nature reviews earth & environment https://doi.org/10.1038/s43017-023-00391-3

Review article Check for updates

Geodynamics of continental rift initiation and evolution

Sascha Brune [1,2] ✉, Folarin Kolawole [3], Jean-Arthur Olive [4], D. Sarah Stamps [5], W. Roger Buck [3], Susanne J. H. Buiter [1,6], Tanya Furman [7] & Donna J. Shillington [8]

Abstract

A continental rift is a nascent plate boundary where the lithosphere is thinned by tectonic activity. Some continental rifts undergo extension to the point that they generate a new ocean basin, whereas others can cease activity altogether. However, the mechanisms that determine rift success or failure remain debated. In this Review, we discuss fundamental rift processes, geodynamic forces and their tectonic interactions and identify the mechanisms that lead to the large variety of rifts on Earth. Rifting initiates through multiscale exploitation of inherited weaknesses, generating dynamic spatiotemporal competition, cessation or localization of rift structures. Progressive thinning of the lithosphere prompts continuous changes in the rift system force balance and prevents a steady-state configuration. Successful continent-scale rifts feature an abrupt and roughly tenfold increase in divergence velocity once the lithosphere is sufficiently weakened. Melt generation during mantle plume impingement can weaken the lithosphere by an order of magnitude, aiding the development of successful rifts. However, at failed rifts, the evolving force balance is dominated by lithospheric strengthening, so that tectonic activity ceases before continental rupture is complete. 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.

Sections

[1]Helmholtz Centre Potsdam — GFZ German Research Centre for Geosciences, Potsdam, Germany. [2]Institute of Geosciences, University of Potsdam, Potsdam-Golm, Germany. [3]Lamont-Doherty Earth Observatory, Columbia University, Palisades, NY, USA. [4]Laboratoire de Géologie, CNRS–Ecole normale supérieure–Paris Sciences & Lettres University, Paris, France. [5]Department of Geosciences, Virginia Tech, Blacksburg, VA, USA. [6]Tectonics and Geodynamics, RWTH Aachen University, Aachen, Germany. [7]Department of Geosciences, Pennsylvania State University, University Park, PA, USA. [8]School of Earth and Sustainability, Northern Arizona University, Flagstaff, AZ, USA. ✉e-mail: brune@gfz-potsdam.de

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 can hence jump or switch between 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.
- 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.
- Mantle plumes simultaneously enhance the forces driving rifting and reduce lithospheric strength by causing magmatic intrusions. Rifts that experience plume impingement appear to always proceed to sea-floor spreading. However, although mantle plumes can aid continental rifting, they are not a requirement, as some rifts proceeded without flood basalt eruptions.
- Rift-related elevated heat flow and permeable normal fault networks facilitate geothermal energy generation and the formation of ore deposits. However, rifts also pose considerable hazards ranging from natural earthquake activity and volcanism to large-scale carbon dioxide degassing.

Introduction

Continental rifts form when the lithosphere thins during tectonic extension. Rift zones manifest through seismic and magmatic activity within a region several tens to hundreds of kilometres 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 have probably affected atmospheric CO_2 concentrations over geological time, particularly during periods of supercontinent break-up[5]. Rift basins hold a strong economical and societal relevance through their geothermal energy potential[6] and as hosts of ore deposits[7,8]. Understanding the links across space and time between the geodynamic processes that control rift initiation and evolution therefore holds relevance for sustainable and safe utilization of rift resources and characterization 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 in the resisting strength can tip the force balance and result in tectonic activity. For example, the impingement of a mantle plume[9] 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[10]. 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[11].

Once a rift has initiated, its dynamics can change markedly 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[12,13]. 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)[14] 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[15]. The largest historical earthquake in central Europe (moment magnitude $M_W \sim 6.5$), for instance, occurred in 1365 in the low-strain Upper Rhine Graben and destroyed the city of Basel, Switzerland[1]. Owing to typically low rift velocities of a few millimetres per year, earthquake recurrence intervals are long (thousands of years), which often lead to limited societal awareness, rendering seismic events particularly devastating.

Although 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 the so-called failed rifts[16]. 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. Rifting stopped in these cases because the strength of the rift eventually exceeded the driving forces, but whether failure resulted from increasing resistance within the rift, waning driving forces or both remains an open question.

In this Review, we focus on key factors and processes affecting rift tectonics and how they evolve over geological timescales, according to various driving forces, resisting factors and weakening processes. Examining the interaction of forces and processes over time is particularly relevant for rifts as they can never achieve a quasi-steady-state configuration that is attainable for plate boundaries such as subduction zones and mature mid-ocean ridges. Instead, at rifts, 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.

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 focused along weak plate boundaries throughout geological history, rather than in continental interiors. Continental rifts, however, are nascent plate boundaries where deformation managed to localize in a previously stable continental interior. As such, resistance has a role equally as important as the driving forces. In the following, we discuss the drivers of extension and how their interplay with static resistance and inherited heterogeneities can guide localization of tectonic deformation during rift initiation (Fig. 1).

Box 1

Conceptual model for the inception and evolution of continental rifts

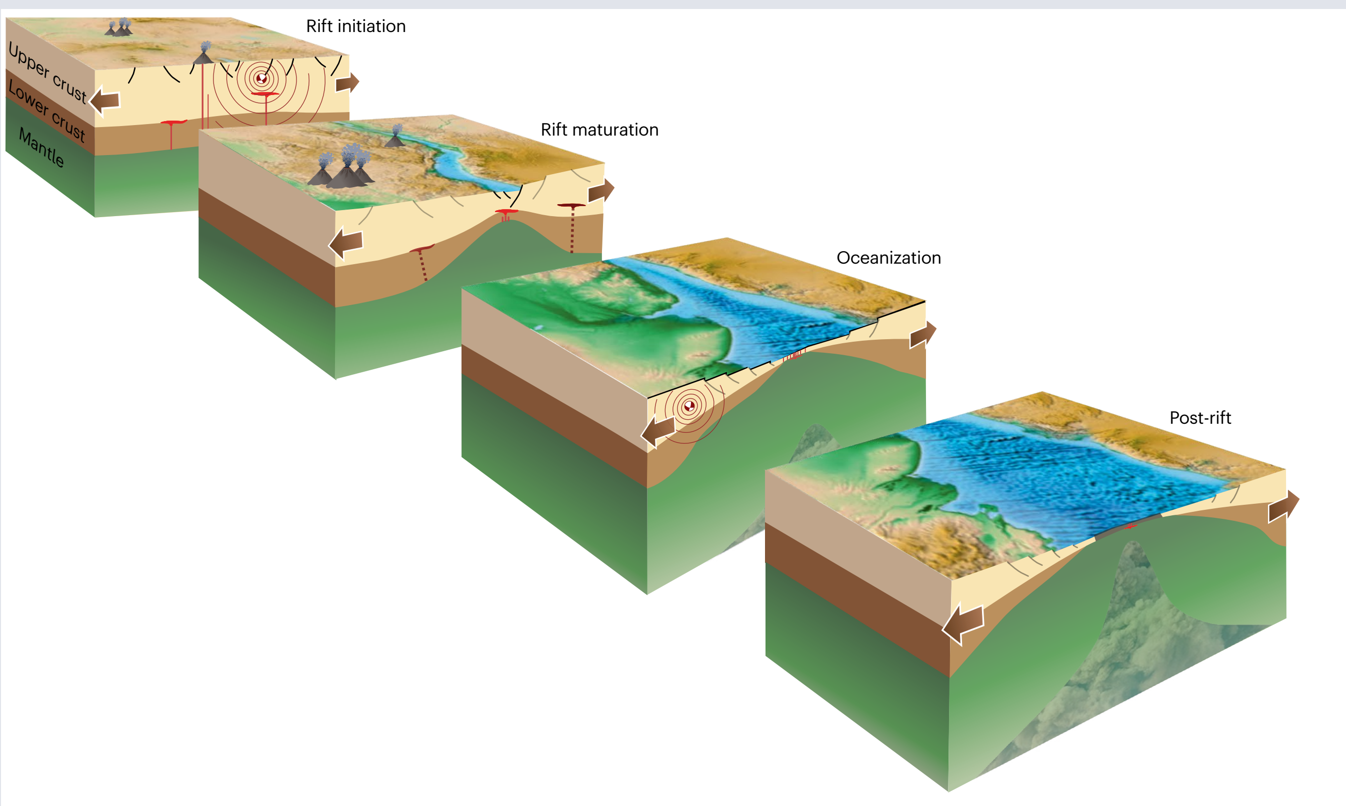
Breaking the continental lithosphere is an intrinsically transient process. Rifting hence progresses continuously, involving several major phases.

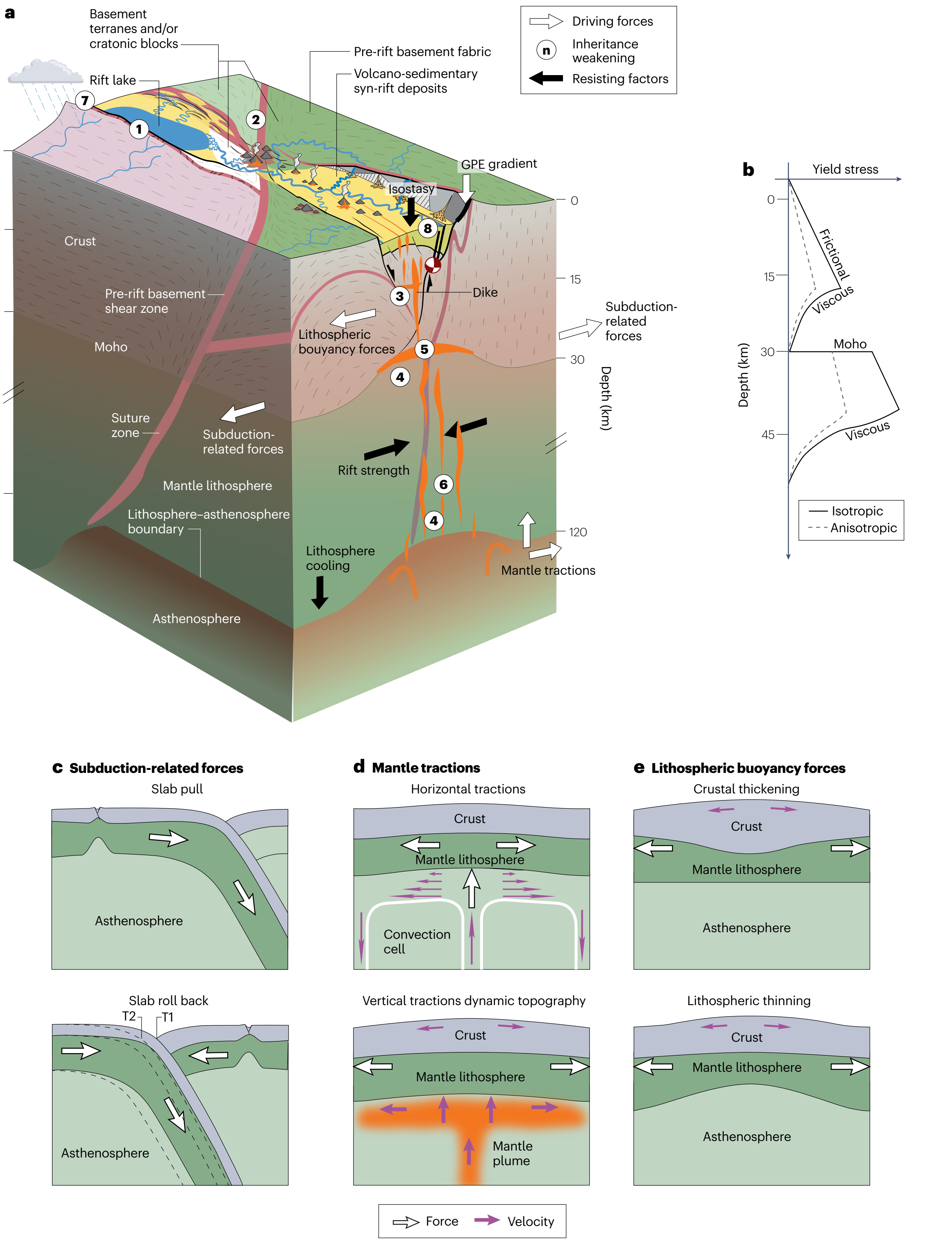
Rift initiation. Rifts localize when tensional stresses exceed the strength of the continental lithosphere. Deformation can be accommodated by brittle faults, ductile shear zones and magmatic dikes. Incipient rifting occurs, for instance, at the Okavango Rift in the south-eastern termination of the East 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 asthenosphere rises in response to lithospheric thinning and causes decompression melting. These melts migrate rapidly through the lithosphere generating dikes, sills and volcanoes.

Oceanization. When the crust is thinned sufficiently, deformation progressively focuses and migrates towards the location of future break-up. 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: proximal margins and distal margins cool at different rates, causing the distal part to subside slightly faster, leading to margin tilting. In contrast, sediments transported by onshore river networks can load the proximal margin inducing subsidence. Additional processes, such as 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.





a
Basement terranes and/or cratonic blocks
Pre-rift basement fabric
Volcano-sedimentary syn-rift deposits
Driving forces
Inheritance weakening
Resisting factors
Rift lake
GPE gradient
Isostasy
Crust
Pre-rift basement shear zone
Moho
Suture zone
Subduction-related forces
Mantle lithosphere
Lithosphere–asthenosphere boundary
Asthenosphere
Dike
Lithospheric bouyancy forces
Subduction-related forces
Rift strength
Lithosphere cooling
Mantle tractions
Depth (km)
0
15
30
120
b
Yield stress
Frictional
Viscous
Moho
Viscous
Depth (km)
0
15
30
45
Isotropic
Anisotropic
c Subduction-related forces
Slab pull
Asthenosphere
Slab roll back
T2
T1
Asthenosphere
d Mantle tractions
Horizontal tractions
Crust
Mantle lithosphere
Convection cell
Vertical tractions dynamic topography
Crust
Mantle lithosphere
Mantle plume
e Lithospheric buoyancy forces
Crustal thickening
Crust
Mantle lithosphere
Asthenosphere
Lithospheric thinning
Crust
Mantle lithosphere
Asthenosphere
Force
Velocity

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. The force balance of a rift is controlled by driving forces, such as subduction-related forces, mantle tractions, buoyancy forces related to gravitational potential energy (GPE) gradients and resisting factors, such as the overall rift strength as well as strengthening processes such as lithosphere cooling. The impact of resisting factors is often alleviated by structural inheritance, where rift segments exploit multiscale inheritance (1) such as a basement shear zone that can segment a rift (2). Weakening processes within faults and shear zones (3) or owing to lithospheric necking and thermal weakening (4) exert decisive control on rift evolution. Other weakening mechanisms derive from melting and shallow melt intrusion (5) as well as metasomatism and lithospheric foundering (6). Exogenic forcing owing to erosion (7) and sedimentation (8) further promotes long-lived faulting. **b**, Strength profile of the crust and lithosphere. The yield stress of pristine, isotopic rocks substantially exceeds the stress that can be sustained by anisotropic fabrics within inherited structures. **c–e**, Simplified representation of major forces that drive continental rifting. **c**, Subduction can contribute major driving forces that directly stretch the subducting plate through slab pull or that extend the upper plate by means of slab roll back. Dashed lines in (part **c**) indicate the future slab position, with T1 and T2 marking the trench positions at time 1 and time 2, respectively. **d**, In the absence of subduction zones, horizontal mantle tractions can directly drive plate divergence, whereas vertical tractions exert indirect influence by generating dynamic topography. **e**, Lithospheric buoyancy forces derive from density heterogeneities within the lithosphere and from topographic gradients, both at the surface of the Earth and the lithosphere–asthenosphere boundary. It is the complex interplay between the depicted rift processes and the geodynamic driving forces that lead to the large variety of rifts on Earth.

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[17]. At the lithosphere scale, the onset of yielding can be conceptualized 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.

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. [18]), 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 subducting plates of the Earth move three to four times faster than plates without attached subducting slabs[19]. How much force subduction can contribute to drive rifting depends, however, on the specific plate tectonic setting.

Intracontinental rifts located on the subducting plate itself can experience tensional stresses that are directly caused by slab pull. An example in which 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[20]. The African continent is separating from Arabia across the Red Sea, whereas the Arabian Plate subducts beneath the Eurasian Plate. The associated slab pull drags the Arabian Plate more forcefully[20], which explains why the Red Sea is rifting about an order of magnitude faster than the rest of East Africa[21]. Another place where rifting is facilitated by subduction is the Woodlark Basin of Papua New Guinea[22]. 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 Ma (ref. [23]).

In some subduction zones, the trench can migrate laterally in the direction opposite to the downgoing motion of the slab. This migration is thought to be the surface manifestation of a retrograde motion of the slab[24], 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[25], such as the Japan Sea, are products of such back-arc rifting[26]. Back-arc rifting does not split the middle of a continent, but breaks off slivers of the margin of the continent, or parts of an island arc. These slivers can later become terranes that eventually attach to another continent during terrane accretion. Many rifts are, however, not connected to subduction zones, and other forces are required to overcome the strength of the plate.

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 in particular.

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[27] (Fig. 1d). However, the magnitude of dynamic topography is difficult to measure owing to incomplete knowledge of asthenosphere viscosity and velocity, and the best ways to estimate it are controversial[28,29]. Straightforward estimates of the amplitude and wavelength of dynamic topography come from oceanic regions where there are good geophysical constraints on isostatic topography. The magnitude of dynamic depth variations in the ocean basins[30] ranges between ±1000 and ±2000 m with maximum values above mantle plumes. Correcting for water loading translates to ±700 and ±1400 m of vertical displacement for continental domains, respectively[30]. 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, owing 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[31,32]. Examples of correlations between rifting and plume impingement include the separation of South America and Africa associated with the Paraná and Etendeka flood basalts ~133 Ma (ref. [33]); the opening of the North Atlantic[34] associated with the North Atlantic Igneous Province between 61 and 56 Ma and the presently active East African Rift System[35]. Although mantle plumes can aid the initiation of continental rifting, they are not a requirement, as, for example, demonstrated by onset of rifting between Iberia and Newfoundland, which occurred 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. As the Pacific plate is the fastest moving plate, it indicates that the drag force is small compared with other forces that could drive plate motions and affect rifting[36]. Horizontal tractions are computed by integrating basal shear stresses over area. If the viscosity beneath the Pacific plate at 100–200 km depths is 10^{19} Pa s, then the shear stress under a plate moving at Pacific plate rates would be in the order of 10^5 Pa, amounting to the total force on the 10,000 km wide Pacific plate of about 1 TN m^{-1}, which is less than a tenth of the slab pull force. However, horizontal tractions do appear to affect plates with thick lithosphere. Plates that include thick cratons move slower across the face of the Earth than plates that do not[37]. This impact of cratons could reflect the increase in mantle viscosity with depth[38], 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 dynamic uplift caused by mantle plumes, such as 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.

Topography and lithospheric structure. Lithospheric buoyancy forces arise from GPE gradients owing to lateral variations in topography and/or lithospheric structure[39] that generate deviatoric stress[40]. 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[41].

In regions where subduction-related forces are negligible, 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[42,43] and lithospheric thickness[44] — but also by vertical mantle tractions[45]. Numerical modelling approaches[46–48] suggest that lithospheric buoyancy forces on the order of 1.5–1.7 TN m^{-1} are the main drivers of the large-scale east-west extension documented by geodetic observations[49]. A similar process drives the broad region of extension of the Basin and Range Province in the western USA[50], with magnitudes of lithospheric buoyancy forces ranging from 0.2 to 2.6 TN m^{-1}. Here, palaeo-elevation 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[51]. The collapse of regions of thick crust has also been suggested as a cause of the broad zones of rifting inferred in West Antarctica[52,53]. These examples illustrate that, in the absence of subduction-related forces, GPE gradients that generate deviatoric tension are important in initiating and maintaining rifting.

We conclude that the most important drivers of rifting are subduction-related forces and GPE gradients. But in many cases, the superposition of available forces appears to be insufficient to overcome the strength of cold, thick continental lithosphere 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.

Exploiting inherited weaknesses

Inherited structures from previous tectonic and/or magmatic events reduce the static lithospheric strength before 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 (Fig. 2). Examples of inherited features include anisotropic rock fabrics — such as foliations (Fig. 2c) and faults from previous 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 multiscale and vary with depth through the lithosphere. At large scales, deformation of the lithosphere is compartmentalized by discrete basement terranes[54,55], with rifts tending to nucleate and develop preferentially in weaker terranes[56]. At smaller scales, new extensional faults (Fig. 2) use favourably oriented fabrics and pre-existing faults.

Rifts are known to develop preferentially in previously deformed lithosphere, as this is now well documented across many Wilson cycles[57,58]. In particular, many rifts develop in orogens created during previous continental collisions, as orogens contain suture zones and other fault systems, and they are often characterized 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. This occurrence 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[59] and the failed Central and West African Rift Systems that formed in the Trans-Saharan Mobile Belt and metacratonized lithosphere. Other examples are the South Atlantic Rift System that reactivated deformation along the Rio Pardo–West Congo Belt[60], the Mid-Continent Rift along the orogenic belts that bound the Superior Craton[61] and the North Atlantic along the Caledonian and Variscan orogens[62].

Within the upper crust, structural inheritance affects the formation of fault patterns as it often leads to the reduction of frictional strength compared with the bulk rock (Fig. 1b). Hence, favourably oriented pre-existing planes of strength contrast will fail and localize deformation before the nucleation of new discontinuities in the pristine portions of the rock[63,64]. This orientation is the reason why major border faults of the East African Rift System reactivated inherited Precambrian structures[67]. However, the exploitation behaviour can 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[65]. Where fabrics or pre-existing structures are not well aligned, faults will cut across them obliquely, sometimes leading to more complex, discontinuous new faults[66,67]. 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[66]. These findings show that structural inheritance can modulate the segmentation of a normal fault network[68,69] and thus has 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

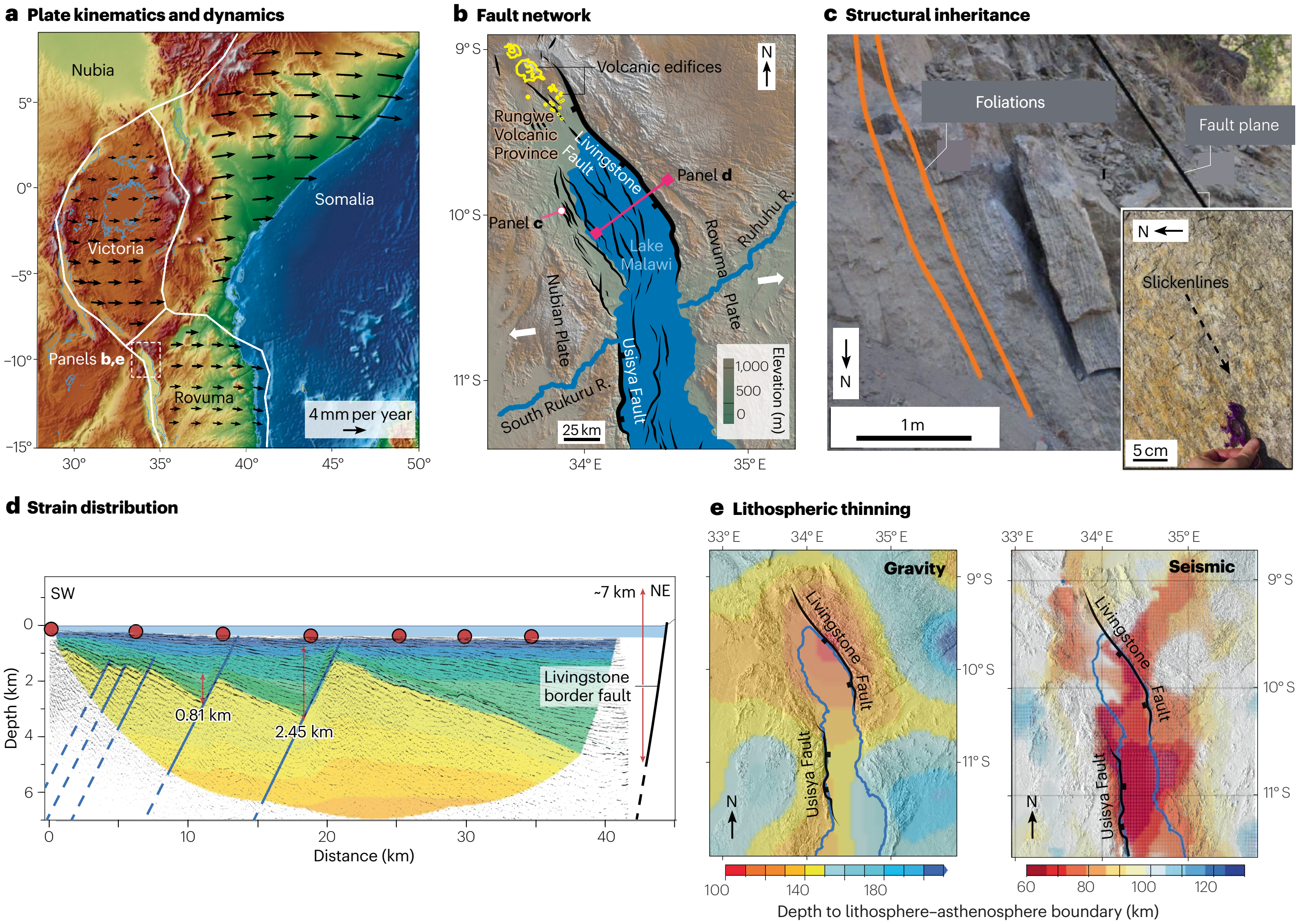


Fig. 2 | Juvenile continental rift structures at the Malawi Rift. a, Topographic map of the East African Rift overlain with plate kinematic data showing plate motions. Area for panel **b** shown with the white box. **b**, Fault network between the Rovuma and Nubian plates, at the north end of Lake Malawi. Extension is accommodated by large border faults (such as the Livingstone and Usisya faults) (bold black lines) and smaller intrabasinal faults[91,184] (thin black lines). **c**, Early-stage rifting commonly features strain localization through exploitation of pre-rift basement structures[65] such as foliations. **d**, Slip accumulation on normal faults (solid black lines) creates spatially variable accommodation space for sediment deposition[139,185] (tilted layers). **e**, Geophysical methods such as gravity (left) and seismic imaging (right) can provide indications of the depth of the lithosphere–asthenosphere boundary. Geophysical data differ in absolute depths of the lithosphere–asthenosphere boundary, but nevertheless show extensive thinning of the lithosphere[186,187]. Below the brittle–ductile transition, extension is accommodated by ductile deformation that thins the lithosphere in an area that can be somewhat wider than the near-surface sediment basin. Active juvenile rifts illustrate the cross-scale interactions between inherited structures and dynamic weakening that control the cessation or localization of rift structures. Panel **c** adapted with permission from ref. [65], Elsevier. Panel **d** adapted from ref. [185], CC BY 4.0 (https://creativecommons.org/licenses/by/4.0/). Part **f** (gravity) adapted from ref. [186], CC BY 4.0 (https://creativecommons.org/licenses/by/4.0/). Panel **f** (seismic) adapted from ref. [187], Springer Nature Limited.

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 at a range of scales is often facilitated by inherited fabrics and structures that are weaker than the surrounding rock and thus localize 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.

Rift weakening

The onset of rifting is governed by yield strength, which is often conceptualized using yield strength envelopes (Fig. 3a). However, the subsequent 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 localize deformation, whereas 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.

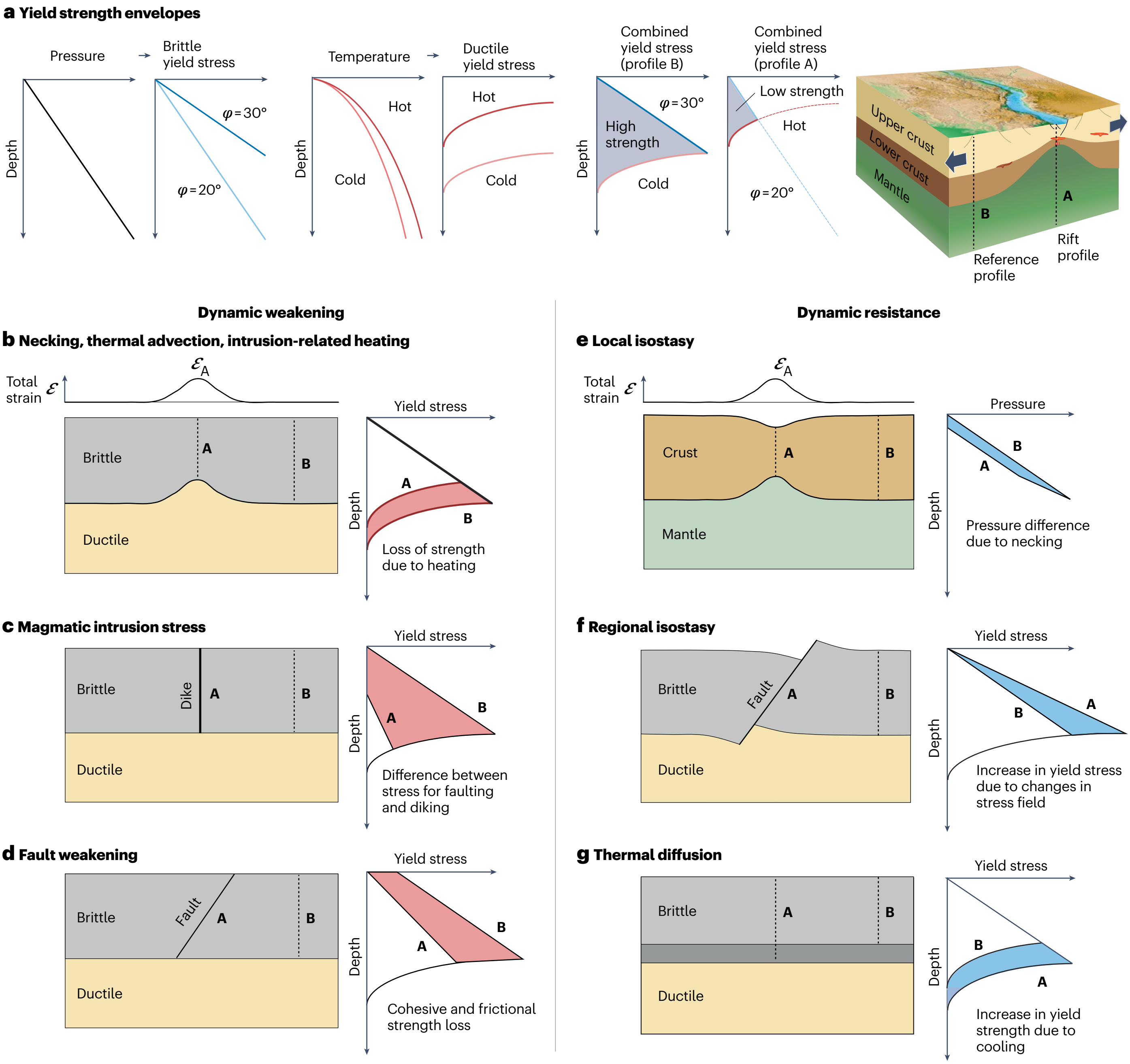


Fig. 3 | Lithospheric strength and response to deformation. Yield strength envelopes illustrate the conceptual strength distribution within the lithosphere with depth and allow for estimating the first-order impact of geodynamic processes on rifting. **a**, Yield strength envelopes are constructed assuming simplified pressure and temperature distributions within homogeneous lithospheric layers. **b**, Thermal weakening owing to necking, thermal advection and magmatic intrusion constitutes a major weakening process. **c**, Dike intrusions can substantially reduce the mechanical stress required to continue rifting. **d**, Chemical and mechanical alteration of fault rocks causes a reduction in cohesion and friction angle of faults. **e**, Isostatic subsidence of thinned crust leads to pressure differences that counteract further thinning. **f**, Fault rotation changes the surrounding stress field, which increases the strength of the fault. **g**, Thermal diffusion cools and strengthens the thinned lithosphere, a process that is particularly important in slow rifts. The relative magnitude of resisting and weakening processes and their interaction with driving forces are highly site-dependent and time-dependent. Panels **b**–**g** adapted with permission from ref. [188], Elsevier.

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. 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 before break-up.

In the brittle upper crust, the growth of faults coincides with a loss of material cohesion and reduction of the friction angle[70]. The latter process is facilitated by the development of a gouge layer where strain-assisted and fluid-assisted reactions can precipitate weak phyllosilicate minerals[71,72], such as serpentine in ultramafic fault rocks[73]. This process requires water to be transported from the surface to the top of the mantle via active faults cutting through the entire crust. Serpentinization has been particularly well documented at the West Iberian rifted margin[74,75] and is thought to occur only during the late stages of rifting when the crust is sufficiently cool to become entirely brittle[76]. The growth of weak minerals, along with elevated pore fluid pressures, causes a reduction in the effective friction coefficient of the fault, lowering the differential 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[77,78]. Very high fluid pressures are, thus, unlikely to have a key role in lowering the brittle strength of rifting lithosphere.

Weakening mechanisms have also been documented in the ductile lower crust[79]. High-temperature, nonlinear ductile creep processes aid strain localization by reducing the effective viscosity of the rock 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[80], making rocks effectively weaker when deforming in the diffusion creep regime[81]. In numerical simulations, the development of a localized, low-viscosity channel in the lower crust enables the position of localized strain in the upper crust to migrate laterally over hundreds of kilometres, which explains the formation of asymmetric (wide and narrow) conjugate margins at the late stages of rifting[82,83].

In order for the aforementioned weakening processes to sustain the localization of strain within a rift zone, they must overcome the action of several key delocalizing phenomena. The first is isostatic adjustment (Fig. 3e), in which 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. Although this force component is usually smaller than the extensional force, it nevertheless opposes continued rifting, which occasionally leads to inversion events within extensional basins[84]. The second is bending of competent lithospheric layers (Fig. 3f), which rotates rift-bounding normal faults to dips less favourable than 60° (refs. [85,86]). This dynamic change happens in a way that more stress is required to maintain fault activity, which ultimately impedes continued deformation[85,87]. Another mechanism is cooling of the lithosphere, which increases the strength of ductile layers by elevating rock viscosity (Fig. 3g). Overall, the competition between localizing and delocalizing processes is strongly modulated by the strength of the rifting lithosphere, which can lead to drastically different modes of continental extension[88].

Rift fault networks pose a considerable seismic threat, especially as 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[89] where earthquakes occur at depths of 40 km or more[15,90]. This deep earthquake depth suggests a predominantly brittle crust and uppermost mantle that are strongly coupled. This configuration promotes a narrow rifting mode, in which 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 (such as, beneath Lake Malawi[91]), they can nucleate earthquakes as large as $M_W \sim 7$.

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[92]. 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 delocalizing effects described earlier that prevent sustained extension in a narrow axis. Wide rifts facilitate the formation of continental core complexes[93], 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 USA[94]. 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 $M_W \sim 6$, as is typical of mid-ocean-ridge settings[95]. If, however, they are able to slip at low angles, the large fault area could accommodate substantially larger earthquakes ($M_W > 7$)[96,97].

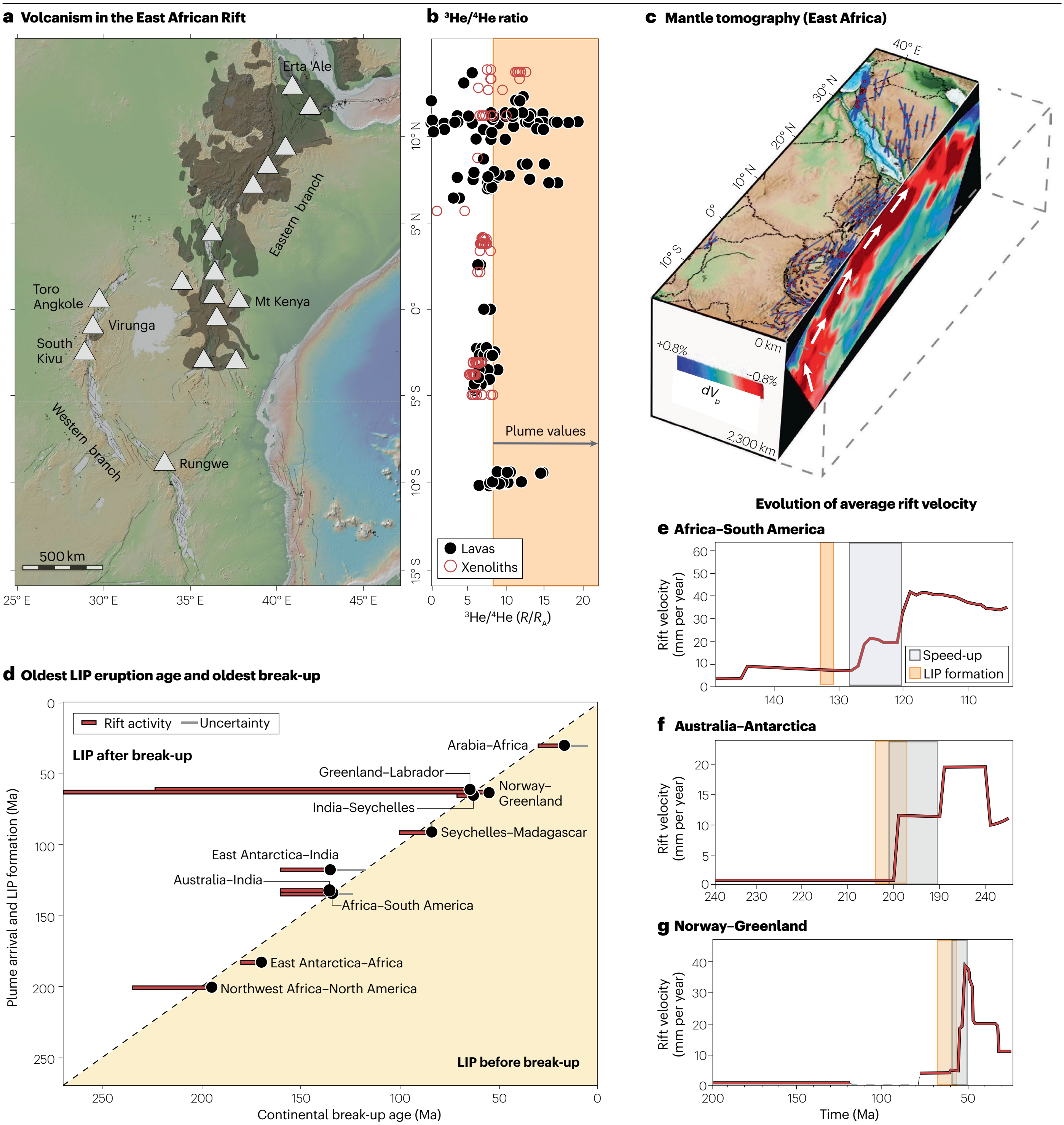
At lithospheric scales, brittle and ductile weakening processes lead to successive focusing of extensional strain within the rift (Box 1). Accumulated thinning, or the so-called necking, of the lithosphere constitutes a large-scale thermal weakening process that replaces cold and strong lithosphere with hot and weak mantle[98]. Numerical models have assessed the relative impact of strain softening versus lithospheric thinning and found that necking dominates the loss of lithospheric strength owing to the highly nonlinear dependence of rock viscosity on temperature[99]. The prominent reduction in lithospheric strength during rift evolution can even generate a feedback on plate kinematics as the loss of the rift strength induces a roughly tenfold acceleration of the involved plates[100], from a few millimetres per year to several centimetres per year (Fig. 4). The strong nonlinearity 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[101]. This process explains the speed-up of North America during Central Atlantic rifting[102], of South America during rifting of the South Atlantic[103] and of Australia during its separation from Antarctica[104].

Melt generation and magmatic weakening

Many major continental break-ups involved extrusion of massive amounts (up to a few billion cubic metres) of basaltic magma in regions called large igneous provinces (LIPs)[105]. Prominent examples are the Parana-Entendeka LIP that occurred close to the centre 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[31]. The question of whether magmatism is a cause or consequence of rifting is hotly debated.

Melt generation in rifts. The generation of magma within the asthenosphere and deep lithosphere of continental rifts can occur in several ways; we describe three mechanisms subsequently.

As stretching thins the lithosphere, the asthenosphere is pulled upwards and can partially melt owing to adiabatic decompression, with the production of higher melt (magma) volumes at greater stretching factors[106]. According to this melting-during-stretching model, large



magma volumes reflect high regional temperatures in the mantle. This model is appealing in its simplicity; although 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, such as the East African Rift[107,108].

Decompression melting can also happen as hot, low-density mantle plumes rise from the lower mantle[109] (Fig. 4c). The plume-driven-melting model for rifting appears viable in many extensional provinces, as dated basalts show that the majority of LIP lavas are extruded after the onset of rifting but before final break-up is achieved. The majority of LIP flood basalts are erupted in less than a million years[34,110], whereas

Fig. 4 | Plume impact on rifting. **a**, East African Rift volcanic fields and major volcanoes[10]. **b**, Mantle plume contributions to volatile release in the East African Rift are depicted in terms of $^{3}He/^{4}He$ ratios in mafic lavas (closed black circles) and mantle xenoliths (open red circles). The shaded region indicates helium isotope ratios from the East African Rift that are enriched compared with normal mid-ocean-ridge basalts. Such elevated ratios are commonly associated with mantle plume impingement. Most analyses are from Miocene and younger samples, but Afar lavas include Oligocene samples, which indicate long-lived plume contributions in this area. **c**, Schematic representation of the mantle plume location beneath East Africa on the basis of seismic tomography[189]. Plume structure at depth is consistent with the distribution of volcanism in panel **a**. **d**, Correlation between the timing of initial plume impingement, inferred by dating lavas of associated large igneous provinces (LIPs), and onset of continental break-up at various rifted margins[58]. The dashed line shows the theoretical relationship if break-up and LIP formation happened simultaneously. In the majority of cases, continental break-up and plume impingement are closely linked. However, many rifts were active (red lines) for variable periods before LIP formation and final break-up occurred. **e**–**g**, Mean syn-rift velocity evolution (red lines) of major ocean-forming rifts[100]. All rifts plotted here show an abrupt increase in extension velocity (grey shading) related to pronounced loss of rift strength after LIP formation (orange shading). The onset of rifting often predates the emplacement of LIPs, but final continental break-up (the formation of a new mid-ocean ridge) clearly post-dates the plume arrival. Plumes weaken the lithosphere and contribute mantle tractions that drive rifting. However, plumes are not a necessity for rift formation as many rifts evolved independently of plume impingement. Panel **c** adapted with permission from ref. [189], Wiley. Panel **d** adapted with permission from ref. [58], Elsevier. Panels **e**–**g** adapted from ref. [100], Springer Nature Limited.

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 such as 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,b). These data therefore support models involving mantle plume contributions since the Oligocene, which agrees with tomographic evidence for a large-scale mantle plume beneath East Africa (Fig. 4c). However, the maximum He isotope ratios measured in lavas and xenoliths from East Africa (19.5 R/R_A) are substantially lower than those observed in lavas from the Iceland or Hawaii plumes (>35 R/R_A)[111]. This discrepancy indicates that plume melts and/or volatiles under East Africa are part of a multicomponent 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 substantially to rift volcanism, particularly in the stages before mid-ocean-ridge style magmatism[112,113]. Silicate metasomatism, which is the transport of plume-derived silicate melts, creates anomalously dense pyroxenitic veins in the lithosphere that can lead to foundering or dripping under far-field stresses[114]. Hydrous and carbonated metasomatism common to young rifts in thick cratonic areas is associated with volatiles released from ancient mobile belts in the deep craton[115,116]. 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[117,118] capable of modifying the shape of the craton edge[119] and/or drip melting[114] enhanced by the presence of fusible metasomes in the lithospheric mantle[120,121]. The associated lavas are dominated by melts of metasomatized mantle lithosphere[121,122], often with no discernible asthenospheric component, and erupt as small magma volumes at high velocity to form monogenetic crater fields (such as Bufumbira and Toro Ankole, Uganda).

Magmatic rift weakening. Regardless of the mechanism through which magma is generated, it can heat and thus 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[123]. This process focuses melt towards regions of thinned lithosphere[124,125], and the latent heat associated with magma intruding near the lithosphere–asthenosphere boundary can drive substantial thermal erosion that thins the lithosphere[126] (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 (for example, 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[127].

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[127], so that the pattern of faulting can change in time. In portions of the East African Rift, results from fieldwork and seismic surveys suggest that magma has steadily moved to shallower depths resulting in a change from shallow faulting to shallow magma intrusion with little fault slip[43], 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[128]. 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.

Surface processes

Basin subsidence and rift shoulder uplift[129] are the primary topographic manifestations of continental rifting. Surface processes, the erosion of topographic highs and sedimentation in topographic lows[130], 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 localization. Erosion and sedimentation, therefore, effectively act as additional weakening processes.

The localizing 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[131,132]. 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[133] (Fig. 3e). In this framework, sedimentation favours the narrow rifting mode, which, for example, characterizes deformation in the highly sedimented Gulf of California. In addition, efficient surface processes delay shifts in strain localization, particularly in weak crust[134,135]. Erosion and sedimentation thus jointly result in longer-lived half-graben faults, which accommodate larger offsets before being abandoned[136,137]. Being heavily modulated by the ever-changing climate of the Earth, surface processes are also inherently transient. Rifts can therefore experience strong fluctuations in sedimentation rate[138] and lake levels[139] on timescales of tens to hundreds thousand years, which can influence fault activity by altering the stress state of the crust[140].

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[141]. Interestingly, lava flows could have a role similar to those of sediments in filling depressions and suppressing relief and acting to focus brittle deformation. Such a mechanism requires very large effusion rates, which are conceivable during the large volcanic events that often accompany continental break-up and result in seaward-thickening igneous wedges, imaged today as seaward dipping reflectors near the ocean–continent transition[142,143].

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, kilometre-sized in the case of magmatic intrusions and up to 100 km width during lithospheric necking. This range of scales also means that small-scale and large-scale strength minima do not always coincide. Rift deformation, therefore, can test out weaknesses on multiple scales, generating complex spatiotemporal 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[62]. Reactivation of these inherited structures during northward propagation of extension resulted in a heavily branched rift system. At 200 Ma, activity of the North Atlantic rift propagated deformation into a sequence of marginal basins[144] 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[145,146]. Competition between neighbouring rift branches involves nonlinear weakening and strengthening feedbacks, which create tipping points that lead to the success of one branch and the abandonment of the other. In the case of the North Atlantic, rift competition took place on two scales: on the basin scale involving the formation of marginal rift basins[144] and 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[147] (Fig. 5c).

Similar to the North Atlantic opening, the present-day East African Rift features an Eastern branch 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 owing to plume deflection by the Tanzania craton[148]. 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 localization 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 such as necking and fault weakening (Fig. 3b–e) and also external processes such as mantle plumes (Fig. 4). Although one rift evolves further into a mid-ocean ridge, the other remains locked in the last deformation state.

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) owing to their temperature-dependent rheology[149]. 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[150] with ~15 mm per year or the Woodlark Rift[151] with up to 30 mm per year 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. However, in slow rifts such as the Rhine Graben[152] or the Rio Grande Rift[153] with only ~1 mm per year divergence velocity, conduction of heat outpaces heat advection[154]. As thinning crust is replaced with rheologically stronger mantle rocks, extensive cooling leads to a gradual increase of rift strength[149,155], so that these rifts will eventually likely evolve into failed rifts.

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[156]. 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 2,000 km from Nigeria to Libya.

The lithospheric weaknesses of failed rifts can be inherited over geological timescales. 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[157] (Fig. 5b). Eventually, extension reactivated in late Cretaceous–Palaeocene times adjacent to the original basin that had cooled and strengthened in the meantime[158]. Other examples can

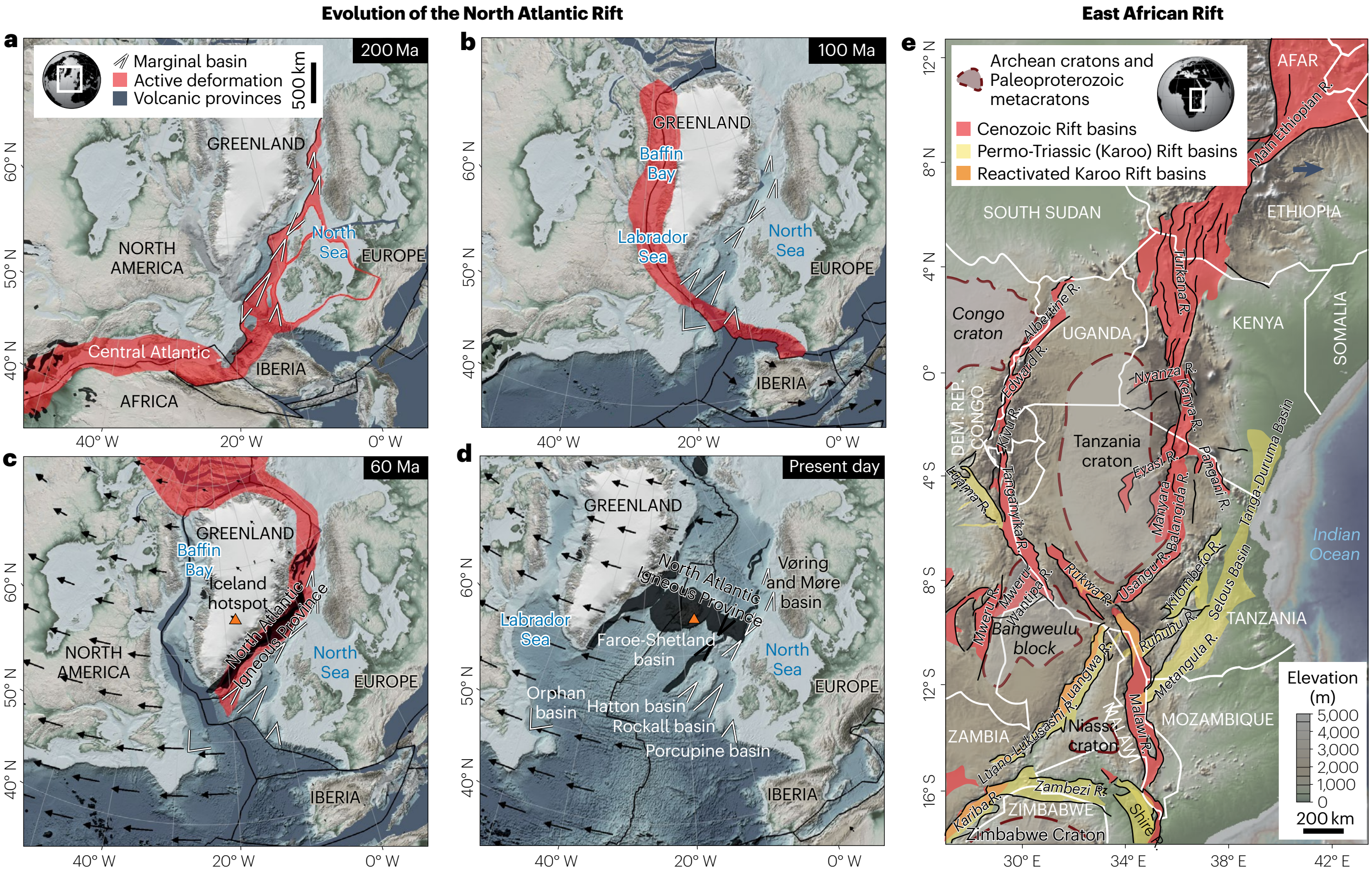


Fig. 5 | Multiscale rift competition in branched rifts. a–d, Plate tectonic history of North Atlantic rifting. Plate reconstructions[190] illustrate the protracted rift history that generated a sequence of marginal basins and failed rifts. The dynamics of the northern North Atlantic is dominated by large-scale rift competition between rift branches east and west of Greenland. The success of the branch between Greenland and Europe (panels **c**,**d**) coincides with the formation of the North Atlantic Igneous Province that is linked to the Iceland plume[147]. 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 impact of weaknesses inherited from previous rifting episodes.

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 (60–50 Ma)[159,160]. This phase of rifting, however, ceased, leading to a seemingly failed rift until the latest phase of extension initiated in middle Miocene times (approximately 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.

Societal and environmental relevance

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[161,162] driving the formation of georesources such as geothermal energy and ore deposits. Geothermal energy is presently produced in major rift basins[163] such as in the Basin and Range, the Kenya Rift and the Rhine Graben but large exploration potentials exist in other rifts as well[6]. 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 owing to shifts towards green technologies and continued population growth[164]. Major sediment-hosted ore deposits have been discovered in ancient rift settings, particularly at intracratonic rifts, failed rifts and rifted continental margins[7]. 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[165]. Rifts can further host epithermal deposits of precious and base metals[166]. These resources form in shallow parts of high-temperature hydrothermal systems within magmatic rift segments such as the Taupo Rift, New Zealand[167]. Despite advances in isolating factors that control the formation of high-grade rift-related ore deposits[168,169], 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[170] of about 440 t km^{-2} per year. 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[115,116], 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 are available shows a substantial variability: 211 t km^{-2} per year in Central Italy[171], 59 t km^{-2} per year in the Taupo Rift[172], New Zealand and 11 t km^{-2} per year in the Eger Rift[173], 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 probably also the associated CO_2 release was about three times larger during Pangaea fragmentation than today[5]. It has been hence suggested that ~60% of the total CO_2 degassing over the past 200 Myr has been derived from rifts; this is three times more than that from mid-ocean ridges and volcanic arcs[174]. As such, rifting has a major role in global carbon cycling and related climate changes over geological timescales.

Summary and future directions

In this Review, a geodynamic perspective has been applied to the investigation of controls on rift localization and evolution (Fig. 6). Continental rifts are a product of driving and resisting forces, modified

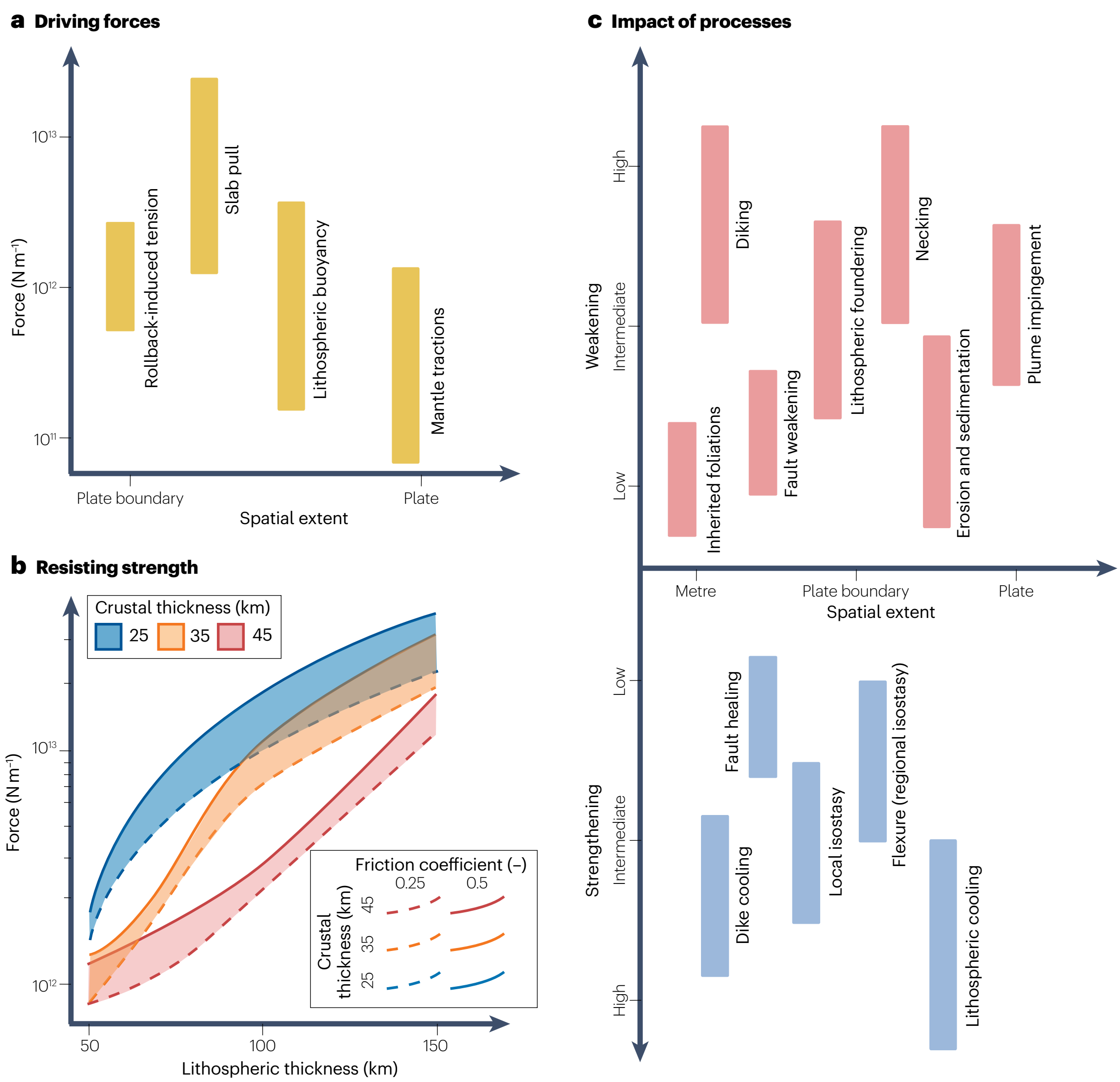


Fig. 6 | Estimates of driving forces, weakening processes and strengthening mechanisms. a, Plate tectonic driving forces estimated for simplified cases. Underlying assumptions and uncertainties are discussed in the main text. **b**, Resisting lithospheric strength as a function of the thickness of the lithosphere. Values are computed analytically by integrating the surface under the yield strength envelope[149] (Fig. 3a) assuming a 1D conductive thermal steady state for a range of realistic friction angles, crustal and lithospheric thicknesses. Solid lines represent unperturbed brittle strength, whereas dashed lines indicate inherited weakness. **c**, Estimated relative impact of weakening and strengthening processes. The given ranges are based on the opinion of the authors accounting for the expected variability between individual cases. Arguments for the magnitude of each process are described in more detail in the main text. For each natural example, the interplay between these factors controls the tectonic evolution of the plate boundary.

Glossary

Basal shear stresses
A stress that is imposed by viscous mantle flow at the base of the lithosphere.

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

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

Distal margins
The near-ocean domain of rifted margins that is characterized by thin continental crust, titled blocks, regions of exhumed mantle or seaward dipping reflectors.

Deviatoric stress
The part of the stress tensor that is related to distortion.

Driving forces
Plate tectonic driving forces result from gravity acting on lateral variations in density.

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

Exploitation
A process by which extensional brittle rift structures (faults and joints) develop along pre-existing strength anisotropies 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 owing to lateral topography and density variations.

Large igneous provinces
(LIPs). Large regions of the crust of the Earth formed by massive volumes of igneous rocks. LIPs are caused by magma generation linked to mantle plumes.

Line force
A force that acts along a line perpendicular to a plate boundary (unit: $N m^{-1}$) and that can be directly compared with estimates of lithospheric strength.

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

Mantle plume
An upwelling in the mantle characterized by high temperature and low density, which is classically depicted with a columnar tail and a mushroom-shaped head.

Mantle tractions
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
Localized thinning of the lithosphere that is often accompanied by a pronounced reduction in rift strength.

Proximal margins
The near-coastal domains of rifted margins that record the early phases of extension, often characterized 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 margins
The edge of a continent that was formed by continental extension (as opposed to continental margins shaped by subduction or transform faulting).

Serpentinization
Serpentinization is a chemical alteration process of ultramafic rocks where olivine, pyroxene and water react to serpentine minerals.

Terranes
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, owing to temperature increase, mechanical damage or increased fluid pressure.

Wilson cycles
Represent the concept that the same plate boundaries are involved repeatedly during plate tectonic history, which implies that inherited plate weaknesses persist over geological times.

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.

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 nonlinearly in phases (Box 1) that can overprint each other as their force equilibrium shifts. 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 as 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[58]. 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[100]. 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.

Major challenges result 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[175]. It is possible that diking and faulting coexist in a given rift if inflation of magma chambers[176] occasionally decreases the stress threshold for diking below that of fault slip[128]. 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 have a key role in determining the frequency of eruptions and fault slip events, which in turn set the long-term partitioning of magmatic and tectonic extension. Monitoring future earthquakes, diking events and aseismic slip as well as modelling their interaction[177,178] can quantify relative magma-tectonic contributions on short time frames. In addition, geochemical characterization, radiometric dating and geophysical imaging are required to determine the timing, origin and volume of magmatism and its cumulative contribution to rift evolution, although the development of new numerical modelling techniques is essential to bridge these short-term 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. A substantial amount of off-fault deformation[179,180] can lead to differences between geologically determined and geodetically observed slip rates. On a larger scale, intraplate deformation at low strain rates[1,181] can modify the amount and direction of extension experienced by a rift. A combination of high-quality campaign or semi-permanent GNSS (Global Navigation Satellite System) observations for short timescales and geological indicators of slip at long timescales[182] are required to examine such changes. This undertaking would be particularly interesting during the inception of new rift branches such as in the southern termination of the East African Rift, in places where rift segments link[64,91] or where they propagate like in the northern Tanzania divergence[49,183].

Over geological timescales, continental rifts evolve where and when lithospheric strength is overcome[149,158]. One should, however, keep in mind that strength minima on a regional scale are not necessarily strength minima on a plate scale (Fig. 6c). Rift branches can emerge by using different strength minima simultaneously. These branches compete 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 and in the force balance of the region. We encourage future research to explore the 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 among 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 sea-floor spreading.

Published online: 07 March 2023

References

1. Poggi, V. et al. Assessing seismic hazard of the East African Rift: a pilot study from GEM and AfricaArray. *Bull. Earthq. Eng.* **15**, 4499–4529 (2017).
2. Biggs, J. et al. Volcanic activity and hazard in the East African Rift Zone. *Nat. Commun.* **12**, 6881 (2021).
3. Dewitte, O. et al. Constraining landslide timing in a data-scarce context: from recent to very old processes in the tropical environment of the North Tanganyika-Kivu Rift region. *Landslides* **18**, 161–177 (2021).
4. Tamburello, G., Pondrelli, S., Chiodini, G. & Rouwet, D. Global-scale control of extensional tectonics on CO_2 earth degassing. *Nat. Commun.* **9**, 4608 (2018).
5. Brune, S., Williams, S. E. & Müller, R. D. Potential links between continental rifting, CO_2 degassing and climate change through time. *Nat. Geosci.* **10**, 941–946 (2017).
6. Jolie, E. et al. Geological controls on geothermal resources for power generation. *Nat. Rev. Earth Environ.* **2**, 324–339 (2021).
7. Leach, D. L. et al. Sediment-hosted lead-zinc deposits in earth history. *Econ. Geol.* **105**, 593–625 (2010).
8. Wilkinson, J. J. in *Treatise on Geochemistry* 2nd edn (ed. Turekian, K.) 219–249 (Elsevier, 2013).
9. Koppers, A. A. P. et al. Mantle plumes and their role in Earth processes. *Nat. Rev. Earth Environ.* **2**, 382–401 (2021).
10. Rooney, T. O. The Cenozoic magmatism of East-Africa: part I — flood basalts and pulsed magmatism. *Lithos* **286**, 264–301 (2017).
11. Dèzes, P., Schmid, S. M. & Ziegler, P. A. Evolution of the European Cenozoic Rift System: interaction of the Alpine and Pyrenean orogens with their foreland lithosphere. *Tectonophysics* **389**, 1–33 (2004).
12. White, R. & McKenzie, D. Magmatism at rift zones: the generation of volcanic continental margins and flood basalts. *J. Geophys. Res. Solid. Earth* **94**, 7685–7729 (1989).
13. Rooney, T. O. The Cenozoic magmatism of East Africa: part V — magma sources and processes in the East African Rift. *Lithos* **360–361**, 105296 (2020).
14. Venzke, E. Global Volcanism Program: volcanoes of the world (v. 5.0.0). *Smithsonian Institution* https://doi.org/10.5479/si.GVP.VOTW5-2022.5.0 (2022).
15. Craig, T. J., Jackson, J. A., Priestley, K. & McKenzie, D. Earthquake distribution patterns in Africa: their relationship to variations in lithospheric and geological structure, and their rheological implications. *Geophys. J. Int.* **185**, 403–434 (2011).
16. Allen, P. A. & Allen, J. R. *Basin Analysis: Principles and Application to Petroleum Play Assessment* (Wiley, 2013).
17. Brace, W. F. & Kohlstedt, D. L. Limits on lithospheric stress imposed by laboratory experiments. *J. Geophys. Res. Solid Earth* **85**, 6248–6252 (1980).
18. Turcotte, D. L. & Schubert, G. *Geodynamics* (Cambridge Univ. Press, 2014).
19. van Summeren, J., Conrad, C. P. & Lithgow-Bertelloni, C. The importance of slab pull and a global asthenosphere to plate motions. *Geochem. Geophys. Geosyst.* **13**, Q0AK03 (2012).
20. Bellahsen, N., Faccenna, C., Funiciello, F., Daniel, J. M. & Jolivet, L. Why did Arabia separate from Africa? Insights from 3-D laboratory experiments. *Earth Planet. Sci. Lett.* **216**, 365–381 (2003).
21. McClusky, S., Reilinger, R., Mahmoud, S., Ben Sari, D. & Tealeb, A. GPS constraints on Africa (Nubia) and Arabia plate motions. *Geophys. J. Int.* **155**, 126–138 (2003).
22. Taylor, B., Goodliffe, A. M. & Martinez, F. How continents break up: insights from Papua New Guinea. *J. Geophys. Res. Solid Earth* **104**, 7497–7512 (1999).
23. Petersen, K. D. & Buck, W. R. Eduction, extension, and exhumation of ultrahigh-pressure rocks in metamorphic core complexes due to subduction initiation. *Geochem. Geophys. Geosyst.* **16**, 2564–2581 (2015).
24. Elsasser, W. M. Sea-floor spreading as thermal convection. *J. Geophys. Res.* **76**, 1101–1112 (1971).
25. Karig, D. E. Origin and development of marginal basins in the western Pacific. *J. Geophys. Res.* **76**, 2542–2561 (1971).
26. Tatsumi, Y., Otofuji, Y.-I., Matsuda, T. & Nohda, S. Opening of the Sea of Japan back-arc basin by asthenospheric injection. *Tectonophysics* **166**, 317–329 (1989).
27. Morgan, W. J. Deep mantle convection plumes and plate motions. *AAPG Bull.* **56**, 203–213 (1972).
28. Molnar, P., England, P. C. & Jones, C. H. Mantle dynamics, isostasy, and the support of high terrain. *J. Geophys. Res. Solid. Earth* **120**, 2014JB011724 (2015).
29. Colli, L., Ghelichkhan, S. & Bunge, H.-P. On the ratio of dynamic topography and gravity anomalies in a dynamic Earth. *Geophys. Res. Lett.* **43**, 2016GL067929 (2016).
30. Hoggard, M. J., White, N. & Al-Attar, D. Global dynamic topography observations reveal limited influence of large-scale mantle flow. *Nat. Geosci.* **9**, 456–463 (2016).
31. Courtillot, V., Jaupart, C., Manighetti, I., Tapponnier, P. & Besse, J. On causal links between flood basalts and continental breakup. *Earth Planet. Sci. Lett.* **166**, 177–195 (1999).
32. Buck, W. R. The role of magma in the development of the Afro-Arabian Rift System. *Geol. Soc. Lond. Spec. Publ.* **259**, 43–54 (2006).
33. Renne, P. R. et al. The age of Paraná Flood Volcanism, rifting of Gondwanaland, and the Jurassic–Cretaceous boundary. *Science* **258**, 975–979 (1992).
34. Courtillot, V. E. & Renne, P. R. On the ages of flood basalt events. *Comptes Rendus Geosci.* **335**, 113–140 (2003).
35. Mulibo, G. D. & Nyblade, A. A. Mantle transition zone thinning beneath eastern Africa: evidence for a whole-mantle superplume structure. *Geophys. Res. Lett.* **40**, 3562–3566 (2013).

36. Richter, F. & McKenzie, D. Simple plate models of mantle convection. *J. Geophys.* **44**, 441–471 (1977).
37. Argus, D. F., Gordon, R. G. & DeMets, C. Geologically current motion of 56 plates relative to the no-net-rotation reference frame. *Geochem. Geophys. Geosyst.* **12**, Q11001 (2011).
38. Yoshida, M. & Yoshizawa, K. Continental drift with deep cratonic roots. *Annu. Rev. Earth Planet. Sci.* **49**, 117–139 (2021).
39. Coblentz, D. D., Richardson, R. M. & Sandiford, M. On the gravitational potential of the Earth's lithosphere. *Tectonics* **13**, 929–945 (1994).
40. Bott, M. H. P. The stress regime associated with continental break-up. *Geol. Soc. Lond. Spec. Publ.* **68**, 125–136 (1992).
41. Zoback, M. L. & Mooney, W. D. Lithospheric buoyancy and continental intraplate stresses. *Int. Geol. Rev.* **45**, 95–118 (2003).
42. Globig, J. et al. New insights into the crust and lithospheric mantle structure of Africa from elevation, geoid, and thermal analysis. *J. Geophys. Res. Solid Earth* **121**, 2016JB012972 (2016).
43. Ebinger, C. J. et al. Crustal structure of active deformation zones in Africa: implications for global crustal processes. *Tectonics* **36**, 3298–3332 (2017).
44. Afonso, J. C. et al. Thermochemical structure and evolution of cratonic lithosphere in central and southern Africa. *Nat. Geosci.* **15**, 405–410 (2022).
45. Kendall, J.-M. & Lithgow-Bertelloni, C. Why is Africa rifting? *Geol. Soc. Lond. Spec. Publ.* **420**, SP420.17 (2016).
46. Moucha, R. & Forte, A. M. Changes in African topography driven by mantle convection. *Nat. Geosci.* **4**, 707–712 (2011).
47. Stamps, D. S., Flesch, L. M., Calais, E. & Ghosh, A. Current kinematics and dynamics of Africa and the East African rift system. *J. Geophys. Res. Solid Earth* **119**, 5161–5186 (2014).
48. Rajaonarison, T. A., Stamps, D. S. & Naliboff, J. Role of lithospheric buoyancy forces in driving deformation in East Africa from 3D geodynamic modeling. *Geophys. Res. Lett.* **48**, e2020GL090483 (2021).
49. Stamps, D. S., Kreemer, C., Fernandes, R., Rajaonarison, T. A. & Rambolamanana, G. Redefining East African rift system kinematics. *Geology* **49**, 150–155 (2021).
50. Jones, C. H., Unruh, J. R. & Sonder, L. J. The role of gravitational potential energy in active deformation in the southwestern United States. *Nature* **381**, 37–41 (1996).
51. Jones, C. H., Sonder, L. J. & Unruh, J. R. Lithospheric gravitational potential energy and past orogenesis: implications for conditions of initial Basin and Range and Laramide deformation. *Geology* **26**, 639–642 (1998).
52. Huerta, A. D. & Harry, D. L. The transition from diffuse to focused extension: modeled evolution of the West Antarctic rift system. *Earth Planet. Sci. Lett.* **255**, 133–147 (2007).
53. Bialas, R. W., Buck, W. R., Studinger, M. & Fitzgerald, P. G. Plateau collapse model for the transantarctic mountains–West antarctic rift system: insights from numerical experiments. *Geology* **35**, 687–690 (2007).
54. Kirkpatrick, J. D. et al. Scale-dependent influence of pre-existing basement shear zones on rift faulting: a case study from NE Brazil. *J. Geol. Soc.* **170**, 237–247 (2013).
55. Hodge, M., Fagereng, Å., Biggs, J. & Mdala, H. Controls on early-rift geometry: new perspectives from the Bilila-Mtakataka Fault, Malawi. *Geophys. Res. Lett.* **45**, 3896–3905 (2018).
56. Petersen, K. D. & Schiffer, C. Wilson cycle passive margins: control of orogenic inheritance on continental breakup. *Gondwana Res.* **39**, 131–144 (2016).
57. Wilson, J. T. Did the Atlantic close and then re-open? *Nature* **211**, 676–681 (1966).
58. Buiter, S. J. H. & Torsvik, T. H. A review of Wilson cycle plate margins: a role for mantle plumes in continental break-up along sutures? *Gondwana Res.* **26**, 627–653 (2014).
59. Daly, M. C., Chorowicz, J. & Fairhead, J. D. Rift basin evolution in Africa: the influence of reactivated steep basement shear zones. *Geol. Soc. Lond. Spec. Publ.* **44**, 309–334 (1989).
60. Maurin, J.-C. & Guiraud, R. Basement control in the development of the early cretaceous West and Central African rift system. *Tectonophysics* **228**, 81–95 (1993).
61. Stein, S. et al. Insights from North America's failed Midcontinent Rift into the evolution of continental rifts and passive continental margins. *Tectonophysics* **744**, 403–421 (2018).
62. Schiffer, C. et al. Structural inheritance in the North Atlantic. *Earth-Sci. Rev.* **206**, 102975 (2020).
63. Osagiede, E. E. et al. Influence of zones of pre-existing crustal weakness on strain localization and partitioning during rifting: insights from analogue modeling using high resolution 3D digital image correlation. *Tectonics* **40**, e2021TC006970 (2021).
64. Kolawole, F., Phillips, T. B., Atekwana, E. A. & Jackson, C. A.-L. Structural inheritance controls strain distribution during early continental rifting, Rukwa rift. *Front. Earth Sci.* **9**, 670 (2021).
65. Dawson, S., Lao-Davila, D., Atekwana, E. & Salam, M. A. The influence of the Precambrian Mughese Shear Zone structures on strain accommodation in the Northern Malawi Rift. *Tectonophysics* **722**, 53–68 (2018).
66. Morley, C. K. Stress re-orientation along zones of weak fabrics in rifts: an explanation for pure extension in 'oblique' rift segments? *Earth Planet. Sci. Lett.* **297**, 667–673 (2010).
67. Kolawole, F. et al. Active deformation of Malawi Rift's North Basin Hinge Zone modulated by reactivation of preexisting Precambrian shear zone fabric. *Tectonics* **37**, 683–704 (2018).
68. Laó-Dávila, D. A., Al-Salmi, H. S., Abdelsalam, M. G. & Atekwana, E. A. Hierarchical segmentation of the Malawi Rift: the influence of inherited lithospheric heterogeneity and kinematics in the evolution of continental rifts. *Tectonics* **34**, 2399–2417 (2015).
69. Manatschal, G., Lavier, L. & Chenin, P. The role of inheritance in structuring hyperextended rift systems: some considerations based on observations and numerical modeling. *Gondwana Res.* **27**, 140–164 (2015).
70. Byerlee, J. Friction of rocks. *Pure Appl. Geophys.* **116**, 615–626 (1978).
71. Escartín, J., Hirth, G. & Evans, B. Effects of serpentinization on the lithospheric strength and the style of normal faulting at slow-spreading ridges. *Earth Planet Sci. Lett.* **151**, 181–189 (1997).
72. Tesei, T., Collettini, C., Carpenter, B. M., Viti, C. & Marone, C. Frictional strength and healing behavior of phyllosilicate-rich faults. *J. Geophys. Res. Solid Earth* **117**, B09402 (2012).
73. Viti, C., Collettini, C., Tesei, T., Tarling, M. S. & Smith, S. A. F. Deformation processes, textural evolution and weakening in retrograde serpentinites. *Minerals* **8**, 241 (2018).
74. Whitmarsh, R. B., Manatschal, G. & Minshull, T. A. Evolution of magma-poor continental margins from rifting to seafloor spreading. *Nature* **413**, 150–154 (2001).
75. Bayrakci, G. et al. Fault-controlled hydration of the upper mantle during continental rifting. *Nat. Geosci.* **9**, 384–388 (2016).
76. Liu, Z. et al. Lateral coexistence of ductile and brittle deformation shapes magma-poor distal margins: an example from the West Iberia-Newfoundland margins. *Earth Planet. Sci. Lett.* **578**, 117288 (2022).
77. Collettini, C. & Sibson, R. H. Normal faults, normal friction? *Geology* **29**, 927–930 (2001).
78. Yuan, X. P., Olive, J.-A. & Braun, J. Partially locked low-angle normal faults in cohesive upper crust. *Tectonics* **39**, e2019TC005753 (2020).
79. Bürgmann, R. & Dresen, G. Rheology of the lower crust and upper mantle: evidence from rock mechanics, geodesy, and field observations. *Annu. Rev. Earth Planet. Sci.* **36**, 531–567 (2008).
80. Ruh, J., Tokle, L. & Behr, W. Grain-size-evolution controls on lithospheric weakening during continental rifting. *Nat. Geosci.* **15**, 585–590 (2022).
81. Hirth, G. & Kohlstedt, D. L. Rheology of the upper mantle and the mantle wedge: a view from the experimentalists. *Geophys. Monogr.* **138**, 83–105 (2003).
82. Brune, S., Heine, C., Clift, P. D. & Pérez-Gussinyé, M. Rifted margin architecture and crustal rheology: reviewing Iberia-Newfoundland, Central South Atlantic, and South China Sea. *Mar. Pet. Geol.* **79**, 257–281 (2017).
83. Tetreault, J. L. & Buiter, S. J. H. The influence of extension rate and crustal rheology on the evolution of passive margins from rifting to break-up. *Tectonophysics* **746**, 155–172 (2018).
84. Le Gall, B., Vétel, W. & Morley, C. K. Inversion tectonics during continental rifting: the Turkana Cenozoic rifted zone, northern Kenya. *Tectonics* **24**, TC2002 (2005).
85. Lavier, L. L., Buck, W. R. & Poliakov, A. N. B. Factors controlling normal fault offset in an ideal brittle layer. *J. Geophys. Res.* **105**, 23431–23442 (2000).
86. Olive, J.-A. & Behn, M. D. Rapid rotation of normal faults due to flexural stresses: an explanation for the global distribution of normal fault dips. *J. Geophys. Res. Solid Earth* **119**, 2013JB010512 (2014).
87. Buck, W. R. Effect of lithospheric thickness on the formation of high- and low-angle normal faults. *Geology* **21**, 933–936 (1993).
88. Buck, W. R. Modes of continental lithospheric extension. *J. Geophys. Res. Solid Earth* **96**, 20161–20178 (1991).
89. Albaric, J., Déverchère, J., Petit, C., Perrot, J. & Le Gall, B. Crustal rheology and depth distribution of earthquakes: insights from the central and southern East African Rift system. *Tectonophysics* **468**, 28–41 (2009).
90. Lavayssière, A. et al. Depth extent and kinematics of faulting in the Southern Tanganyika Rift, Africa. *Tectonics* **38**, 842–862 (2019).
91. Ebinger, C. J. et al. Kinematics of active deformation in the Malawi Rift and Rungwe Volcanic Province, Africa. *Geochem. Geophys. Geosyst.* **20**, 3928–3951 (2019).
92. Hamilton, W. Crustal extension in the Basin and Range Province, southwestern United States. *Geol. Soc. Lond. Spec. Publ.* **28**, 155–176 (1987).
93. Whitney, D. L., Teyssier, C., Rey, P. & Buck, W. R. Continental and oceanic core complexes. *Geol. Soc. Am. Bull.* **125**, 273–298 (2013).
94. Gans, P. B. & Gentry, B. J. Dike emplacement, footwall rotation, and the transition from magmatic to tectonic extension in the Whipple Mountains metamorphic core complex, southeastern California. *Tectonics* **35**, 2016TC004215 (2016).
95. Olive, J. & Escartín, J. Dependence of seismic coupling on normal fault style along the Northern Mid-Atlantic Ridge. *Geochem. Geophys. Geosyst.* **17**, 4128–4152 (2016).
96. Wernicke, B. Low-angle normal faults and seismicity: a review. *J. Geophys. Res. Solid Earth* **100**, 20159–20174 (1995).
97. Biemiller, J., Gabriel, A.-A. & Ulrich, T. The dynamics of unlikely slip: 3D modeling of low-angle normal fault rupture at the Mai'iu Fault, Papua New Guinea. *Geochem. Geophys. Geosyst.* **23**, e2021GC010298 (2022).
98. Chenin, P., Schmalholz, S. M., Manatschal, G. & Karner, G. D. Necking of the lithosphere: a reappraisal of basic concepts with thermo-mechanical numerical modeling. *J. Geophys. Res. Solid Earth* **123**, 5279–5299 (2018).
99. Brune, S., Popov, A. A. & Sobolev, S. V. Modeling suggests that oblique extension facilitates rifting and continental break-up. *J. Geophys. Res.* **117**, B08402 (2012).
100. Brune, S., Williams, S. E., Butterworth, N. P. & Müller, R. D. Abrupt plate accelerations shape rifted continental margins. *Nature* **536**, 201–204 (2016).
101. Ulvrova, M. M., Brune, S. & Williams, S. Breakup without borders: how continents speed up and slow down during rifting. *Geophys. Res. Lett.* **46**, 1338–1347 (2019).

102. Kneller, E. A., Johnson, C. A., Karner, G. D., Einhorn, J. & Queffelec, T. A. Inverse methods for modeling non-rigid plate kinematics: application to mesozoic plate reconstructions of the Central Atlantic. *Comput. Geosci.* **49**, 217–230 (2012).
103. Heine, C., Zoethout, J. & Müller, R. D. Kinematics of the South Atlantic rift. *Solid Earth* **4**, 215–253 (2013).
104. Whittaker, J. M., Williams, S. E. & Müller, R. D. Revised tectonic evolution of the Eastern Indian Ocean. *Geochem. Geophys. Geosyst.* **14**, 1891–1909 (2013).
105. Ernst, R. E. *Large Igneous Provinces* (Cambridge Univ. Press, 2014).
106. McKenzie, D. & Bickle, M. J. The volume and composition of melt generated by extension of the lithosphere. *J. Petrol.* **29**, 625–679 (1988).
107. Ebinger, C. J. & Ibrahim, A. Multiple episodes of rifting in Central and East Africa: a re-evaluation of gravity data. *Geol. Rundsch.* **83**, 689–702 (1994).
108. Hendrie, D. B., Kusznir, N. J., Morley, C. K. & Ebinger, C. J. Cenozoic extension in northern Kenya: a quantitative model of rift basin development in the Turkana region. *Tectonophysics* **236**, 409–438 (1994).
109. Campbell, I. H. & Griffiths, R. W. Implications of mantle plume structure for the evolution of flood basalts. *Earth Planet. Sci. Lett.* **99**, 79–93 (1990).
110. Blackburn, T. J. et al. Zircon U-Pb geochronology links the end-Triassic extinction with the Central Atlantic Magmatic Province. *Science* **340**, 941–945 (2013).
111. Jackson, M. G., Konter, J. G. & Becker, T. W. Primordial helium entrained by the hottest mantle plumes. *Nature* **542**, 340–343 (2017).
112. Schilling, J.-G., Kingsley, R. H., Hanan, B. B. & McCully, B. L. Nd-Sr-Pb isotopic variations along the Gulf of Aden: evidence for Afar mantle plume–continental lithosphere interaction. *J. Geophys. Res. Solid. Earth* **97**, 10927–10966 (1992).
113. Rooney, T., Furman, T., Bastow, I., Ayalew, D. & Yirgu, G. Lithospheric modification during crustal extension in the Main Ethiopian Rift. *J. Geophys. Res. Solid Earth* **112**, B10201 (2007).
114. Furman, T., Nelson, W. R. & Elkins-Tanton, L. T. Evolution of the East African rift: drip magmatism, lithospheric thinning and mafic volcanism. *Geochim. Cosmochim. Acta* **185**, 418–434 (2016).
115. Foley, S. F. & Fischer, T. P. An essential role for continental rifts and lithosphere in the deep carbon cycle. *Nat. Geosci.* **10**, 897–902 (2017).
116. Muirhead, J. D. et al. Displaced cratonic mantle concentrates deep carbon during continental rifting. *Nature* **582**, 67–72 (2020).
117. King, S. D. & Anderson, D. L. Edge-driven convection. *Earth Planet. Sci. Lett.* **160**, 289–296 (1998).
118. Njinju, E. A., Stamps, D. S., Neumiller, K. & Gallager, J. Lithospheric control of melt generation beneath the Rungwe volcanic province, East Africa: implications for a plume source. *J. Geophys. Res. Solid. Earth* **126**, e2020JB020728 (2021).
119. Currie, C. A. & van Wijk, J. How craton margins are preserved: insights from geodynamic models. *J. Geodyn.* **100**, 144–158 (2016).
120. Rooney, T. O., Nelson, W. R., Dosso, L., Furman, T. & Hanan, B. The role of continental lithosphere metasomes in the production of HIMU-like magmatism on the northeast African and Arabian plates. *Geology* **42**, 419–422 (2014).
121. Pitcavage, E., Furman, T., Nelson, W. R., Kalegga, P. K. & Barifaijo, E. Petrogenesis of primitive lavas from the Toro Ankole and Virunga Volcanic Provinces: metasomatic mineralogy beneath East Africa's Western Rift. *Lithos* **396–397**, 106192 (2021).
122. Rosenthal, A., Foley, S. F., Pearson, D. G., Nowell, G. M. & Tappe, S. Petrogenesis of strongly alkaline primitive volcanic rocks at the propagating tip of the western branch of the East African Rift. *Earth Planet. Sci. Lett.* **284**, 236–248 (2009).
123. Sparks, D. W. & Parmentier, E. M. Melt extraction from the mantle beneath spreading centers. *Earth Planet. Sci. Lett.* **105**, 368–377 (1991).
124. McKenzie, D. The generation and compaction of partially molten rock. *J. Petrol.* **25**, 713–765 (1984).
125. Sleep, N. H. Lateral flow of hot plume material ponded at sublithospheric depths. *J. Geophys. Res. Solid Earth* **101**, 28065–28083 (1996).
126. Havlin, C., Parmentier, E. M. & Hirth, G. Dike propagation driven by melt accumulation at the lithosphere–asthenosphere boundary. *Earth Planet. Sci. Lett.* **376**, 20–28 (2013).
127. Bialas, R. W., Buck, W. R. & Qin, R. How much magma is required to rift a continent? *Earth Planet. Sci. Lett.* **292**, 68–78 (2010).
128. Olive, J.-A. & Dublanchet, P. Controls on the magmatic fraction of extension at mid-ocean ridges. *Earth Planet. Sci. Lett.* **549**, 116541 (2020).
129. Weissel, J. K. & Karner, G. D. Flexural uplift of rift flanks due to mechanical unloading of the lithosphere during extension. *J. Geophys. Res. Solid Earth* **94**, 13919–13950 (1989).
130. Leeder, M. R. & Gawthorpe, R. L. Sedimentary models for extensional tilt-block/half-graben basins. *Geol. Soc. Lond. Spec. Publ.* **28**, 139–152 (1987).
131. Burov, E. & Cloetingh, S. Erosion and rift dynamics: new thermomechanical aspects of post-rift evolution of extensional basins. *Earth Planet. Sci. Lett.* **150**, 7–26 (1997).
132. Burov, E. & Poliakov, A. Erosion and rheology controls on synrift and postrift evolution: verifying old and new ideas using a fully coupled numerical model. *J. Geophys. Res. Solid Earth* **106**, 16461–16481 (2001).
133. Bialas, R. W. & Buck, W. R. How sediment promotes narrow rifting: application to the Gulf of California. *Tectonics* **28**, TC4014 (2009).
134. Olive, J.-A., Behn, M. D. & Malatesta, L. C. Modes of extensional faulting controlled by surface processes. *Geophys. Res. Lett.* **41**, 6725–6733 (2014).
135. Olive, J.-A., Malatesta, L. C., Behn, M. D. & Buck, R. W. Sensitivity of rift tectonics to global variability in the efficiency of river erosion. *Proc. Natl Acad. Sci. USA* **119**, e2115077119 (2022).
136. Theunissen, T. & Huismans, R. S. Long-term coupling and feedback between tectonics and surface processes during non-volcanic rifted margin formation. *J. Geophys. Res. Solid Earth* **124**, 12323–12347 (2019).
137. Neuharth, D. et al. Evolution of rift systems and their fault networks in response to surface processes. *Tectonics* **41**, e2021TC007166 (2022).
138. McNeill, L. C. et al. High-resolution record reveals climate-driven environmental and sedimentary changes in an active rift. *Sci. Rep.* **9**, 3116 (2019).
139. Lyons, R. P., Scholz, C. A., Buoniconti, M. R. & Martin, M. R. Late quaternary stratigraphic analysis of the Lake Malawi Rift, East Africa: an integration of drill-core and seismic-reflection data. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **303**, 20–37 (2011).
140. Xue, L., Moucha, R. & Scholz, C. A. Climate-driven stress changes and normal fault behavior in the Lake Malawi (Nyasa) Rift, East Africa. *Earth Planet. Sci. Lett.* **593**, 117693 (2022).
141. de Sagazan, C. & Olive, J.-A. Assessing the impact of sedimentation on fault spacing at the Andaman Sea spreading center. *Geology* **49**, 447–451 (2020).
142. Buck, W. R. The role of magmatic loads and rift jumps in generating seaward dipping reflectors on volcanic rifted margins. *Earth Planet. Sci. Lett.* **466**, 62–69 (2017).
143. Morgan, R. L. & Watts, A. B. Seismic and gravity constraints on flexural models for the origin of seaward dipping reflectors. *Geophys. J. Int.* **214**, 2073–2083 (2018).
144. Péron-Pinvidic, G. & Manatschal, G. The final rifting evolution at deep magma-poor passive margins from Iberia-Newfoundland: a new point of view. *Int. J. Earth Sci.* **98**, 1581–1597 (2009).
145. Neuharth, D., Brune, S., Glerum, A., Heine, C. & Welford, J. K. Formation of continental microplates through rift linkage: numerical modeling and its application to the Flemish Cap and Sao Paulo Plateau. *Geochem. Geophys. Geosyst.* **22**, e2020GC009615 (2021).
146. King, M. T. & Welford, J. K. Advances in deformable plate tectonic models: 2. Reconstructing the southern North Atlantic back through time. *Geochem. Geophys. Geosyst.* **23**, e2022GC010373 (2022).
147. Steinberger, B., Bredow, E., Lebedev, S., Schaeffer, A. & Torsvik, T. H. Widespread volcanism in the Greenland–North Atlantic region explained by the Iceland plume. *Nat. Geosci.* **12**, 61 (2019).
148. Koptev, A. et al. Contrasted continental rifting via plume–craton interaction: applications to Central East African Rift. *Geosci. Front.* **7**, 221–236 (2016).
149. Kusznir, N. J. & Park, R. G. The extensional strength of the continental lithosphere: its dependence on geothermal gradient, and crustal composition and thickness. *Geol. Soc. Lond. Spec. Publ.* **28**, 35–52 (1987).
150. Gawthorpe, R. L. et al. Tectono-sedimentary evolution of the Plio-Pleistocene Corinth rift, Greece. *Basin Res.* **30**, 448–479 (2018).
151. Tregoning, P. et al. Estimation of current plate motions in Papua New Guinea from global positioning system observations. *J. Geophys. Res. Solid Earth* **103**, 12181–12203 (1998).
152. Tesauro, M., Hollenstein, C., Egli, R., Geiger, A. & Kahle, H.-G. Continuous GPS and broad-scale deformation across the Rhine Graben and the Alps. *Int. J. Earth Sci.* **94**, 525–537 (2005).
153. Berglund, H. T. et al. Distributed deformation across the Rio Grande Rift, Great Plains, and Colorado Plateau. *Geology* **40**, 23–26 (2012).
154. Heckenbach, E. L., Brune, S., Glerum, A. C. & Bott, J. Is there a speed limit for the thermal steady-state assumption in continental rifts? *Geochem. Geophys. Geosyst.* **22**, e2020GC009577 (2021).
155. van Wijk, J. W. & Cloetingh, S. A. P. L. Basin migration caused by slow lithospheric extension. *Earth Planet. Sci. Lett.* **198**, 275–288 (2002).
156. Heine, C. & Brune, S. Oblique rifting of the equatorial Atlantic: why there is no Saharan Atlantic Ocean. *Geology* **42**, 211–214 (2014).
157. Faleide, J. I. et al. Structure and evolution of the continental margin off Norway and the Barents Sea. *Episodes* **31**, 82–91 (2008).
158. Naliboff, J. & Buiter, S. J. H. Rift reactivation and migration during multiphase extension. *Earth Planet. Sci. Lett.* **421**, 58–67 (2015).
159. Foster, D. A. & Gleadow, A. J. W. Structural framework and denudation history of the flanks of the Kenya and Anza Rifts, East Africa. *Tectonics* **15**, 258–271 (1996).
160. Torres Acosta, V. et al. Cenozoic extension in the Kenya Rift from low-temperature thermochronology: links to diachronous spatiotemporal evolution of rifting in East Africa. *Tectonics* **34**, 2367–2386 (2015).
161. Faulkner, D. R. et al. A review of recent developments concerning the structure, mechanics and fluid flow properties of fault zones. *J. Struct. Geol.* **32**, 1557–1575 (2010).
162. Rowland, J. V. & Sibson, R. H. Structural controls on hydrothermal flow in a segmented rift system, Taupo Volcanic Zone, New Zealand. *Geofluids* **4**, 259–283 (2004).
163. Lund, J. W. & Toth, A. N. Direct utilization of geothermal energy 2020 worldwide review. *Geothermics* **90**, 101915 (2021).
164. Ali, S. H. et al. Mineral supply for sustainable development requires resource governance. *Nature* **543**, 367–372 (2017).
165. Rodriguez Piceda, C. et al. Lithospheric density structure of the southern Central Andes constrained by 3D data-integrative gravity modelling. *Int. J. Earth Sci.* **110**, 2333–2359 (2021).
166. Tosdal, R. M., Dilles, J. H. & Cooke, D. R. From source to sinks in auriferous magmatic-hydrothermal porphyry and epithermal deposits. *Elements* **5**, 289–295 (2009).
167. Rowland, J. V. & Simmons, S. F. Hydrologic, magmatic, and tectonic controls on hydrothermal flow, taupo volcanic zone, New Zealand: implications for the formation of epithermal vein deposits. *Econ. Geol.* **107**, 427–457 (2012).
168. Hoggard, M. J. et al. Global distribution of sediment-hosted metals controlled by craton edge stability. *Nat. Geosci.* **13**, 504–510 (2020).

169. Lawley, C. J. M. et al. Data-driven prospectivity modelling of sediment–hosted Zn–Pb mineral systems and their critical raw materials. *Ore Geol. Rev.* **141**, 104635 (2022).
170. Lee, H. et al. Massive and prolonged deep carbon emissions associated with continental rifting. *Nat. Geosci.* **9**, 145–149 (2016).
171. Frondini, F. et al. Carbon dioxide degassing from Tuscany and Northern Latium (Italy). *Glob. Planet. Change* **61**, 89–102 (2008).
172. Seward, T. M. & Kerrick, D. M. Hydrothermal CO_2 emission from the Taupo Volcanic Zone, New Zealand. *Earth Planet. Sci. Lett.* **139**, 105–113 (1996).
173. Weinlich, F. H. et al. An active subcontinental mantle volatile system in the western Eger rift, central Europe: gas flux, isotopic (He, C, and N) and compositional fingerprints. *Geochim. Cosmochim. Acta* **63**, 3653–3671 (1999).
174. Wong, K. et al. Deep carbon cycling over the past 200 million years: a review of fluxes in different tectonic settings. *Front. Earth Sci.* **7**, 263 (2019).
175. Jones, J. R., Stamps, D. S., Wauthier, C., Saria, E. & Biggs, J. Evidence for slip on a border fault triggered by magmatic processes in an immature continental rift. *Geochem. Geophys. Geosyst.* **20**, 2515–2530 (2019).
176. Pagli, C. et al. Shallow axial magma chamber at the slow-spreading Erta Ale Ridge. *Nat. Geosci.* **5**, 284–288 (2012).
177. Wright, T. J. et al. Magma-maintained rift segmentation at continental rupture in the 2005 Afar dyking episode. *Nature* **442**, 291–294 (2006).
178. Calais, E. et al. Strain accommodation by slow slip and dyking in a youthful continental rift, East Africa. *Nature* **456**, 783–787 (2008).
179. Karabacak, V., Ring, U. & Uysal, I. T. The off-fault deformation on the North Anatolian Fault zone and assessment of slip rate from carbonate veins. *Tectonophysics* **795**, 228633 (2020).
180. Herbert, J. W., Cooke, M. L., Oskin, M. & Difo, O. How much can off-fault deformation contribute to the slip rate discrepancy within the eastern California shear zone? *Geology* **42**, 71–75 (2014).
181. Stamps, D. S., Saria, E. & Kreemer, C. A geodetic strain rate model for the East African Rift System. *Sci. Rep.* **8**, 732 (2018).
182. Muirhead, J. D. et al. Evolution of upper crustal faulting assisted by magmatic volatile release during early-stage continental rift development in the East African Rift. *Geosphere* **12**, 1670–1700 (2016).
183. Mana, S., Furman, T., Turrin, B. D., Feigenson, M. D. & Swisher, C. C. III Magmatic activity across the East African North Tanzanian Divergence Zone. *J. Geol. Soc.* **172**, 368–389 (2015).
184. Williams, J. N. et al. The Malawi active fault database: an onshore–offshore database for regional assessment of seismic hazard and tectonic evolution. *Geochem. Geophys. Geosyst.* **23**, e2022GC010425 (2022).
185. Shillington, D. J. et al. Controls on rift faulting in the North Basin of the Malawi (Nyasa) Rift, East Africa. *Tectonics* **39**, e2019TC005633 (2020).
186. Njinju, E. A. et al. Lithospheric structure of the Malawi Rift: implications for magma-poor rifting processes. *Tectonics* **38**, 3835–3853 (2019).
187. Hopper, E. et al. Preferential localized thinning of lithospheric mantle in the melt-poor Malawi Rift. *Nat. Geosci.* **13**, 584–589 (2020).
188. Buck, W. R. in *Treatise on Geophysics* 2nd edn Vol. 6 (eds Schubert, G. & Watts, A.B) 325–379 (Elsevier, 2015).
189. Bagley, B. & Nyblade, A. A. Seismic anisotropy in eastern Africa, mantle flow, and the African superplume. *Geophys. Res. Lett.* **40**, 1500–1505 (2013).
190. Müller, R. D. et al. A global plate model including lithospheric deformation along major rifts and orogens since the triassic. *Tectonics* **38**, 1884–1907 (2019).

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

Competing interests
The authors declare no competing interests.

Additional information
Peer review information *Nature Reviews Earth & Environment* thanks the anonymous reviewers for their contribution to the peer review of this work.