A review of structural inheritance in rift basin formation

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ABSTRACT

In the context of rift basin formation, structural inheritance describes the influence of pre-existing basement structures on new, rift-related structures, including faults. Examples of basin features influenced by inheritance include rift localisation and segmentation at the plate scale, as well as variations in the geometries, orientations, and kinematics of individual rift-related faults. Given that continental rifts commonly form in pre-deformed crust, structural inheritance is likely to be the norm, not the exception. As such, structural inheritance has implications for reconstructing the paleotectonic history of rifts, investigating seismic hazards, and understanding the fluid transport and storage capabilities of natural fracture systems in the context of geo-energy and ore deposits. Our review of the literature shows that inheritance is driven by several mechanisms, which include frictional reactivation and local re-orientation of the far-field strain and/or stress. Here we highlight how insights from field observations, geophysics, and analogue and numerical models can be used to classify these mechanisms in terms of hard-linked and soft-linked inheritance. We demonstrate how different inheritance mechanisms can produce different geometric and kinematic relationships between pre-existing basement structures and rift-related faults, and that these mechanisms can be active at different depths within the same rift. Our aim is to provide a framework for recognising various expressions of structural inheritance and their underlying mechanism(s) in natural rifts, so that we can better interpret basement structures under cover and are equipped with additional constraints for understanding the multi-stage evolution of basement-influenced rift basins worldwide.

1. INTRODUCTION

Continental rifts commonly form in crust with a pre-existing structural framework. Given the appropriate stress orientations, stress magnitudes, fluid pressures, and thermal gradients of the crust, these structures exert a control on the development of rifts and rift-related structures (e.g., Wilson, 1966; Morley, 1995; Krabbendam and Barr, 2000; Schumacher, 2002; Chenin and Beaumont, 2013; Fossen et al., 2016; Schiffer et al., 2018, 2020). This **structural inheritance** operates at multiple scales in both frictional and viscous layers of the crust and influences the location, geometry, and internal architecture of rift basins (examples in Table 1). Lithospheric-scale weaknesses up to thousands of kilometres in length span the frictional-viscous transition and localise and segment entire rift systems (McConnell, 1972; Daly et al., 1989; Versfelt and Rosendahl, 1989; Wheeler and Karson, 1989; Piqué and Laville, 1996; Doré et al., 1997; Holdsworth et al., 2001a; Miller et al., 2002; Gibson et al., 2013; Molnar et al., 2017; Peace et al., 2018b; Heron et al., 2019). They also modulate the evolution of the lithosphere and magmatic budget during rifting, in some cases up to the continental break-up and passive margin stages (e.g., Dunbar and Sawyer, 1989; Buiter and Torsvik, 2014; Petersen and Schiffer, 2016; Gouiza and Paton, 2019; Gouiza and Naliboff, 2021). In the brittle crust, weaknesses at length scales of tens of kilometres or less influence the orientations and distributions of basin-bounding and basin-internal faults, some of which are misoriented with respect to inferred plate motion directions (e.g., Morley et al., 2004; Lyon et al., 2007; Wilson et al., 2010).

Many studies invoking the influence of pre-rift basement structures on rift-related faults pay particular attention to frictional **reactivation**, which describes structures that have accommodated displacement during multiple discrete (i.e., intervals >1 Ma) deformation events (sensu Holdsworth et al., 1997). These structures include lithological contacts (Ashby, 2013; Holdsworth et al., 2020; Wedmore et al., 2020a), faults or fault zones (McCaffrey, 1997; Holdsworth et al., 2001b), older rift fabrics (Whipp et al., 2014; Bladon et al., 2015; Duffy et al., 2015; Deng et al., 2017a; McHarg et al., 2019; Phillips and McCaffrey, 2019; Wang et al., 2021), and shear zones and the metamorphic fabrics within them (Kirkpatrick et al., 2013; Bird et al., 2014; Phillips et al., 2016; Fazlikhani et al., 2017; Kolawole et al., 2018; Peace et al., 2018a; Heilman et al., 2019; Wedmore et al., 2020b). Such focus on reactivation has resulted in the synonymous use of the terms *structural inheritance* and *reactivation*; in some cases, the two terms are used interchangeably (e.g., Corti et al., 2007; Manatschal et al., 2015). This usage is problematic, as it does not acknowledge other mechanisms by which pre-rift basement structures influence rift-related deformation. It also suggests that there is limited or no influence of pre-existing structures in areas where we have no evidence of reactivation (as defined by Holdsworth et al., 1997).

A number of field and modelling studies point to subtle mechanisms, such as the re-orientation of strain (Corti et al., 2013b; Philippon et al., 2015; Hodge et al., 2018; Samsu et al., 2021) and stress (Bell, 1996; Morley, 2010; Tingay et al., 2010a), through which pre-rift basement structures can influence deformation localisation and distribution without exhibiting significant (i.e., observable) signs of repeated slip or displacement on pre-existing structures themselves. It is important to distinguish between inheritance mechanisms, as they have different implications for the kinematics of rift-related faults, the interpretation of far-field paleo-extension and paleo-stress directions, and hard and/or soft linkage between pre-rift basement structures and rift-related faults (Figure 1).

In this paper, we explore the multifaceted ways in which heterogeneous basement rocks can control fault system geometry and influence deformation in the overlying cover rocks during rift basin formation in magma-poor rifts. In this context, we consider the crust to be rheologically layered with a thermally controlled transition separating an upper frictional and lower viscous regime (sensu Handy et al., 2007). A more complete picture of different inheritance mechanisms can help us clarify the mechanical controls on rift and basin evolution and open up opportunities to: (i) "uncover" the characteristics of buried basement rocks that are commonly poorly imaged in geophysical surveys; (ii) better understand the structural evolution of rifted margins; and (iii) provide additional constraints on plate reconstructions. Understanding the potential impact of pre-existing basement structures on younger fault networks also has implications for seismic hazards assessment (e.g., Fonseca, 1988; Wedmore et al., 2020b; Hecker et al., 2021), geothermal energy (e.g., Schumacher, 2002; Bertrand et al., 2018), mineral exploration (e.g., Rowland and Sibson, 2004), and CO_2 and nuclear waste storage (e.g., Barton and Zoback, 1994; Andrés et al., 2016).

In the first part of this review, we synthesise research on structural inheritance during rift basin formation. We focus on the impact of inheritance on faults that strike obliquely to the inferred far-field extension direction but formed or are active during rifting. Observations from field and geophysical data are presented first, followed by insights from analogue and numerical models and geomechanical theory (Section 2). In the Discussion, we propose two mechanisms that underpin all instances of structural inheritance, namely reactivation and strain re-orientation (Section 3.1). We also suggest a classification of observed basement-cover relationships into hard-linked and soft-linked inheritance and interpret the mechanisms behind them (Section 3.2). Finally, we apply this framework to a natural rift system (i.e., the East African Rift System; Section 3.3) and discuss the implications of structural inheritance for natural resources exploration (Section 3.4).

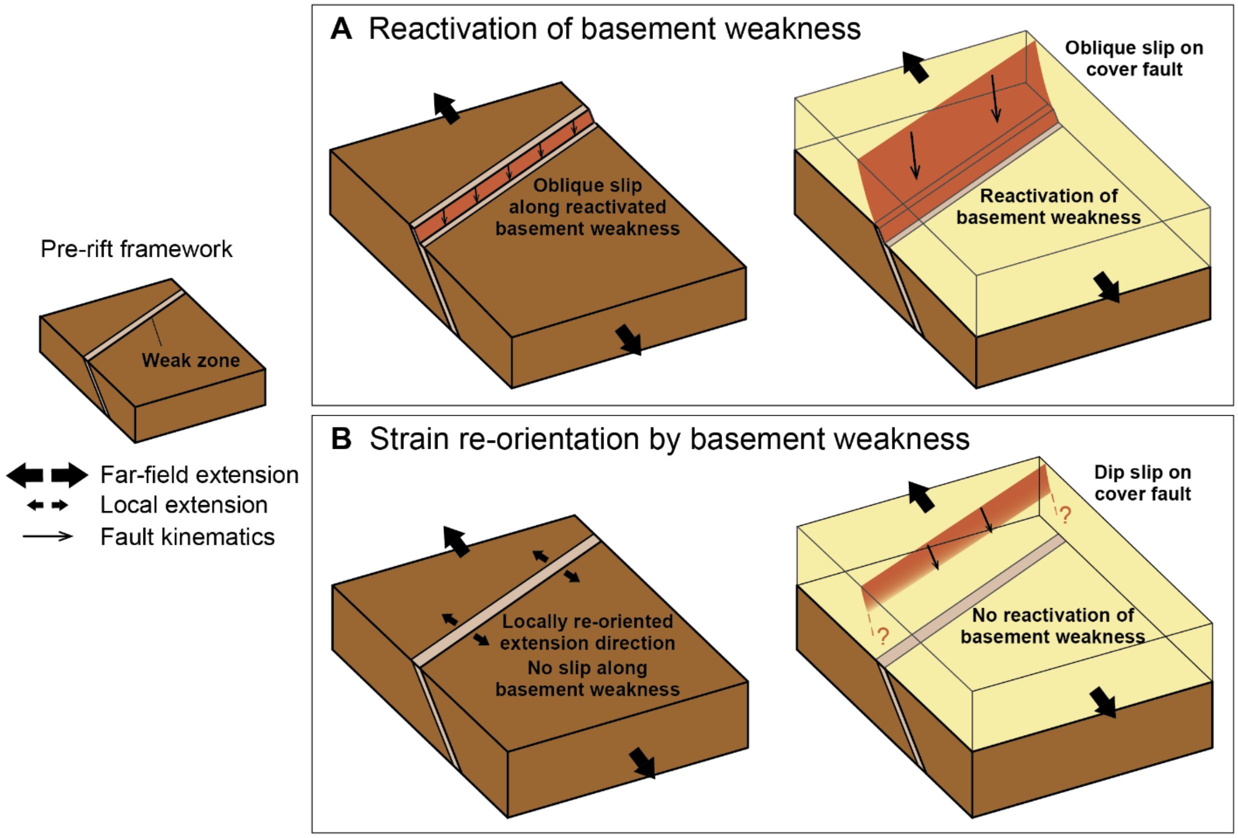

Figure 1 A pre-existing basement weakness, striking obliquely to the far-field extension direction, can influence the orientation and kinematics of rift-related faults (e.g., normal faults) in the overlying cover. Different inheritance mechanisms can have different impacts on fault kinematics and linkage between a rift-related cover fault and basement weakness, despite the two exhibiting geometric similarity in map view in both scenarios. Reactivation of the basement weakness (A) is commonly associated with oblique slip along the basement weakness and cover fault (e.g., Morley, 1995; Corti et al., 2007), in addition to hard linkage between the two structures (Phillips et al., 2016). A second mechanism through which the basement weakness can influence rift-related faulting is local re-orientation of the far-field strain (B), which may result in dip-slip kinematics along the cover fault (Philippon et al., 2015).

Here we refer to **basement** as any rock underlying the sedimentary **cover**, which we define as basin fill related to the current rift-related episode. The basement can comprise multiple units overlying one another and may include basin fill from previous rift episodes. While we acknowledge that inheritance can impact all kinds of brittle structures (e.g., faults, joints, dykes, etc.) during and after rifting, here we focus on the influence of pre-rift (a.k.a. pre-existing) basement structures on rift-related faults. Pre-existing basement structures include discrete discontinuities (e.g., the contact between a dyke and surrounding host rock; a fault or fault zone; the boundary between two rheologically distinct terranes), pervasive strength anisotropies (e.g., penetrative metamorphic or rift fabrics from previous collisional or extensional events), and lithospheric-scale weaknesses in the viscous lower crust or lithospheric mantle (Table 1).

Table 1 Natural examples of pre-existing basement features, their influence on subsequent deformation in the overlying cover, and the resulting geometries which can be observed.

Pre-existing feature	Scale	Mechanism	Observation(s)	Example(s)
Rheological boundary (crustal-scale)				
Dyke-host rock contact		Reactivation	Geometric similarity	Lower Shire Graben, southern Malawi Rift (Wedmore et al., 2020); NW Scotland (Holdsworth et al., 2020)
Discrete fabric				
Fault or cemented fault zone		Reactivation	Geometric similarity	North Sea Rift (Deng et al., 2017a)
		Stress perturbation	Local stress re-orientation	North Sea Rift (Bell, 1996; Yale, 2003)
Pervasive strength				
Metamorphic foliation within shear zone	10s of km	Stress perturbation	3D transtensional strain	North Sea Rift (Reeve et al., 2015; Osagiede et al., 2020); East African Rift (Morley et al., 2010 and references therein)
	10s of km	Reactivation	Geometric similarity	Northern Thailand (Morley et al., 2004); North Sea Rift (Færseth et al., 1995; Phillips et al., 2016; Fazlikhani et al., 2017); offshore New Zealand (Muir et al., 2000; Collanega et al., 2019; Phillips & McCaffrey, 2019); onshore New Zealand (Villamor et al., 2017); East African Rift System (Laó-Dávila et al., 2015; Dawson et al., 2018); East Greenland Rift System (Rotevatn et al., 2018); onshore and offshore Norway (Fossen & Hurich, 2005; Fossen et al., 2014; Fossen et al., 2016)
Rift fabric	10s of km	Reactivation	Non-colinear normal fault sets	Barmer Basin, NW India (Bladon et al., 2015); Horda Platform, North Sea (Whipp et al., 2014; Duffy et al., 2015); Bohai Bay Basin, eastern China (Wang et al., 2021); Great South Basin, offshore New Zealand (Phillips & McCaffrey, 2019); Northern Carnavron Basin, Western Australia (McHarg et al., 2019)
Rheological boundary (lithospheric-scale)				
Terrane boundary	10s of km	Barrier to fault propagation	Fault splay	Great South Basin, offshore New Zealand (Phillips & McCaffrey, 2019); Sudan Rift System (Daly et al., 1989)
Lithospheric weakness				
Orogenic structures, including shear zones (i.e., terrane boundary, suture zone, orogenic belt, mobile belt)	10s−1000s of km	Strain localisation	Rift localisation & segmentation	North Atlantic (Wilson, 1966; Piqué & Laville, 1996; Doré et al., 1997; Buiter & Torsvik, 2014; Petersen & Schiffer, 2016; Schiffer et al., 2018); South Atlantic (Krabbendam & Barr, 2000); East African Rift System (McConnell, 1972; Daly et al., 1989; Versfelt & Rosendahl, 1989; Wheeler & Karson, 1989; Smith and Mosley, 1993; Ring, 1994; Theunissen et al., 1996; Morley, 2010; Kolawole et al., 2018; Muirhead & Kattenhorn, 2018; Wedmore et al., 2020); Gondwana break-up along Pan African mobile belts (Sykes, 1978); Karoo Rift Basins (Daly et al., 1989); Baikal Rift (Petit et al., 1996); West Greenland Rift System (Heron et al., 2019); offshore southern Norway (Færseth et al., 1995; Phillips et al., 2016); onshore New Zealand (Rowland and Sibson, 2004); offshore New Zealand (Muir et al., 2000; Mortimer et al., 2002; Phillips & McCaffrey, 2019); Australian Southern Margin (Miller et al., 2002; Gibson et al., 2013)
Pervasive anisotropic crystalline fabric within lithospheric mantle	10s−1000s of km		Rift localisation	North, Central & South Atlantic (Vauchez et al., 1997; Tommasi & Vauchez, 2001)

2. EXAMPLES AND INSIGHTS FROM NATURE AND MODELS

2.1. Observations of basement-cover interactions from geophysical data and outcrops

In this section, we summarise the spatial, geometric, and kinematic relationships between basement structures and cover faults based on observations from natural rifts.

2.1.1. <u>Spatial co-location and geometric similarity</u>

Structural inheritance in rift basins is typically recognised through two observations in map view: *spatial co-location* and *geometric similarity* between rift-related faults and older pre-rift structures (Figure 2). **Spatial co-location** refers to the occurrence of rift-related faults above or in the vicinity of a pre-existing structure. **Geometric similarity** (sensu Holdsworth et al., 1997) is characterised by rift-related faults with traces that trend parallel to pre-rift structures (or, more rigorously, they have the same strike and dip directions in three dimensions). Both types of observations have been made at a variety of scales (Figure 3).

Plate-scale structural inheritance (a.k.a. **tectonic inheritance**; Wilson, 1966; Thomas, 2006; Audet and Bürgmann, 2011; Buiter and Torsvik, 2014; Petersen and Schiffer, 2016) is demonstrated by the co-location of pre-existing crustal or lithospheric-scale weaknesses with younger rift systems (Figure 3a). It is widely acknowledged that the exploitation of these mechanical (including thermal) weaknesses promotes rift initiation and propagation, so that younger rifts commonly follow the trends of terrane boundaries and highly deformed orogenic belts (see observations of inheritance of “Lithospheric weakness” in Table 1). We explicitly differentiate this *activation* by tectonic inheritance from *reactivation* of discrete structures. For example, most Phanerozoic rift basins in Africa are located within Proterozoic orogenic belts (Daly et al., 1989). These belts may be linked to deep-seated weaknesses in the lithospheric mantle (Tommasi and Vauchez, 2001), which localise rifting in the overlying crust (Versfelt and Rosendahl, 1989). Here orogenic structures are broadly co-located with, and geometrically similar to, historical seismic events (Fairhead and Henderson, 1977; Sykes, 1978; Craig et al., 2011).

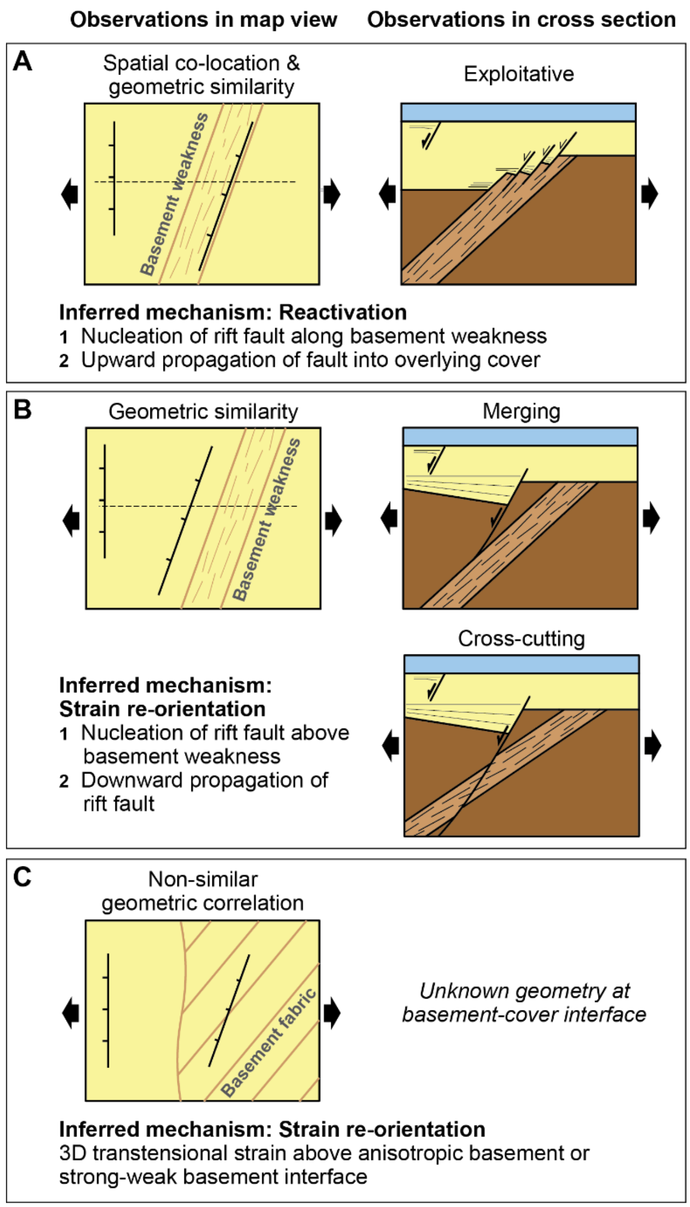

Figure 2 Geometric relationships between pre-existing basement structures and rift-related normal faults in map view and cross section (exploitative, merging, and cross-cutting relationships after Phillips et al., 2016). The inferred inheritance mechanisms behind these relationships are reactivation (A) and strain re-orientation (B and C). Thick arrows indicate the inferred regional extension direction during rifting.

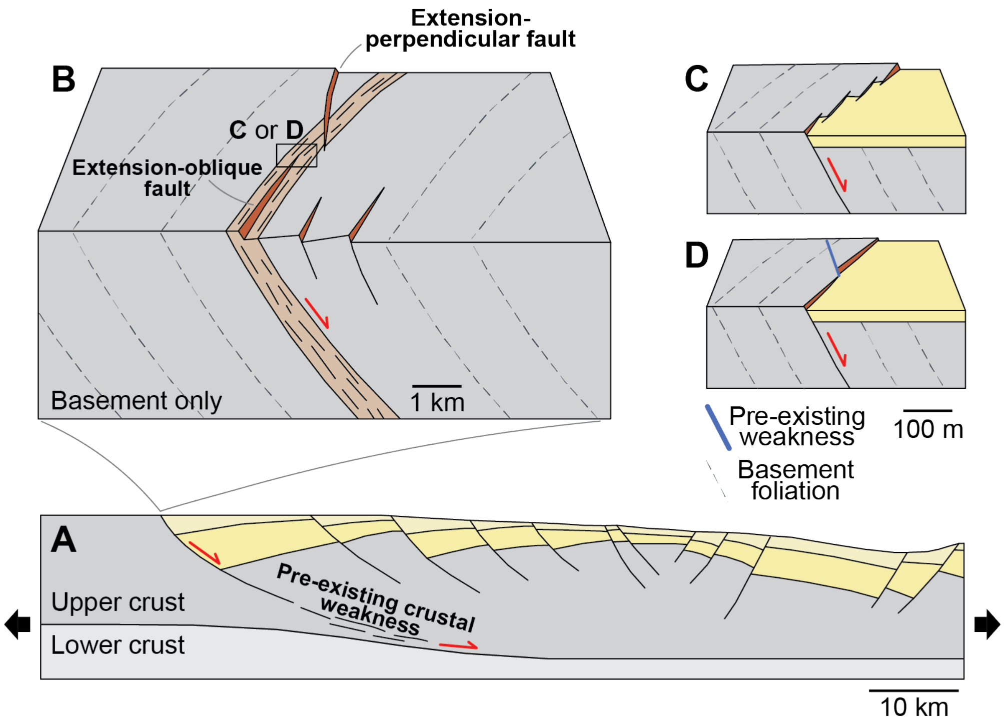

Figure 3 The activation of a pre-existing crustal weakness influences rift-related cover faults at a range of scales. Tectonic inheritance contributes to rift localisation (A), while basin to sub-basin scale inheritance results in spatial co-location and geometric similarity between pre-rift structures (e.g., basement shear zones and foliation) and rift-related normal faults (B). At the outcrop scale, some minor faults can deviate from the main fault and crosscut basement foliation, reflecting the complex architecture of a fault zone (C) or reactivation of other pre-existing structures, such as dykes (D). Thick arrows indicate the inferred regional extension direction during rifting.

Spatial and geometric indicators of structural inheritance are also common at the basin and sub-basin scale (Figure 3b). Several km-long basin-bounding and basin-internal faults commonly exhibit spatial co-location and geometric similarity with pre-rift structures in the underlying basement rocks (e.g., Laó-Dávila et al., 2015; Fazlikhani et al., 2017; Dawson et al., 2018; Collanega et al., 2019; Wedmore et al., 2020b). For example, the strike of the Terrane Boundary Fault in the Great South Basin (offshore South Island, New Zealand), which formed during Late Cretaceous extension (i.e., Gondwana break-up), is parallel to the Devonian–Cretaceous Terrane Boundary Shear Zone (Mortimer, 2004) that separates the Median Batholith and Western Province terranes. Seismic reflection data show that the fault links with the crustal-scale (or potentially lithospheric-scale; e.g., Muir et al., 2000; Mortimer et al., 2002) Terrane Boundary Shear Zone at depth (Phillips and McCaffrey, 2019). Farther north in the Taranaki Basin (offshore North Island, New Zealand), the Terrane Boundary Shear Zone and an adjacent boundary between the Median Batholith and Eastern Province terranes coincide spatially with the traces of the Cenozoic Cape Egmont and Taranaki fault systems (Muir et al., 2000). Onshore, currently active N-S striking faults of the Taupō Rift (North Island, New Zealand) are parallel to terrane boundaries and bedding in the Mesozoic basement which crops out to the west and east of the rift (Villamor et al., 2017; Milicich et al., 2021).

In offshore southern Norway, the bounding faults of the Stavanger Basin, which formed during Permo-Triassic rifting, are parallel to and co-located with the Devonian Stavanger Shear Zone; interpreted seismic reflection data also show that the two structures are hard-linked (Phillips et al., 2016). Farther north in onshore Norway, the Lærdal-Gjende fault system is co-located and hard-linked with the Devonian Hardangerfjörd Shear Zone, which is associated with a major Moho offset (Færseth et al., 1995; Fossen and Hurich, 2005; Fossen et al., 2014). Similarly, Permo-Triassic basin-bounding rift faults in offshore Scotland exhibit geometric similarity with deep crustal reflectors in the underlying crystalline basement (Wilson et al., 2010 and references therein). From aerial imagery and field observations from northeastern Brazil, Kirkpatrick et al. (2013) found that the regional, km-scale traces of Mesozoic rift faults trend parallel to Precambrian Brasiliano shear zones and crustal-scale anomalies interpreted from magnetic and gravity data.

In the Upper Rhine Graben in central Europe, Bertrand et al. (2018) observed geometric similarity between magnetic foliation in magmatic basement rocks (Skrzypek et al., 2014) and km-scale horst and graben-bounding faults. There is a spatial co-location between Proterozoic NE-trending basement structures of the Jemez Lineament and a rotation in fault strike in the Rio Grande Rift (United States of America) from N-NNE to ENE-E (Aldrich, 1986; Chapin et al., 2004). A growing body of work (discussed in detail in Section 3.3) provides compelling evidence that basin-scale deformation in the active East African Rift System is controlled by structural inheritance (see also a recent review on the Main Ethiopian Rift by Corti et al., 2022).

2.1.2. Hard-link relationships between basement and cover structures

Phillips et al. (2016) illustrated three types of two-dimensional (2D) cross sectional relationships between a pre-existing basement shear zone and younger rift-related faults - *exploitative*, *merging*, and *cross-cutting* relationships - based on seismic reflection data from offshore southern Norway (Figure 2). These relationships were identified from basement shear zones and rift-related faults that have the same strike and dip direction. They used kinematic analyses of mapped faults to determine the history of fault activity and infer whether or not basement shear zones were reactivated as the younger faults formed.

Phillips et al. (2016) showed that an **exploitative** relationship is characterised by a throughgoing structure that links the basement and cover structures. Based on the physical linkage between shear zone-internal reflectors and younger rift faults, they inferred that the rift faults initiated in relatively weak mylonitic rocks (within broader shear zones) in the basement and propagated upwards into the cover. A similar observation of such an exploitative relationship was made by Collanega et al. (2019) in the Taranaki Basin (offshore North Island, New Zealand), except that it was the damage zone above a pre-existing structure that was reactivated. These two case studies suggest that reactivation (of a discrete pre-existing basement structure) is the underlying mechanism behind an exploitative basement-cover relationship (Figure 2a).

Merging and cross-cutting relationships (Figure 2b) differ from an exploitative relationship in that the rift-related fault nucleates above the basement structure and propagates downwards until they intersect. In a **merging relationship**, the rift-related fault nucleates within the hanging wall of the basement structure and – because of its steeper dip – merges at depth with the more gently dipping basement structure (Phillips et al., 2016). In a **cross-cutting relationship**, the rift-related fault inherits the strike and dip direction of the basement structure but offsets it at depth. For example, the Mesozoic rift-related Dombjerg Fault (East Greenland Rift System) locally strikes parallel to the more gently dipping Caledonian Kildedalen Shear Zone but offsets it farther down-dip (A. Rotevatn et al., 2018). Phillips et al. (2016) attribute merging and cross-cutting relationships to strain re-orientation by a pre-existing basement structure that has been reactivated (Figure 2b).

2.1.3. Non-similar geometric correlation

The influence of basement heterogeneities has also been invoked in areas where geometric similarity is not observed, but instead: (i) the strikes of coeval normal faults vary within a relatively small area (on the order of several km) (e.g., Morley et al., 2004; Reeve et al., 2015), (ii) normal fault strikes are oblique to the strikes of basement structures and not perpendicular to the inferred regional extension direction (Figure 2c) (Donath, 1962; Samsu et al., 2019), and/or (iii) fault strikes vary across areas that overlie different basement terranes (Wilson et al., 2010). In the Gippsland Basin (southeast Australia), rift-related fault orientations vary above two different basement domains. These different orientations were previously attributed to a change in rifting directions, but may instead reflect the influence of basement fabrics that are highly oblique to the inferred rifting direction (Samsu et al., 2019 and references therein). By comparing onshore fault orientations within the North Coast Transfer Zone (Scotland), Wilson et al. (2010) showed that changes in complex fault patterns in cover rocks coincide with basement terrane boundaries, reflecting localised changes in three-dimensional (3D) transtensional strain in response to rifting. In the absence of evidence of basement reactivation, which is normally inferred from geometric similarity and other kinematic or chronostratigraphic indicators (Holdsworth et al., 1997), non-similar geometric correlations may indicate the subtle influence of basement rocks with fabrics that are oblique to far-field rifting directions.

2.1.4. Basement strength variations impacting fault propagation and distribution

In addition to its influence on fault orientations, lateral variations in basement rock strength control the propagation and distribution of rift-related normal faults. For example, the NW trending Southern Sudan Rift terminates abruptly against the NE trending Central African fault zone, which follows the trace of the steep Late Proterozoic Foumban Shear Zone (Daly et al., 1989). Farther south, the northern tip of the western branch of the East African Rift System (trending NE at this locality) terminates against the NW trending, Mesoproterozoic Aswa Shear Zone (Katumwehe et al., 2015; Saalmann et al., 2016).

At the basin scale, the propagation of a normal fault can be inhibited by a relatively strong basement block when the lateral boundary between the weak and strong units is at a high angle to the fault strike. In the Great South Basin (offshore South Island, New Zealand), Phillips and McCaffrey (2019) documented two styles of normal faulting near the Terrane Boundary Fault, which separates the dominantly sedimentary Western Province from the dominantly plutonic Median Batholith. Within the hanging wall of the Terrane Boundary Fault, a large-displacement normal fault within an Upper Cretaceous sedimentary unit splays into a system of low-displacement segments as it approaches a structural high in the basement comprising granitic material.

In other cases, where a propagating normal fault is geometrically similar to a pre-existing crustal weakness, structural inheritance can facilitate fault tip propagation and segment linkage (e.g., Walsh et al., 2002; Jackson and Rotevatn, 2013; Rotevatn et al., 2018). This leads to the establishment of the final normal fault length at a relatively early stage (10-20%; Rotevatn et al., 2019) of its displacement history; possibly within a few earthquake cycles (Hecker et al., 2021).

2.1.5. Scale dependence

The degree of influence of basement structures on cover fault orientations varies with the size of the basement structure and the scale of observation. Lithospheric and crustal-scale weaknesses localise the initiation and propagation of rifts and cause along-strike variations in rift geometry (refer to examples in Section 2.1.1 and Table 1). The activation of >1 km-long or km-wide weaknesses in lower levels of the crust can result in cover faults which are spatially co-located and geometrically similar with the reactivated weaknesses (Figure 3). Seismic reflection data from offshore areas (Phillips et al., 2016, 2021a; Collanega et al., 2019) show that this correlation results from hard linkage between relatively shallow faults and deep, pre-existing weaknesses (e.g., faults or mylonitic foliation within shear zones).

Rift-related faults show a more complex geometry at the sub-km scale (Figure 3c-d). Outcrop observations suggest that the influence of crustal-scale basement weaknesses is less prominent at the meters scale, which has been attributed to the increasing complexity of fault zone architecture as faults accommodates greater amounts of strain (Kirkpatrick et al., 2013). While fault initiation is controlled by grain-scale anisotropy, the coalescence of fault segments and the development of mechanically isotropic fault core rock (i.e., cataclasite and fault gouge) with increasing displacement results in complex networks of subsidiary brittle structures (Childs et al., 2009; Kirkpatrick et al., 2013; Deng et al., 2017b; Williams et al., 2022). The main fault zone therefore maintains an orientation that is parallel to the basement weakness, while secondary faults deviate from the main fault trend, cross-cutting local basement foliation in some places (Figure 3c).

Collanega et al. (2019) documented the depth dependence of inheritance from seismic reflection data from the Taranaki Basin, where extension-oblique, intrabasement shear zones were reactivated during Plio-Pleistocene rifting. Here, rift-related faults nucleated either from intrabasement shear zones (possibly Mesozoic in age; Muir et al., 2000) or within damage zones above Late Miocene reverse faults. Collanega et al. (2019) suggested that the upward propagation of these faults into the overlying cover during rifting was controlled by the obliquity of the shear zone to the regional extension direction. Shear zones that strike perpendicular to the extension direction are hard-linked to rift-related faults that extend into the upper levels of the cover. In contrast, shear zones that strike more obliquely to the extension direction are hard-linked to rift-related faults that are restricted to the lower levels of the cover. A similar depth and orientation-dependent influence on fault orientation and linkage was observed along the reactivated, basement-rooted Parihaka Fault farther north in the Taranaki Basin (Giba et al., 2012).

Collanega et al. (2019) also showed that the potential for a reactivated basement weakness to localise strain and locally re-orient the far-field stress or strain can be modulated by the size of the basement weakness, i.e., the width of the weak zone. 100 m-wide zones comprising multiple discrete, mylonite-bearing faults can be reactivated and propagate upwards to shallow depths when they are nearly perpendicular to the regional extension direction. In contrast, ~1 km-wide shear zones can exert a stronger influence, reaching shallow depths, even if they are more oblique to the extension direction. The type of linkage between basement shear zones and younger faults also changes with depth. Deeper faults are hard-linked with, and geometrically similar to, the reactivated shear zone (Figure 4). Shallower faults are perpendicular to the regional extension direction but form an *en échelon* arrangement parallel to basement shear zones (cf. soft linkage between basement fault and cover fault in Phillips et al., 2021). Linkage between deep and shallow fault segments creates an up-sequence rotation of fault strike to define a sigmoidal fault geometry in cross section, and a twisted fault surface in 3D (Figure 4).

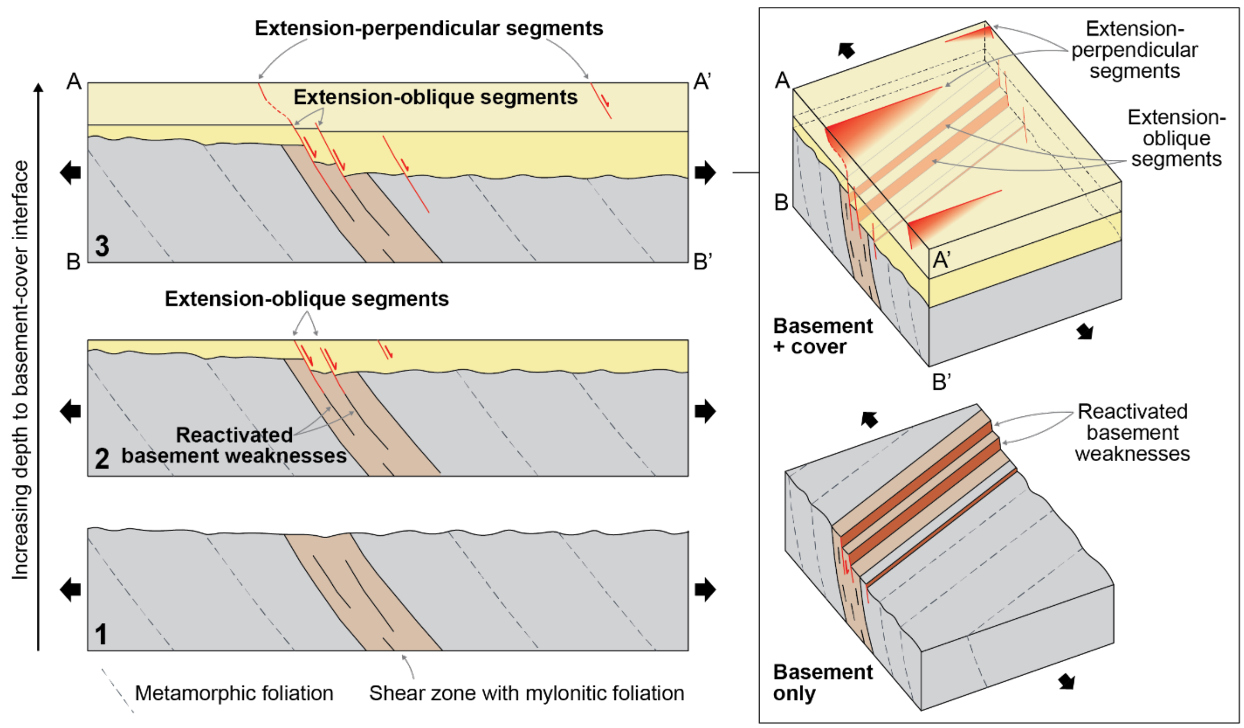

Figure 4 Conceptual illustration of depth dependence of inheritance, based on observations by (Collanega et al., 2019). Numbers 1–3 indicate the progression of rifting and increasing thickness of cover rocks. Faults near the basement-cover interface are geometrically similar with and rooted in reactivated basement weaknesses (e.g., shear zone, metamorphic foliation). Shallower faults are perpendicular to the regional extension (black arrows), suggesting that they are influenced to a lesser degree by the basement weaknesses.

2.1.6. Stress re-orientation and coupling

In the brittle part of the crust, local perturbations of the regional stress field can be attributed to lateral variations in the elastic properties of rocks (Bell, 1996 and references therein). Deviation from the far-field or regional stress trajectories occurs because the principal stresses are deflected when they cross an interface between two materials with contrasting elastic properties (e.g., Zhang et al., 1994; Morley, 2010; Zang and Stephansson, 2010) (Figure 5). Stress perturbations around a pre-existing weakness or discontinuity (e.g., a fault or damage zone) have been observed in photoelastic, numerical, and rock deformation experiments. In the field, they are reflected by curving of joints near a pre-existing fault (Cruikshank and Aydin, 1995; Muirhead and Kattenhorn, 2018; Samsu et al., 2020) or variable strain axes and kinematics of secondary, smaller-scale fault populations around a larger fault (Riller et al., 2017). Based on in-situ horizontal stress measurements, Yale (2003) suggested that faults are likely to perturb the regional stress field where: (i) differential horizontal stress is small (i.e., maximum horizontal stress is not much greater than minimum horizontal stress, usually correlating with low tectonic stress); (ii) large faults are present, which separate the study area into distinct fault blocks; or (iii) active faulting is present.

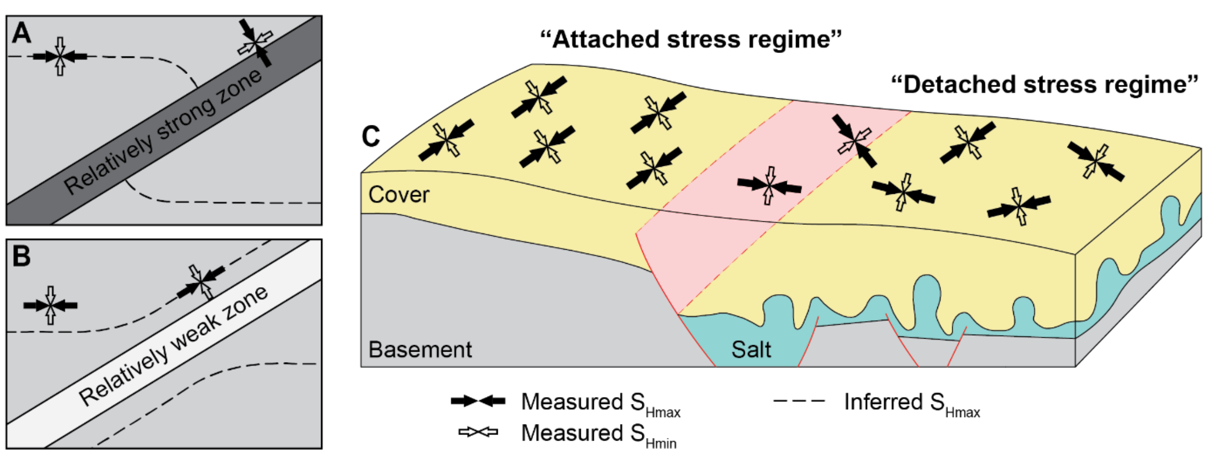

Figure 5 Map view illustration of the re-orientation of maximum horizontal stress (S_{Hmax}) near a mechanically strong (A) and weak (B) zone (modified after Bell, 1996). S_{Hmax} trajectories are perpendicular to the trend of a strong zone and parallel to the trend of a weak zone. (C) Stress coupling between basement and cover rocks occurs in an "attached stress regime" (sensu Bell, 1996). Stress decoupling occurs when there is an intervening, mechanically weak layer (e.g., evaporites, overpressured shales, and low-angle faults). All figures modified after Bell (1996).

We can invoke basement influence on rift faults in the overlying cover by considering stress coupling between the two units. Bell (1993, 1996) proposed that stress orientations in a sedimentary unit can "exhibit the signature" of underlying rocks, provided that there is no intervening weak unit that acts as a mechanical detachment. He refers to this coupled system as an "attached stress regime" (Figure 5c) and draws on an example from the Labrador Shelf (offshore eastern Canada), where the in-situ maximum horizontal stress measured from borehole breakouts is similarly oriented to the focal mechanism P-axis of a 1971 earthquake with an epicentre located in the basement. In an attached stress regime, we can infer that any deflection or deviation of the far-field stress within the basement rocks would be transferred onto the directly overlying sedimentary unit(s) and accommodated by new brittle structures (i.e., faults and other fractures) during rifting. Such mediation of stresses from basement to cover is apparent in the Lake Rukwa area of the East African Rift System. Here, Morley (2010) suggested that the NW-SE trending Precambrian basement foliation has locally rotated the regionally dominant N-S maximum horizontal stress, resulting in similar NW-SE trends for the present-day maximum horizontal stress direction (Delvaux and Barth, 2010) and Cenozoic rift-related faults. In contrast, where a mechanically weak unit (e.g., shale-dominated unit, salt, or low-angle fault) is present between basement and cover rocks, the in-situ maximum horizontal stresses tend to show more variable orientations ("detached stress regime" in Figure 5c).

2.2. Insights on inheritance mechanisms from analogue and numerical modelling

Analogue and numerical modelling allow us to study rift and basin-forming processes over geological timescales in 2D and 3D, complementing observations from natural rifts (e.g., Allemand and Brun, 1991; Brun, 1999; Corti, 2012; Brune et al., 2017; Molnar et al., 2020). Modelling also gives us the flexibility to examine the influence of various model parameters separately (e.g., rheological layering, obliquity, geometry of inherited weakness) as well as their combined effects (Zwaan et al., 2016, 2021a, 2021b; Zwaan and Schreurs, 2017). The modelling approach is especially useful for distinguishing between the relative contributions of oblique rift kinematics and inherited structures in shaping rift basins, as each on its own can create a transtensional system, resulting in faults that are oblique to the inferred paleo-extension direction. In this section, we summarise insights from analogue and numerical models on the mechanical interaction between extension-oblique pre-rift structures and rift-related faults. Here we focus on inheritance-driven obliquity, where the model includes pre-rift weaknesses that strike obliquely to the bulk extension direction (Table 2), as opposed to boundary-driven obliquity, where bulk extension is oblique to the model boundaries (e.g., Tron and Brun, 1991; Keep and McClay, 1997; Autin et al., 2010).

The analogue models we discuss are simplified, multi-layer, crustal or lithospheric-scale experiments comprising a brittle upper crust, ductile lower crust, and in some examples a ductile lithospheric mantle. In this case, ‘ductile’ is defined as exhibiting spatially continuous deformation at the scale of observation, and the ductile materials simulate deformation in the viscous layers of the lithosphere, although the rheology may be different (Wang, 2021). The brittle upper crust is modelled with granular materials that exhibit frictional behaviour and follow the Mohr-Coulomb failure criterion (e.g., Hubbert, 1951; Mandl et al., 1977; Davy and Cobbold, 1991; Schellart, 2000). The ductile lower crust and lithospheric mantle are modelled using materials that behave viscously under the experimental strain rates. The yield strength profile of the models resembles natural lithospheric strength profiles, which include (but are not limited to) relatively strong upper crust and lithospheric mantle layers and an intervening, weak lower crust. Pre-existing structures are implemented as two rheologically distinct blocks, discrete weak zones, or pervasive anisotropies (see Table 2 for an overview). Based on these models, we discuss the impact of crustal strength variations and inherited weaknesses – with different strike orientations – on deformation localisation and partitioning. We also highlight why these models helped us to better understand how mechanical heterogeneities in the basement influence the local 3D strain field.

Table 2 Pre-rift crustal and mantle/lithospheric heterogeneities in nature, their influence on rift to fault-scale deformation, and their implementation in analogue and numerical models. 'Discrete' and 'pervasive' denote the distribution of the heterogeneities at the scale of the model.

Relevance to natural feature	Influence on deformation	Implementation in analogue model	Reference(s)	Comparable numerical models
Crustal heterogeneity (discrete)				
Fault or fault zone in upper crust	Fault orientation & kinematics	Zone of dilation in granular layer	Corti et al. (2007)	Deng et al. (2017b, 2018)
Shear zone in upper crust	Fault orientation & kinematics	Ductile weak zone in granular layer	Brun & Tron (1993); Osagiede et al. (2021)	
		Cylindrical PDMS seeds at granular-ductile layer interface	Zwaan et al. (2015, 2020; 2021a; 2021b); Molnar et al. (2019, 2020)	
Shear zone in lower crust	Fault orientation & kinematics; transfer zone width & orientation	Ductile weak zone in ductile layer	Corti (2008); Corti et al. (2004, 2013)	
		Velocity discontinuity (rigid basal plates)	Acocella et al. (1999)	
		Velocity discontinuity (rigid basal plate or plastic sheet separated by rubber sheet)	Withjack & Jamison (1986); McClay & White (1995)	
Pre-existing rift (thinned crust & strong lithospheric mantle)	Orientation & distribution of linkage structures between rift segments	Thin granular upper crust and ductile lower crust layers	Brune et al. (2017)	Brune et al. (2017)
Two rheologically distinct blocks	Fault spacing & propagation	Adjacent blocks with different ductile layer viscosities or thicknesses	Corti et al. (2013a); Beniest et al. (2018); Samsu et al. (2021)	Phillips et al. (2021b)
Crustal heterogeneity (pervasive)				
Fabric in upper crust	Fault orientation, length & kinematics	Zone of dilation in granular layer	Bellahsen & Daniel (2005)	
		Brushed plaster at model base underneath granular layer	Chattopadhyay & Chakra (2013); Ghosh et al. (2020)	
Fabric in ductile crust	Fault spacing, orientation, kinematics & propagation	Laterally alternating weak/normal zones in ductile layer	Samsu et al. (2021)	
Mantle heterogeneity (discrete)				
Shear zone in lithospheric mantle, terrane boundary, suture zone, orogenic belt, or thermally weakened zone in lithospheric mantle	Strain (rift) localisation; border fault orientation, length & kinematics; basin asymmetry & subsidence	Velocity discontinuity (basal plastic sheet)	Tron & Brun (1991); Brun & Tron (1993); Michon & Merle (2000); Corti et al. (2007); Zwaan et al. (2021a; 2021b)	Tommasi et al. (2009); Tommasi & Vauchez (2015); Petersen & Schiffer (2016)
		Weak zone in lithosphere/lithospheric mantle	Agostini et al. (2009); Molnar et al. (2017, 2018, 2019)	
Pre-existing rift basin	Barrier to fault propagation; fault localisation at edges	Strong zone in lithosphere/lithospheric mantle	Autin et al. (2013)	

2.2.1. <u>The influence of lithospheric and crustal strength variations on strain distribution</u>

The strength of the continental lithosphere and the rheology of its constituent layers exert first-order controls on the localisation of extension during rifting (Sokoutis et al., 2007). The presence of a high-strength lithospheric mantle in "cold and strong" lithosphere favours the end-member narrow rift (e.g., East African Rift System), while extension of a "hot and weak" lithosphere leads to a wide rift (e.g., Basin and Range province) (see Buck, 1991 and Brun, 1999 for a thorough discussion). Strain distribution in the cover is attributed to the coupling between the brittle and ductile layers in the model, which depends on the applied strain rate as well as the mechanical layering (i.e., thermal structure) of the lithosphere (Kusznir and Park, 1986; Cowie et al., 2005; Wijns et al., 2005; Zwaan et al., 2021b). This distribution determines where basin-bounding faults and accommodation space are created during rifting.

Modelling shows that rifting of heterogeneous lithosphere results in varying styles of deformation above different lithospheric or crustal blocks. Extension of two laterally juxtaposed crustal domains of different integrated strengths results in earlier strain localisation and more widely spaced and higher displacement faults in the weaker domain (Phillips et al., 2021b; Samsu et al., 2021). When the domain boundary strikes perpendicular to the extension direction, greater strain localisation above the weaker domain leads to rift margin asymmetry (Corti et al., 2013a; Beniest et al., 2018).

2.2.2. <u>Activation of inheritance mechanisms depending on the orientation, depth and rheology of weaknesses: modelling insights on reactivation vs. strain re-orientation</u>

The (re)activation of a pre-existing weakness in a model lithosphere facilitates strain localisation and partitioning and influences the orientation of rift-related faults. Whether or not a weakness is prone to (re)activation depends on several factors: (i) its obliquity with respect to the bulk extension direction, which is defined as the angle α measured between the strike of the weakness and the orthogonal to the extension direction (e.g., Agostini et al., 2009); and (ii) its strength relative to surrounding rocks, which may depend on depth (i.e., the rheology of the host layer). A hierarchy of inheritance is apparent in models where discrete weaknesses of variable orientations are present in more than one layer and compete with each other to localise strain. Weaknesses in the strong upper crust and lithospheric mantle layers are preferentially (re)activated over a weakness in the relatively weak lower crust (Chenin and Beaumont, 2013; Molnar et al., 2020). When weaknesses are present in both the upper crust and lithospheric mantle, their relative contributions to strain localisation are determined by the degree of mechanical coupling between the two layers. This coupling depends on kinematic boundary conditions (e.g., rate of divergence) and the thickness and strength of the intervening, weak lower crust (Zwaan et al., 2021b).

Models show that discrete and pervasive weaknesses in the brittle upper crust (Brun and Tron, 1993; Bellahsen and Daniel, 2005; Corti et al., 2007; Chattopadhyay and Chakra, 2013; Ghosh et al., 2020; Zwaan and Schreurs, 2020; Zwaan et al., 2021a, 2021b; see Table 2) undergo frictional reactivation (discussed from a geomechanics perspective in Section 2.3). This reactivation influences the orientations of rift-related faults (i.e., they show geometric similarity with the weaknesses) as well as their kinematics, growth, and linkage. Weaknesses that are at a high angle to the extension direction ($\alpha < 30°$) are reactivated in a normal sense, exhibiting mainly dip-slip kinematics and accommodating most of the extensional strain. Weaknesses that are at a lower angle to the extension direction are reactivated with a greater strike-slip component of movement; these give rise to secondary, extension-oblique faults that link the primary, extension-perpendicular faults (Bellahsen and Daniel, 2005; Osagiede et al., 2021).

Weaknesses in the ductile layer allow the model crust or lithosphere to locally thin more rapidly during extension, facilitating fault and graben localisation in the brittle upper crust above the weakness (Corti, 2004). By implementing a single weak zone in the ductile lithospheric mantle, some analogue models have explored the effects of weakness obliquity on fault orientations and strain localisation and partitioning, with implications for understanding inheritance at the scale of rift systems, + zone in the ductile lithospheric mantle (Corti et al., 2007; Corti, 2008, 2012; Agostini et al., 2009) or lower crust (Corti, 2004). These models show that ductile weaknesses oriented at high angles to the extension direction (α < 45°) guide the map-view strike of rift segments (Corti, 2008; Agostini et al., 2009), consistent with numerical modelling of the influence of pre-existing mantle weaknesses (e.g., Tommasi et al., 2009; Tommasi and Vauchez, 2015; Petersen and Schiffer, 2016). Ductile weaknesses at low angles to the extension direction (α > 45°) determine where rifts are segmented and transfer zones form (e.g., Molnar et al., 2020). Deformation in the overlying brittle layer is characterised by an *en échelon* pattern of faults at the boundary between the weak and normal-strength lithosphere or crust (i.e., boundary faults) in map view. The strike of individual faults is between the main rift trend (i.e., the trend of the linear weakness) and the bulk extension direction in both low and high-obliquity rifts (Agostini et al., 2009; Corti et al., 2013b) (Figure 6). The obliquity of these faults reflects the relative contributions of extension perpendicular to the rift trend and shear parallel to the rift trend (Withjack and Jamison, 1986). Early interpretations based on fault orientations alone suggested that these faults have oblique-slip kinematics (Agostini et al., 2009), which is a response to transtension under the imposed bulk extension boundary condition. Similar studies later showed that the extension-oblique faults actually display dip-slip kinematics (Corti et al., 2013b; Philippon et al., 2015), highlighting local slip re-orientation that is associated with the underlying ductile weakness. These insights from modelling support the interpretation of extension-oblique faults with dip-slip kinematics in some natural settings, including in the Main Ethiopian Rift (Philippon et al., 2015 and references therein), the Baikal Rift (Petit et al., 1996), and the Rukwa Basin in the East African Rift System (Delvaux et al., 2012). They also imply that strain re-orientation – an inheritance mechanism that is different to reactivation – can occur when a pre-existing ductile weakness is oblique to the far-field extension direction.

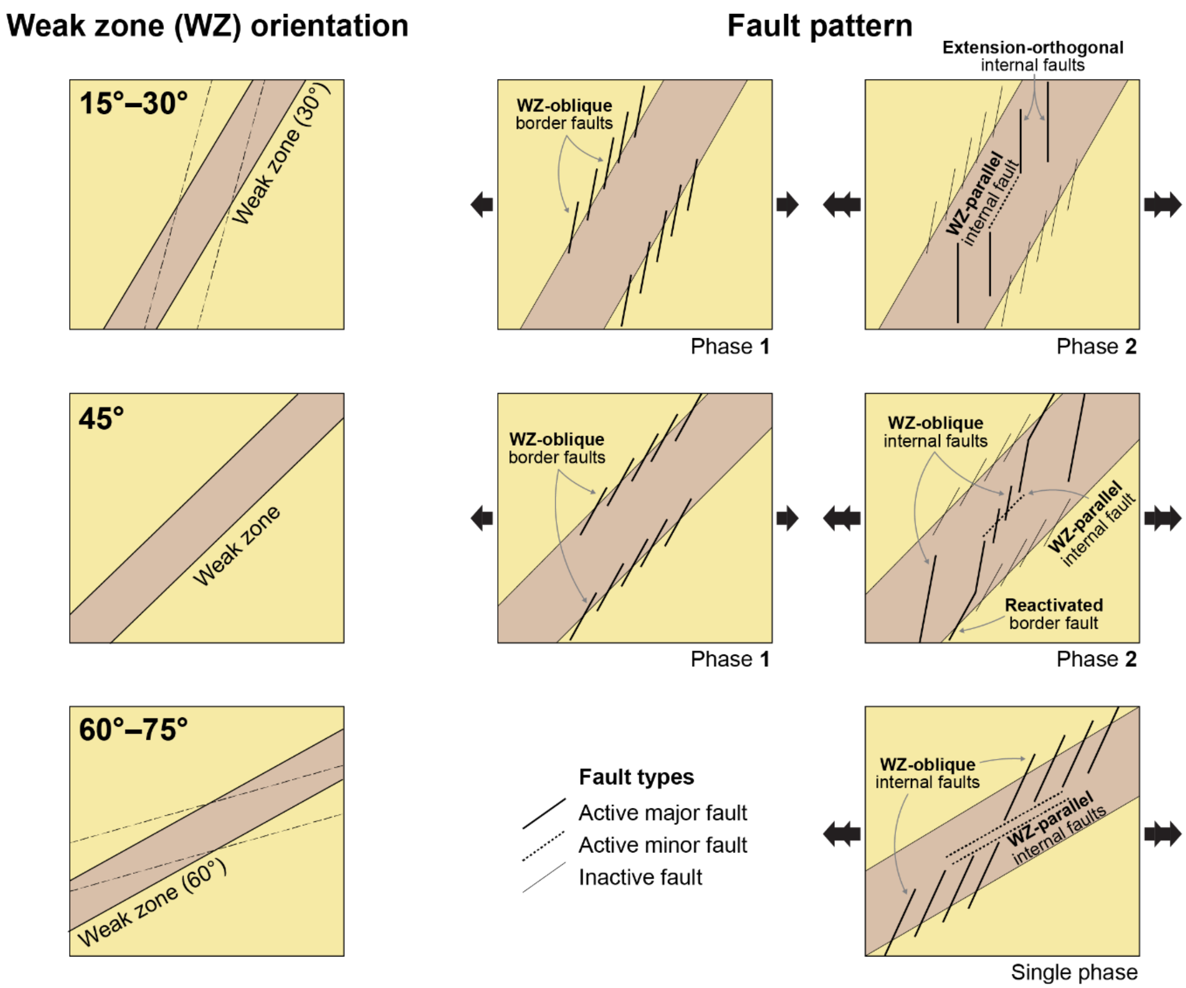

Figure 6 Map-view summary of analogue experiments showing differences in the evolution and pattern of rift-related faults above a ductile weak zone (WZ) that is oblique to the imposed extension direction (modified after Agostini et al., 2009). Thick black arrows indicate the extension direction. Major faults are oblique to the extension direction and weak zone irrespective of weak zone obliquity.

Pervasive anisotropies in the ductile layer may re-orient the far-field strain over a wider area, creating a complex pattern of extension-oblique faults in the brittle upper crust over the entire anisotropic domain. Samsu et al. (2021) presented an analogue experiment in which the orthorhombic fault pattern above a lower crustal anisotropy, oriented 45° to the imposed extension, is distinct from normal faults that developed above a homogeneous lower crust in the adjacent domain. They inferred that the interaction between the anisotropy and the imposed bulk extension direction resulted in localised rotation of the horizontal maximum stretching direction and non-Andersonian faulting above the anisotropic domain. The distinct fault patterns that developed above two different basement domains in this experiment are comparable with onshore fault pattern variations in the North Coast Transfer Zone (Scotland) which is underlain by basement rocks with different rheology and fabric orientations (Wilson et al., 2010). The non-coaxial vs. coaxial deformation patterns above different basement domains in these crustal-scale analogue models are also comparable with outcrop-scale observations of strain partitioning in the Carboniferous Northumberland Basin (northeast England) (De Paola et al., 2005). Here, minor faulting in the hanging wall of the reactivated Fathom Fault is partitioned into an extension-dominated regime in dolostones and wrench-dominated transtension in quartz-rich sandstones. De Paola et al. (2005) suggested that this lithological control on fault kinematics may be attributed to differences in the Poisson's ratio of different rock types.

Samsu et al. (2021) suggested that strain re-orientation in their experiments resulted from the strength contrast between the strong and weak domains – or in the case of an anisotropic basement, the contrast between the alternating strong and weak layers that make up the anisotropy. The strong and weak domains resist extension to different degrees, so that the difference in the rate of thinning across their boundary results in shearing at the strong-weak interface. This results in localised re-orientation of the 3D strain field at: (i) the vertical interface between strong and weak crustal domains; and (ii) above basement domains which are sufficiently anisotropic (i.e., where the width of the alternating strong and weak layers is below a critical threshold).

Similarities between crustal-scale models and basin to outcrop-scale field observations suggest that during rifting, strain re-orientation operates at a range of length scales and requires sufficient contrast in the mechanical properties of the extended heterogeneous medium. Future experiments involving anisotropy with various orientations, different thicknesses and strength ratios of the alternating strong-weak layers, and different kinematic boundary conditions, will increase our understanding of the influence of ductile basement fabrics on upper crustal deformation at the scale of individual basins and fault systems.

2.3. Geomechanical considerations of structural inheritance

One of the mechanisms of structural inheritance that we discuss in this review – reactivation – requires some degree of movement on pre-existing planes of weakness (e.g., pre-existing faults, shear zones, pervasive fabrics, or lithological contacts). For strain re-orientation, it is less clear whether such movement is required and in which situations; nevertheless, strain re-orientation has been observed to occur in conjunction with reactivation. In this section we review the conditions under which pre-existing planes of weakness in basement rocks are expected to undergo frictional failure and potentially contribute to inheritance.

The tendency for frictional slip to occur on planes of weakness (as opposed to new faults nucleating in intact rock; Byerlee, 1978; Butler et al., 1997; Holdsworth et al., 1997, 2001b) in the dominantly elastic upper crust is governed by the local stress state, the orientation of the planes of weakness (Jaeger and Cook, 1979; Morris et al., 1996; Lisle and Srivastava, 2004), and the effective frictional strength of these weaknesses relative to that of surrounding intact rock. Reduced effective frictional strength in pre-existing planes of weakness have been attributed to both pore fluid pressures in excess of hydrostatic fluid pressure, and lower cohesion or frictional properties due to grain-scale features such as: (a) interconnected zones of phyllosilicates or clays that provide low-friction, sliding surfaces (Byerlee, 1978; Holdsworth, 2004; Collettini et al., 2009, 2019); (b) grain boundary alignment and compositional banding (e.g., in gneisses; Shea and Kronenberg, 1993; Kirkpatrick et al., 2013); and (c) high grain boundary porosity and/or low intergranular cohesion in non-foliated rocks (e.g., non-cohesive cataclasites; Sibson, 1977).

For a rift basin forming in an idealised Andersonian normal fault stress regime (Anderson, 1905) with a vertical maximum principal compressive stress (σ_1), the fault population will be dominated by normal faults that strike perpendicular to the horizontal minimum principal compressive stress (σ_3) and dip between 58-68° (Collettini and Sibson, 2001). Much of our understanding of the geomechanics of inheritance in rift settings relates to deviations from this idealised scenario, in which pre-existing structures either strike obliquely to σ_3 and/or have a dip ≠ 58-68° (e.g., Etheridge, 1986; Ranalli and Yin, 1990; Massironi et al., 2011; Williams et al., 2019). Here, we examine 2D and 3D approaches to understand how slip occurs along pre-existing planes of weakness in such cases where the dip or strike is not optimally oriented for failure and frictional sliding.

2.3.1. Influence of frictional strength and fluid pressure on basement structures

In the case of a pre-existing plane of weakness that strikes orthogonal to σ_3, the plane of weakness contains the intermediate principal compressive stress (σ_2). Therefore, a 2D analysis can provide insights into the conditions required for reactivation, depending on the dip of the plane of the weakness. This analysis follows Sibson (1985), who defined an effective stress ratio for slip to occur;

$$R = \frac{\sigma_1'}{\sigma_3'} = \frac{(1+\mu_s cot\ \theta\ )}{(1-\mu_s tan\ \theta\ )} \qquad (1)$$

where μ_s is the static coefficient of friction, which is a measure of the friction between two cohesionless stationary surfaces that are in contact (Byerlee, 1978), and σ_1' and σ_3' are the maximum and minimum effective principal stresses (σ_1'= $\sigma_1 - P_f$ and σ_3'= $\sigma_3 - P_f$; where P_f is the pore fluid pressure). For pure dip-slip normal faulting in the Andersonian regime, θ is related to the dip of the fault or plane of weakness ($\theta = 90 - \delta$, where δ is dip).

The approach in Equation 1 hinges on the application of a simple linear failure criterion for cohesionless faults (Amonton's Law) containing σ_2' (the intermediate effective principal stress). Alternative failure criteria can be applied that incorporate the important effect of σ_2' (e.g., Extended Griffith, Drucker-Prager; Murrell, 1963; Haimson, 2006); nevertheless, the analysis provides a useful base case. For example, it can be used to identify three non-optimal failure regimes (Leclère and Fabbri, 2013; Figure 7 and Figure 8a), which can be compared against the R factor for an optimally oriented surface (R = R*; Figure 7a): (1) "Favourably oriented" surfaces have dips that deviate from the optimal orientation for failure by ~15° (e.g., 1-1.5R*, Figure 7b). (2) "Unfavourably oriented" surfaces have higher R values (e.g., R > 1.5R*) with dips misoriented by up to 25-30° from the optimal orientation (Figure 7c). (3) "Severely misoriented" surfaces are ones that require pore fluid pressures that are greater than σ_3 to reactivate (i.e., R is negative, Figure 7d).

A second outcome of Equation 1 is the significant influence of friction on the reactivation potential of pre-existing planes of weakness in basement rocks. Laboratory measurements have yielded friction coefficients as low as 0.2 for wet phyllosilicate and clay-rich fault gouges (Moore and Lockner, 2004; Numelin et al., 2007), compared to 0.5-0.85 for most other Earth materials tested in wet and dry conditions (e.g., Byerlee, 1978; Jaeger et al., 2007). This means that for structures rich in graphite, phyllosilicates, or clays (especially illite-smectite, chlorite, talc, biotite, and pyrophyllite), such as pre-existing basement faults, the range of dips that are optimally or favourably oriented for failure expands considerably (providing clay/phyllosilicate/graphite content >30-60%; Numelin et al., 2007). Basement faults with coefficients of friction as low as 0.3 are therefore able to reactivate, even if their dips are as low as 25° (e.g., low angle normal faults; e.g., Healy, 2009; Massironi et al., 2011; Demurtas et al., 2016; Singleton et al., 2018) or nearly subvertical, without requiring elevated fluid pressure (Figure 8b).

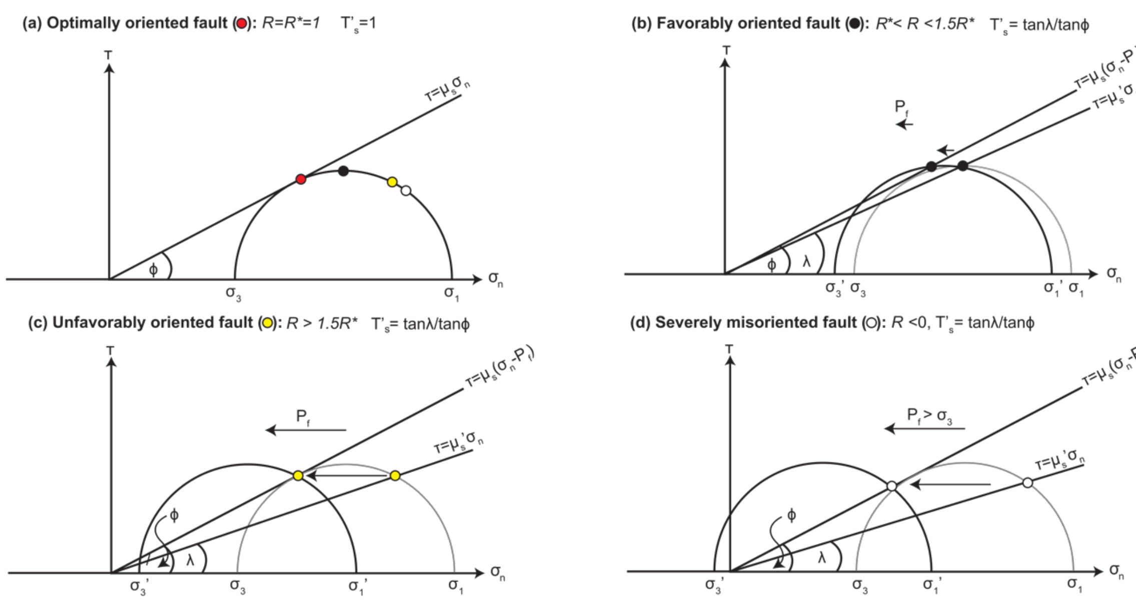

Figure 7 Illustration in Mohr Space (i.e., plots of shear stress (τ) versus normal stress (σ_n)) for reactivation analysis of cohesionless faults that contain σ_2. (a) The orientation of four faults with different reactivation potential as quantified by their stress ratio ($R = \sigma_1'/\sigma_3'$) and normalized slip tendency (T'_s; Lisle and Srivastava, 2004). For the case of the optimally oriented fault ($\theta = 30°$), R is denoted R^* and $T'_s = 1$. In (b)-(d), R and T'_s are schematically depicted for the reactivation of (b) favourably oriented, (c) unfavourably oriented, and (d) severely misoriented fault. In cases (b)-(d) a reduction in frictional strength (μ_s) and/or increase in pore fluid pressure (P_f) is required for fault reactivation. Modified after Williams et al. (2019).

Pore fluid factor diagrams (Figure 8c; Cox, 2010) are a useful alternate graphical method and can be used to assess the role of friction, fluid pressure, and stress on the potential for frictional sliding of basement structures. Based on this approach, Cox (2010) suggested that for the more unfavourably oriented normal faults (e.g. ~30° dip) with a typical frictional strength ($\mu = 0.75$), there is only a narrow range of differential stress in which it is possible for slip to occur on such faults, and that elevated pore fluid pressures are a pre-requisite. However, using the same approach, Figure 8c shows that when basement structures with low frictional strength are present, there is a wider range of stress conditions under which they are capable of sliding, with or without the presence of elevated fluid pressures. Figure 8b also implies that sliding on these lower friction structures can occur over a range of fault dip angles. These pre-existing weaknesses are likely to be the first structures to begin sliding at the inception of rifting and therefore have the potential to become important controls on subsequent basin geometries.

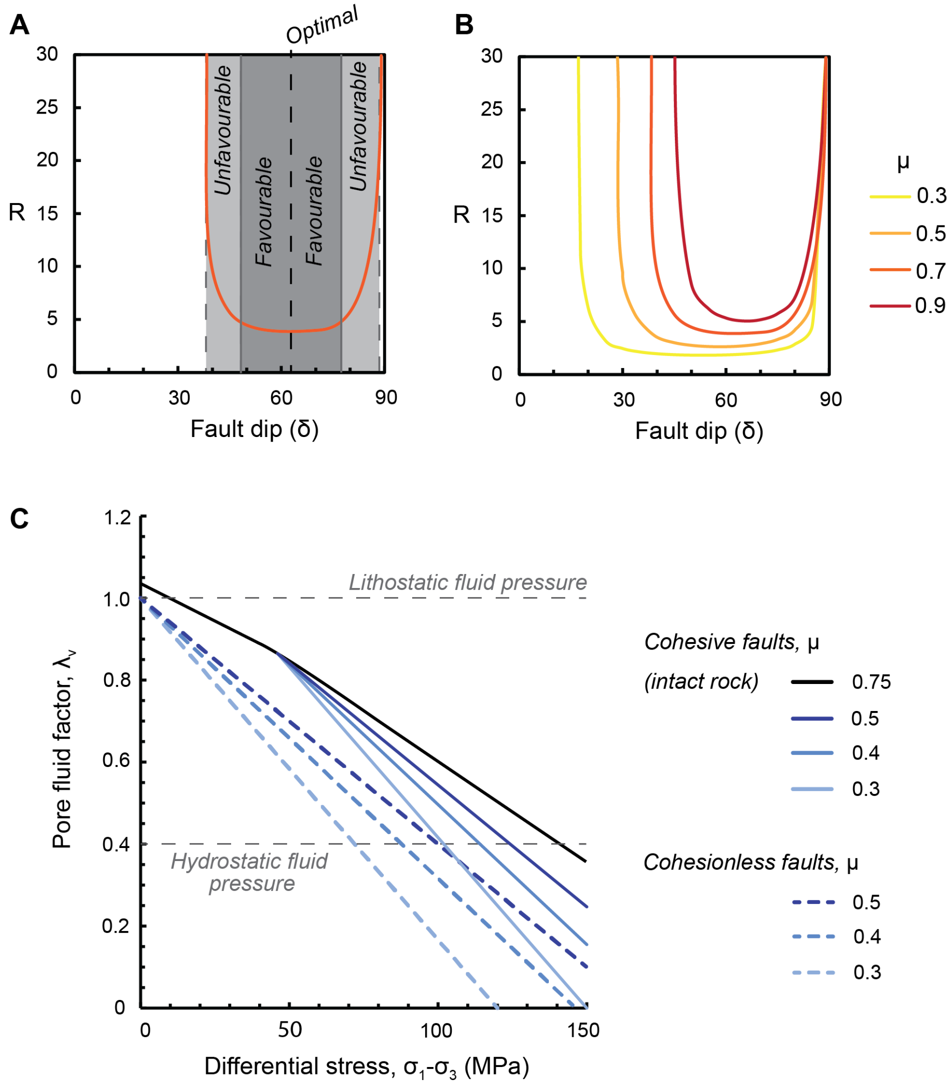

Figure 8 (A) Stress ratio required for frictional reactivation vs. fault dip for a cohesionless normal dip-slip fault (μ_s=0.7), in an Andersonian regime. Shown are the dip angles for the optimal fault orientation for (re)activation and the range of fault dips that are favourably oriented for failure. Also shown is the range of unfavourably oriented fault dips that require elevated fluid pressures to fail. (B) Reactivation potential for faults with low coefficients of static friction. The potential for failure of favourably oriented faults extends over a much wider range of dips. (C) Pore fluid factor-stress diagram (Cox, 2010), for normal dip-slip faults (10 km depth) that are optimally oriented for each value of static friction coefficient. Basement stress states increasing from near hydrostatic fluid pressure and low differential stress have the potential to reactivate pre-existing structures with lower coefficients of static friction, relative to intact rock, over a wide range of stress and fluid pressure conditions.

2.3.2. Extension of geomechanical analyses to 3D and implications for sedimentary basins

A 3D approach is required to examine the mechanics of structural inheritance when pre-existing basement structures strike obliquely to σ_3 (i.e., the plane of weakness does not contain σ_2) (Williams et al., 2019). For example, 'slip tendency analysis' (Morris et al., 1996; Lisle and Srivastava, 2004) provides a method to estimate the potential of a cohesionless plane of weakness to activate under any stress state (Figure 7). More recently, Leclère and Fabbri (2013) introduced a 3D solution for the effective stress ratio, *R*, for reactivation that accounts for cohesion and does not assume an Andersonian stress regime. Relatively few studies have used this 3D approach to investigate the reactivation potential of pre-existing structures that strike obliquely to the regional extension direction. Williams et al. (2019) applied the 3D solution of Leclère and Fabbri (2013) to the southern Malawi Rift to show that weaknesses with a non-optimal strike (at an angle >50° to the trend of σ_3) can still reactivate, even if there is only a small difference in the strength between intact rock and moderately dipping basement weaknesses ($\mu \sim 0.7$ and 0.55-0.65 for intact rock and weakness respectively) or a small increase in pore fluid pressure (effective pore fluid factor ~ 0.1–0.3, i.e., sub-hydrostatic pore fluid pressures), when they have favourable dip of ~50-60° and a stress shape ratio of ~0.5. Hence, a wide range of strike and dip orientations and fluid pressure conditions are likely under which weak basement surfaces can be reactivated.

The geomechanical considerations above enable us to be more specific about the conditions in which inheritance is likely to be triggered. Basement rocks comprising metamorphosed and deformed carbonaceous shales have the potential to contain graphite-rich planes of weakness (implying low μ_s = 0.2-0.4). Likewise, crystalline basement subjected to hydrothermal alteration or containing ultramafic rocks (ophiolites or Archean greenstone) may host phyllosilicate- and talc-rich planes of weakness.

It is also well established that many sedimentary basins develop elevated pore fluid pressures below a critical depth due to compaction disequilibrium and other mechanisms (Suppe, 2014; Zhang, 2019). Given that basinal fluids are known to infiltrate into the basement (Yardley et al., 2000; Gleeson et al., 2003), there are situations in which basement rocks may be saturated in hydrothermal fluids at elevated pressure. As such, elevated fluid pressures and low friction planes of weakness are common in basement rocks, meaning that structural inheritance during basin formation should be the norm rather than the exception.

3. DISCUSSION

3.1. Reactivation and strain re-orientation are the two main structural inheritance mechanisms in rift basin formation and faulting

2D and 3D observations from outcrops, geophysical data, and analogue and numerical models provide insight into the interactions between pre-rift basement rocks and rift-related structures, as summarised in Section 2. This synthesis suggests that two mechanisms underpin the control of basement structures on fault orientations and geometry in the overlying basin: **reactivation** and **strain re-orientation** (Figure 9). Notably both of these mechanisms can be active during the same rift phase, either along different basement structures within the same basin, or along the same basement structure at different depths.

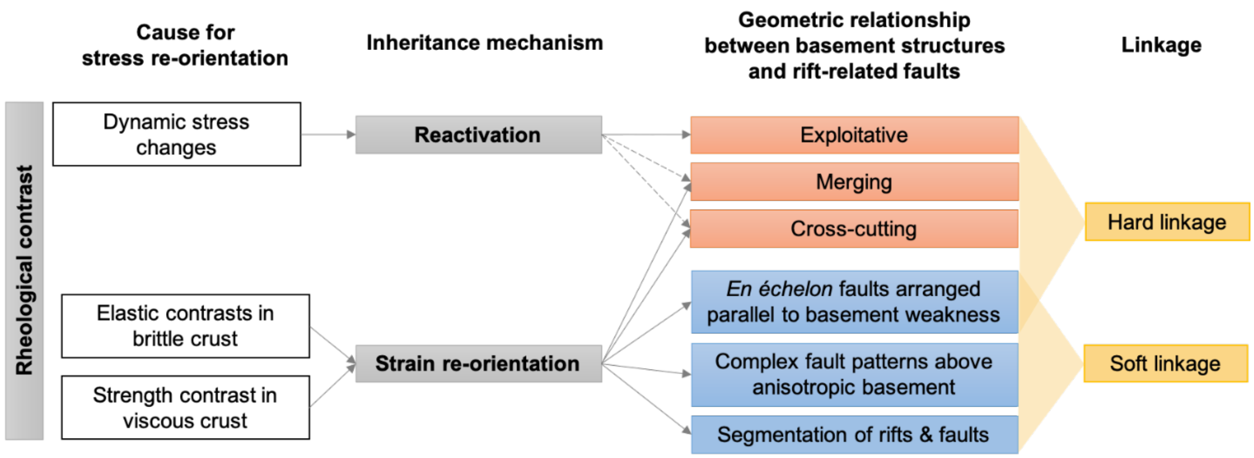

Figure 9 Classification of inheritance mechanisms (reactivation and strain re-orientation), their drivers, and their expressions. Rheological contrast is a pre-requisite for both mechanisms, which may be associated with local perturbations of the far-field stress. Such stress perturbation can be driven by dynamic stress changes due to movement along a pre-existing weakness or by elastic and strength contrasts in the brittle and viscously deforming crust, respectively. The two inheritance mechanisms can be identified based on the 2D and 3D geometric relationships between pre-rift basement structures and rift-related faults, which have been further grouped into hard and soft linkage (referred to as hard-linked and soft-linked inheritance, respectively, in the text).

In the context of this discussion, reactivation implies slip along a pre-existing weak surface in the brittle crust. During rifting, a pre-existing basement weakness can be a nucleation site from which a new fault propagates into the overlying cover (e.g., Collanega et al., 2019) (Figure 2a). Strain re-orientation refers to non-coaxial deformation patterns, which result from partitioning of the far-field extensional strain at the interface between two mechanically distinct rock units (e.g., adjacent basement domains). Bulk stretching of anisotropic basement – where the anisotropy is oblique to the far-field stretching direction – is accommodated by different rates of extension in the strong and weak rocks, resulting in shearing along the strong-weak boundary (e.g., Samsu et al., 2021). Strain re-orientation can occur in conjunction with movement along a pre-existing structure (i.e., reactivation) (Phillips et al., 2016).

Both reactivation and strain re-orientation may be associated with local perturbations of the far-field stress around a pre-existing structure (i.e., local stress re-orientation) (Section 2.1.6). Stress re-orientation has been observed and modelled in three scenarios: (i) dynamic stress changes due to movement along a fault or shear zone (Barton and Zoback, 1994); (ii) deflection of stress trajectories due to elastic contrasts in the brittle crust (Figure 5a and Figure 5b) between a pre-existing structure and its host rock or between two adjacent basement domains (Bell, 1996; de Joussineau et al., 2003; Morley, 2010); and (iii) strength contrasts in the viscous crust that results in strain re-orientations (Samsu et al., 2021) which may be transmitted as stress re-orientations to an overlying mechanically coupled brittle crust. We therefore suggest that rheological contrast is a prerequisite for inheritance, which must be present for reactivation and/or strain re-orientation to occur. In sedimentary basins, a locally rotated stress field within the basement can be transferred to the overlying basin fill when the two units are mechanically coupled. If, however, an intervening weak layer separates basement and cover, then stress coupling between the basement and cover rocks is limited (Figure 5c).

Reactivation and strain re-orientation can occur concurrently during a single rift episode but result in different spatial and geometric relationships between the pre-existing basement structures and rift-related faults (Figure 9). Applied to the offshore southern Norway study area of Phillips et al. (2016), the distinction between these mechanisms and the extent of their influence may explain why rift-related faults exploit, merge with, or cross-cut different basement shear zones within the same basin. Strain re-orientation can also explain the occurrence of rift-related faults that form above – but do not exhibit geometric similarity with – a pre-existing basement weakness or anisotropy. The range of geometric relationships associated with strain re-orientation emphasises that reactivation is not synonymous with structural inheritance, of which strain re-orientation is a common and important mechanism.

Whether reactivation or strain re-orientation occurs may depend on whether a pre-existing structure behaves in a frictional or viscous manner. Discrete, pre-existing weaknesses in the brittle upper crust undergo frictional reactivation when they are appropriately oriented relative to the far-field stress (Section 2.3). On the other hand, weak zones with relatively low viscosity in the viscous lower crust or lithospheric mantle influence deformation by localising and/or re-orienting strain, as demonstrated by brittle-ductile analogue models (Section 2.2.2). In natural rifts, this distinction may be used to determine whether the inherited weakness lies in the frictional or viscous crustal regimes (cf. Holdsworth et al., 2001a).

3.2. Hard and soft-linkage classification

The natural examples presented in Section 2.1 demonstrate an array of geometric relationships between pre-existing basement structures and rift-related faults (Figure 2). We suggest that these relationships can be classified into hard- and soft-linked inheritance, to denote whether a rift-related fault is physically connected to a pre-existing basement structure (Figure 9): **Hard-linked inheritance** is characterised by a physical link between a pre-existing basement structure and a younger, rift-related fault in the sedimentary cover. **Soft-linked inheritance** is defined as the apparent influence of a pre-existing basement structure on a rift-related fault, where a physical link between the two is not observed. We note that this classification does not describe the genetic relationship between the pre-existing and rift-related structures or the inheritance mechanism (Figure 9).

3.2.1. Hard-linked inheritance

A shallow fault and a deeper, pre-existing structure are classified as hard-linked if they have identical strike and dip direction and an exploitative, merging, or cross-cutting relationship (sensu Phillips et al., 2016; Section 2.1.2) (Figure 9). This classification is based on observations made in cross section, where there is a physical connection between the rift-related fault and the pre-existing structure. An exploitative relationship (Figure 2a) relies on reactivation of a pre-existing weakness, associated with previous fault or shear zone-forming processes. A merging relationship (Figure 2b) involves strain re-orientation and may involve some degree of reactivation; in the example documented by Phillips et al. (2016), the basement structure did not exhibit significant rift-related displacement at the scale of observation. Similarly, hard-linked inheritance may apply to examples of the cross-cutting relationship where the rift-related fault inherits the strike and dip direction of the basement structure (Rotevatn et al., 2018).

The relative contributions of strain re-orientation and reactivation mechanisms to the merging and cross-cutting relationships described above are exemplified by rift-related normal faults that inherit the strike of a nearby, pre-existing basement shear zone with a gentler dip than the rift-related normal faults (e.g., Fazlikhani et al., 2017; Rotevatn et al., 2018; Osagiede et al., 2020). Geometric similarity between the rift-related faults and a pre-existing shear zone may be initiated by strain re-orientation, for example caused by a grain-scale mechanical anisotropy that formed in the same deformation event as the shear zone (Kirkpatrick et al., 2013). Such a pervasive fabric records distributed strain, and movement along the weak layers within the fabric enables strain re-orientation. Evidence from in-situ stress measurements and earthquake focal mechanisms suggest that such strain re-orientation is associated with local re-orientation of the far-field maximum (and minimum) horizontal stress trajectories by the mechanically weak zones (Bell, 1996; Morley, 2010).

Geometric similarity, in association with hard-linked inheritance, may be limited to deeper cover units, which are closer to the pre-existing basement structure. This depth-dependent relationship is demonstrated by rift-related faults that are segmented and oblique to a pre-existing basement structure in the upper part of the cover but appear to merge into a continuous structure striking parallel to the basement structure at depth (Deng et al., 2017b; Collanega et al., 2019). The shallow faults exhibit an *en échelon* arrangement that strikes parallel to the basement structure and perpendicular to the far-field extension direction (Figure 4). In-situ stress measurements show that changes in fault strike with increasing vertical distance from a pre-existing basement weakness can be attributed to local stress re-orientation by the basement structure, which is mechanically weak compared to the surrounding rock (Tingay et al., 2010b). Near the surface, the far-field stresses have a greater influence than at depth (Yale, 2003), contributing to shallow rift-related faults that strike perpendicular to the minimum horizontal far-field stress.

3.2.2. Soft-linked inheritance

The *Soft-linked inheritance* classification applies when rift-related faults are spatially co-located with a mapped basement structure, but the fault strikes are oblique to both the basement structure and inferred regional extension direction (e.g., Samsu et al., 2019) and/or we cannot observe a hard linkage between the basement structure and rift-related fault in cross section (e.g., Phillips et al., 2021). Analogue experiments suggest that such 'misoriented' faults form when the pre-existing basement structures behave in a viscous as opposed to frictional manner during rifting, regardless of whether they are discrete or pervasive. Models with a discrete weakness in the ductile lower crust, which strikes ≤45° relative to the extension direction, demonstrate the formation of extension-oblique faults in the overlying crust (Agostini et al., 2009; Corti et al., 2013b) (Figure 6). Dip-slip kinematics and slip re-orientation along the extension-oblique faults (Philippon et al.,

2015) reflect strain re-orientation near the boundary between the weaker and stronger lower crust domains and in the overlying upper crust. Strain re-orientation can also occur across a wider area when the analogue ductile basement is mechanically anisotropic, with pervasive, closely spaced weaknesses that are oblique to the far-field extension direction (Samsu et al., 2021). The anisotropy gives rise to transtensional strain, even under boundary conditions that simulate orthogonal rifting, resulting in sets of non-Andersonian faults with traces that are oblique to the anisotropy in map view.

The presence of a relatively weak, ductile layer (e.g., clay, salt) can decouple the basement from cover units, as observed in seismic reflection data and through in-situ stress measurements (Bell, 1996). Where an intervening weak layer is present between two cover units, this decoupling effect can result in a combination of soft-linked and hard-linked inheritance. In this case, the basement weakness is geometrically similar to rift-related faults below the mechanically weak layer but geometrically dissimilar to an *en échelon* fault array above the weak layer (e.g., Jackson and Rotevatn, 2013; Roche et al., 2020; Phillips et al., 2021a).

The spatial co-location of inferred, deep-seated basement structures with transfer zones (a.k.a. accommodation zones) that separate distinct rift segments and basins (e.g., Rowland and Sibson, 2004; Fossen et al., 2016) can also be described as soft-linked inheritance. The absence of cover faults parallel to the inferred basement structures imply that the basement structures were not directly reactivated and are not hard-linked to any of the rift-related faults. Therefore, we can infer that the basement structure, which strikes at a high angle to the main rift trend, is not favourably oriented for reactivation but locally re-orients the far-field extension direction. While it is widely acknowledged that such pre-existing structures contribute to continental-scale rift segmentation, there is scope for further exploring the relationship between soft-linked inheritance and the formation of sub-basin-scale relay structures (Fossen and Rotevatn, 2016).

3.3. Applying the structural inheritance framework to the East African Rift System

The East African Rift System (EARS) is the pre-eminent example of an active continental rift, exhibiting all stages of continental rift evolution from nascent seafloor spreading in Afar to incipient faulting in the Okavango Rift of Botswana (Figure 10a; McConnell, 1972; Chorowicz, 2005; Ebinger, 2005; Macgregor, 2015; Daly et al., 2020). Hard- and soft-linked structural inheritance has influenced the evolution of the EARS at many spatial scales (and also magma emplacement mechanisms and volcanic activity; see Corti et al., 2022 for examples from the Main Ethiopian Rift). A famous example of plate-scale inheritance is present south of the Main Ethiopian Rift, where the Eastern and Western Branches of the EARS wrap around the relatively rigid Archean Tanzanian craton and spatially co-locate with orogenic belts that formed during the progressive Proterozoic amalgamation of the African Continent (Figure 10a; Daly et al., 1989; Versfelt and Rosendahl, 1989; Nyblade and Brazier, 2002; Corti et al., 2007). New geodetic and geologic evidence indicates that this is just one of several cases of plate-scale structural inheritance where the EARS bifurcates to co-locate with pre-existing weak zones around Archean cratons; in this context, the EARS represents inheritance of relatively weak, deformed Proterozoic lithosphere bounding more rigid blocks (Figure 10a; Daly et al., 2020; Stamps et al., 2021; Wedmore et al., 2021).

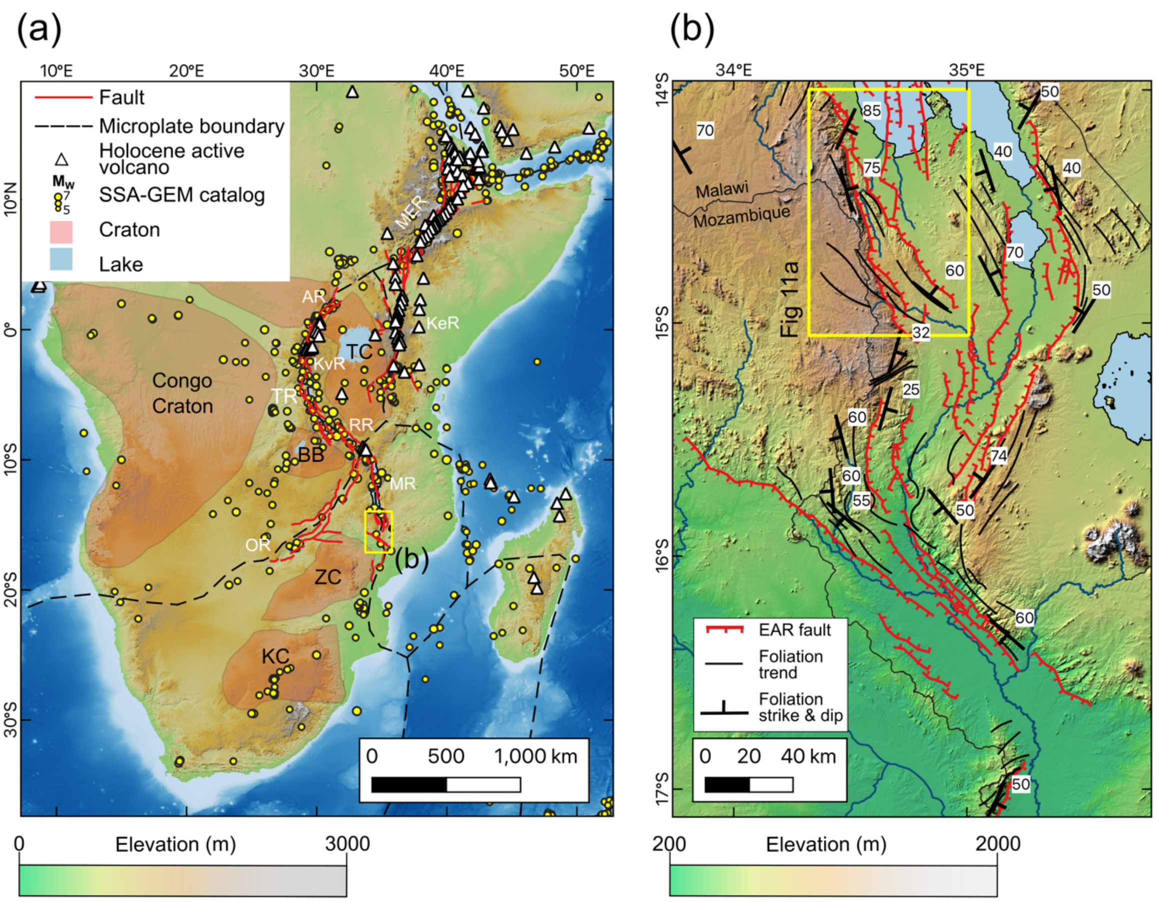

Figure 10 Plate and basin scale structural inheritance in the East African Rift System (EARS). (a) Distribution of active faults (Hodge et al., 2018; Daly et al., 2020; Styron and Pagani, 2020), microplate boundaries (Wedmore et al., 2021), Holocene volcanoes (Global Volcanism Project, 2013), earthquake events ($M_W \geq 5$, 1875-2015) from the Sub-Saharan Africa Global Earthquake Model (SSA-GEM) catalogue (Poggi et al., 2017), and Archean-Paleoproterozoic cratons in the EARS (Van Hinsbergen et al., 2011). MER; Main Ethiopian Rift; KeR, Kenya Rift; AR, Albertine Rift; KvR, Kivu Rift; TR; Tanganyika Rift RR, Rukwa Rift; MR, Malawi Rift; OR, Okavango Rift; TC; Tanzanian Craton; BB, Bangweulu Block; ZC, Zimbabwe Craton; KC, Kaapvaal Craton. (b) Surface traces of faults and foliation in southern Malawi (Williams et al., 2019, 2021b; Kolawole et al., 2021a). Foliation trends and strike and dip measurements from (Williams et al., 2019 and references therein). (a) Underlain by GTOPO30 Digital Elevation Model (DEM) and (b) by Shuttle Radar Topography Mission DEM (Sandwell et al., 2011).

At the ~10-100 km scale, the EARS can be divided along strike into distinct half-graben or graben basins. It is debated whether the along-strike extent of these basins is guided by the intrinsic strength of the lithosphere or pre-existing rift-perpendicular structures (Ebinger et al., 1987;

Ebinger, 1989; Upcott et al., 1996; Katumwehe et al., 2015; Laó-Dávila et al., 2015; Heilman et al., 2019; Scholz et al., 2020; Corti et al., 2022); nevertheless, in either case the patterns of faulting within these basins can be linked to structural inheritance. The most commonly described examples are fault traces that are subparallel to surface foliations (i.e., 'geometric similarity') imparted by Proterozoic orogenic events in East Africa, as observed, for example, in the Malawi (Ring, 1994; Laó-Dávila et al., 2015; Dawson et al., 2018; Hodge et al., 2018; Kolawole et al., 2018; Williams et al., 2019; Scholz et al., 2020; Shillington et al., 2020; Wedmore et al., 2020b), Albertine (Katumwehe et al., 2015), Kivu (Smets et al., 2016), Kenya (Smith and Mosley, 1993; Robertson et al., 2016; Muirhead and Kattenhorn, 2018), the Main Ethiopian (Corti et al., 2022), Okavango (Kinabo et al., 2007), and Rukwa Rifts (Morley, 2010; Delvaux et al., 2012; Heilman et al., 2019; Kolawole et al., 2021b). East Africa also experienced a Mesozoic (or 'Karoo') phase of rifting related to Gondwana fragmentation. The segmentation and orientation of Karoo structures was also affected by Proterozoic structures and fabrics (Ring, 1994; Delvaux, 2001; Paton, 2006; Bingen et al., 2009). In turn, Karoo structures have been reactivated during East African rifting (Castaing, 1991; Macgregor, 2015; Daly et al., 2020; Wedmore et al., 2020b) or have influenced rift segmentation (Accardo et al., 2018), depending on their orientation relative to the regional stresses.

The 3D geometric relationship between faults and fabrics in the EARS at depth is uncertain; however, it is likely that where fabrics are moderately dipping (e.g., southern Malawi Rift; Wedmore et al., 2020) relationships are exploitative (sensu Phillips et al., 2016), whilst merging or cross-cutting relationships are present in regions where the fabrics are subvertical (e.g., northern Malawi Rift; Dawson et al., 2018; Kolawole et al., 2018). Geometric similarity is not ubiquitous within the EARS. For example, faults in the Tanganyika Rift and central Malawi Rift are not parallel to surrounding crustal fabrics (Muirhead et al., 2019; Scholz et al., 2020), and the degree to which metamorphic fabrics influence fault orientations can change as rift extension progresses (Muirhead and Kattenhorn, 2018; Nutz et al., 2022). Furthermore, local variations in fabric orientation may disrupt geometric similarity at the scale of an individual fault, resulting in faults that locally cross-cut fabrics, along-strike fault segmentation, and/or faults that are Z-shaped in plan-view due to scarps that are continuous across perpendicular bends (Figure 11; Hodge et al., 2018; Corti et al., 2022).

When invoking a mechanism for the geometric similarity in the EARS it is not always clear if: (1) fabrics are non-optimally oriented to the regional extensional direction, but reactivate because they are frictionally weak and/or incohesive (i.e., hard-linked exploitative reactivation, Figure 2a; (Dawson et al., 2018; Wedmore et al., 2020), (2) fabrics are optimally oriented to the regional extension direction and so geometric similarity is a coincidence (Baker et al., 1972; Smith and Mosley, 1993), or (3) non-optimally oriented fabrics are actively rotating the local (i.e., at the scale of the fault; Twiss and Unruh, 1998) extension direction, so that foliation-parallel faults exhibit dip-slip displacement despite striking oblique to the regional extension direction (Tingay et al., 2010b; Corti et al., 2013b, 2022; Muirhead and Kattenhorn, 2018; Williams et al., 2019). Analogue models imply that the latter is an example of soft-linked structural inheritance as these local extension directions reflect deep-seated (not surface or shallow) weaknesses in the crust (Corti et al., 2013b; Philippon et al., 2015). Therefore, mechanisms (1) and (3) could apply at different depths to the same fault (Hodge et al., 2018; Wedmore et al., 2020b).

Structural inheritance in the EARS is not limited to geometric similarity. For example, as rift extension proceeds in the EARS, it is typically observed that strain migrates from large basin-bounding faults to a network of smaller intrabasin faults in the rift valley (Ebinger, 2005). Early localisation of rift-related strain onto a border fault may be facilitated by the exploitation of a discrete pre-existing weakness (e.g., a terrane boundary or basement viscous shear zone; Wheeler and Karson, 1989; Katumwehe et al., 2015; Scholz et al., 2020; Wedmore et al., 2020b; Kolawole et al., 2021b). Alternatively, pervasive lateral heterogeneities in the crust (e.g., a wide anastomosing shear zone) can promote distributed deformation involving migration of extensional strain from the basin boundary to intrabasin faults (Kolawole et al., 2018, 2021b; Wedmore et al., 2020a). Normal fault lengthening by exploitation of pre-existing fabrics (Section 2.1.4; Walsh et al., 2002) may account for why faults in the EARS achieved their full length at an early stage of their displacement accumulation (Vétel et al., 2005; Accardo et al., 2018; Corti et al., 2019; Ojo et al., 2022) and exhibit narrow fault damage zones and large single-earthquake displacement-to-length ratios (Figure 11) (Hodge et al., 2020; Williams et al., 2022). An important observation at the $<$ 100 km scale is therefore that although the EARS is a classic example of plate-scale structural inheritance and co-location with relatively weak lithosphere (e.g., Versfelt and Rosendahl, 1989), this inheritance is not synonymous with reactivation. Instead, reactivation is limited to where discrete, relatively weak structures are available in an orientation that is favourable for reactivation as the dominant (frictional or viscous) deformation mechanism (Wheeler and Karson, 1989; Kolawole et al., 2018; Heilman et al., 2019). However, at depths or locations where such structures are not available, the orientations and dimensions of new structures are locally variable and likely related to basement anisotropy (Hodge et al., 2018; Williams et al., 2019; Wedmore et al., 2020b).

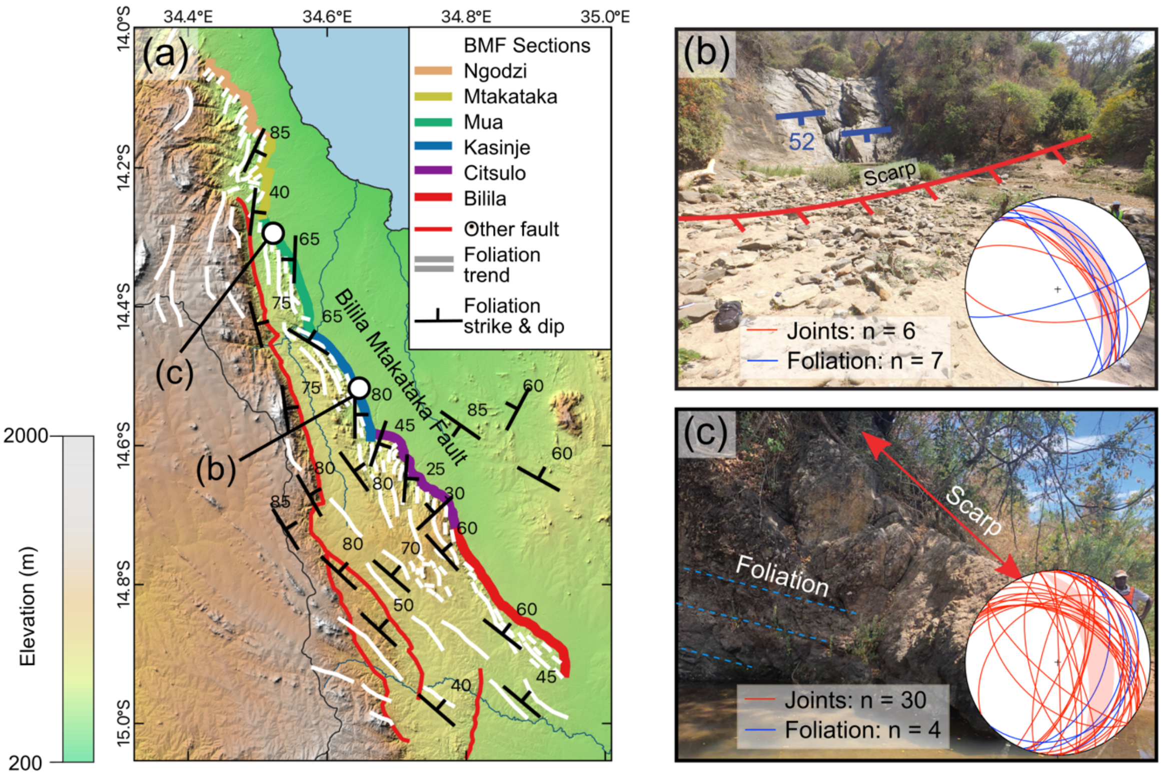

Figure 11 Fault-scale structural inheritance in the EARS using the example of the Bilila-Mtakataka Fault (BMF) in the southern Malawi Rift (Jackson and Blenkinsop, 1997; Hodge et al., 2018). (a) Map of the BMF and surrounding surface foliation orientations (Hodge et al., 2018 and references therein). The relative orientation of the fault, surrounding foliation, and along strike minima in fault scarp height have been used to divide it into the shown segments (Hodge et al., 2018). Extent of figure shown in Figure 10b. Field examples of where the BMF is (b) parallel and (c) oblique to the surrounding foliation (Williams et al., 2022). Equal area stereonets indicate relative orientations of foliation and joints in the fault's surrounding damage zone, shaded area indicates local trend of BMF scarp and a range of plausible fault dips (40-65°; Hodge et al., 2018; Stevens et al., 2021; Williams et al., 2022).

Seismic hazard investigation is a practical application for the study of structural inheritance in the EARS. For example, the influence of structural inheritance on the distribution of extensional strain between border and intrarift faults will dictate whether regions of high seismic hazard are in the basin interior or along the margin(s) (Dawson et al., 2018; Wedmore et al., 2020a; Williams et al., 2021b). Pre-existing crustal fabrics may also modulate the along-strike segmentation of faults, and this can affect whether faults rupture in relatively frequent, moderate-magnitude segmented fault ruptures, or more uncommon whole-fault ruptures (Biasi and Wesnousky, 2016; Hodge et al., 2018; Wedmore et al., 2020b).

3.4. Implications of structural inheritance for natural resources

The activation of pre-existing weaknesses during rifting has wide ranging implications from a geo-energy and minerals exploration perspective, some of which are highlighted in this section. The extraction of geothermal energy and the formation of hydrothermal mineral systems rely on hydrothermal fluid circulation, which is affected by the spatiotemporal pattern of deformation. For example, heat and mass transfer from deep sources to shallower depths are facilitated by convection through networks of open fractures (e.g., Rowland and Simmons, 2012). These fractures are especially important in rocks with low primary porosity and permeability, such as crystalline basement. At the plate scale, tectonic inheritance controls the location and lithology of sedimentary basins as well as the preservation of fluid pathways between deep heat and metal sources and shallow reservoirs (e.g., Hoggard et al., 2020). At the basin scale, rifting can bring about favourable stress conditions for reactivating pre-existing faults as permeable fractures.

The Upper Rhine Graben is an example of a deep geothermal system where the most permeable reservoir is located at the top of the granitic basement (e.g., Vidal and Genter, 2018; Glaas et al., 2021). NNW-SSE to N-E striking Variscan faults and fabrics in the basement may have been reactivated in a normal sense during the basin's multiphase history of Cenozoic regional shortening and extension (e.g., Schumacher, 2002), contributing to thick (i.e., wide-aperture) fractures with enhanced permeability (Glaas et al., 2021). Bertrand et al. (2018) show that in the Upper Rhine Graben, pre-existing faults, fabrics, and lithologies can impact fracture patterns at certain scales but not in others, demonstrating that the impact of inheritance on flow modelling is scale dependent. Structural mapping, with an aim to understand the multi-scale controls of inheritance on fractures, has been applied to other geothermal systems, including in France (Dezayes et al., 2010), Mexico (Norini et al., 2019), and the UK (Yeomans et al., 2020).

In the Taupō Volcanic Zone (New Zealand), structural inheritance appears to control shallow geothermal systems and the formation of hydrothermal gold and silver deposits at <2,000 m depths (i.e., epithermal ore deposits). Upwelling of hot water plumes is enhanced at the intersections of mapped or inferred structures, including faults, basement-controlled transfer zones, and caldera boundaries (Rowland and Sibson, 2004; Rowland and Simmons, 2012; Villamor et al., 2017; Milicich et al., 2021). Rowland and Sibson (2004) associated the segmentation of the NNE-SSW trending rift system, via so-called accommodation zones, with WNW-ESE trending basement structures interpreted from geophysics. The same geometric relationship has also been inferred for the similarly NNE-SSW trending Coromandel Volcanic Zone farther north (Bahiru et al., 2019). Rowland and Simmons (2012) suggested that such basement structures may be physically linked with potentially permeable shear zones in the lower crust, enabling fluid transport across the brittle-ductile transition zone (Cox et al., 2001), though at present there is no evidence of hard linkage between the basement structures and shallower rift-related structures.

Inheritance of lithospheric boundaries and crustal-scale weaknesses during rifting contributes to the formation of world-class sediment-hosted base metal deposits (i.e., copper, lead, zinc, and nickel) near craton edges (e.g., Mount Isa, Australia; Gibson et al., 2016; Hoggard et al., 2020). Here, the long-lived nature of the craton edge is attributed to focusing of deformation over multiple extensional and contractional events. Rifting of thick cratonic lithosphere facilitates the formation of deep, widely spaced faults that remain active (through reactivation) for up to 100 Myr, extending the time window for mineralisation (Allen and Armitage, 2012) and the distribution of basin fill lithologies that are conducive to mineralisation (Hoggard et al., 2020 and references therein). In addition, reactivation of crustal structures maintains focused fluid flow between the deep and shallow levels of basins (Gibson et al., 2016).

Our understanding of tectonic inheritance and fault/fabric reactivation have been integrated into studies on rift evolution and basin formation in the context of paleotectonic reconstructions (e.g., Gouiza and Paton, 2019; Heron et al., 2019) and hydrocarbon exploration (e.g., Morley, 1995; Lyon et al., 2007; Whipp et al., 2014). Studies on the East African Rift System have also demonstrated the role of inherited structures in modulating seismic hazard (refer to Section 3.3). The examples we discuss here suggest that there is scope for applying our knowledge of lithospheric and crustal-scale controls of inheritance to geothermal and mineral systems, which can contribute to successful exploration and development of geothermal energy, base metals, and critical metals. Basement structure characterisation and quantification of the reactivation potential of pre-existing fractures (including faults) under in-situ stresses must also be implemented in geothermal drilling projects (Deichmann and Giardini, 2009; Diehl et al., 2017), in addition to exploring the feasibility of geological storage of CO_2 (e.g., Andrés et al., 2016) and nuclear waste (e.g., Barton and Zoback, 1994) to support the current energy transition.

4. CONCLUSION

Compositional heterogeneities and mechanical discontinuities in pre-deformed crust can locally alter the local stress or strain field. This structural inheritance, which has been observed from the plate scale down to the outcrop scale, is probably the norm rather than exception, and it influences the location and geometries of entire rift systems, basins, and faults during rifting. Lithospheric and crustal-scale zones of weakness facilitate localised thinning, contributing to the formation and propagation of rifts along inherited higher strain belts. At the same scale, accommodation zones are spatially co-located with inferred deep-seated structures that strike at high angles to the main rift. The boundaries and evolution of rift basins are controlled by boundary faults that, if inconsistent with the far-field strain, may indicate exploitation of a reactivated basement weakness and/or reflect a locally re-oriented strain field above a deep, potentially viscously deforming structure. Individual faults at the sub-basin scale may also exploit weak surfaces in the basement or a pre-existing rift fabric. Some faults may appear to be unaffected by inheritance and reflect the far-field regional extension, suggesting that the vertical and lateral extents of stress and strain re-orientation are limited by the depth, size, orientation, and strength (relative to the surrounding rocks) of the pre-existing structure.

Based on our review of previous field and modelling work, we have distilled observations of structural inheritance into two key mechanisms: reactivation and strain re-orientation. One or both of these mechanisms are activated when extension affects two mechanically distinct basement terranes or occurs in the presence of an anomalously weak (or strong) structure within relatively homogeneous-strength basement rock. Reactivation is generally associated with geometric similarity and hard linkage between the pre-existing structure and the younger, rift-related fault. These observations are consistent with many of the expressions of inheritance found around the world, most notably in the Atlantic and the East African Rift System (Table 1). However, strain re-orientation is invoked when it comes to certain observations of soft linkage, particularly where a rift-related fault is oblique to a pre-existing structure (e.g., Tingay et al., 2010b; Giba et al., 2012; Collanega et al., 2019; Phillips et al., 2021a). While strain re-orientation is not as readily recognisable as reactivation, further understanding of the conditions under which strain re-orientation applies can help us explain the presence of complex fault patterns in rift basins, better constrain the far-field paleostrain in ancient rifts, and more confidently map basement structures under cover.

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