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Key Points:

- We test the idea that rifts of a given age share some common features
- We discuss the relationships between the style of rifts and their development
- We discuss the respective role of mantle in rifting trigger

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Style of rifting and the stages of Pangea breakup

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Abstract Pangea results from the progressive amalgamation of continental blocks achieved at 320 Ma. Assuming that the ancient concept of “active” versus “passive” rifting remains pertinent as end-members of more complex processes, we show that the progressive Pangea breakup occurred through a succession of rifting episodes characterized by different tectonic evolutions. A first episode of passive continental rifting during the Upper Carboniferous and Permian led to the formation of the Neo-Tethys Ocean. Then at the beginning of Triassic times, two short episodes of active rifting associated to the Siberian and Emeishan large igneous provinces (LIPs) failed. The true disintegration of Pangea resulted from (1) a Triassic passive rifting leading to the emplacement of the central Atlantic magmatic province (200 Ma) LIP and the subsequent opening of the central Atlantic Ocean during the lowermost Jurassic and from (2) a Lower Jurassic active rifting triggered by the Karoo-Ferrar LIP (183 Ma), which led to the opening of the West Indian Ocean. The same sequence of passive then active rifting is observed during the Lower Cretaceous with, in between, the Parana-Etendeka LIP at 135 Ma. We show that the relationships between the style of rifts and their breakdown or with the type of resulting margins (as magma poor or magma dominated) are not straightforward. Finally, we discuss the respective role of mantle global warming promoted by continental agglomeration and mantle plumes in the weakening of the continental lithosphere and their roles as rifting triggers.

1. Introduction

Continental rifting results from extensional deformation of the continental lithosphere leading to its rupture and potentially subsequent formation of a new oceanic domain [Şengör and Burke, 1978; Khain, 1992; Ziegler, 1992a]. Rifting plays consequently a major role in the evolution of the continental lithosphere and, in particular, in the so-called Wilson cycle describing the cyclical opening and closing of oceans during the Earth's history. It is classically considered that a continental rifting process ends when oceanic accretion begins. However, the rifting does not always continue up to its end, and many rifts become inactive before their oceanization remaining “rifted basins” and not continental margins. In this case, the terminology of “failed rift” or “aborted rift” is generally used. Conversely, a “successful” rift is indeed made up by the conjugate continental passive margins of an oceanic domain. Both passive margins and continental rifts store major hydrocarbon provinces [i.e., Roberts and Bally, 2012].

Two different processes have been proposed to explain rifting: either horizontal extensional forces acting at the lateral boundaries of the system [Mc Kenzie, 1978] or a heat source acting at the bottom of the lithosphere [Burke and Dewey, 1973]. In the first case, the required far-field stresses are essentially related to the “slab-pull” mechanism [Lithgow-Bertelloni and Richards, 1998]; in the second, a connection with a mantle plume is investigated. These two mechanisms, which can lead to the lithosphere breaking, correspond to the ancient concepts of “passive” and “active” rifting, respectively [Şengör and Burke, 1978] and can develop in different geodynamic settings [Şengör and Natal'in, 2001]. We are aware that numerous rifts display a tectonic evolution combining passive and active mechanisms in one order or the reverse [Ziegler and Cloetingh, 2004; Huismans et al., 2001; Merle, 2011]. However, for a first-order analysis, we assume that this classification remains useful. Referring to the geodynamic context and following Merle [2011], we regard as active the so-called plume-related rifts and as passive the mountain-related rifts. By contrast, subduction-related rifts are difficult to classify using this terminology because of numerous interfering processes. However, the initial mechanism is the slab pull, and their mechanism is consequently rather “passive.” Other parameters such as the rate of extension, the geothermal gradient, the rheological profile of the

lithosphere, or the structural heritage can explain specific features of rift systems [Buck, 1991; Hopper and Buck, 1996; Brun, 1999; Gueydan and Précigout, 2014].

Besides, it is known that during Earth's history, rift systems did not develop continuously but throughout successive rifting episodes of specific ages (during the Upper Carboniferous-Permian, Triassic-Lower Jurassic, Upper Jurassic-Lower Cretaceous, etc.) [Ziegler, 1992b; Şengör and Natal'in, 2001; Roberts and Bally, 2012]. Such a discontinuous evolution suggests an alternation of periods during which the Earth is rather "in extension" with periods rather "in compression," as at the moment [Zoback, 1992] or quiescent. This is particularly well illustrated in Africa and along the margins of this continent [Burke et al., 2003; Burke and Gunnell, 2008]. The aim of the present paper is to test the idea that rifts of a given age share some common features at global scale. In this case, we guess that the knowledge of their common specificities can help understanding the deep mechanisms triggering the rifting and can be useful for exploration of frontier rifts and passive margins.

In order to address this question, we propose to examine the successive phases of Pangea breakup by the end of Paleozoic up to the end of Lower Cretaceous times in order to compare their specific characteristics. For this purpose we will (1) assess the geological parameters allowing categorization of the various rifting modes; (2) present the main features of the Pangea configuration and how the Permo-Carboniferous and Permo-Triassic rifting episodes are integrated within this specific pattern; and (3) try to characterize the Triassic-Lower Jurassic and Upper Jurassic-Lower Cretaceous rifting, which led to the almost complete disintegration of Pangea and marked the onset of convergence between the Gondwanian and Laurasian domains. Finally, the major trends of rifting processes through times will be discussed, and the consequences will be addressed.

2. Main Parameters Allowing a Classification of Rift Systems, Definition of the Style of Rifting

The question addressed here is neither to propose a new classification of rift systems nor discussing in detail the existing classification [i.e., Şengör and Natal'in, 2001; Merle, 2011] and the thermomechanical models [i.e., Huismans and Beaumont, 2003]. Our approach is mainly pragmatic. The purpose is to define the first-order rift attributes that can be useful to better understand the mechanisms. The main questions are as follows:

1. Is there a prerift (or rift-onset) unconformity, and is this unconformity postdating an important sedimentary hiatus?

If the answer to these two questions is positive then one can suppose that doming and erosion have occurred before rifting (Figure 1). Such features are considered as typical of "active rifting" [Şengör and Burke, 1978; Campbell, 2007]. As emphasized by Franke [2013], the presence of a prerift unconformity is not the intrinsic signature of a mantle plume. It is the reason why we believe that the main point is to put forward evidence for prerift uplift. It is consequently useful to verify if the prerift unconformity is accompanied by an important truncation at a regional scale indicative of a prerift uplift. The signature of this uplift can be a sedimentary hiatus or a rapid change of the paleoenvironment from deep marine to subaerial [Campbell, 2007]. An evident bias happens in the case of mountain-related rifting because the basal unconformity could be a postorogenic erosion surface and not a rift-onset unconformity (Figure 1).

2. Is the rifting associated with massive volcanic flows?

A subsidiary but crucial question is to know if volcanism occurred before, during, or after rifting. In an active rifting setting, volcanism is expected to occur before rifting and just after the doming [Campbell, 2007] (Figure 1). In a "passive rifting" setting, volcanism, if any, is basically a consequence of lithospheric extension and is expected to occur during and/or after rifting (Figure 1). The occurrence of volcanic flows during the latest stage of rifting leads to the development of "magma-dominated margins" [Reston, 2009], which were previously called "volcanic-rifted margins" [Roberts et al., 1979; White and McKenzie, 1989] or "volcanic passive margins" [Geoffroy, 2003]. It is worth noting that such margins are not the foregone consequence of active rifting but seem independent from the rifting modes.

3. Where do rifts develop?

The question is to know if the rifts are localized by preexisting structures or if they ignore the structural heritage. To validate this criterion, it is assumed that the structural inheritance must be relatively young (tens of millions of years) and not too old (several hundreds of millions of years). Otherwise, it should be

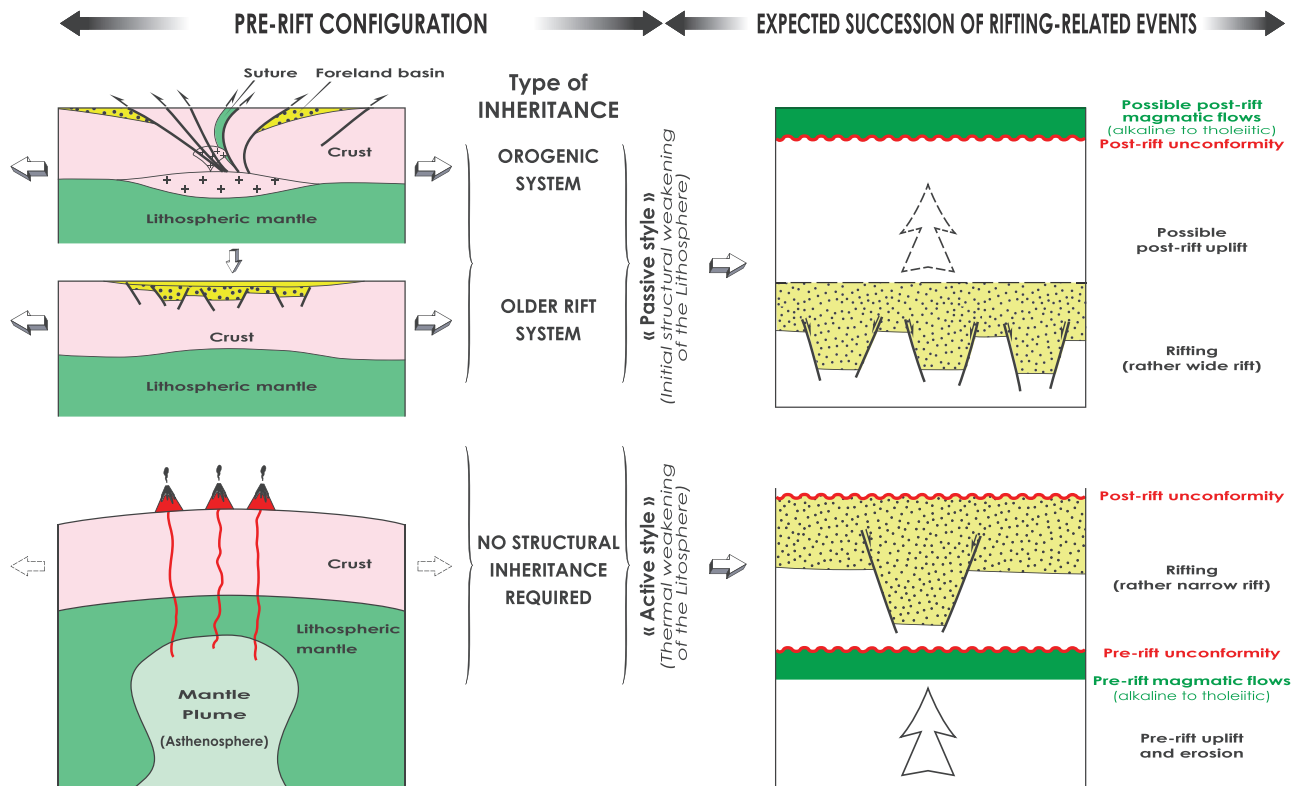


Figure 1. Sketch illustrating the main parameters allowing the distinction between active and passive rifting styles. The left part of the diagram illustrates the prerift configuration; the right part indicates the succession of events occurring during the rifting process for passive and active settings, respectively. It is worth noting that initially this terminology referred to the role (active or passive) of the asthenospheric mantle [Şengör and Burke, 1978]. Even if we conserve the same terms, the meaning is not the same. In the case of an active style, the heat source (the plume) is the precursor but not, strictly speaking, the trigger of the rifting. As it is demonstrated that extensional tectonic forces (even weak) are required to explain rift development above a plume [Burov and Gerya, 2014], the key point is the nature of the factor controlling the weakening of the lithosphere and allowing the rift localization. If the weakness has a mechanical origin (structural inheritance), the style is considered as passive; if the weakness has a thermal origin, the style is considered as active.

always possible to find a structural heritage. In the case of clear and young localization, the structural objects on which the rifts system develops could be, for instance, a collapsing mountain belt or a previous rift system (Figure 1). In both cases, the expected mechanism is rather passive rifting. In the literature, it is often indicated that the duration of rifting stages is highly variable for both abortive and successful rifts. For instance, Ziegler and Cloetingh [2004] note durations from 10 Myr (Provençal-Ligurian Basin) to almost 300 Myr (Norwegian-Greenland Sea). However, apparent long-lived rift systems rather result from a succession of rifting episodes located in the same sites and separated by postrift sag periods. In such case, the older rifts must be considered as an inheritance impacting the development of the younger ones. Finally, we can also consider that, for an active rifting, if the plume head does not really represent an inheritance, it is a kind of precursor of the rift contributing to its localization (Figure 1).

4. Is the rifting expressed by a narrow rift or a wide rift zone?

Buck [1991] considers that a typical narrow rift has a width on the order of 100 km, whereas a wide rift zone can reach more than 800 km. From the theoretical approach of this author, the factors controlling the width of the rift are the thickness of the crust, the heat flow, and the strain rate. Recently, Gueydan and Précigout [2014] have shown that the modalities of rheological coupling between crust and mantle play also an important role. Typically, the development of a wide rift zone requires a thick and hot crust deformed at low strain rate. Numerical modeling by Van Wijk and Cloetingh [2002] confirms that basin migration (i.e., development of a new basin adjacent to a previous one leading to the progressive widening of the deformed domain) is a direct consequence of slow lithospheric extension. Analog modeling by Brun [1999] corroborates that spreading of the continental crust giving birth to a wide rifts occurs in a thickened lithosphere with maximum strength at the bottom of the brittle upper crust. By contrast,

necking giving birth to narrow rifts occurs in a stable lithosphere with maximum strength below the Moho in the lithospheric mantle. So it is expected that wide rift zones develop preferentially over young mountain belt and can even be the signature of continental collision-related rifts [Şengör and Natal'in, 2001] also called mountain-related rifts [Merle, 2011].

In order to compare the principal rift systems studied in the paper, we have reported in Table 1 the main systems and their major geological features. From our analysis, a passive style is mainly characterized by (1) the existence of a strong heritage and (2) volcanism (if any) occurring after rifting. Secondary criteria are the frequent development of wide and long-lived rift systems and the absence of prerift uplift (no hiatus between the prerift and the synrift sediments). It is worth noting that in the case of wide rift systems, extensive activity can “jump” from one site to another. So the activity should be considered for the system as a whole and not for an individual rift within the system. An active style is characterized by (1) the existence of a prerift uplift, (2) volcanism before rifting, and (3) the absence of necessary structural inheritance (except the plume head itself). A secondary criterion is the development of narrow and short-lived rather than wide and long-lived rift systems. Finally, it is worth emphasizing that, for a given system, the rifting style can change through time [Merle, 2011].

3. Pangea Configuration and Style of Permo-Carboniferous and Permian-Early Triassic Rifting

Pangea was formed by progressive agglomeration of crustal blocks starting at the end of Neo-Proterozoic (see a review in Stämpfli *et al.* [2013]). In this configuration, which has existed several times throughout Earth's history, all continents form a unique assemblage occupying a single hemisphere. It is also recognized that the external outline of the continental hemisphere follows a geodesic great circle remaining close to the magnetic poles [Le Pichon and Huchon, 1984]. As the surface of continents is smaller than the surface of oceans (about 40%), an oceanic domain is continuously preserved within the continental hemisphere. This domain is the Tethys (see a review in Ricou [1994]), which, through time, always roughly presented an isocles triangle shape pointing westward.

It is generally considered that Pangea existed for approximately 120 Myr from the mid-Carboniferous, about 320 Ma ago, until the Lower Liassic, about 195 Ma ago, i.e., between the final collision of Laurussia and Gondwana [Matte, 2001] and the onset of central Atlantic accretion [Sahabi *et al.*, 2004; Labails, 2007; Labails *et al.*, 2010]. However, the premises of Pangea disintegration actually began earlier. By the latest Carboniferous and during the Early Permian, rift systems developed in numerous places (Figure 2) but especially (1) in Laurussia over the Variscan and Caledonian Belts [Burg *et al.*, 1994; Oftedal *et al.*, 2005; Faleide *et al.*, 2008; Štolfova and Shannon, 2009]; (2) within Gondwana along the present eastern border of Arabia and Africa, where they constitute the Karoo I system [Delvaux, 2001; Tankard *et al.*, 2009]; (3) along the northern border of Gondwana, i.e., along the Paleo-Tethys southern margin from Arabia to Australia [Stämpfli *et al.*, 1991; Langhi and Borel, 2005; Chauvet *et al.*, 2009; Korsch *et al.*, 2009] (Figure 2); and finally (4) within East Gondwana in the present Indian [e.g., Kumar, 1985], Antarctic [Harrowfield *et al.*, 2005], and Australian continents, where numerous rift basins also register important strike-slip motions [e.g., Song and Cawood, 2000]. These four rift systems are challenging because they are either buried at depth and hidden by more recent rifts (systems 1, 2, and 3) or inverted and involved in the Alpine orogeny (a minor part of system 2 and system 3). It is generally acknowledged that the first system expresses the collapse of the Caledonian-Variscan Belt [Burg *et al.*, 1994] even though Olausson *et al.* [1994] argue that the North Sea and Oslo rifts also originate from a far effect of the Variscan orogeny, while the second rift system results from a far effect of the Cape Fold Belt [Delvaux, 2001]. These systems did not trigger oceanic accretion and constitute consequently failed systems. On the contrary, the third system was successful giving birth to the Neo-Tethys Ocean [Ricou, 1994] and the coeval detachment of the Cimmerian blocks from Gondwana [Şengör, 1990] during the Late Permian. A few other Permian rift systems are also known elsewhere in Pangea, like for instance in the Cordilleran orogen of South America [Limarino and Spaletti, 2006; Sempere, 2002; Javadi *et al.*, 2011] and along the northern margin of the Paleo-Tethys Ocean [Stämpfli and Kozur, 2006] (Figure 2). Both are on the sites of potential back-arc basins and are interpreted in the same way.

The opening of the Neo-Tethys, during the Permian, did not really alter the Pangea configuration because all continental masses remained confined in the same hemisphere. In fact, a single crustal ribbon (the Cimmerian

Table 1. The First-Order Parameters Allowing the Definition of the “Rifting Style” of the Main Rift Systems Studied in the Paper^a

Name and Age of the Rift Systems	Existence of an Evident Rift-onset	Existence of Magmatic Intrusions/Extrusions and Timing (Pre-, Syn-, and Post)	Presence of Preexisting Structures and Nature of These Structures	Width (Wide Versus Narrow) and Duration (Long Lived Versus Short Lived) of the Rift System	Future of the Rift (Failed or Successful Oceanization)	Dominant Rifting Style: Active Versus Passive
Western Europe rift system (Upper Carboniferous-Lower Permian)	An unconformity exists when the rifts are over the Variscan Belt but it is a postorogenic erosion surface	Synrifting and locally postrifting magmatism	(1) the Variscan Belt and (2) the Devonian rifts over the Caledonian Belt	Wide and rather long lived but difficult to evaluate due to possible jumps from one rift to an adjacent one	Failed	Passive
Karoo I rift system (Upper Carboniferous-Permian)	No general unconformity, no evidence of prerift uplift	Globally no magmatism	The rift system follows several branches of the Panafrikan Belt but this is an old heritage	Rather wide and rather long lived (≥ 30 Myr)	Failed	Passive
West-Siberian rift system (Lower Triassic)	Evidence of prerift uplift and erosional unconformity	Prerift magmatism (Siberian LIP)	No evident heritage	Wide and short lived (± 20 Myr)	Failed	Active
Central Atlantic rift system (Triassic)	An unconformity generally exists but it is a post-Variscan erosion surface.	Post-rift magmatism (CAMP LIP)	Variscan-Appalachian orogenic belt	Wide and long lived (≥ 50 Myr)	Successful	Passive
Karoo II rift system (Lower Jurassic)	Evidence of prerift uplift and erosional unconformity	Prerift magmatism (Karoo-Ferrar LIP)	The rift system is partially superimposed to the Karoo I system	Narrow and short lived (≤ 20 Myr)	Successful	Active
Austral segment of the South Atlantic rift system (Upper Jurassic-Lower Cretaceous)	No general unconformity, no evidence of prerift uplift	Postrift magmatism (Parana-Etendeka LIP)	The rift system is superimposed over older rifts, which are themselves over a branch of the Panafrikan Belt	Wide and long lived (≥ 50 Myr)	Successful	Passive
Equatorial and central segments of the South Atlantic rift system (Lower Cretaceous)	Evidence of prerift uplift and erosional unconformity.	Prerift magmatism (Parana-Etendeka LIP)	The central Segment follows a branch of the Panafrikan Belt but this is an old heritage. No heritage for the equatorial segment	Narrow and short lived (≤ 20 Myr)	Successful	Active

^aMajor references: (1) for the western Europe rift system: *Burg et al.* [1994], *Offedal et al.* [2005], *Faleide et al.* [2008], and *Štollfova and Shannon* [2009]; (2) for the Karoo I rift system: *Delvaux* [2001] and *Tankard et al.* [2009]; (3) for the West Siberian rift system: *Czamanske et al.* [1998], *Tanton and Hager* [2000], and *Reichow et al.* [2009]; (4) for the central Atlantic rift system: *Tankard and Welsink* [1989], *Withjack et al.* [1998], *Withjack and Schlische* [2005], *Leleu and Hartley* [2010], and *Redfern et al.* [2010]; (5) for the Karoo II rift system: *Reeves* [2002, 2014] and *Geiger et al.* [2004]; (6) for the South Atlantic rift systems (central and austral segments): *Jackson et al.* [2000], *Szatmari* [2000], *Moulin et al.* [2010], *Heine et al.* [2013], and *Franke* [2013].



Figure 2. Sketch map showing the Earth at 280 Ma (Lower Permian) with the major rift systems developed during the Carboniferous and Permian and the major Paleozoic Orogenic Belts (Caledonian, Variscan, Uralian, and Gondwanide). Note that numerous rifts are superimposed to the orogenic belts, in particular in western Europe and along the future North Atlantic oceanic domain. Along the northern border of Gondwana the rift system, which led to the opening of Neo-Tethys and the drifting of the Cimmerian Domain, is mostly inferred from paleomagnetic data [Besse *et al.*, 1998]; its geometry is indicative. The Karoo I system, at right angle, follows a branch of the Panafrican Belt and foreshadows the West Indian Ocean. The background of the map is from “Plate tectonic maps and continental drift animations by C. R. Scotese, PALEOMAP Project (www.scotese.com).”

Domain) has been pulled out from the southern margin of the Paleo-Tethys Ocean during the Late Permian and was already accreted to its northern margin by Late Triassic, thus fully restoring a Pangea configuration as a consequence of the Cimmerian (or Indosinian) orogeny [Zanchi *et al.*, 2009]. Ricou [1994] showed that the Tethys configuration, with a southern passive margin and a northern active margin, was fundamentally unstable. As the surface of the oceanic domain remains unchanged, the mid-oceanic ridge is expected to be quite rapidly involved in the subduction process. So maintaining a Tethys configuration requires a southward periodic shift of the spreading ridge. Such a shift is likely responsible for the Neo-Tethys rifting, which occurred during the latest Carboniferous/Early Permian [Chauvet *et al.*, 2009].

Gutiérrez-Alonso *et al.* [2008] propose a unified model in which the Neo-Tethys rifting and, more generally, the whole Permian rifting distributed over Pangea should be the result of slab-pull taking place following the subduction of the Paleo-Tethys mid-oceanic ridge and preceding the drifting giving birth to Neo-Tethys Ocean (their concept of “self-subduction”) (Figure 3). This slab-pull model, which is typically a “passive-rift” model (Table 1), gives an elegant solution to explain (1) the place where the Neo-Tethys rift develops cutting through older structures, (2) the major strike-slip component of the Late Paleozoic East-Gondwanian rifts due to far-field intraplate stresses also induced by the slab-pull phenomenon, and (3) the ribbon shape of the Cimmerian domain. Associated to this rifting, the Panjal Traps (Figure 3) occurred at 289 ± 3 Ma (Early Permian) (see a review in Shellnutt *et al.* [2012]), i.e., during the late phases or after the Neo-Tethys rifting [Chauvet *et al.*, 2009] in agreement with what is expected for a passive-rifting model.

The Gutiérrez-Alonso *et al.* [2008] model (Figure 3) could also possibly explain the set of Carboniferous to Permian rifts, which develop over the Caledonian Belt and subsequent Devonian rifts along the future northern Atlantic [Séranne and Séguret, 1987; Oftedal *et al.*, 2005; Faleide *et al.*, 2008; Štollfová and Shannon, 2009] and converge southward through the North Sea [Heeremans *et al.*, 2004; Hansen *et al.*, 2007] to join the Variscan Belt of western Europe at the apex of the Paleo-Tethys [Burg *et al.*, 1994]. In contrast, the model fails to explain

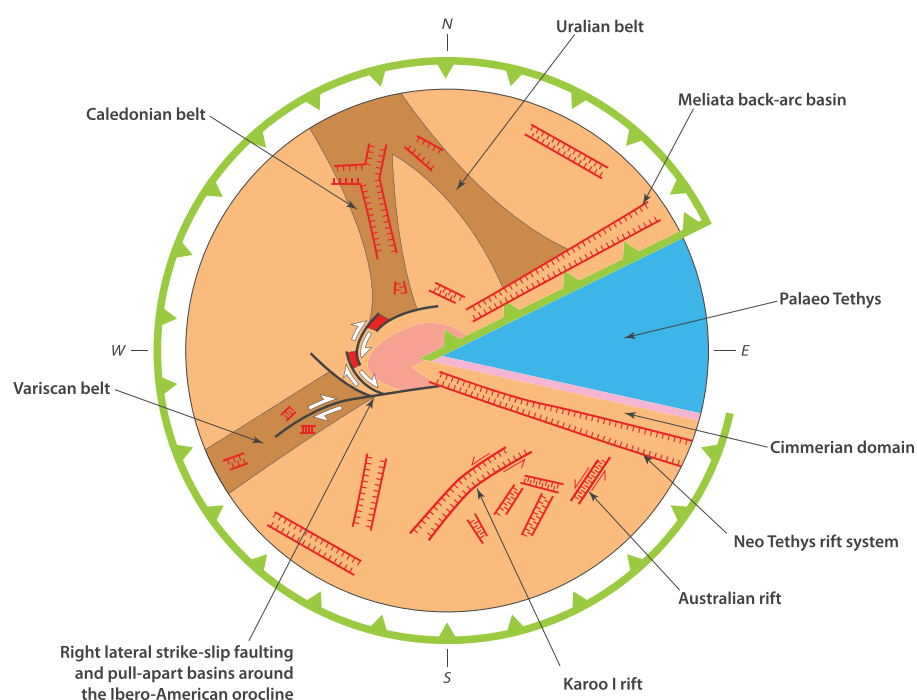


Figure 3. The Gutiérrez-Alonso *et al.*'s [2008] model modified to take into account the actual geometry of the Karoo I system. This conceptual model suggests that at 270 Ma Pangea suffered a self-subduction mechanism explaining the development of extensional deformation within the whole and unique continental plate. Comparison with the Figure 2 is useful.

the onset of the Karoo rift system, which develops obliquely to the Cape Fold Belt, element of the Gondwanide Belt. For these rifts, which follow inherited Panafrican structures, we can invoke a passive rifting response to the buildup of orogeny-related intraplate stresses in the same manner as the European Cenozoic rift system have been developed in front of the Pyrenees and Alps [Dèzes *et al.*, 2004]. It is worth noting that the Karoo rift system has the same age and is connected to the Neo-Tethys system through rifts located in Madagascar [Geiger *et al.*, 2004; Schandelmeier *et al.*, 2004] and Somalia [Ghebreab, 1998]. We can add that the Permian Panjal traps occurred at the exact place where the two systems converged (Figure 2).

Although they belong to different settings and mechanisms, the Late Carboniferous-Early Permian rifts pertain in their entirety to the category of passive rifts. Accordingly, when the rifting is associated with magmatism, as in western Europe and northwest India (Panjal), it does not mark the rifting onset but rather accompanies and follows rift development. Regarding inheritance, the location of the North Atlantic and West Europe rifts is clearly controlled by the Caledonian-Variscan Belts (Figure 2); accordingly, the rift zone is wide rather than narrow. On the contrary, a major role of heritage is less straightforward for the other systems whose location seems rather to be chiefly explained by the geodynamic context at the considered time.

After the Permo-Carboniferous rifting and the opening of Neo-Tethys, a new episode of failed rifting (Figure 4), associated to the Siberian Traps, occurred in western Siberia during the latest Permian and the Lower Triassic, with the magmatic episode lasting from approximately 262 Ma to approximately 242 Ma, with an undergoing paroxysmal activity at 252 Ma [Renne and Basu, 1991; Courtillot *et al.*, 1999; Reichow *et al.*, 2009; Nikishin *et al.*, 2002; Dobretsov, 2005; Kuzmichev and Goldyrev, 2007; Ivanov *et al.*, 2013]. The rift basins are buried below a very thick Meso-Cenozoic cover, and their tectonostratigraphic evolution is not very well constrained even though it appears obvious that the branches of this West Siberian rift system follow the tectonic trends of the older Altaian and Uralian orogens. However, the authors converge to consider that uplift and basaltic flows occurred before the rifting in agreement with an active rifting process (Table 1). More precisely, an erosional unconformity, indicative of uplift, is evidenced at the bottom of the basaltic flows, but this uplift is brief and immediately followed by synmagmatic subsidence [Czamanske *et al.*, 1998; Tanton and Hager, 2000]. Accordingly, the rift system itself is demonstrated to be short lived, that means with a duration lesser than 20 Myr [Nikishin *et al.*, 2002; Saunders *et al.*, 2005]. At

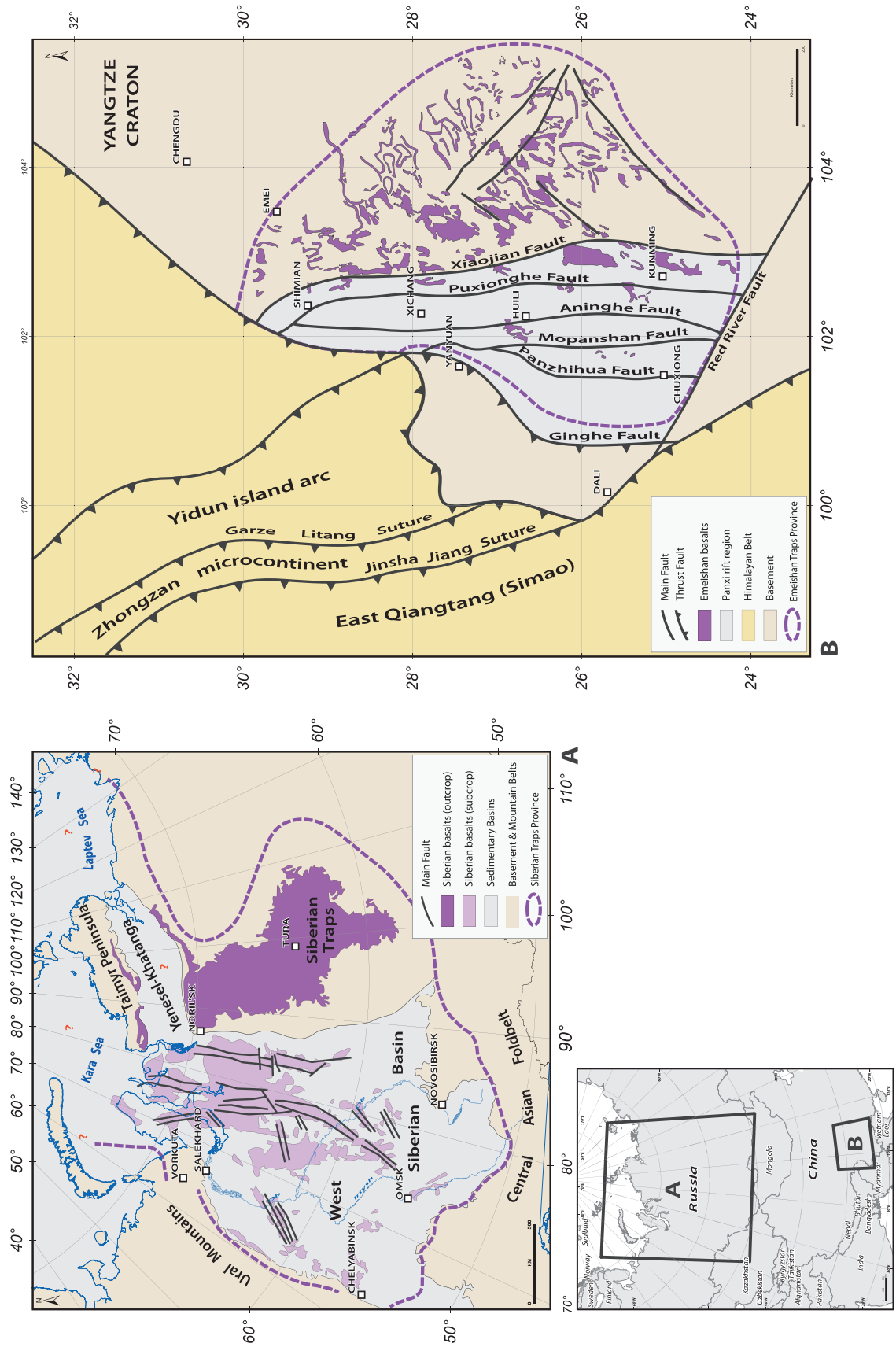


Figure 4. Sketch maps showing the (a) West Siberian (modified from Reichow *et al.* [2009]) and (b) Panxi (modified from Munteanu *et al.* [2013]) rift systems and the associated Siberian and Emeishan large igneous provinces (LIPs). Both are failed rift systems.



Figure 5. Sketch map showing the Earth at 190 Ma (Lower Jurassic), i.e., at the time of the drifting onset in central Atlantic. The central Atlantic rift system (Triassic) is superimposed to the southern branch of the Variscan Belt. To the north, a diverging branch will subsequently form the Alpine Tethys (see *Frizon de Lamotte et al.* [2011]). The CAMP basaltic flows postdate the rifting episode. At 190 Ma, the Cimmerian blocks (see Figure 2) are accreted to Laurasia forming the Cimmerian (or Indosinian) orogenic system. Between West and East Gondwana, the Karoo II system (Lower Jurassic) is partly superimposed to the Karoo I system (see Figure 2). However, the two systems are diverging southward. The Karoo-Ferrar large igneous province was emplaced between Karoo I and Karoo II rift systems. The background of the map is from Plate tectonic maps and continental drift animations by C. R. Scotese, PALEOMAP Project (www.scotese.com).

approximately the same time (258 Ma), the Emeishan basalts, located in southwestern China, could be also associated to rifting [*Courtillot et al.*, 1999; *Song et al.*, 2004; *Munteanu et al.*, 2013]; however, the so-called Panxi rift [*Munteanu et al.*, 2013, Figure 4] is difficult to characterize because of severe subsequent deformation due to the Himalayan orogeny.

Considering the Upper Carboniferous to Lower Triassic period as a whole, we note that a long episode of passive rifting responsible for numerous rifts scattered all over Pangea and likely triggered by slab pull. It was preceded by a short episode of active rifting clustered in western Siberia and possibly in the Chinese Emeishan province. Only one oceanic domain, the Neo-Tethys, developed during this period. It is clearly correlated with a passive rifting episode, whereas the two systems related to active episodes remained failed systems.

4. Pangea Breakup and Style of Triassic to Lower Jurassic Rifting

Contrasting with the Siberian and Emeishan events, the rifting, which began at the same time (i.e., uppermost Permian-Lower Triassic) along the present central Atlantic, is a long-lasting rifting episode. Processes and controls on the development of this rift system are still debated [*Tankard and Welsink*, 1989; *Withjack et al.*, 1998; *Withjack and Schlische*, 2005; *Leleu and Hartley*, 2010; *Redfern et al.*, 2010]. Nevertheless, a control by the Variscan orogenic system is clearly established: (1) the rift system is located over the Appalachians and the Mauritanides, which both are elements of this system and (2) from a chronological point of view, rifting follows immediately the Permian-Early Triassic collapse of the Variscan Belt (Table 1 and Figure 5).

The rifting style is complex with the development of a broad rift zone including several subbasins with highly variable geometries. Each subbasin displays similar basin fill evolution with a continental sand-prone lower unit followed by an upper mud-prone unit with lacustrine and marginal marine environments [*Leleu and Hartley*, 2010]. Everywhere in the area, the end of Triassic is marked by a major magmatic event known as the central

Atlantic magmatic province (CAMP) (Figure 5) sealing the rift system and preceding the drifting in the northern central Atlantic segment [Marzoli *et al.*, 1999; Knight *et al.*, 2004; Verati *et al.*, 2007; Nomade *et al.*, 2007]. In the basins of the southern segment, erosional truncation indicates a Late Triassic but pre-CAMP uplift phase [Withjack *et al.*, 1998], likely of thermal origin, i.e., thermal doming preceding the basalts extrusion. Withjack *et al.* [1998] suggested that uplift (interpreted as a basin inversion) was due to seafloor spreading and ridge push processes. They consequently supposed that seafloor spreading became very early in the southern segment, but this has not been demonstrated by recent studies [Sahabi *et al.*, 2004; Labails *et al.*, 2010]. Further to the north, a Late Triassic unconformity within the Nova Scotian [Welsink *et al.*, 1989] and Moroccan basins [Hafid, 2000; Hafid *et al.*, 2006] might be related to the same cause, i.e., thermal doming, but with lesser amplitude [Leleu *et al.*, 2014].

The timing of the CAMP magmatic event following a long rifting period and more local doming as well as the width of the rift system and the role of heritage (location over a young orogenic system) are consistent with a passive mechanism initiated by lithospheric extensional stresses causing mantle uplift and subsequent magmatic production. Such transition from passive to active rifting has been modeled by Huismans *et al.* [2001]. Accordingly, geochemical constraints from CAMP basaltic dikes of the southeastern American Margin argue in favor of an upper mantle source for the magma [Callegaro *et al.*, 2013]. Seafloor spreading began during earliest Jurassic in the southern central Atlantic [Sahabi *et al.*, 2004; Labails *et al.*, 2010; Kneller *et al.*, 2012] and propagated northeastward toward the Tethys realm [Frizon de Lamotte *et al.*, 2011].

Along the eastern side of Africa (Figure 5), it is currently acknowledged that the breakup of Gondwana leading to the opening of the Somali and Mozambique Basins have occurred in the Middle Jurassic (171 Ma) [Malod *et al.*, 1991] but the tectono-stratigraphic context associated with this event has long been poorly understood. The conjugate margins, exposed along the coast of Tanzania, Kenya, and Somalia on the one side and Madagascar on the other side, permit to specify the framework [Reeves *et al.*, 2002]. The main question is the duration of the rifting event, which preceded the drifting. As indicated above, a rifting episode, called here Karoo I, began during the Upper Carboniferous in this area along a branch of the former Panafrican Belt. In Madagascar, authors have suggested that rifting continued without disruption up to the Early Jurassic [Montenat *et al.*, 1996; Piqué *et al.*, 1999]. However, such a scenario does not explain the important sedimentary hiatus occurring during the uppermost Triassic and Liassic nor the rift axis migration occurring at the same time [Geiger *et al.*, 2004]. It is the reason why a two-step scenario is now favored. The first step is the poly-phased (Carboniferous to Triassic) Karoo I failed rifting and associated pull-apart basins described above [Delvaux, 2001]. The second step, called here Karoo II, is a short-lived episode that occurred during the Toarcian and Aalenian just before the breakup [Geiger *et al.*, 2004]. Moreover, the two systems diverge south of Tanzania (Figure 6). The old Karoo I system, overall trending NE-SW, has been abandoned, and a new N-S line of fracture, close to the Davie Fracture Zone, developed along the southwestern border of Madagascar while the Karoo II system propagated southward toward the Mozambique Channel [Malod *et al.*, 1991; Reeves, 2014]. In the places where the two rift systems are superimposed, it is worth noting that the Karoo II system is narrower than the Karoo I (Figure 6).

Between the Karoo I and II systems, the discontinuity and sedimentary hiatus observed in Madagascar from Late Triassic to Toarcian [Geiger *et al.*, 2004], i.e., from 200 to 183 Ma, might be the signature of a doming. It is worth noting that the massive extrusion of the Karoo-Ferrar basalts (Figure 5) precisely occurred at the end of this period (182 Ma [Riley *et al.*, 2004]) further to the south in Mozambique, South Africa, and Antarctica. So doming and basaltic extrusions sign together the end of Karoo I failed rift system and the onset of Karoo II system, which subsequently led to continental breakup and the occurrence of fully oceanic crust.

In summary, rifting occurring during the Triassic and Lower Jurassic period displays very different configurations (Figure 5 and Table 1). The central Atlantic rift system is a typical example of passive system developed over a preexisting weak zone, the recently formed Variscan Belt, and developed before the massive magmatic extrusion of the CAMP. Karoo II rift system is an active system in which the massive magmatic extrusion (the Karoo-Ferrar LIP) predates the rifting. However, and contrarily to what is expected for an active system, a heritage (the previous Karoo I system) is evident for the part of Karoo II system running from Tanzania to Arabia. It is worth noting that this segment superimposed to a preexisting weak zone is located outside the Karoo-Ferrar large igneous province. It is as if this major heritage was used to channel the northward propagation of the rift system toward the Tethys realm (Figure 5). Interestingly, the two major rift systems

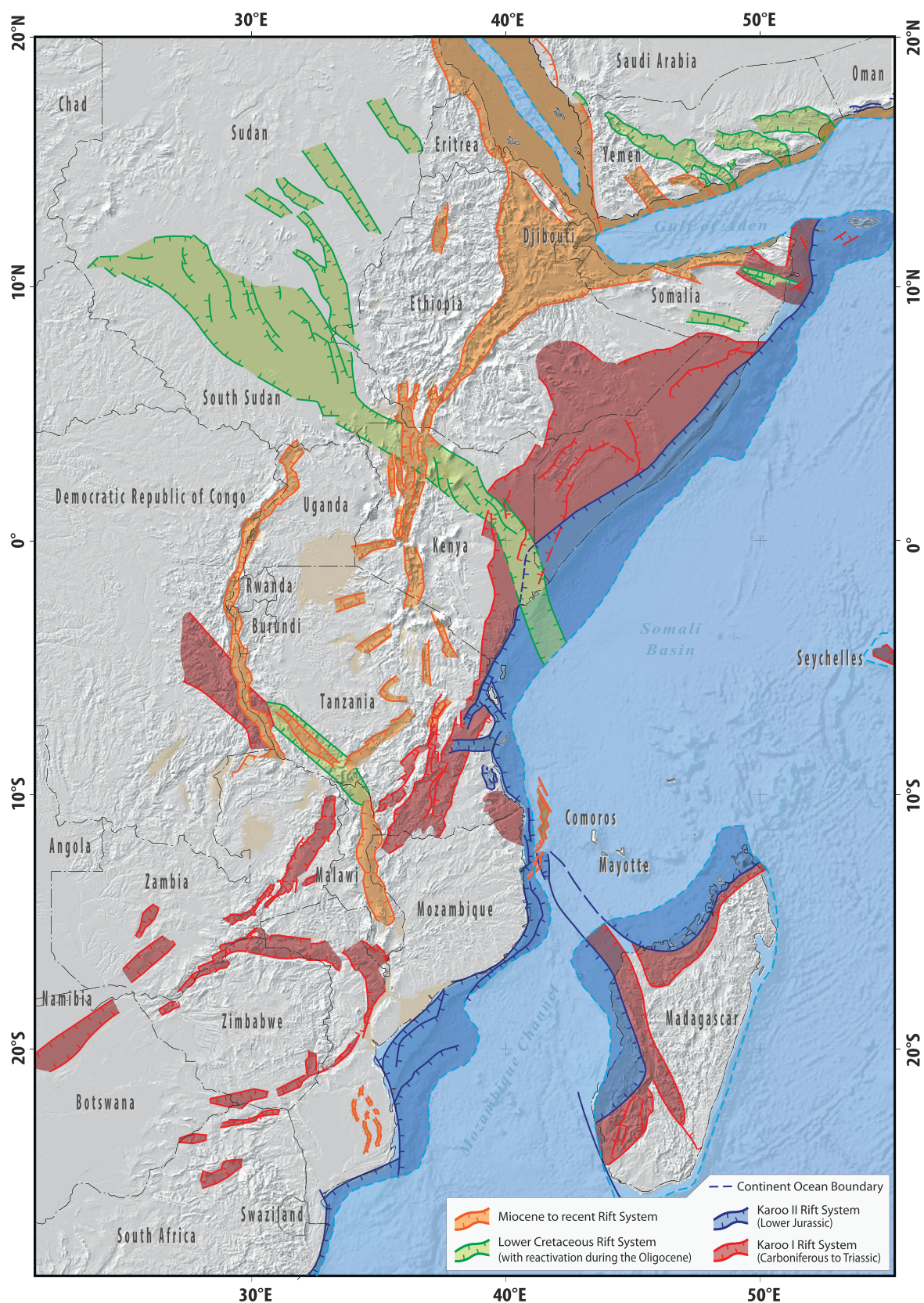


Figure 6. Schematic map illustrating the cross-cutting relationships between the four major rift systems developed in eastern Africa by the end of Carboniferous. The rift systems have been drawn using the tectonic map of Africa at 1:10 M [Milesi et al., 2010].

functioning during the considered time (Triassic to Lower Jurassic) were successful and gave birth to the central Atlantic and West Indian (i.e., Mozambique and Somali Basins) Oceans, respectively.

5. Pangea Breakup and Style of Upper Jurassic to Lower Cretaceous Rifting

An important episode of rifting responsible for the final scattering of Pangea initiated by the Late Jurassic (145 Ma) and climaxed during the Lower Cretaceous before the Late Aptian (110 Ma). This rifting episode led to the opening of the following oceans: South Atlantic [Jackson *et al.*, 2000; Szatmari, 2000; Moulin *et al.*, 2010; Heine *et al.*, 2013; Franke, 2013], central Indian [Gaina *et al.*, 2007], and the Bay of Biscay together with the southern part of the North Atlantic located between the Iberia-Newfoundland conjugate margins [Montadert *et al.*, 1979; Boillot, 1984; Tucholke *et al.*, 2007; Péron-Pinvidic *et al.*, 2007; Tugend *et al.*, 2014].

During the same period, the development of an important quantity of failed intracontinental rifts is observed especially in Africa and Arabia [Guiraud and Maurin, 1992; Moulin *et al.*, 2010; Frizon de Lamotte *et al.*, 2011] but also in the region of the subsequent northern Atlantic [Geoffroy, 2003], South America [Marquillas *et al.*, 2005], India [Lal *et al.*, 2009], Antarctica [Lisker *et al.*, 2005], Australia [Falvey and Mutter, 1981], or Mongolia [Johnson, 2004], indicating that continental areas were already separated by oceanic domains (Figure 7). Finally, this period is also the one during which two important back-arc systems, the Black and South Caspian Seas, developed in relation with the subduction of Neo-Tethys below Eurasia [Nikishin *et al.*, 2003; Brunet *et al.*, 2009] (Figure 7). The occurrence of three major LIPs, the Parana-Etendeka, the mid-Pacific Mountains, and the Kerguelen-Rajmahal climaxing at 133 Ma, 122 Ma, and 118 Ma, respectively [Courtillot and Renne, 2003], is recognized during the considered time span.

For the purpose of this paper, which is distinguishing the first-order processes active during rifting, it seems useful to distinguish the Upper Jurassic and the Lower Cretaceous rift systems even if geographically and chronologically the transition between them is progressive (Figure 7). It seems indeed that rifting starting during the Upper Jurassic is generally superimposed over older rift systems. This is in particular the case for the Pyrenean-Bay of Biscay and Galicia-Newfoundland systems developed over Permian-Triassic rifts [Bixel and Lucas, 1987; Oliva-Urcia *et al.*, 2012; Uphoff, 2005; Soares *et al.*, 2012; Kullberg *et al.*, 2013a, 2013b] or for the Lambert Graben in Antarctica [Lisker *et al.*, 2005] and the South Atlantic system developed over Carboniferous-Permian rifts [Stollhofen *et al.*, 2000]. This evident heritage associated with the absence of prerift uplift and scarcity of prerift magmatism suggests that a passive mode of rifting was dominant during this period (Table 1).

By contrast, the rift systems initiated during the Lower Cretaceous do not follow preexisting weak zones. This is in particular the case for the sub-Sahara system, which exhibits a star configuration, centered about the present Benoué Basin acting as a quintuple junction (Figure 8). The basement, consisting in Pan-African and older rocks, is cut out at random emphasizing the lack of evident heritage and suggesting that doming and related erosion of preexisting basins preceded rifting (Figure 8). Each branch of the system develops as a deep and long isolated rift. This configuration of “narrow rift” contrasts with the one observed in central Atlantic domain during the Triassic.

The analysis of relationships with the Parana-Etendeka LIP is also instructive. On the one hand, it is established that the austral segment of the South Atlantic rift propagated northward up to the Santos “buffer” [Moulin *et al.*, 2010] and that drifting started by the Hauterivian [Moulin *et al.*, 2010; Franke, 2013]. However, the precise timing between the LIP emplacement and the rifting remains unclear; it has been proposed either that drifting occurred exactly when the massive extrusion of the Parana-Etendeka basalts occurred (132 Ma) or maybe short time before or even immediately after [Stica *et al.*, 2014]. On the other hand, it has been recently demonstrated that in the equatorial and central segment of the South Atlantic, and more generally in central Africa, rifting began just after the Parana-Etendeka extrusion and propagated southward from the Gulf of Guinea to the Santos Basin and the Parana Etendeka LIP [Franke, 2013; Chabouraud *et al.*, 2013]. So the Parana-Etendeka LIP appears as synrift to postrift by reference to the austral segment of South Atlantic but ante-rift by reference to the central segment. This is in agreement with other evidences (presence or absence of inheritance), suggesting that austral segment rifting results rather from a passive mechanism whereas central segment rifting results rather from an active mechanism (Table 1).

The end of Lower Cretaceous is a pivotal time in the Earth’s history [Burke *et al.*, 2003]. Even if the scattering of Pangea is not completely achieved (separation of India from Madagascar and Australia from Antarctica

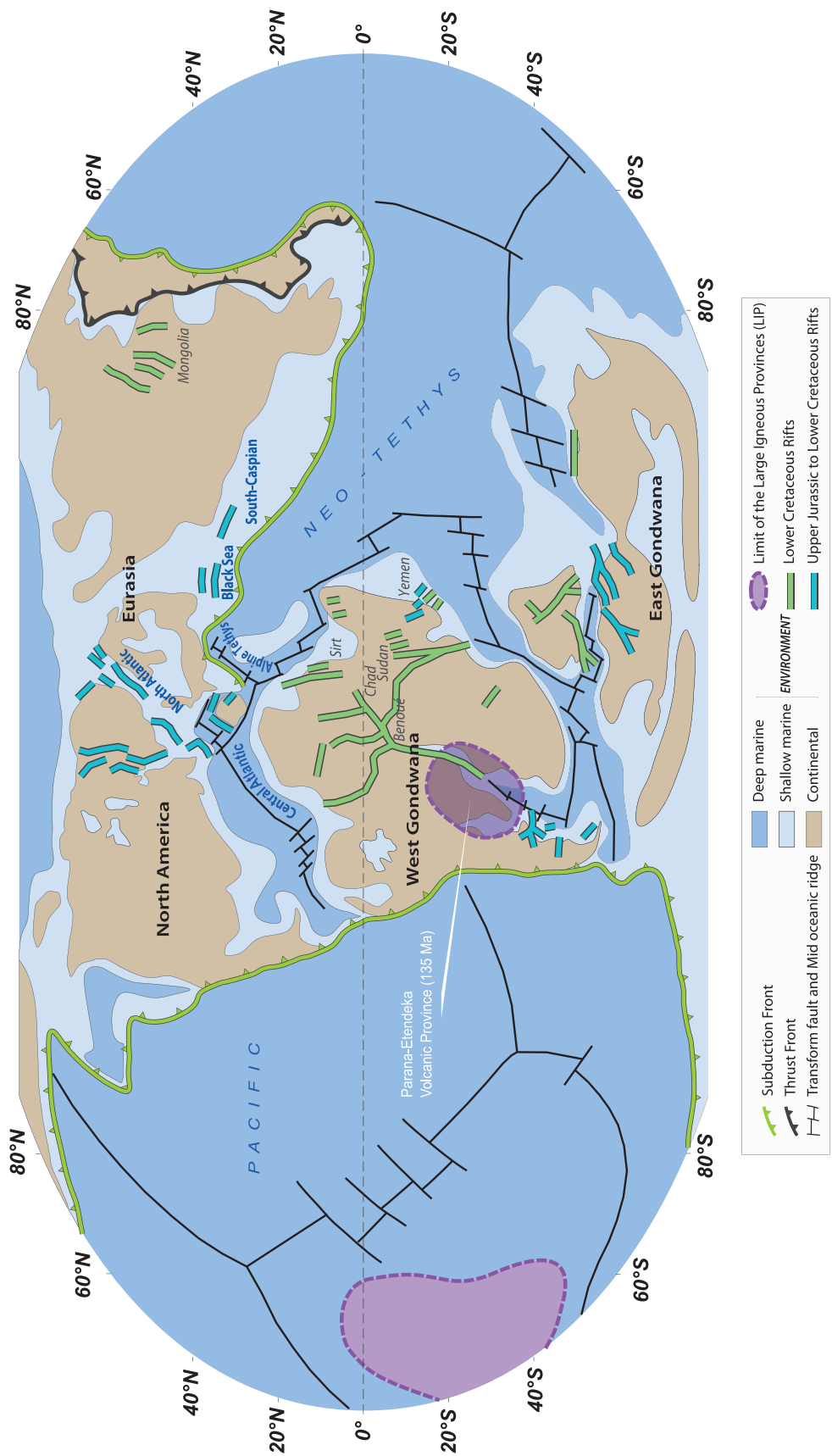


Figure 7. Sketch map showing the Earth at 120 Ma (Lower Cretaceous). The map shows the main Upper Jurassic and Lower Cretaceous rift system and the Parana-Etendeka associated large igneous province. The position of mid-Pacific LIP is also shown. The background of the map is from Plate tectonic maps and continental drift animations by C. R. Scotese, PALEOMAP Project (www.scotese.com).

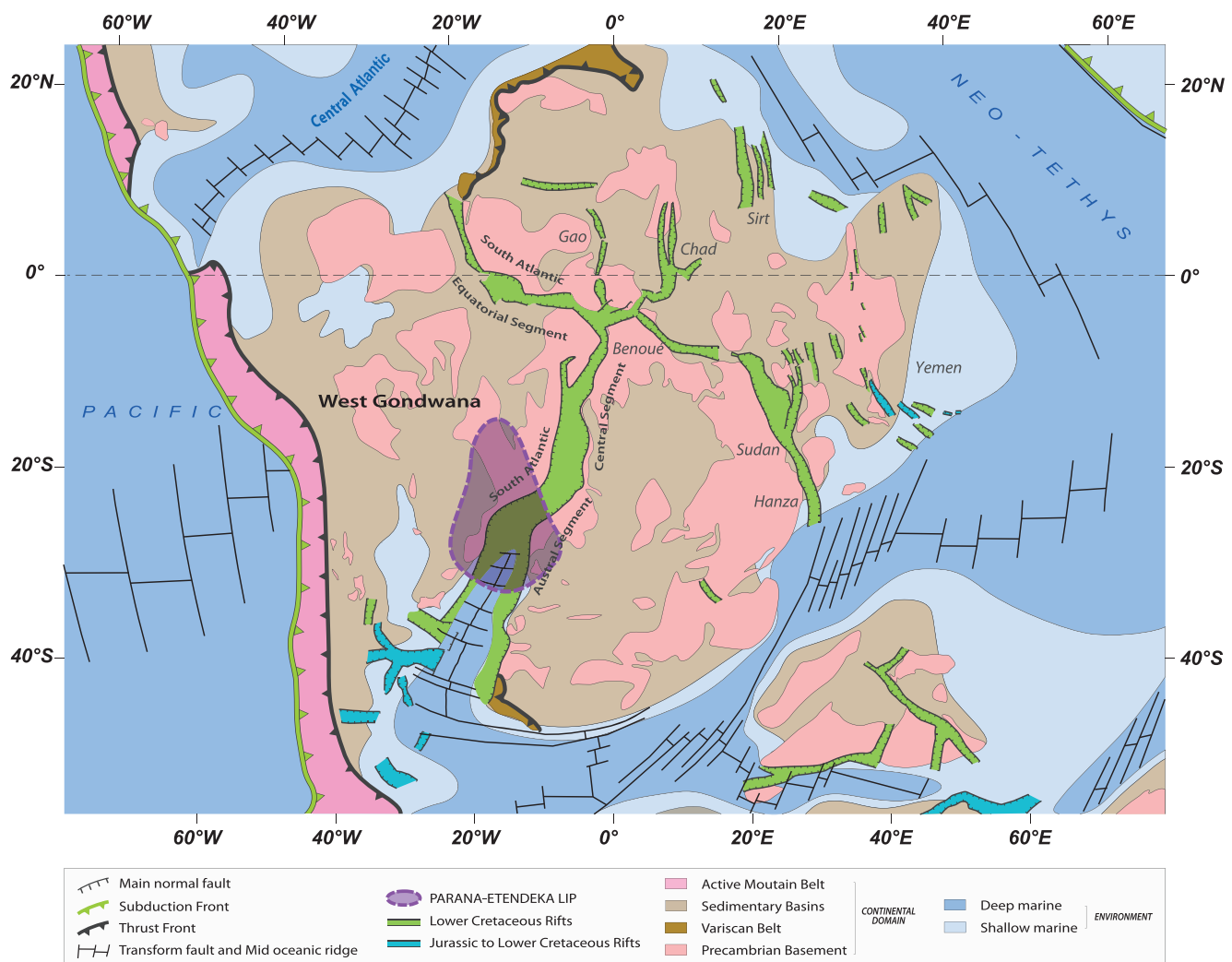


Figure 8. Map at 120 Ma (detail) showing how the African and South American rift systems cross out directly the Precambrian basement. Such geometry suggests that important doming-related erosion preceded the rifting and scrubbed the prerift cover. The background of the map is from Plate tectonic maps and Continental drift animations by C. R. Scotese, PALEOMAP Project (www.scotese.com).

occurred during the Upper Cretaceous), the opening of South and equatorial Atlantic signs the onset of the Neo-Tethys closure and convergence between the former Gondwanian blocks and Eurasia. As for the previous considered period, it seems that rifting followed first a “passive mode” during the Upper Jurassic up to the Hauterivian and passed transitionally to an active mode by the Hauterivian. The modalities of this transition are not completely understood and have to be investigated.

6. Discussion

The close association between rifting, occurrence of flood basalts, and LIPs has long been highlighted by different authors [Morgan, 1971; Burke and Dewey, 1973; White and McKenzie, 1989; Courtillot *et al.*, 1999; Burke and Cannon, 2014; Buiter and Torsvik, 2014]. These authors underline the major role played by the mantle, in a passive or active manner, during the history of continental breaking. Our geological approach of Pangea breakup completes previous work and allows discussing different aspects: (1) the pertinence of the concepts of active versus passive rifting as end-members of complex processes, (2) the reality of changes in rifting mechanisms through time from Carboniferous to Lower Cretaceous, (3) the relationships between the mode of rifting and its successful (or not) result as an oceanic domain and with the nature (magma rich versus magma poor) of the resulting margins, and finally (4) the possible causes of the changes in rifting mechanisms.

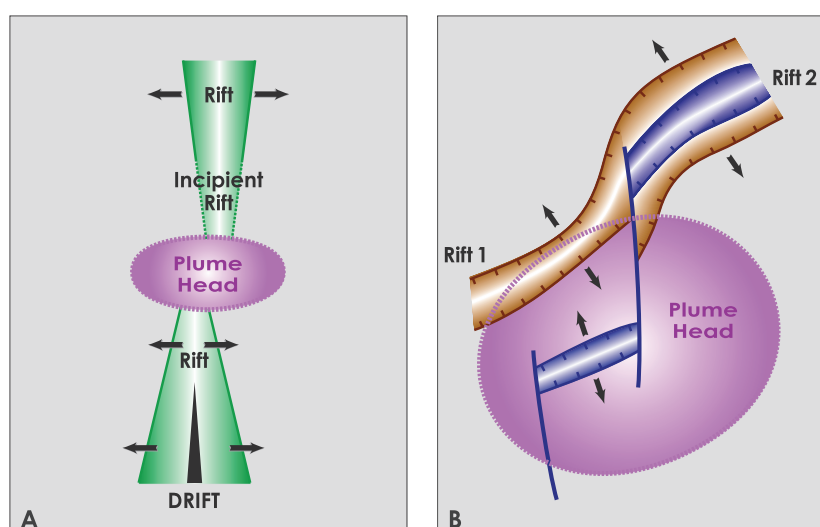


Figure 9. Schematic maps (not to scale) illustrating the relationships between plumes and rifts in the (a) South Atlantic Domain and in the (b) West Indian Domain. In both cases, the map illustrates the situation few time after the plume eruption at 135 Ma and 183 Ma, respectively. Figure 9a (inspired by the Figure 16 of Franke [2013]) displays that at 135 Ma, the plume eruption marks the onset of drifting south of the plume head and the southward propagation (toward the plume) south of the plume head. Figure 9b suggests that at 183 Ma, the Karoo II rift initially situated on the site of the Karoo I rift jumped southward toward the plume head.

6.1. Relevance of the Concept of Active Versus Passive Rifting

Our analysis permits to support the idea that the ancient classification of active and passive rifting remains useful to characterize the rifting style (Figure 1). Simple cases are illustrated by the central Atlantic and West Siberian provinces, typical examples of passive and active rifting, exhibiting postrift and prerift magmatism, respectively (Table 1). More complex configurations are observed in the South Atlantic and West Indian areas, where the style of rifting changes through time and space by reference to the Parana-Etendeka and Karoo-Ferrar LIPs, respectively (Figure 9). In both cases, a “chicken-and-egg” dilemma occurs: are the LIPs postdating the first stage or rather predating the second stage?

In the South Atlantic, we have considered that south of the Parana-Etendeka, the rifting is pre-LIP emplacement and rather passive, but north of the LIP, the rifting is post-LIP emplacement and rather active (Table 1 and Figure 9a). Everything happens as if on the one hand, the southern part of the rift system triggered the magmatism excess (perhaps as a consequence of a rift propagation barrier [Koopmann *et al.*, 2014] suggested by the dog-leg shape of the rift (Figure 8)), and on the other hand, as if the progressive opening of the northern part was pulled toward the LIP (Figure 9a). The same process happened during the Miocene when rifting in the Gulf of Aden (considered as the typical example of active rifting [Merle, 2011]) progressed toward the Afar LIP [Fournier *et al.*, 2010].

Between the East and West Gondwana, Karoo I rifting event is considered rather passive whereas Karoo II is considered rather active (Table 1), and the transition between the two events is marked by the emplacement of the Karoo-Ferrar LIP. However, the geographic configuration (Figure 9b) is very different from the one observed in South Atlantic because the plume head is not located at the rift extremity but away from the Karoo I system. The precise relative chronology of the different segments of the Karoo II system is not completely known, but we can propose that Karoo II first developed over Karoo I and subsequently advanced toward the LIP.

The fact, noticed in the South Atlantic and in the West Indian as well, that the plume head exerts an apparent attraction on the rifts' propagation and is able to deviate a rift trajectory can be understood if the hot spot acts as a weak zone. In this case, however, the active asthenospheric upwelling cannot be considered as the cause of the rift development but only as the cause of the weak spot, which itself acts as a precursor of the rift. This is in agreement with the Burov and Gerya's [2014] models showing that extensional tectonic forces (even weak) in the lithosphere are required to really trigger the rifting. However, we do not share

the idea that, during rifting, the late occurrence of plume material is more or less a rule as suggested by *Buiter and Torsvik* [2014]. A plume is able to localize a rift (our conception of what is an active rift; Figure 1) and not only to help the final continental breakup. In this respect, a comparison between the central Atlantic and the central segment of the South Atlantic is illustrative (Table 1). On the contrary, it is known that uplift above asthenosphere upwelling can develop without rift growth. It is, for instance, the situation of the “Morocco Hot Line” during the Neogene [e.g., *Frizon de Lamotte et al.*, 2009; *Michard et al.*, 2013] and of the Gondwanian Arches during the Late Devonian [*Frizon de Lamotte et al.*, 2013].

6.2. Changes of Rifting Style Through Time (Table 1)

The geological analysis of the different stages of Pangea fragmentation shows that the modes of rifting are different and discontinuous through time. The Carboniferous-Permian period is globally characterized by the development of passive rift systems leading to the formation of a huge quantity of rifts in various contexts (back arc, slab pull, and orogenic collapse). These rifts are localized by preexisting weaknesses of the lithosphere (Caledonian-Variscan Belt, previous rift systems (in particular Devonian rifts)) and/or by stress concentration in particular geodynamic settings (back arcs relative to the subduction of Panthalassa around Pangea or of Paleo-Tethys beneath the North-Tethys active margin; slab-pull exerted by Paleo-Tethys). Only one of these rift systems has been successful giving birth to the Neo-Tethys, but without alteration of the Pangea configuration, all continents remain in the same hemisphere.

This period has been followed by a short episode of active rifting and the development of the Panxi rift system (associated with Emeishan LIP) and the West Siberian rift system (associated with the Siberian LIP) at the very end of the Permian and during the Lower Triassic. Contrasting to this episode, the central Atlantic Triassic rifting, which marks the onset of Pangea breakup, is typically passive and directly controlled by structures inherited from the collapse of underlying Variscan Belt. Then a new stage of active rifting, associated to the Early Jurassic Karoo-Ferrar basalts, led to the separation of Madagascar and Antarctica from Africa. The final disintegration of Pangea, marked by the opening of South Atlantic and East Indian Oceans, is preceded by a succession of passive then active rifting episodes during the Upper Jurassic and Lower Cretaceous, the transition of which being underlined by the emplacement of the Parana-Etendeka LIP. This last period is also characterized by the development of many intracontinental rifts.

This alternation of passive and active rifting episodes during the Pangea breakup asks about the mechanisms. For the Lower Cretaceous active rifting, responsible for many failed and successful systems, it is interesting to note the coincidence with the occurrence of the mid-Cretaceous “superplume,” which led to a significant worldwide increase of spreading rate from oceanic ridges [*Larson*, 1991a; *Seton et al.*, 2009] between 120 and 80 Ma. For *Larson* [1991a], this superplume originated at about 125 Ma near the core-mantle boundary that means approximatively at the transition between passive to active rifting observed in South Atlantic. The other superplume advocated by *Larson* [1991b] lasted from 323 to 248 Ma (Upper Carboniferous to Lower Triassic) and corresponds to the long period of Permian passive rifting, which preceded the Siberian and Emeishan massive basalt extrusions and associated active rifting episodes. So the relationships between the so-called superplumes and the active rifting episodes are not straightforward. Pangea breakup seems to result from long (typically more than 50 Myr) periods of passive rifting interrupted by short (typically 20 Myr) periods of active rifting.

6.3. Relationships Between the Style of Rifting and the Rift Development

Our analysis suggests that the development of a rift system as a failed or successful system does not depend on the style of rifting (i.e., active or passive; Table 1). There are examples of successful rift systems resulting from passive mode (Neo-Tethys, central Atlantic, and austral segment of South Atlantic) or active mode (central segment of South Atlantic and West India). Reversely, there are examples of failed rift systems in active modes (Siberian rifts and central African rift system) as well as in passive mode (Karoo I). It is easy to understand that a passive rift system can become inactive at each stage of its development if the stress field changes within the lithosphere [*Ziegler and Cloetingh*, 2004]. The reason why all active rift systems is not associated to a new oceanic basin is more complex. In the case of central and equatorial segments of South Atlantic successful rifts versus central Africa failed rifts, the dynamics of rift evolution can explain the selection of one branch rather than the other [*Heine and Brune*, 2014]. By contrast, the causes of the Siberian or Panxi rift system failure after a short period of activity remain puzzling.

If we focus on successful rift systems, it is interesting to investigate the relationships between style of rifting (active versus passive) and type of resulting margins (“magma-poor” versus magma-dominated margins). From the compilation of different margin types done by *Reston* [2009], it appears that here again the relations are not straightforward. For instance, the margins of central Atlantic, which typically result from passive rifting, are magma poor between Morocco and Nova Scotia but magma dominated further to the south. Considering the different segments of the South Atlantic Ocean, the austral segment, which derives from passive rifting, exhibits typical magma-dominated margins, whereas the central segment, which derives from active rifting, shows magma-poor (or “intermediate”) margins. It appears consequently that the importance of melt production during rifting, which determines the type of resulting margins (either magma rich or magma poor), is not a direct consequence of the style of rifting but rather from the efficiency of rift propagation barriers [*Koopmann et al.*, 2014].

6.4. How to Explain the Changes in Rifting Mechanisms Through Time?

Given that continental plates act as thermal insulators and that continental agglomeration promotes mantle global warming and subsequent large-scale melting [i.e., *Coltice et al.*, 2007], a particular global weakness of the continental lithosphere of Pangea is expected. Late Carboniferous-Permian rifting appears consequently as the direct consequence of (1) this weakness and (2) the tension forces described by *Gutiérrez-Alonso et al.* [2008]. These two factors result from the Pangea configuration and are independent from mantle plumes. Interestingly, we note that the onset of Pangea breakup at 195 Ma ensues a typical passive rifting along central Atlantic (Figure 5), whereas the two former active rift systems acting at the Permian-Triassic boundary (Siberian and Panxi) remained failed rifts (Figure 4). It is as if these two almost synchronous LIPs had ensured the transition between a period characterized by very diffuse rifting toward a period characterized by more localized rifting, which will be finally successful.

By the Jurassic, the consequence of the progressive Pangea fragmentation is the decrease of the “global warming” effect and the relative increase of the “plumes” effect. As a matter of fact, the passive rifting observed during the Upper Jurassic and the lowermost Cretaceous is more confined than the one observed during Permian. It has been successful in the Galicia-Newfoundland segment of North Atlantic as well as in the austral segment of South Atlantic but was immediately followed, in South Atlantic, by the Parana-Etendeka LIP, which triggered the final disintegration of Pangea (Figure 6). A rapid examination of the more recent oceanic basins openings show that most of them are resulting from active rifting postdating a plume [i.e., *Courtillot et al.*, 1999]. It is the case for the Red Sea and Gulf of Aden related to the Ethiopian Traps (30 Ma), for the North Atlantic Ocean related to the Greenland Traps (59–60 Ma), and for the two steps of the West Indian Ocean opening related to the Deccan (65 Ma) and Madagascar (88 Ma) Traps, respectively. So a necessary condition for the development of wide passive rift systems seems the existence of large continents (ideally a unique continent). This suggests that passive rifting is triggered by the weakening of the lithosphere (the global warming effect) rather than by an important increase of forces applied at the limits of the continental lithosphere.

7. Conclusion

In this contribution, we try to provide a comprehensive overview of the successive rifting episodes leading to the disintegration of Pangea. We show that the localization of a continental rift can always be explained by an existing weakness corresponding either to a structural inheritance (young mountain belt and previous rift) or to an erupting mantle plume. These two distinct processes, alternating through time, are highlighted using the classical binary categories of active and passive rifting. However, it is worth noting that, in the case of active rifting and contrarily to the classic assumption, we consider the plume as the precursor and not the trigger of the rifting process (Figure 1).

It is established that rifting can enhance asthenosphere upwelling and subsequent volcanic eruptions (as in the case of central Atlantic Triassic system), ensuring the transition from passive to active rifting mode. It is also known that uplift above asthenosphere upwelling can develop without rift growth. It is finally demonstrated that far-field stresses (even very weak) are required to explain strain localization and subsequent rifting above a plume head. So to account for the wide development of rift systems, it is essential to envisage that at the time of rifting the interiors of the plates affected by the rifts were globally “in extension.” It is not the present situation. On the contrary, the Earth’s lithosphere is currently “in compression” due to the dominance

of mountain building and “Ridge Push” effect acting at the Oceanic Spreading Centers. During other periods, typically the ones illustrated in this paper (Carboniferous-Permian, Triassic-Lower Jurassic, and Lower Cretaceous), it is realistic to consider that the dominance of the slab-pull effect explains the general extension observed on the Earth’s surface. In this extensional context the zones of the continental lithosphere weakened by the impingement of Plumes or by structural inheritance can evolve as active or passive rift systems, respectively.

Currently, the questions relative to the global solid Earth’s system are mainly addressed by workers using numerical modeling or by geochemists. With this paper, we wish to convince that, even at this scale, a geological approach is useful to progress in the understanding of the deep Earth. Geological arguments are powerful to draw long-distance correlations between discontinuities, which, in a given place, are often composite due to the stack of several events. Geological arguments are essential to sort out the complex heritage, which, through time, shaped “the face of the Earth.”

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