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Geodynamics of continental rift initiation and evolution

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Key Points

- Continental rifting is an intrinsically transient process that thins the lithosphere through distinct successive phases from inception to breakup. The structural evolution is controlled by the competition between geodynamic drivers, resisting factors and weakening processes.
- Rifting proceeds where lithospheric strength is lowest. Strength minima on a local scale are not necessarily strength minima on a plate scale. Rift deformation hence 'tries out' weaknesses on multiple scales, subsequently generating competing structures, migration and cessation of tectonic activity.
- Failed rifts should be considered dormant rather than dead as rift-induced weaknesses can get reactivated even after hundreds of millions of years if the local force balance changes.
- Mantle plumes simultaneously enhance the forces driving rifting and reduce lithospheric strength by causing magmatic intrusions. Rifts that experienced plume impingement appear to always proceed to seafloor spreading.
- Lithospheric thinning and magmatism are intimately coupled. If dikes cut through the lithosphere, they efficiently heat and weaken the rift, potentially reducing its resistance by an order of magnitude.

Abstract

A rift is a nascent plate boundary where continental lithosphere is extended and possibly broken. In the geologic past, rifting played a major role in shaping the surface of our planet, while at present, continental rifts are of societal relevance by hosting key georesources such as geothermal energy and ore deposits. This Review discusses fundamental rift processes, geodynamic forces and their tectonic interactions over geological time scales, and identifies the mechanisms that lead to the large variety of rifts on Earth. Linking forces and processes is particularly relevant for rifts, as progressive thinning of the lithosphere prevents a steady-state configuration and prompts continuous changes in their force balance. Rifting initiates through multi-scale exploitation of inherited weaknesses, which generates dynamic spatiotemporal competition, cessation or localisation of rift structures. Successful continent-scale rifts feature an abrupt and roughly ten-fold increase of divergence velocity once the lithosphere is sufficiently weakened. We infer that mantle plumes exert major control on breakup, through dynamic topography and by generating melts that can weaken the lithosphere by an order of magnitude. Outstanding future challenges include unravelling where magmatism is a cause or a consequence of rifting, isolating the tipping points that separate successful from failed rifting, and deciphering the interaction of rift tectonics with fluid flow during georesource formation and volatile release.

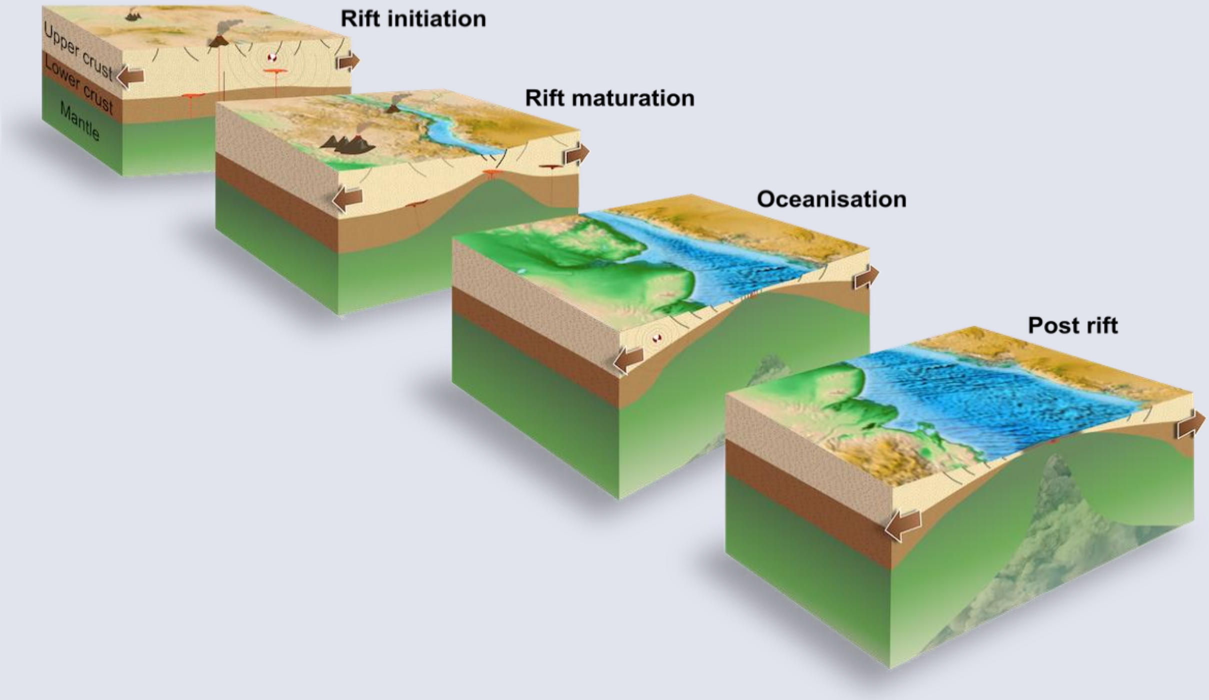
1. Introduction

Continental rifts form when the lithosphere thins during tectonic extension. Rift zones manifest through seismic and magmatic activity within a region that can be several 10s to 100s km wide, resulting in considerable hazards such as earthquakes[1], volcanism[2] and landslides[3]. Rifts are also responsible for large-scale CO_2 degassing[4] and likely affect atmospheric CO_2 concentrations, particularly during periods of supercontinental break-up. Rift basins hold a strong economical and societal relevance through their geothermal energy potential[5] and as hosts of ore deposits[6,7]. Understanding the links across space and time between the geodynamic processes that control rift initiation and evolution therefore holds relevance for sustainable and safe utilisation of rift resources and characterisation of rift-related hazards.

The initiation and evolution of continental rifting is a result of the interplay between tectonic extensional forces and the mechanical resistance of the continental lithosphere (Box 1). In most continental areas, tectonic driving forces are smaller than lithospheric strength, resulting in continental interiors that remain stable over hundreds of millions of years. However, an increase in driving forces or a decrease of the resisting strength can tip the force balance and result in tectonic activity. For example, the impingement of a mantle plume[8] beneath the lithosphere can create topographic deflection as well as elevated mantle temperatures thereby generating melts that weaken the lithosphere, as is inferred along the East African Rift[9]. Another tipping point could result from a change in surrounding plate boundary configurations, such as a transition from subduction to continental collision. For instance, the collision of the Adriatic plate with Eurasia that formed the European Alps also generated the Rhine Graben and Eger Rift in Central Europe[10].

Once a rift has initiated, its dynamics can change dramatically through time. Thinning of the lithosphere results in upwelling of hot asthenosphere that leads to melt generation and volcanism. The generated magmatic volumes and volcanic eruption rates are modulated by mantle temperatures, volatile content, and divergence rate. Magma-poor rifts (such as the central Malawi Rift, East Africa, without eruptive volcanic centres) can thus turn into magma-rich rifts (like in Ethiopia with more than 50 Holocene volcanoes[11]) if rift velocity or mantle temperature increase. Conversely, young rifts with thick brittle crustal layers and large border faults are prone to generating larger earthquakes than mature rifts where the crust has been extensively thinned. The largest historical earthquake in central Europe (M_W~6.5), for instance, occurred in 1365 in the low-strain Upper Rhine Graben and destroyed the city of Basel, Switzerland[1]. Due to typically low rift velocities of a few mm yr^{-1}, earthquake recurrence intervals are long (thousands of years), which often leads to limited societal awareness, rendering seismic events particularly devastating.

While some continental rifts undergo extension to the point that they generate a new ocean basin flanked on either side by rifted continental margins (Box 1), other rifts can cease activity altogether, forming so-called failed rifts[12]. Prominent examples of rift failure include the North Sea, the West African Rift System and the Midcontinent Rift underlying the Great Lakes of North America. It is clear that rifting stopped in these cases because the rift's strength eventually exceeded the driving forces, but whether failure resulted from increasing resistance within the rift, waning driving forces, or both, remains an open question so far.



Box 1 | Conceptual model for the inception and evolution of continental rifts.

Breaking the continental lithosphere is an intrinsically transient process. Rifting hence progresses continuously, involving several major phases.

Rift initiation. *Rifts localise when tensional stresses exceed the strength of the continental lithosphere. Deformation can be accommodated by brittle faults, ductile shear zones, and magmatic dikes. Incipient rifting occurs, for instance, at the Okavango Rift in the south-eastern termination of the East African Rift and in the Central European Eger Rift. Structures inherited from previous deformation episodes often facilitate and guide deformation.*

Rift maturation*. Neighbouring faults compete and ultimately coalesce into an array of dominant faults. The strike-perpendicular extent of a rift can vary from less than a hundred kilometres in narrow rifts, such as the Tanganyika Rift in East Africa, to several hundred kilometres in wide rifts, like in the Basin and Range region of North America. Slip along major faults and ductile thinning of the lower crust causes hanging wall subsidence, which creates sedimentary basins. Hot, sublithospheric mantle (i.e. asthenosphere) rises in response to lithospheric thinning and causes decompression melting. These melts migrate rapidly through the lithosphere generating dikes, sills, and volcanoes.*

Oceanisation*. When the crust is thinned sufficiently, deformation progressively focusses and migrates towards the location of future breakup. Enhanced decompression melting at a successively larger depth range intensifies magmatic emplacement in the rift centre. Eventually, magmatic segments often accommodate most extensional deformation, such as in the Afar region in East Africa or the Woodlark Rift in southeastern Papua New Guinea. When the continental lithosphere is separated and replaced by upwelling sublithospheric mantle and basaltic melt intrusions, the transition to mid-ocean spreading takes place.*

Post rift*. Once the newly formed mid-ocean ridge accommodates all plate divergence, the former rift turns into a rifted continental margin. This margin, however, continues to deform: on the one hand, proximal and distal parts of the margin cool at different rates, causing the distal part to subside slightly faster, leading to margin tilting. On the other hand, sediments transported by onshore river networks can load the proximal margin inducing subsidence. Additional processes, like along-shore sediment transport or glacial loading and unloading as in the North Atlantic, continue to shape rifted margins long beyond the original rift phase.*

In this Review, we focus on key factors and processes affecting rift tectonics and how they evolve on geological time scales by providing a modular review of the driving forces, resisting factors and weakening processes. Our approach that examines the interaction of forces and processes is particularly relevant for rifts, since in contrast to plate boundaries like subduction zones and mature mid-ocean ridges, rifts can never achieve a quasi steady-state configuration. Instead, the balance between driving forces and resisting strength evolves continuously, shaping the characteristic normal faults, sedimentary basins and magmatic features documented in presently active rifts, failed rifts and rifted margins worldwide.

2. Rift initiation

Breaking strong continental lithosphere requires overcoming large resistance forces. In stable interiors of tectonic plates, resisting factors are larger than the drivers of deformation, which is why tectonic motion is focussed along weak plate boundaries throughout geological history, rather than in continental interiors. Continental rifts, however, are nascent plate boundaries where deformation managed to localise and where resistance can play a role equally as important as the driving forces. In the following, we discuss drivers of extension and how their interplay with static resistance and inherited heterogeneities can guide tectonic deformation during rift initiation (Fig. 1).

2.1 Breaking stable lithosphere

Forming a rift requires bringing the entire continental lithosphere to its yield point, which at low temperatures involves brittle fractures and faults, and at higher temperatures requires the activation of ductile creep processes[13]. At the lithosphere scale, this can be conceptualised as applying a driving force that matches or exceeds the strength of the plate, which includes crust and mantle lithosphere (Fig. 1). The possible driving forces of rifting result from subduction, mantle convection, and lithospheric buoyancy.

2.1.1 Subduction-related forces

Subduction is potentially the largest driver of continental extension. All plate-driving forces result from lateral variations in density within the Earth, but the largest density gradients in the upper mantle are associated with subduction of cold lithosphere. Subduction produces a downward force often termed slab-pull (Fig. 1c). This line force is related to the negative buoyancy of a subducting oceanic plate and can be as large as ~30 TN m^{-1} (Ref.[14]) depending on the depth range of subduction and the age of the plate being subducted. The pull of subduction zones on the attached plates is thought to be a major driver of plate tectonics, which is inferred from the observation that Earth's subducting plates move 3 to 4 times faster than plates without attached subducting slabs[15]. How much force subduction can contribute to drive rifting depends, however, on the specific plate tectonic setting.

Intra-continental rifts located on the subducting plate itself can experience tensional stresses that are directly caused by slab pull. An example where subduction kinematically facilitates rifting is the Afro-Arabian Rift system. The East African part of the broader Afro-Arabian Rift system does not seem to be directly affected by subduction, but the Red Sea is. The African continent is separating from Arabia across the Red Sea while the Arabian Plate subducts beneath the Eurasian Plate. The associated slab pull drags the Arabian Plate more forcefully[16], which explains why the Red Sea is rifting about an order of magnitude faster than the rest of East Africa[17]. Another place where rifting is facilitated by subduction is the Woodlark Basin of Papua New Guinea[18]. There, the initiation of a new subduction zone located a few hundred kilometres to the north occurred roughly synchronously with the initiation of Woodlark rifting 10 Myr ago[19].

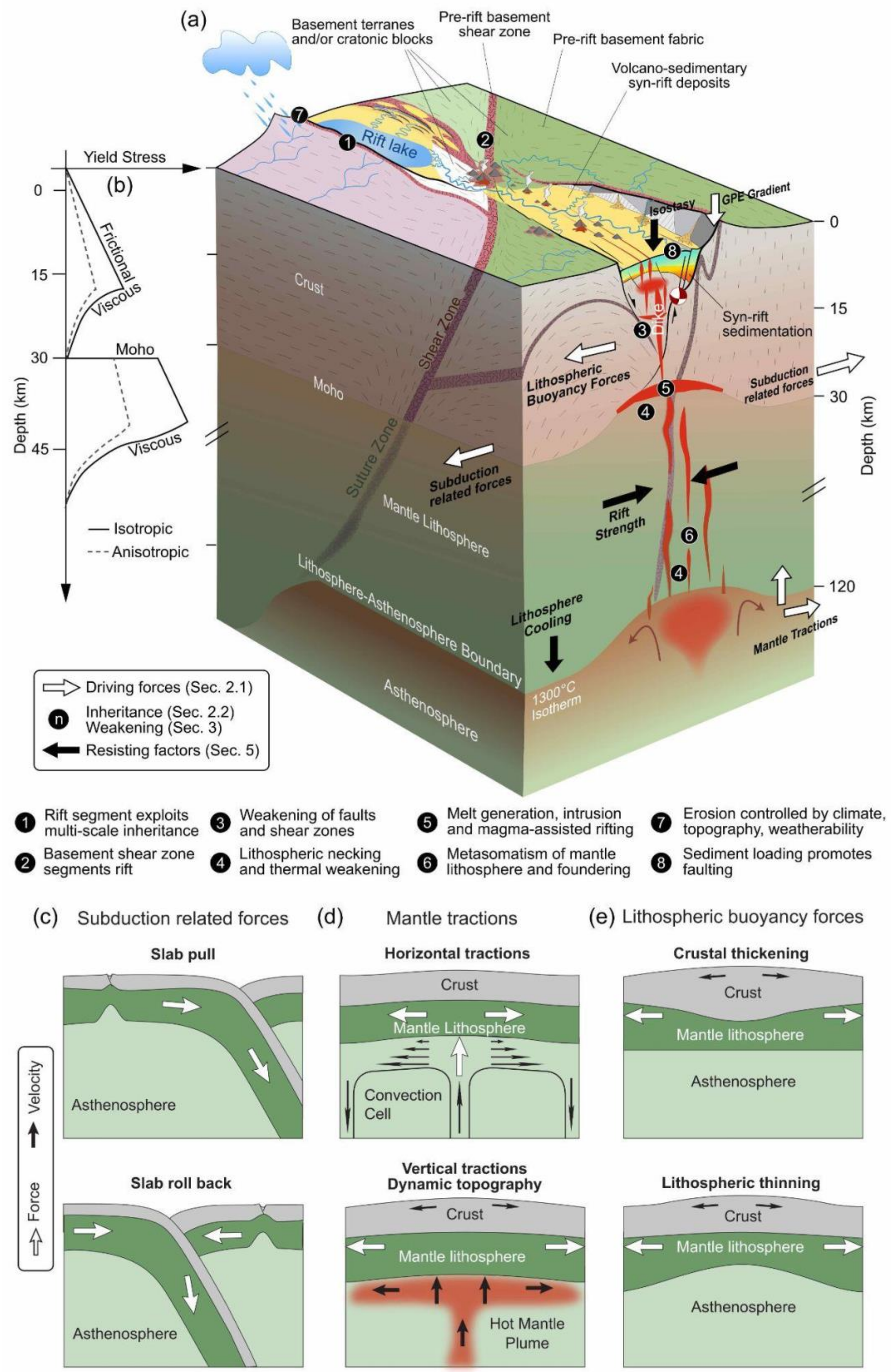


Fig. 1 | Driving forces, resisting factors, and weakening processes that accompany rifting. (a) *Continental rifts are affected by a multitude of processes ranging from plate-tectonic driving forces to crustal faulting and surface processes.* ***(b)*** *Strength profile of the crust and lithosphere, showing the relatively lower yield strength imposed by the presence of mechanical anisotropy.* ***(c-e)*** *Simplified representation of major forces that drive continental rifting.*

In some subduction zones, the trench can migrate laterally in the direction opposite to the downgoing motion of the slab. This is thought to be the surface manifestation of a retrograde motion of the slab, termed trench retreat. If trench retreat is faster than upper plate advance, the upper plate undergoes extension forming a back-arc rift (Fig. 1c). Many marginal basins around the Pacific[20] such as the Japan Sea are products of such back-arc rifting[21]. Since arcs are typically located within a few hundred kilometres of a trench, this type of rifting does not split the middle of a continent, but breaks off slivers of the continent's margin, or parts of an island arc. These slivers can later become terranes that eventually attach to another continent. Many rifts are, however, not connected to subduction zones and other forces are required to overcome the strength of the plate.

2.1.2 Mantle convection and mantle plumes

Thermal and chemical density variations in the mantle generate convective flow. Where asthenosphere motion interacts with the lithosphere, it causes both vertical and horizontal mantle tractions that contribute to driving plate tectonics in general and rifting specifically.

Vertical tractions deflect the lithosphere from its isostatic equilibrium causing dynamic topography. Dynamic uplift in rift settings is inferred over many regions where present and past mantle plumes have been identified[22] (Fig. 1d). However, the magnitude of dynamic topography is difficult to measure due to incomplete knowledge of asthenosphere viscosity and velocity, and the best ways to estimate it are controversial[23,24]. Straightforward estimates of the amplitude and wavelength of dynamic topography come from oceanic regions where there are good geophysical constraints on isostatic topography[25]. The magnitude of dynamic depth variations in the ocean basins[25] ranges between ±1000 and ±2000 m with maximum values above mantle plumes. Correcting for water loading translates to ±700 and ±1400 m for continental domains, respectively. The extensional force related to dynamic uplift results from gravitational potential energy (GPE) gradients and is proportional to elevation and the thickness of the lithosphere that is elevated. For 1 km of dynamic elevation, for instance due to a mantle plume and a continental lithospheric thickness of 100 km, the extensional force is about 3 TN m^{-1}.

An equally important part of plume-impingement is plume-related melting of mantle rocks and subsequent magma ascent. This process is a very efficient means to heat up and therefore weaken the lithosphere that in some cases might be responsible for rift initiation[26,27]. Examples of correlations between rifting and plume impingement include the separation of South America and Africa associated with the Paraná and Etendeka flood basalts[28] that erupted ~133 Myr ago, the opening of the North Atlantic[29] associated with the North Atlantic Igneous Province between 61-56 My ago, as well as the presently active East African Rift System[30]. While mantle plumes can aid continental rifting, they are not a requirement, as for example, demonstrated by the breakup between Iberia and Newfoundland, which proceeded without flood basalt eruptions.

Horizontal tractions at the base of the plate result from viscous drag caused by the movement of plates over the asthenosphere, or from lateral spreading of mantle upwellings such as plumes. However, the magnitude and even the sign of horizontal tractions are disputed. Because the Pacific plate is the largest on Earth it should be the one that is slowed down most by basal viscous drag. The fact that the Pacific plate is the fastest moving plate indicates that the drag force is small compared to other forces that could drive plate motions and affect rifting[31]. Horizontal tractions are computed by integrating basal shear stresses over area. If the viscosity beneath the Pacific plate at 100 km to 200 km depths is 10^{19} Pas, then the shear stress under a plate moving at Pacific plate rates would be of order 10^5 Pa, amounting to the total force on the 10,000 km wide Pacific plate of about 1 TN m^{-1}, i.e., less than a tenth of the slab pull force.

Horizontal tractions, however, do appear to affect plates with thick lithosphere. Plates that include thick cratons move slower than plates that do not[32]. This impact of cratons could reflect a strong increase in mantle viscosity with depth[33], which impedes thick cratonic roots to plough through the strong deeper mantle.

In summary, mantle convective motions are most likely to initiate and drive rifting through the uplift caused by mantle plumes like in the East African Rift. Horizontal tractions on plates with thick lithosphere can affect lithospheric stress, but are more likely to resist than to drive rifting.

2.1.3 Topography and lithospheric structure

Lithospheric buoyancy forces arise from GPE gradients due to lateral variations in topography and/or lithospheric structure[34] that generate deviatoric stress[35]. The topography-related component of this force has been discussed in the previous section. The component that is caused by lithospheric structure variations is due to heterogeneities in crustal and lithospheric thickness (Fig. 1e), temperature, composition, and density. From a global perspective, lithospheric buoyancy forces range from approximately 0.5 to 5.1 TN m^{-1} for areas in deviatoric tension[36].

In regions where subduction-related forces play a minimal role, such as in Africa, Antarctica, and the Basin and Range, lithospheric buoyancy forces dominate the force balance. Along the East African Rift, there is long-wavelength topography stretching from Ethiopia to South Africa that is to a large degree supported by isostasy - with distinct variations in crustal[37,38] and lithospheric thickness[39] - but also by vertical mantle tractions[40]. Numerical modelling approaches[41–43] suggest lithospheric buoyancy forces on the order of 1.5 - 1.7 TN m^{-1} are the main driver of the large-scale east-west extension documented by geodetic observations[44]. A similar process drives the broad region of extension of the Basin and Range Province in the western United States[45] with magnitudes of lithospheric buoyancy forces ranging from 0.2 - 2.6 TN m^{-1}. Here, paleoelevation data have been used to estimate a wide range of past GPE gradients that suggest lithospheric buoyancy forces result primarily from crustal thickening that ultimately produces extensional stresses over a large region[46]. The collapse of regions of thick crust has also been suggested as a cause of the broad zones of rifting inferred in West Antarctica[47,48]. These examples illustrate that, in the absence of subduction-related forces, GPE gradients that generate deviatoric tension have played an important role in initiating and maintaining rifting.

We conclude that the most important drivers of rifting are subduction-related forces and gravitational potential energy gradients. But in many cases, the superposition of available forces appears to be insufficient to overcome the strength of cold, thick continental lithosphere in order to initiate rifting. Exploitation of inherited weaknesses and dynamic softening mechanisms, such as magmatic intrusions and lithospheric necking, are therefore necessary to locally weaken the lithosphere so that continental rupture can occur.

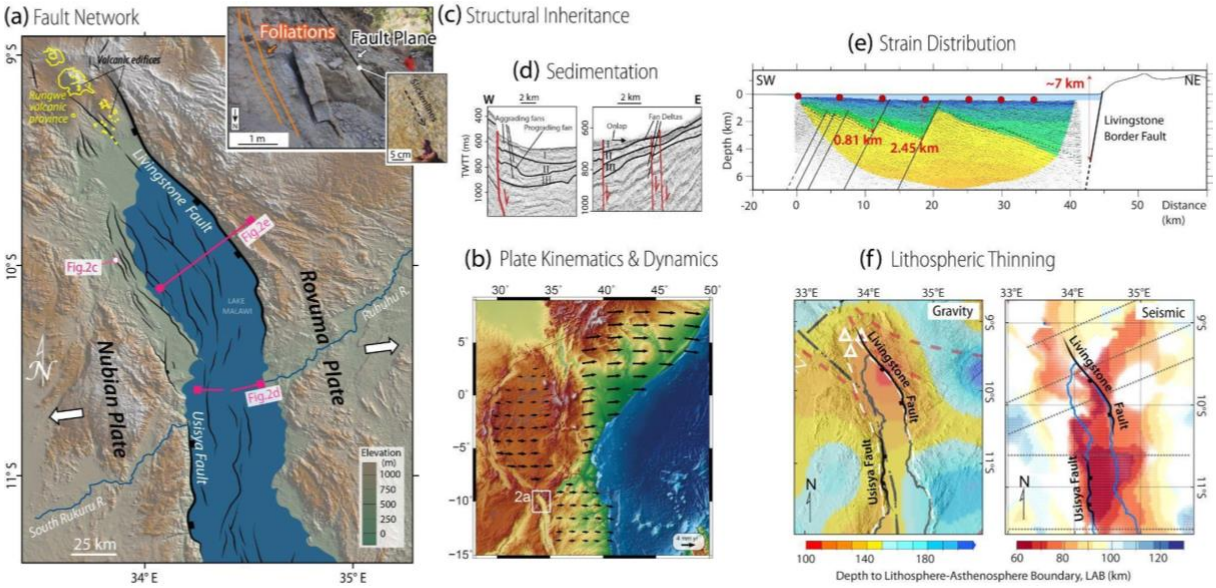


Fig. 2 | Juvenile continental rift structures exemplified by observations and data from the Malawi Rift. *(a,b) Near the surface, extension is accommodated by large border faults and smaller intrabasinal faults*[49,50]*. (c) Early-stage rifting commonly features strain localization through exploitation of pre-rift basement structures*[51]*. (d,e) Slip accumulation on normal faults creates spatially variable accommodation space for sediment deposition*[52,53] *(f) At depth, extension is accommodated by ductile deformation that thins the lithosphere in an area that can be somewhat wider than the near-surface sediment basin. Geophysical data differ in absolute LAB depths, but nevertheless show extensive thinning of the lithosphere*[54,55]*.*

2.2 Exploiting inherited weaknesses

Inherited structures from previous tectonic and/or magmatic events reduce the static lithospheric strength prior to the inception of rifting (Fig. 1b) and thereby affect the geometry of the nascent rift on scales ranging from an individual fault to extension throughout the lithosphere. Examples of inherited features include anisotropic rock fabrics such as foliations (Fig. 2c) and faults from prior tectonic events, suture zones from previous collisions, regions of metasomatized mantle and variations in the thickness and composition of the crust and lithosphere. Such pre-rift heterogeneities in strength are multi-scale and vary with depth through the lithosphere. At large scales, deformation of the lithosphere is compartmentalised by discrete basement terranes, i.e. geological provinces that vary in age, thermal evolution, composition, tectonic history, and rheology[56,57]. Rifts nucleate and develop preferentially in weaker terranes[58]. At smaller scales, new extensional faults (Fig. 2) exploit favourably oriented fabrics and pre-existing faults.

The fact that rifts develop preferentially in previously deformed lithosphere is now well documented across many Wilson cycles[59,60]. In particular, many rifts develop in orogens created during previous continental collisions. This fact is evidenced by examples worldwide such as the localization of the East African Rift (Fig. 2) in the orogenic belts that surround the Tanzania Craton[61], the Central and West African Rift Systems in the TransSaharan Mobile Belt and metacratonised lithosphere, the South Atlantic Rift System along the Rio Pardo-West Congo Belt[62], the Mid-Continent Rift along the orogenic belts that bound the Superior Craton[63], and the North Atlantic along the Caledonian and Variscan orogens[64]. Orogens are favourable locations for rifting because they contain suture zones and other fault systems, and they are often characterised by thickened, hot and weak crust. These inherited structures remain weak for hundreds of million years, so that continents collide and tear apart repeatedly along roughly the same boundaries.

Within the upper crust, structural inheritance affects the formation of fault patterns as it often leads to the reduction of frictional strength compared to the bulk rock (Fig. 1b). Hence, favourably-oriented pre-existing planes of strength contrast will fail and localise deformation prior to the nucleation of new discontinuities in the pristine portions of the rock[65,66]. The exploitation behaviour can, however, be very complex as faults form as a result of 3D anisotropies interacting with the 3D stress field. Field observations show that newly formed fault surfaces can align with the fabrics along-strike and down-dip[51]. Where fabrics or pre-existing structures are not well aligned, faults will cut across them obliquely, sometimes leading to more complex, discontinuous new faults[67,68]. In other cases, the low stiffness of the shear zones can also cause a local re-orientation of the principal compressive stresses to facilitate exploitation of the shear zones by normal faulting, as observed in the East African Rukwa Rift[67]. Although the integrated strength of the lithosphere is the first-order control on the length and displacement of basin-bounding faults that define segmentation[69], segmentation can be modulated by structural inheritance[70,71], thus, playing a vital role both for the long-term evolution of a rift and its short-term seismicity.

In summary, driving forces and pre-existing strength heterogeneities of the lithosphere conspire to shape rift initiation. Of all drivers of extension, subduction zones provide the largest source of horizontal stress. In the absence of subduction, we find that lithospheric buoyancy forces can superpose to overcome the resisting strength of the lithosphere, particularly in regions with large dynamic topography like East Africa. Continental extension is often facilitated by inherited fabrics and structures at a range of scales that are weaker than the surrounding rock and thus localise extensional deformation when favourably oriented. Even when unfavourably oriented, these remnant structures can modulate segmentation, influence rift fault complexity and locally perturb the stress field.

3. Rift weakening

Once a continental rift has formed, its geodynamic evolution is controlled by a range of processes that either weaken (Fig. 3b-d) or strengthen (Fig. 3e-g) the rifting lithosphere: strain softening, magmatic weakening and surface processes help to localise deformation while isostasy and cooling lead to strengthening of the rift. It is the dynamic interaction between these processes that generates the complex surface expressions during rift maturation.

3.1 Mechanical weakening and modes of rifting

Mechanical weakening is a cross-scale process ranging from grain-scale deformation to lithospheric-scale thermal weakening. To initiate rifting, the differential stresses must overcome the yielding thresholds of both shallow, brittle upper crust and deep, ductile lower crustal materials. The mode of rifting is, thereby, a function of the lithospheric yield strength distribution (Fig. 3). Crucially, deformation alters the internal state of rocks in a manner that lowers the strength of the lithosphere, which promotes the continuation of rifting and leads to a speed-up of extension prior to break-up.

In the brittle upper crust, the growth of faults coincides with a loss of material cohesion and reduction of the friction angle[72]. The latter process is facilitated by the development of a gouge layer where strain- and fluid-assisted reactions can precipitate weak phyllosilicate minerals[73,74], such as serpentine in ultramafic fault rocks[75]. This process requires water to be transported from the surface to the top of the mantle via active faults cutting through the entire crust. Serpentinisation has been particularly well documented at the West Iberian rifted margin[76,77] and is thought to occur only during the late stages of rifting when the crust is sufficiently cool to become entirely brittle[78]. The growth of weak minerals, along with elevated pore fluid pressures, causes a reduction in the fault's effective friction coefficient, lowering the differential

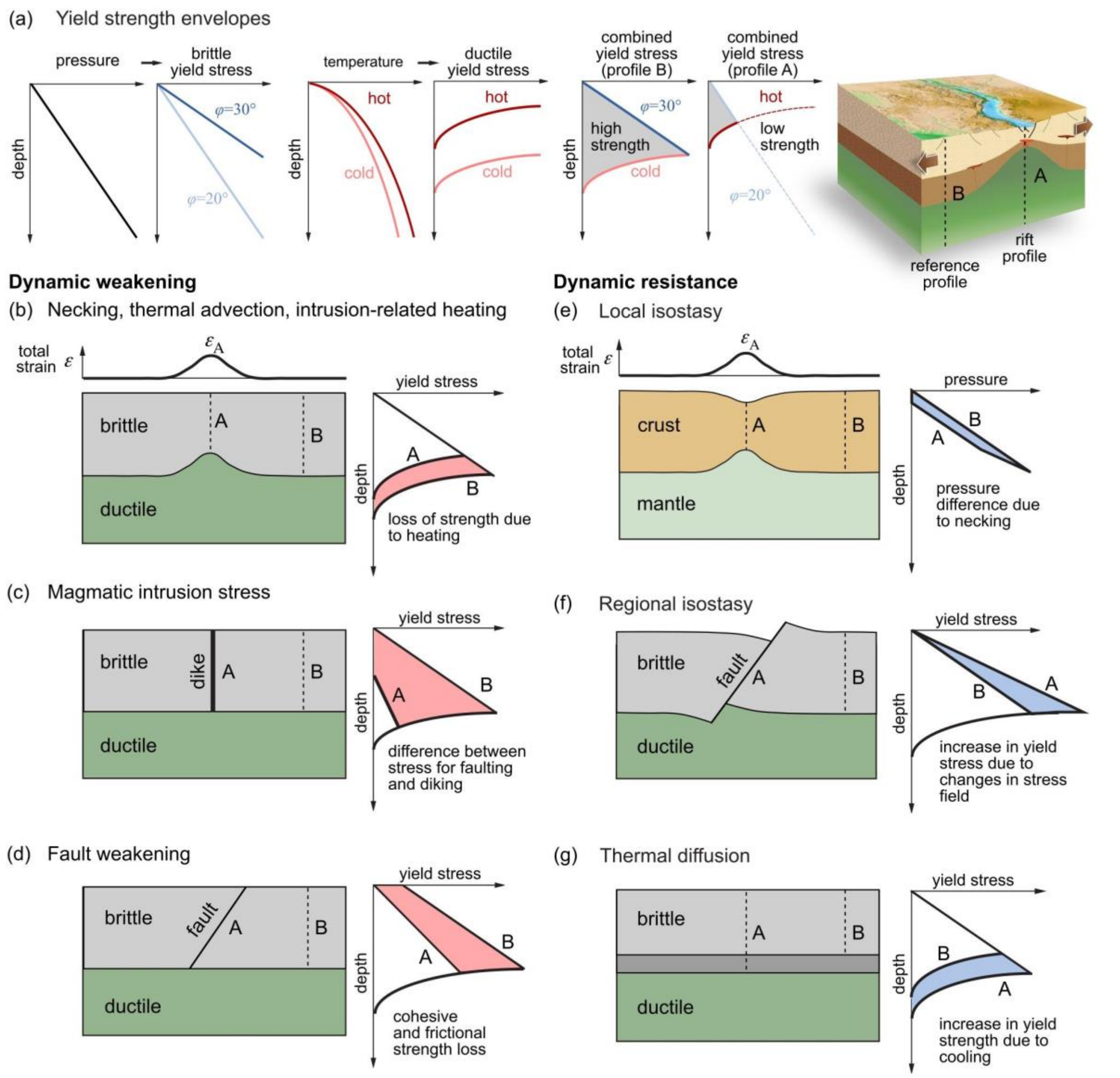


Fig. 3 | Lithospheric strength and response to deformation. *Yield strength envelopes illustrate the conceptual strength distribution within the lithosphere and allow for estimating the first-order impact of geodynamic processes on rifting.* ***(a)*** *Simplified representation of yield strength envelopes.* ***(b-d)*** *Dynamic processes facilitating deformation.* ***(e-g)*** *Dynamic processes impeding deformation. The relative magnitude of resisting and weakening processes as well as their interaction with driving forces is site and time-dependent. Panels b-g are based on Ref.* [51].

stress needed to drive slip. It is important to note that —unlike in convergent settings— super-hydrostatic fluid pressures can easily lead to net tensile stresses in a rifting context, prompting hydro-fracturing that decreases fluid pressures down to a hydrostatic state[79,80]. Very high fluid pressures are, thus, unlikely to play a key role in lowering the brittle strength of rifting lithosphere.

Weakening mechanisms have also been documented in the ductile lower crust[81]. High-temperature, non-linear ductile creep processes aids strain localization by reducing the rock's effective viscosity as its deformation rate increases. Lithospheric mantle can be weakened substantially by metasomatism, which is particularly relevant to enable rifting in otherwise strong cratonic lithosphere. High strain rates also reduce grain size[82], making rocks effectively weaker when deforming in the diffusion creep regime[83]. In numerical simulations the development of a localised, low-viscosity channel in the lower crust enables the position of localised strain in the upper crust to migrate laterally over hundreds of km, which explains the formation of asymmetric (wide and narrow) conjugate margins at the late stages of rifting[84,85].

In order for the above weakening processes to sustain the localisation of strain within a rift zone, they must overcome the action of several key delocalising phenomena. The first is isostatic adjustment (Fig. 3e), where crustal thinning generates pronounced surface depressions. Within a mature rift basin with a deep valley, GPE gradients across the rift flanks put the rift centre into relative compression. While this force component is usually smaller than the extensional force, it nevertheless opposes continued rifting, which occasionally leads to inversion events within extensional basins[86]. The second is bending of competent lithospheric layers (Fig. 3f), which rotates rift-bounding normal faults to dips less favourable than 60° (refs [87,88]). This dynamic change happens in a way that more stress is required to maintain fault activity, which ultimately impedes continued deformation[87,89]. Another mechanism is cooling of the lithosphere, which increases the strength of ductile layers by elevating rock viscosity (Fig. 3g). Overall, the competition between localising and delocalising processes is strongly modulated by the strength of the rifting lithosphere, which can lead to drastically different modes of continental extension[90].

Rift fault networks pose a considerable seismic threat, especially since rifts are often highly populated places. Maximum earthquake magnitudes, however, depend on the tectonic makeup of the rifting lithosphere. Some active rifts, such as the Balangida segment of Northern Tanzania or Lake Tanganyika, occur in strong cratonic lithosphere[91] where earthquakes occur at depths of 40 km or more[92,93]. This suggests a predominantly brittle crust and uppermost mantle that are strongly coupled. This configuration promotes a narrow rifting mode, where deformation remains focused along a narrow axis straddled by half-graben structures. Because the border faults of such rifts can root to depths in excess of 25 km (e.g., beneath Lake Malawi[50]), they can nucleate earthquakes as large as M_W~7. By contrast, extension in the Basin and Range Province affects a hotter lithosphere in which crust and mantle deformation are decoupled by a low-viscosity lower crust[94]. This configuration promotes the growth of a wide rift, which instead of having a clearly defined axis, distributes extension on a series of half-graben and horst structures across hundreds of kilometres. This wide distribution results from the delocalising effects described above that prevent sustained extension in a narrow axis. Wide rifts facilitate the formation of continental core complexes[95], which are topographic domes capped by low-angle, very large offset (>10 km) fault surfaces. Classic examples of such structures include the Whipple Mountains in the Southwestern United States[96]. The seismogenic potential of these large-offset faults remains debated. If they represent the rotated footwall of a steep fault cutting through a thin brittle crust, earthquake sizes are limited to M_W~6, as is typical of mid-ocean ridge settings[97]. If, however, they are able to slip at low angles, the large fault area could accommodate substantially larger earthquakes[98,99] (M_W>7).

At lithospheric scales, brittle and ductile weakening processes lead to successive focussing of extensional strain within the rift (Box 1). Accumulated thinning, or so-called necking, of the lithosphere constitutes a large-scale thermal weakening process that replaces cold and strong lithosphere with hot and weak mantle[100]. Numerical models have assessed the relative impact of strain softening versus lithospheric thinning and found that necking dominates the loss of lithospheric strength due to the highly non-linear dependence of rock viscosity on temperature[101]. The prominent reduction in lithospheric strength during rift evolution can even generate a feedback on plate kinematics as the loss of rift strength induces a roughly ten-fold acceleration of the involved plates[102], from a few mm yr^{-1} to several cm yr^{-1}. (Fig. 4d-f) The strong non-linearity of necking generates a tipping point that marks the transition from a state where the rift velocity is controlled by rift strength to a state where the divergence rate is governed by the large-scale force budget of the involved plates[103]. This process explains the speed-up of North America during Central Atlantic rifting[104], of South America during rifting of the South Atlantic[105], and of Australia during its separation from Antarctica[106].

3.2 Melt generation and magmatic weakening

Many major continental break-ups involved extrusion of massive amounts (up to a few billion cubic meters) of basaltic magma in regions called Large Igneous Provinces[107] (LIPs). Prominent examples are the Parana-Entendeka LIP that occurred close to the center of rifting of South America from Africa, and the Ethiopian-Yemeni LIP located at the triple junction of the Gulf of Aden, the Red Sea and the Main Ethiopian rift[26]. The question of whether magmatism is a cause or consequence of rifting is hotly debated. The generation of magma within the asthenosphere and deep lithosphere of continental rifts can occur in several ways; we describe three models below.

3.2.1 Melt generation in rifts

As stretching thins the lithosphere, the asthenosphere is pulled upwards and can partially melt due to adiabatic decompression, with the production of higher melt (i.e., magma) volumes at greater stretching factors[108]. According to this "melting-during-stretching" model, great magma volumes reflect high regional temperatures in the mantle. This model is appealing in its simplicity, but while it is applicable to mid-ocean ridge systems it is not adequate for continental rifts where abundant volcanism is observed in regions where stretching factors do not reach values sufficient to support adiabatic decompression melting (e.g., the East African Rift[109,110]).

Decompression melting can also happen as hot, low-density mantle plumes rise from great depths in the mantle[111]. The "plume-driven-melting" model for rifting appears viable in many extensional provinces, as dated basalts show that almost all LIP lavas are extruded before break-up is achieved (Fig. 4c). The majority of LIP flood basalts are erupted in less than a million years[29,112] while rift stretching at observed rates would take many millions of years to thin the lithosphere and drive partial melting of the mantle. Geochemical measurements support a role for mantle plumes in generating magma in places of ongoing rifting like East Africa. The $^{3}He/^{4}He$ signature of olivine and pyroxene in mafic lavas and mantle xenoliths is the least controversial indicator of mantle plume contributions to mafic volcanism (Fig. 4a). The shaded region of Fig. 4b indicates helium isotopic values that are enriched over values measured in normal mid-ocean ridge basalts. These data therefore support models involving mantle plume contributions, which agrees with tomographic evidence for a large-scale mantle plume beneath East Africa (Fig. 4c). Most analyses are from Miocene and younger samples, but Afar lavas include Oligocene eruptives which suggest long-lived plume contributions in this area. However, the maximum He isotopic values measured in East Africa (19.5 R/R_A) are substantially lower than those observed globally (~50 R/R_A), indicating that plume melts/volatiles are part of a multi-component mixture of mantle sources supporting East African Rift volcanism.

The third way rift magma can form is through melting of the lithospheric mantle. Rocks of the lowermost lithosphere can melt through elevated mantle potential temperatures induced by the impingement of mantle plumes and contribute significantly to rift volcanism, particularly in the stages prior to mid-ocean ridge style magmatism[114,115]. Silicate metasomatism, e.g., transport

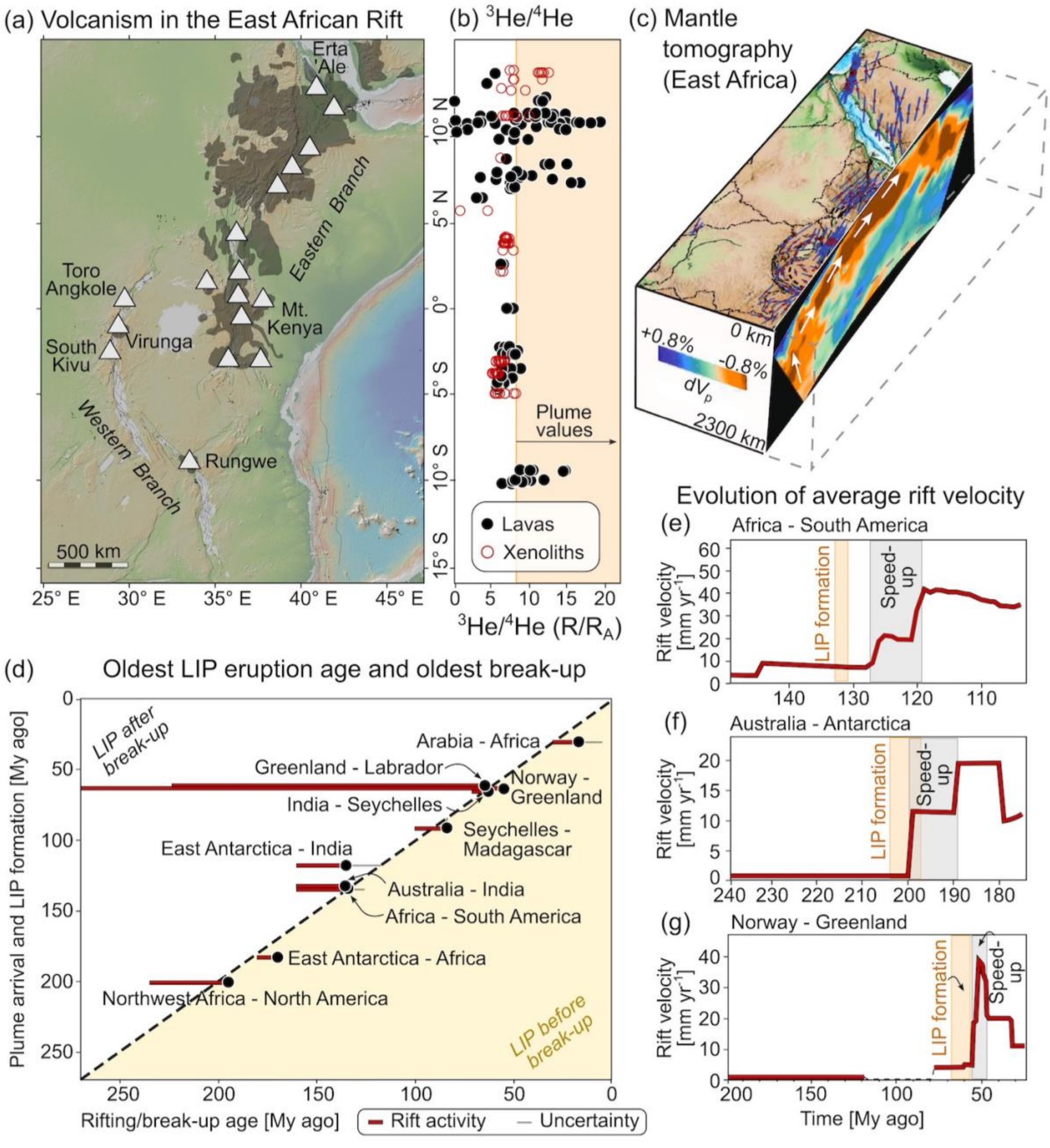


Fig. 4 | Plume impact on rifting. (a) *East African Rift volcanic fields and major volcanoes[9].* ***(b)*** *Mantle plume contributions to volatile release in the East African Rift are depicted in terms of $^3He/^4He$ signature in mafic lavas (closed black circles) and mantle xenoliths (open red circles).* ***(c)*** *Schematic representation of the mantle plume location beneath East Africa based on seismic tomography[113]. Plume structure at depth is consistent with the distribution of helium isotopic values in (b).* ***(d)*** *Correlation between plume impingement and continental break-up[60]. Black dots display the close relationship between oldest ages of large igneous province (LIP) eruption and oldest ages of break-up. Many rifts have been active (red lines) before flood basalts were emplaced.* ***(e-g)*** *Mean syn-rift velocity evolution (red lines) of major ocean-forming rifts[102]. All rifts show an abrupt increase in extension velocity related to drastic loss of rift strength. Formation of LIPs clearly often postdates onset of rifting but occurs directly before break-up.*

of plume-derived silicate melts, creates anomalously dense pyroxenitic veins in the lithosphere that can lead to foundering or dripping under far-field stresses[116]. Hydrous and carbonated metasomatism common to young rifts in thick cratonic areas are associated with volatiles released from ancient mobile belts in the deep craton[117,118]. Lithospheric thinning that results from foundering, whether driven by brittle or ductile processes, creates locally steep topography along the lithosphere-asthenosphere boundary. The associated rheological and thermal contrast induces small-scale convection at lithospheric edges[119,120] capable of modifying the shape of the craton edge[121], and/or drip melting[116] enhanced by the presence of fusible metasomes in the lithospheric mantle[122,123]. The associated lavas are dominated by melts of metasomatised mantle lithosphere[123,124], often with no discernible asthenospheric component, and erupt as small magma volumes at high velocity to form monogenetic crater fields (e.g., Bufumbira and Toro Ankole, Uganda).

3.2.2 Magmatic rift weakening

Regardless of the mechanism through which magma is generated, it can heat and so weaken the lithosphere. Buoyant melts are thought to ascend by porous flow faster than the mantle upwells, and to accumulate beneath permeability barriers, which in the mantle can closely align with the lithosphere-asthenosphere boundary[125]. This process focuses melt towards regions of thinned lithosphere[126,127] and the latent heat associated with magma intruding near the lithosphere-asthenosphere boundary can drive significant thermal erosion that thins the lithosphere[128] (Fig. 1). If dikes cut through the entire lithosphere, then the lithosphere can be very efficiently heated and weakened (Fig. 3c). Models show that intrusion of relatively small amounts of magma (i.e. less magma than is extruded in LIPs) can heat and thereby weaken the lithosphere by an order of magnitude and thus allow rifting to continue even without continuous magma input[129].

An open question, prompted by data from East Africa and elsewhere, is how tectonic faulting and fairly abundant magmatism can coexist in the same area. Rifts are likely to be places where both the thickness of the lithosphere and the amount of magma available change as the rift evolves[129] so that the pattern of faulting can change in time. In portions of the East African Rift, fieldwork and seismic studies suggest that magma has steadily moved to shallower depths resulting in a change from shallow faulting to shallow magma intrusion with little fault slip[38], or erupted from great depth thus bypassing shallow crustal chambers. Models of extensional processes suggest that there must be periods where no magma is available to fill dikes so that the stress can build to the level needed for fault offset to accumulate[130]. These periods can be far shorter than the lifespan of a fault, but it is not clear what controls the variation in the availability of magma.

3.3 Surface processes

Basin subsidence and rift shoulder uplift[131] are the primary topographic manifestations of continental rifting. Surface processes, i.e., erosion of topographic highs and sedimentation in topographic lows[132], continuously act to level this relief. This redistribution of mass affects both the stress and thermal state of the lithosphere, in a manner that enhances strain localisation. Erosion and sedimentation, therefore, effectively act as additional weakening processes.

The localising effect of surface processes has long been postulated on the basis of numerical simulations. Early numerical models of lithosphere extension coupled with landscape evolution showed that topographic redistribution modulates lower crustal flow by altering lateral pressure gradients[133,134]. Sedimentation (Fig. 2d,e) facilitates this process by warming the geotherm through "thermal blanketing", thereby reducing the effective viscosity of the lower crust. In a mechanical sense however, sedimentation alleviates the dynamic resistance that develops through isostatic thinning of the continental crust[135] (Fig. 3e). In this framework, sedimentation favours the narrow rifting mode, which, for example, characterises deformation in the highly sedimented Gulf of California. In addition, efficient surface processes delay shifts in strain localisation, particularly in weak crust[136,137]. Erosion and sedimentation thus jointly result in longer-lived half-graben faults which accommodate larger offsets prior to being abandoned[138,139]. Being heavily modulated by the Earth's ever-changing climate, surface processes are also inherently transient. Rifts can therefore experience dramatic fluctuations in sedimentation rate[140] and lake levels[52] on times scales of 10s to 100s ky, which can influence fault activity by altering the stress state of the crust[141].

Similar concepts hold true in rifting contexts subjected to both surface processes and magmatic accretion. Intense sedimentation has, for example, been proposed to enhance the characteristic spacing of normal faults at the magmatically robust Andaman Sea spreading centre[142]. Interestingly, lava flows could play a role similar to that of sediments in filling depressions and suppressing relief and so act to focus brittle deformation. Such a mechanism requires very large effusion rates, which are conceivable during the large volcanic events that often accompany continental breakup and result in seaward-thickening igneous wedges that can be imaged today as seaward dipping reflectors near the ocean-continent transition[143,144].

4. Rift propagation and competition

The weakening and resisting processes discussed in the previous sections each apply to a specific range of spatial scales – from centimetres in the case of foliations and kilometre-sized magmatic intrusions up to 100 km width of lithospheric necking. This range of scales also means that small and large-scale strength minima do not always coincide. Rift deformation, therefore, ‘tries out’ weaknesses on multiple scales and generates complex spatio-temporal patterns involving the formation of competing structures, cessation or migration of tectonic activity and reactivation of deformation. On the lithospheric scale, this process is evidenced by branched rifts, where secondary rift segments propagate away from the main rift.

The opening of the North Atlantic was guided by weaknesses inherited from the Appalachian and Caledonian orogens[64]. Reactivation of these inherited structures during northward propagation of extension resulted in a heavily branched rift system. 200 My ago, activity of the North Atlantic rift propagated deformation into a sequence of marginal basins[145] such as the Orphan basin, the Porcupine basin, the Faroe-Shetland basin, and the North Sea basin among others (Fig. 5a-d). These rift arms were separated from the main rift by competent continental ribbons and rotating microcontinents[146,147]. Competition between neighbouring rift branches involves non-linear weakening and strengthening feedbacks (Sec. 2 and 3) that create tipping points which lead to the success of one branch while the other gets abandoned. In the case of the North Atlantic, rift competition took place on two scales: (1) on the basin scale involving the formation of marginal rift basins[145] and (2) on the plate scale, where the rift in the Labrador Sea separating Greenland and North America competed with the northernmost North Atlantic rift. At 100 Ma, the Labrador Sea attracted all of the deformation (Fig. 5b). But when the Iceland plume induced magmatism and GPE gradients between Greenland and Europe, the force balance eventually tipped towards this rift branch[148] (Fig. 5c).

Similar to the North Atlantic opening, the present-day East African Rift features an Eastern and a Western branch that currently compete with each other. The entire rift system is underlain by a large mantle plume anomaly that extends further southward into the lower mantle (Fig. 4c). The Eastern branch however exhibits more extensive magmatism, possibly due to plume deflection by the Tanzania craton[150]. We speculate that melting beneath the Eastern branch induces stronger weakening that can ultimately lead to a decisive advantage during rift competition with the Western branch. Both rift arms in their turn, exhibit several rift basins branching off the main rift, like the Mweru, Nyanza, Eyasi and Pangani Rift (Fig. 5e). These rifts of more than 200 km length can be considered the present-day equivalent to the marginal basins of the North Atlantic.

Rift localisation and propagation often create neighbouring branches that compete with each other. Success of one branch necessarily reduces extensional stress in the other, which leads to rift failure. Weakening can thereby involve internal processes like necking and fault weakening (Fig. 3b-e) but also external processes like mantle plumes (Fig. 4). While one rift evolves further into a mid-ocean ridge, the other remains locked in the last deformation state.

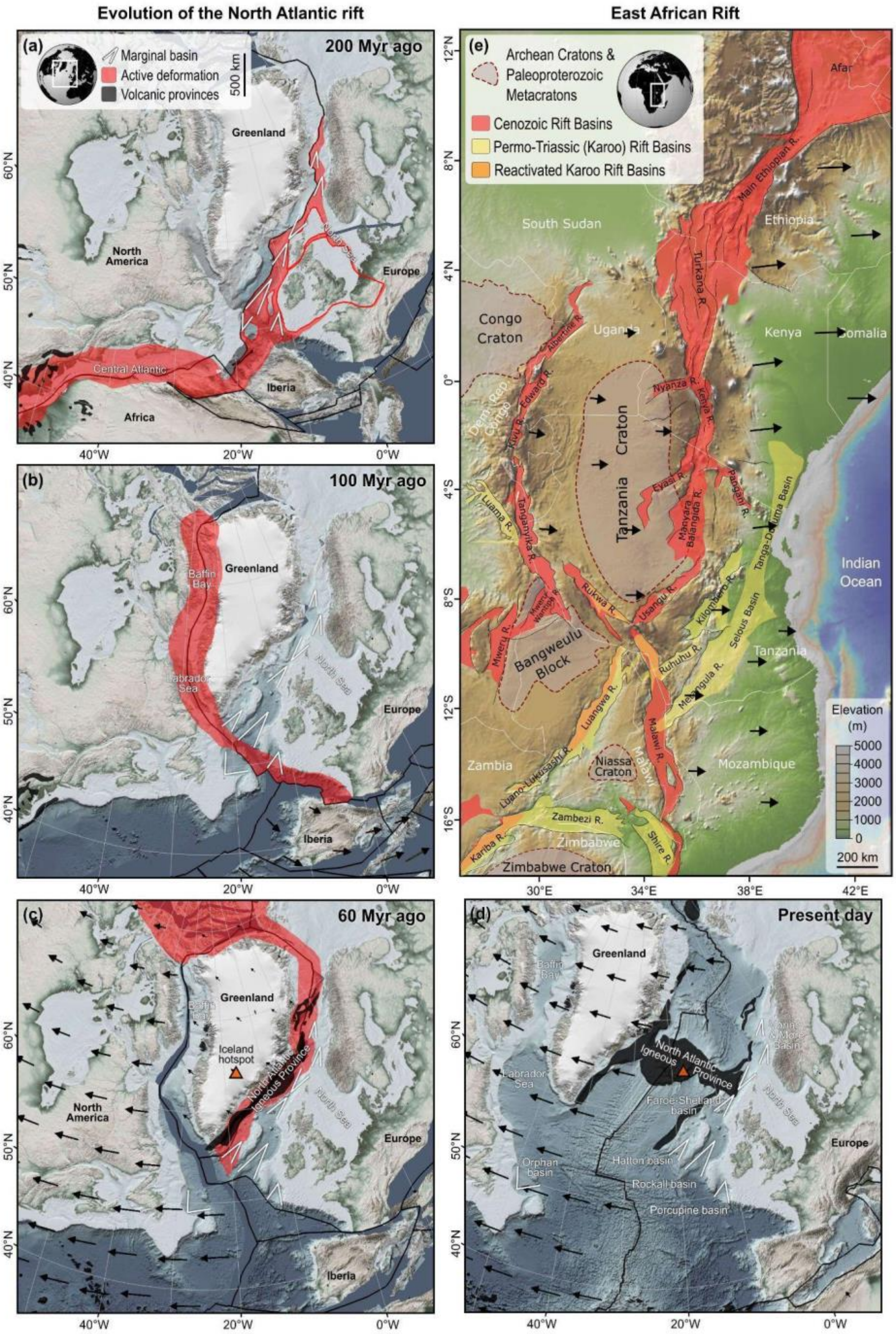


Fig. 5 | Multi-scale rift competition in branched rifts. (a-d) *Plate tectonic history of North Atlantic rifting. The plate reconstruction*[149] *illustrates the protracted rift history that generated a sequence of marginal basins and failed rifts. The dynamics of the northern North Atlantic is dominated by large-scale rift competition between rift branches east and west of Greenland. The success of the branch between Greenland and Europe coincides with the formation of the North Atlantic Igneous Province that is linked to the Iceland plume*[148]*. Plate motions are given relative to a fixed Eurasian plate. Reconstructed present-day topography is shown for orientation.* ***(e)*** *Present-day competition between the Eastern and Western rift branches of the East African Rift. Southern rift segments reactivated some but not all Permo-Triassic rift basins illustrating the limited impact of weaknesses inherited from previous rifting episodes.*

5. Rift resistance and failure

Whether a rift finally generates an ocean basin or extension ceases before achieving continental rupture depends on the competition between strength evolution and driving forces. Failed rifts can form because strengthening processes successively gain impact or if the competition with another rift branch is lost and the driving forces decrease below the strength of the losing branch.

The most impactful process leading to strengthening is lithospheric cooling. It is enabled through conductive heat loss, which increases the viscosity of ductile domains (Fig. 3g) due to their temperature-dependent rheology[151]. The impact of conductive heat loss is thereby regulated by the extension velocity of the rift. Fast rifting such as in the Gulf of Corinth[152] with ~15 mm yr^{-1} or the Woodlark Rift[153] with up to 30 mm yr^{-1} leads to fast heat advection that cannot be counteracted by conductive cooling. High temperatures within the rift lead to rheological weakening and enhance partial melting, both of which decrease rift strength and thus enhance tectonic activity. In slow rifts like the Rhine Graben[154] or the Rio Grande Rift[155] with only ~1 mm yr^{-1} divergence velocity, however, conduction of heat outpaces heat advection[156]. As thinning crust is replaced with rheologically stronger mantle rocks, extensive cooling leads to a gradual increase of rift strength[151,157] so that these rifts will likely evolve into failed rifts eventually.

The aforementioned arguments on the interaction between weakening, strengthening, and driving processes assume an internal force balance that controls rift failure. However, in a branched rift system (Fig. 5), individual rift segments can also compete with each other. If one rift segment is weaker and attracts more deformation, the surrounding stress field is changed such that the extensional force at a competing rift branch gets reduced. Rift competition therefore acts as an external process controlling the success of a rift segment. Such competition, for instance, occurred during the opening of the North Atlantic between the rift basins east and west of Greenland (Fig. 5a-d). Another example is the Cretaceous competition between the Equatorial Atlantic Rift and the West African Rift[158]. The success of the Equatorial Atlantic led to contemporaneous cessation of extension in West Africa, which now constitutes one of the largest failed rifts worldwide stretching over 2000 km from Nigeria to Libya.

The lithospheric weaknesses of failed rifts can be inherited over geological time scales. If plate tectonic changes eventually increase or reorient the tensional forcing, failed rifts can get reactivated. One such example is the Norwegian-Greenland rift, which ceased in the Early Cretaceous after the prominent Møre and Vøring basins off mid-Norway were formed[159] (Fig. 5b). Eventually, extension reactivated in late Cretaceous-Paleocene times adjacent to the original basin that had cooled and strengthened in the meantime[160]. Other examples can be found in the southern East African Rift that partially reactivated Permo-Triassic rift basins in the Rukwa, Karonga and Luangwa rifts (Fig. 5e). In the Turkana rift and other rift segments of northern Kenya, thermochronological data suggest extensional activity in early Cenozoic times[161,162] (60-50 Ma). This phase of rifting, however, ceased, leading to a seemingly failed rift until the latest phase of extension initiated in middle Miocene times (ca. 25-15 Ma). Therefore, failed rifts should actually be considered as dormant rifts that can eventually reactivate once they experience a sufficiently large tensional stress field.

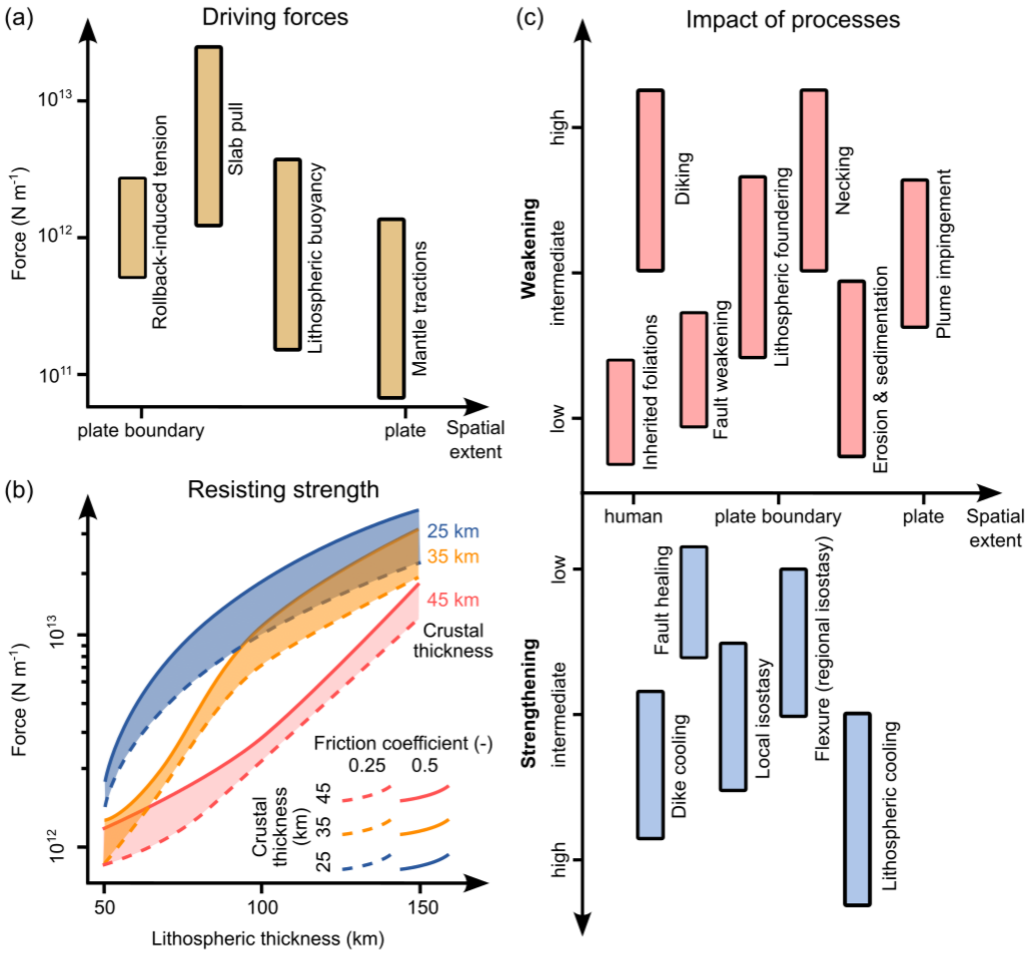


Fig. 6 | Estimates of driving forces, weakening processes and strengthening mechanisms. (a) *Plate tectonic driving forces.* ***(b)*** *Resisting lithospheric strength as a function of the thickness of the lithosphere. Values are computed assuming a conductive thermal steady-state for a range of realistic crustal thicknesses. Solid lines represent unperturbed brittle strength while dashed lines indicate inherited weakness.* ***(c)*** *Estimated relative impact of weakening and strengthening processes.*

5. Societal and environmental relevance of rifts

As a consequence of lithospheric thinning and magmatism, the thermal gradient in rift basins is anomalously high. In conjunction with the extensional stress field of normal fault networks, rifts therefore provide ideal conditions for widespread hydrothermal fluid circulation[163,164] driving the formation of georesources such as geothermal energy and ore deposits. Geothermal energy is presently produced in the Basin and Range[165], the Kenya Rift[166], and the Rhine Graben[167] but large exploration potentials exist in other rifts as well[5]. Considering that millions of people live close to active rifts worldwide, geothermal exploitation has a strong potential for local and sustainable energy generation.

The economic demand for base metals such as copper, lead, and zinc is accelerating due to shifts towards green technologies and continued population growth[168]. Major sediment hosted ore deposits have been discovered in ancient rift settings, particularly at intra-cratonic rifts, failed rifts and rifted continental margins[6]. These tectonic settings provide a favourable environment for metal leaching in deep basin layers, hydrothermally driven upward migration of fluids and finally ore formation within shallow sedimentary strata[169]. Rifts can further host epithermal deposits of precious and base metals[170]. These resources form in shallow parts of high-temperature hydrothermal systems within magmatic rift segments such as the Taupo Rift, New Zealand[171]. Despite recent advances in isolating factors that control the formation of high-grade rift-related ore deposits[172,173], further process understanding is needed for locating new deposits, particularly if they are buried under shallow sedimentary cover.

In addition to georesource formation, fluid flow along rift faults also constitutes a pathway to release deeply sourced CO_2 (Ref. [4]). In particular, the Magadi basin in Southern Kenya exhibits an extremely high CO_2 flux[174] of about 440 t km^{-2} yr^{-1}. The majority of CO_2 is notably released along rift faults and not at volcanoes. However, because of its proximity to the old lithosphere of the Tanzania craton[117,118], the Southern Kenya Rift is likely not representative of rifts in general. Indeed, when comparing regional CO_2 flux densities at rifts where such data is available shows a significant variability: 211 t km^{-2} yr^{-1} in Central Italy[175] , 59 t km^{-2} yr^{-1} in the Taupo Rift[176], New Zealand, and 11 t km^{-2} yr^{-1} in the Eger Rift[177], Czech Republic. It is so far unclear whether this high variability is due to different source regimes, carbon transport pathways or even data density and extrapolation uncertainties. Nevertheless, the global extent of rift systems and likely also the associated CO_2 release was about 3 times larger during Pangaea fragmentation than today[178]. It was hence suggested that ~60% of total CO_2 degassing over the last 200 million years have been derived from rifts; this is 3 times more than that from mid-ocean ridges and volcanic arcs[179]. Rifting can therefore play a major role in global carbon cycling and related climate changes over geological time scales.

6. Summary and future directions

In this review we apply a geodynamic perspective to investigating controls on rift localisation and evolution (Fig. 6). We describe continental rifts as a product of driving and resisting forces, modified by both weakening and strengthening processes that ultimately lead to rift success (the plate breaks) or failure (the rift is abandoned). Two immediate conclusions can be drawn from this perspective: First, that continental rifting is not a steady-state process, but instead rifts evolve non-linearly in phases (Box 1) that can overprint each other as their force equilibrium shifts, and second that rift evolution is a spatial and temporal scale-dependent competition between the driving and resisting forces and their modifiers.

The balance between tectonic driving forces and lithospheric strength is different for each rift and often difficult to quantify, particularly for past rifting phases. But since the dynamic feedback relationships between all involved factors are the same for all rifts, we can nevertheless deduce some overarching rules for the evolution of rifts in general. One such rule is that rifts experiencing plume impingement are very likely to become successful and ultimately form ocean basins[60]. This plume effect results from the simultaneous plume-induced increase of driving force and magma-induced decrease of rift strength. Another general rule is that successful continent-scale rifts feature a prominent abrupt acceleration once the lithosphere is sufficiently weakened[102]. This behaviour marks the transition from a state where the extension velocity is limited by rift strength to a state where the velocity is controlled by the plate-scale force budget. Concerning failed rifts, we infer that lithospheric weaknesses can be inherited over hundred millions of years so that failed rifts actually constitute dormant rifts that possibly reactivate when the tensional driving forces become sufficiently large.

A major challenge results from the vastly different temporal scales involved in breaking a continent. Fault slip and diking events occur on a scale from seconds to months, but achieving continental rupture requires million of years. A key issue here is the relationship between magmatic activity and fault-related extension[180]. It is possible that diking and faulting co-exist in a given rift if inflation of magma chambers[181] occasionally decreases the stress threshold for diking below that of fault slip[130]. In young rifts however, cratonic extension is accompanied by deep-seated carbonatitic volcanism that is fed by magma that traverses the entire lithosphere without creating a crustal magma chamber. In this framework of intertwined volcanic and seismic cycles, magmatic influx and lithospheric thickness play a key role in determining the frequency of eruptions and fault slip events, setting the long-term partitioning of magmatic and tectonic extension. Studying future earthquakes, diking events and aseismic slip as well as their interaction[182,183] can quantify relative magma-tectonic contributions on short time frames. In addition, geochemical characterisation, radiometric dating and geophysical imaging are required to determine the timing, origin and volume of magmatism and its cumulative contribution to rift evolution, while the development of new numerical modelling techniques is essential to bridge these short-and long-term tectono-magmatic processes.

Another cross-scale aspect of rifting relates to the spatial distribution of deformation within fault zones and adjacent fault blocks. Recent studies have revealed a significant amount of off-fault deformation[184,185] leading to differences between geologically determined and geodetically observed slip rates. On a larger scale, intraplate deformation at low strain rates[1,186] can modify the amount and direction of extension experienced by a rift; a combination of high quality campaign or semi-permanent GNSS observations for short time scales and geological indicators of slip at long time scales[187] are required to examine such changes. This undertaking would be particularly interesting during the inception of new rift branches like in the southern termination of the East African Rift, in places where rift segments link[50,66] or where they propagate like in the northern Tanzania divergence[44,188] .

Over geological time scales, continental rifts evolve where and when lithospheric strength is overcome[151,160]. One should, however, keep in mind that strength minima on a regional scale are not necessarily strength minima on a plate scale (Fig. 6c). Rifts 'try out' different strength minima simultaneously, until one system takes over and the other becomes dormant. We argue that the causes of rift abandonment are to be found both in the local interaction of weakening and strengthening processes, as well as, in the force balance of the region. We encourage future studies to search for tipping points in the force balance by studying not only the successful systems, but also the systems that failed.

Future research should aim at identifying the temporal evolution of crustal weakening or strengthening processes, as this will improve our understanding of stress build-up that determines seismic hazard in rift regions. Advancing monitoring techniques of fault-related volatile release are required to quantify the connection between fault strength and the subsurface flow of fluids and volatiles, also allowing for more precise estimates of rift-induced CO_2 degassing. A more profound knowledge of the interaction between fault networks, sedimentary processes and fluid flow will lead to better understanding of rift-related geothermal energy systems and the formation of strategic mineral deposits needed for green energy technologies. Finally, interweaving of modelling and observational efforts is required to bridge the temporal and spatial scales of rift processes and how they evolve from inception to seafloor spreading.

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Author contributions
All authors contributed to the scientific input, writing and editing of the manuscript.

Competing interests
The authors declare no competing interests.

7. Glossary

Basal shear stress: A stress that is imposed by viscous mantle flow at the base of the lithosphere. It is defined by the vertical gradient of horizontal velocity times the viscosity at the lithosphere-asthenosphere boundary.

Core complex: Structures where metamorphosed lower crustal rocks are exhumed to the surface along long-offset normal faults.

Differential stress: The difference between the maximum and minimum principal stresses.

Distal margin: The near-ocean domain of rifted margins. Characterised by thin continental crust, titled continental blocks, regions of exhumed lithospheric mantle or seaward dipping reflectors.

Deviatoric stress: The part of the stress tensor that is related to distortion. The complementary part of the stress tensor that is related to volume change is called hydrostatic stress.

Driving forces: Plate tectonic driving forces result from gravity acting on lateral variations in density. These variations are caused by the topography of Earth's surface or of internal layers and by thermal or compositional heterogeneities like subducting slabs or mantle plumes.

Dynamic topography: The component of Earth's surface topography that is generated by mantle flow.

Exploitation: A process by which extensional brittle rift structures (faults and joints) develop along pre-existing strength anisotropies of a ductile regime that was inherited from an older compressional tectonic event.

Gravitational potential energy (GPE): The energy of an object owing to its position in a gravitational field. GPE gradients constitute a force that emerges due to lateral topography and density variations.

Line force: A force that acts along a line perpendicular to a plate boundary (Unit: N m^{-1}). Line forces are used in simplified concepts where the plate boundary is assumed to be homogeneous along-strike. They can be directly compared to estimates of lithospheric strength.

Lithospheric strength: The vertical integral of the maximum differential stress (i.e., the yield stress) between Earth's surface and the lithosphere-asthenosphere boundary. Unit: N m^{-1}.

Mantle plume: An upwelling in the mantle characterised by higher temperature and lower density relative to the adjacent mantle. Classically depicted with a columnar tail and a mushroom-shape head. The depth of origin of mantle plumes is debated, with the deepest origin placed at the core-mantle-boundary.

Mantle traction: The force per area exerted by mantle flow along the base of a plate. A vector variable with units of stress (MPa).

Metasomatism: The chemical alteration of a rock by hydrothermal and other fluids.

Necking: Localised thinning of the lithosphere. Necking is often accompanied by a drastic reduction in rift strength.

Proximal margin: The near-coastal domains of rifted margins that record the early phases of extension. Often characterised by sedimentary basins and steep normal faults.

Resisting factors: Factors that oppose tectonic deformation. Resistance can be exerted statically or through dynamic processes.

Rifted continental margin: The edge of a continent that was formed by continental extension (as opposed to continental margins shaped by subduction or transform faulting).

Serpentinisation: Serpentinisation is a chemical alteration process of ultramafic rocks where olivine, pyroxene and water react to serpentine minerals. This process affects tectonic deformation by decreasing frictional rock strength and increasing its volume.

Terrane: A crustal fragment that has been broken off its original plate. If accreted to another plate, terranes feature distinctly different properties than adjacent crust owing to their different geological histories.

Weakening processes: Processes that reduce the strength of the lithosphere for instance due to temperature increase, mechanical damage, or increased fluid pressure. Weakening processes are often interlinked by non-linear feedbacks.

Wilson cycle: Represents the concept that the same plate boundaries are involved repeatedly during plate tectonic history, which implies that inherited plate weaknesses persist over geological times. Named after geologist Tuzo Wilson.

Yield: The maximum differential stress that a material can sustain before it deforms by brittle fracture or ductile flow.

Yield strength envelopes: A diagram of the maximum differential stress that the lithosphere can withstand as a function of depth. This stress is a threshold for either brittle failure or ductile flow, depending on which mechanism can be more easily activated as a function of temperature, pressure, deformation rate and lithology. The total strength of the lithosphere (measured in N m^{-1}) is the area enclosed by the strength envelope.

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