and internal dynamics of these sediments control the chemical maturation and migration of organic matter [Allen and Allen, 2013]. Additional complexities in sedimentary dynamics arise from the occurrence of salt, which can form diapirs, walls, or provide weak subhorizontal detachment layers [see Hudec and Jackson, 2007 for a review]. Earth's largest passive margin salt basins are located in the Central South Atlantic [Karner and Gambôa, 2007], the Gulf of Mexico [Fort and Brun, 2012] and the Red Sea [Bosworth et al., 2005]. The sediment load interacts with the lithospheric basement by generating basin-wide isostatic subsidence, which creates further accommodation space and leads to elastic bending of the plate [Driscoll and Karner, 1994; Watts et al., 2009]. In rifted margins with very weak crust, such as the Gulf of Thailand and South China Sea, post-rift sedimentation can even generate tectonic deformation by inducing horizontal ductile crustal flow [Morley and Westaway, 2006; Clift et al., 2015].

5. NATURAL HAZARDS

Strong earthquakes and volcanic eruptions are a hallmark of subduction zones. Nevertheless, a considerable natural hazard potential exists at rifts and rifted margins, where events tend to occur at a lesser severity and frequency. While passive margins earthquakes are rarely tsunamigenic, the large volumes of sediments, especially within formerly

glaciated regions, can result in large submarine mass movements, which can trigger tsunamis.

5.1. Syn-rift earthquakes and volcanism

Rifting is typically accompanied by high-frequency low-magnitude seismicity [Ibs-von Seht et al., 2008; Ebinger et al., 2010], especially in magmatic rifts where dike intrusions are thought to prevent the buildup of elastic strains inhibiting large earthquakes. However, several earthquakes with magnitudes larger than 6 have been documented in the East African rift during the last century: the $M_S=7.3$ Kasanga earthquake in Tanzania 1910, the Subukia earthquake of 1928 ($M_S=6.9$), the 1966 Tooro earthquake in Uganda 1966 ($M_S=6.1$), the 1960 Ethiopian Awasa earthquake ($M_S=6.1$) and the 1989 Dobi graben event ($M_S=6.5$) in Ethiopia [Midzi et al., 1999; Zielke and Strecker, 2009] pointing out a missing element in our knowledge about rift seismicity. Other rift zones besides the East African rift have also experienced large earthquakes such as the Rhine Graben, where a $M_W\sim 6.5$ event destroyed the city of Basel [Meghraoui et al., 2001]. This earthquake constitutes the largest historical seismic event in central Europe. The Shanxi and Weihe rifts, which lie south-west of Beijing, have been active since the Pliocene and are characterized by exceptionally strong earthquake activity [Xu and Ma, 1992]. Here, the 1303 Hongdong earthquake ($M=8.0$) caused over 470,000 casualties, while the 1556 Huaxian earthquake measured at $M=8.3$ and killed about 830,000 people, rendering it the deadliest

earthquake in human history [Liu *et al.*, 2007]. The mechanisms that govern large rift earthquakes like these are largely still unknown, but they seem to occur within slowly deforming continental interiors, they feature complex fault systems, large recurrence times, and migrating seismic activity [Liu *et al.*, 2011].

Rift volcanoes are sites where mantle fluids and volatiles are released to the surface. Many active volcanoes in the East African rift system have caused catastrophic degassing events, explosive eruptions, and lava flows. The deep waters of Lake Kivu contain high concentrations of dissolved carbon dioxide and methane, posing a significant hazard should the gas undergo catastrophic release. Similar conditions arose in lakes Monoun and Nyos (Cameroon), where catastrophic gas release events in 1984 and 1986 caused more than 1700 casualties [Kling *et al.*, 1987; Schmid *et al.*, 2002]. In 2007, the Oldoinyo Lengai volcano in the northern Tanzania Divergence generated an earthquake swarm and set off a major episode of explosive ash eruptions, leading to the formation of an ash plume that measured several kilometers high [Baer *et al.*, 2008]. In 2002, the city of Goma (Democratic Republic of Congo) in the western branch of the East African rift system was partially destroyed by a volcanic eruption along the southern flank of Mount Nyiragongo [Tedesco *et al.*, 2010], which generated two massive lava flows [Chirico *et al.*, 2008; Favalli *et al.*, 2008]. Two of the largest well-documented basaltic eruptions in Africa occurred in Afar: the Alu-Dalafilla eruption in 2008 caused no known deaths or damage since it erupted onto an unpopulated salt plain [Pagli *et al.*, 2012]. However, the 2011 eruption of the Nabro volcano, an off-rift volcano adjacent to the Afar rift [Hamlyn *et al.*, 2014] killed several people and injected more than 1.5 megatons of sulfur dioxide into the stratosphere, making this eruption the largest since the 1991 Pinatubo eruption [Bourassa *et al.*, 2012].

5.2. Rifted margin seismicity

Despite the often-used term “passive margins”, rifted continental margins release tectonic stress via earthquakes [e.g. Heidbach *et al.*, 2010]. In the North Atlantic region, seismic activity is caused by lithospheric adjustment that followed the removal of continental ice sheets after the last glaciation [Arvidsson, 1996]. The vertical motion associated with this lithospheric readjustment can often reactivate basement faults inherited from the rift phase, triggering earthquakes with magnitudes up to $M=7.3$, as in Baffin Bay, 1933 [Stein *et al.*, 1979]. This same process causes large earthquakes in high southern latitudes, such as the 1998 Antarctic earthquake, which had a moment magnitude of 8.1 [Tsuboi *et al.*, 2014].

Additionally, passive margin deformation may be influenced by far-field plate boundary forces such as

ridge push [Bott, 1991], basal drag [Yamato *et al.*, 2013], or compressive stress resulting from a lateral buoyancy gradient between continents and ocean [Artyushkov, 1973]. These forces manifest in the interior of the plate and activate zones of intraplate weakness [Hillis *et al.*, 2008]. In accordance with the Wilson Cycle, reverse-fault passive margin earthquakes may also be interpreted as precursors of margin transformation into an active subduction zone, and this has been proposed for the Atlantic margins of America [Nikolaeva *et al.*, 2011; Marques *et al.*, 2013], and south-west Europe [Duarte *et al.*, 2013].

5.3. Submarine landslides and tsunami hazard

Another natural hazard is related to the accumulation of large quantities of sediments at rifted margins. Sediment relocation occurs gradually via submarine channels or bottom currents, but can also occur abruptly in submarine slumps, debris flows or turbidity currents [Leynaud *et al.*, 2009; Korup, 2012]. Hundreds of submarine landslides have been described at active and passive margins worldwide [e.g. Moore *et al.*, 1994; McAdoo *et al.*, 2000; Tinti *et al.*, 2004; Brune *et al.*, 2010a, 2010b, 2010c; Løvholt *et al.*, 2012; Schwab *et al.*, 2012, 2014; Harbitz *et al.*, 2013; Clarke *et al.*, 2014; Hubble *et al.*, 2016] and some have produced large tsunamis. In 1929 for instance, a magnitude $M_S=7.2$ earthquake struck the continental shelf of Newfoundland, disrupting 200 km³ of slope sediments and triggering the largest submarine landslide in Canada’s history. The sediments disintegrated into a turbidity current, which carried mud and sand up to 1000 km from the source area. Several submarine telegraph cables were broken by the resultant turbidity current, which enabled subsequent estimations of slide speed (60–100 kilometers per hour). The slide generated a tsunami that killed 28 people [Fine *et al.*, 2005]. In the rifted margins of the Mediterranean, landslide-generated tsunamis damaged coastal structures and claimed casualties in the Corinth Gulf, Greece in 1963 [Papadopoulos *et al.*, 2007], and near Nice, France in 1979 [Assier-Rzadkiewicz *et al.*, 2000].

Certain conditions must be fulfilled in order to induce submarine slope failure. First, a large amount of sediments must be available developing a critical slope and, secondly, a trigger has to induce the mass movement. The first condition is fulfilled near river mouths and within submarine fans with high sedimentation rates, but also in the oceanic melting regions of continental glaciers. Slopes may destabilize for two main reasons: due to an increase of the applied stresses (because of gravitational loading by sediments or accelerations of an earthquake) or if the inner strength decreases (by means of increasing excess pore pressure) [e.g. Hampton *et al.*, 1996]. If a slope is relatively unstable, a landslide can theoretically be initiated by any small perturbation, although in most cases, the trigger is an earthquake. Some of Earth’s

largest submarine landslides have been identified at passive margins. In particular, formerly glaciated shelf regions, like the northern North Atlantic margins, are prone to large landslides: the interlayering of glacial and interglacial sediments combined with enhanced seismicity due to postglacial rebound facilitate slope instability [Bryn *et al.*, 2005]. Another important factor for destabilizing slopes appears to be gas hydrate dissociation as it generates high pore water overpressure [Micallef *et al.*, 2009].

Norway's continental slopes have undergone several catastrophic failures (Fig. 9). The best studied mass movement is the Storegga slide, which took place 8200 years ago and dislocated 2400 km³ of sediment [Bondevik *et al.*, 2005]. With an affected surface area of nearly 100,000 m², this single slide was as large as

Iceland or Portugal [Hafliðason *et al.*, 2004]. Other events offshore Norway include the Trænadjupet slide (~500 km³ [Laberg *et al.*, 2002]), the Bear Island Slide (~1100-1400 km³ [Leynaud *et al.*, 2009]), the Fugløy Bank slide (volume currently unknown [Leynaud *et al.*, 2009]), the early holocene Andøya Slide (~900 km³ [Laberg *et al.*, 2000]) and the Hinlopen slide north of Svalbard (~1350 km³ [Hogan *et al.*, 2013]).

Tsunami deposits of the Storegga event have been found on the Faroe Islands, Scottish coastlines, and in Norwegian lake sediments. Run-ups as high as 20m were reported on the Shetland Islands, [Bondevik *et al.*, 2005] illustrating the tsunami potential of these submarine mass movements. Several places along the Norwegian margin have not yet failed following the

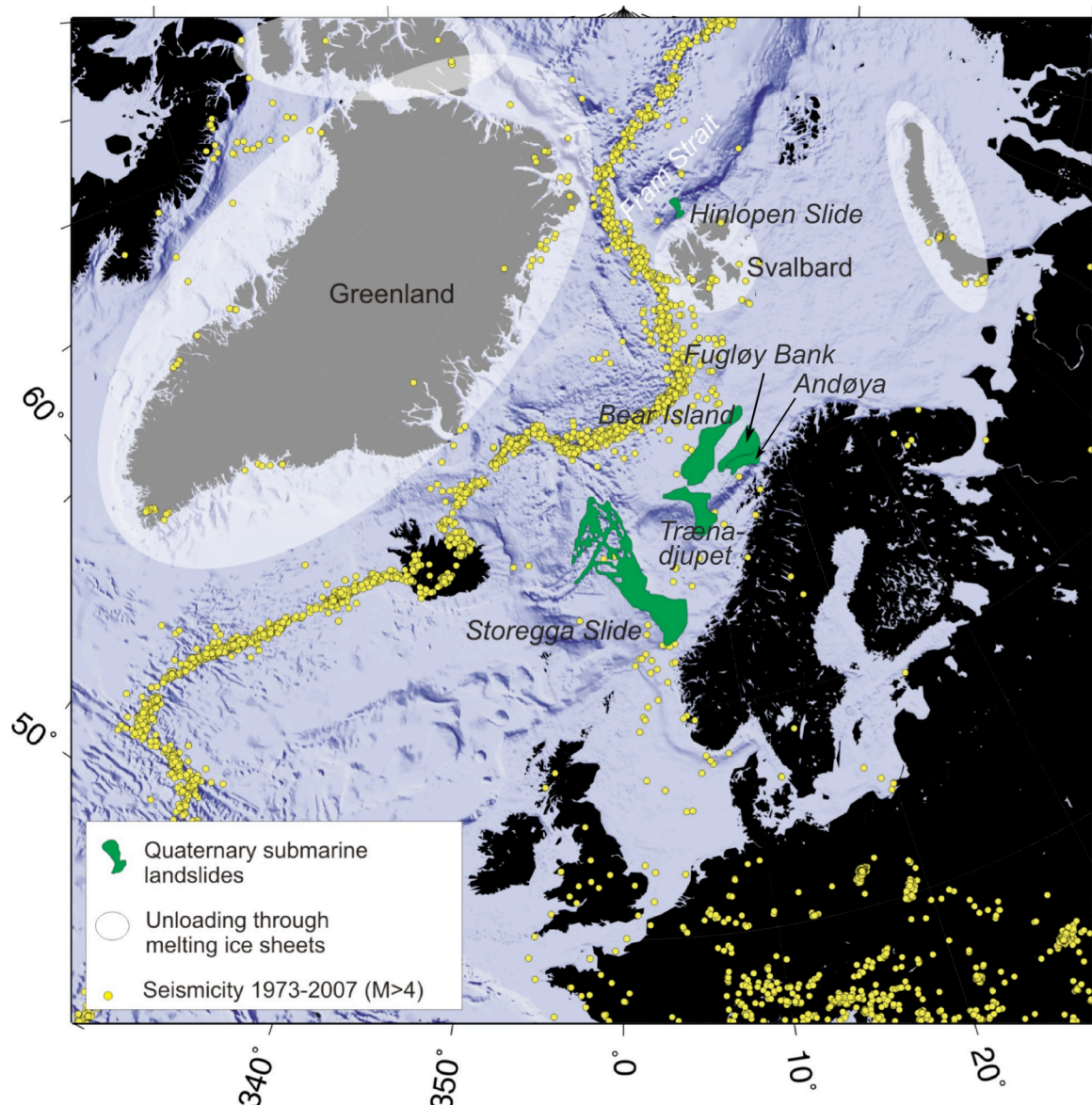


Fig. 9: Submarine slope failures at north-east Atlantic rifted margins.

Mapped landslides are shown in green and seismic activity in yellow. Areas of unloading ice shields and expected post-glacial rebound are highlighted. Modified after Berndt *et al.* [2009]. Shaded relief based on the International Bathymetric Chart of the Arctic Ocean (IBCAO).

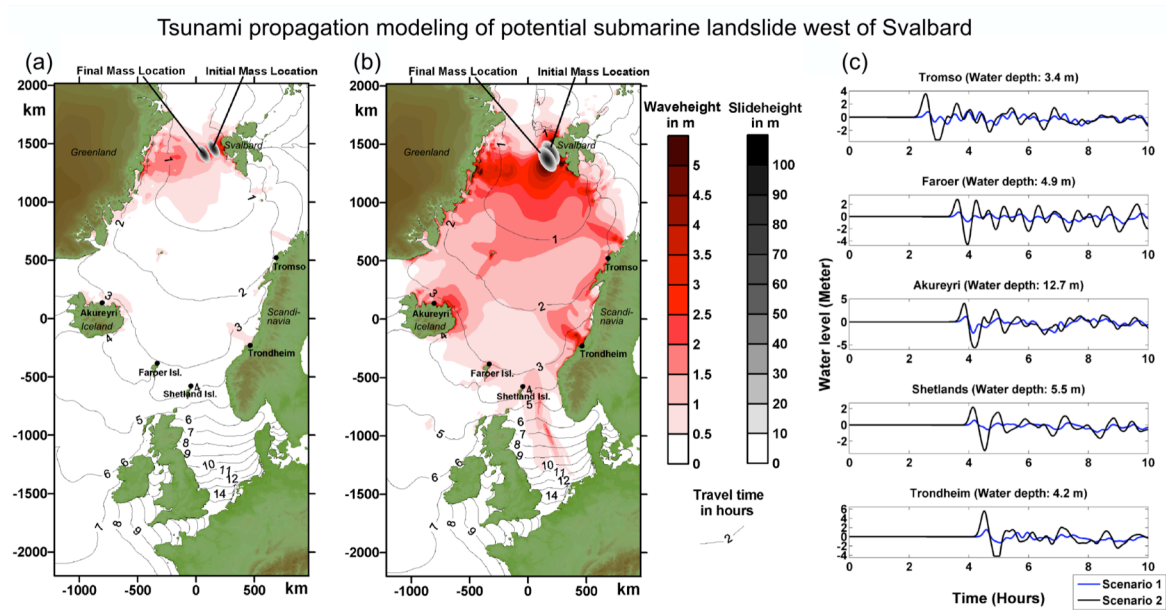


Fig. 10: Tsunami model for two possible landslide scenarios west of Svalbard.

(a,b) Assumed landslide locations, size and slide direction is depicted with gray color scale. Tsunami maximal wave height is shown in red and arrival times (in hours) after slope failure in black contours. Scenario (a) involves a slide of 500 km^3 moving westward, while scenario (b) consists of 1000 km^3 slide that moves south-eastward. Note elevated wave heights in the North Sea, offshore Trondheim, and north of Tromsø that are generated by tsunami focusing due to passive margin bathymetry. While both setups are realistic, the resulting tsunami can be (a) minor or (b) hazardous. (c) Time series of virtual tide gauge stations. Figure modified after Berndt et al. [2009].

last glaciation. One location offshore Svalbard features gas hydrate-bearing sediments that are sensitive to recent Arctic warming [Vogt and Jung, 2002]. It has thus been proposed as a likely site for a future submarine slope failure [Berndt et al., 2009]. Numerical modeling studies have demonstrated that a submarine landslide west of Svalbard could trigger a tsunami capable of reaching northwest Europe (Fig. 10). New monitoring techniques are urgently needed for continental slopes which are susceptible to this threat [Brune et al., 2009], as early-warning capacity currently is limited for such hazards. The likelihood of such an event is currently unknown, but geophysical surveys west of Svalbard are endeavoring to determine the current stability of gas hydrates in the region [Berndt et al., 2014].

6. SUMMARY AND OUTLOOK

Continental rifts and rifted margins are complex tectonic features affected by a large spectrum of geological processes, such as plate driving forces, localization of crustal deformation, partial melting, and dike emplacement which operate at different temporal and spatial scales. The tectonic structure of continental rifts is governed by inherited features of previous deformation episodes, the rheology of the lithosphere, the rate and direction of extension, the amount of partial melting, and the availability of sediments. These factors vary from one rift to another and geological observation, geophysical surveying and

numerical modeling elucidate the relative importance of these processes. It is clear that large earthquakes pose a significant geohazard at rift systems and at rifted margins where they also have the potential for generating devastating landslide tsunamis. The likelihood of these events, however, is difficult to assess since they occur in slowly deforming regions and feature large recurrence times.

Future investigations require three-dimensional observation and modeling techniques to address key questions surrounding rift dynamics: How does rift segmentation evolve over time and how does it link to passive margin segmentation? What are the causes and consequences of rift competition between two rift branches? What controls rift architecture during the transition from rifting to stable sea-floor spreading? A more comprehensive understanding of the processes that shape continental rifts and passive margins will be vital for future georesource evaluation and will also empower us to devise new geohazard assessment and mitigation strategies.

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