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Role of rift maturity on the architecture and shortening distribution in mountain belts

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Abstract

Many relic of passive margin are found in mountain belts sometimes together with pristine exhumed mantle. The reasons why these structures are preserved from tectonic reworking and whether the piece of mantle that accompanies them is exhumed during extension or collision remains a question. Using thermo-mechanical modelling we find on the one hand that, the strain distribution in the orogenic wedge is controlled by the original position of crustal domains across the rifted margin. In the upper plate, the structural relationships acquired during the rifting phase between the upper mantle and surrounding units are never reactivated during compression due to crustal strain localization. Mantle exhumation on the other hand can occur either during the collision or rifting phase depending on the thermal regime at the onset of the inversion. Applying these general rules to the Pyrenees, we show that exhumed mantle domain in the former passive margin could not exceed a width of 50 km to 70 km; otherwise obduction would have occurred due to the high thermal regime at the onset of the inversion.

1. Introduction

The impact of inherited mechanical properties of the lithosphere is recognized to be a major driver in the evolution of orogenic belts (Beaumont et al., 2009; Calignano et al., 2015a; Chenin et al., 2017; Heron et al., 2016; Jourdon et al., 2017; Manatschal et al., 2015; Velasque et al., 1989; Yamato et al., 2008). Geological, geophysical and numerical modelling studies show for instance that age-dependent properties of the lithospheric mantle and crust control first-order patterns of shortening distribution, cyclical evolution and seismicity in orogens (Audet and Bürgmann, 2011; Mouthereau et al., 2013). While lateral heterogeneities in the lithospheric mantle control the location of intracontinental orogenic belt (Calignano et al., 2015b; Heron et al., 2016), the local mechanical coupling between crust and mantle or within the crust controls the tectonic style (Jourdon et al., 2017; Lafosse et al., 2018).

In the recent years, an increasing number of studies have found evidence that structure and timing of tectonic evolution of Alpine-type orogens largely inherit the properties and architecture of former rifted margins. For instance, the onset of fast exhumation and abrupt changes in the thermal structure of orogens point to the accretion of stronger, less thinned segment of rifted margins, as shown in the Zagros (Ballato et al., 2010; Mouthereau et al., 2012), Taiwan (Mesalles et al., 2014) or the Pyrenees (Mouthereau et al., 2014; Vacherat et al., 2016, 2014). An important step forward in the exploration of these relationships have been reached thanks to new high-resolution seismological imaging techniques showing the nature of the mountain root and lateral segmentation of an orogenic system can preserve first-order tectonic features of a rift system (Wang et al., 2016), which confirm former inferences from field geology in the Pyrenees (e.g. Jammes et al., 2009; Lagabrielle and Bodinier, 2008).

However, we still have a poor understanding of the key physical processes at play during the earliest stages of collision, when the distal part of a rifted margin is accreted to form mountain

46 ranges. Only a limited number of orogens preserve markers of this transient stage in a
47 continuum of laterally evolving geodynamic settings from subduction to collision. For
48 instance, the complex structure of high-pressure units in the Alps (Beltrando et al., 2014) and
49 the formation of non-metamorphic ophiolitic *mélange* in the Alps or the Pyrenees (Beltrando
50 et al., 2014; Lagabriele and Bodinier, 2008; Müntener et al., 2009) display end-members
51 expression of inverted hyper-extended domain of magma-poor margins. Here, we focus on the
52 role played by the size, number and vergence of detachment zones, thermal maturity and the
53 overall architecture of a rifted margin on the accretion of the hyper-extended domain
54 preserved in collisional belts. Using numerical models we try to get a better understanding on
55 how, where and when rift inherited structures are reactivated throughout the collisional
56 orogenic process and whether exhumed mantle units preserved in mountain belts related to
57 rifting, orogeny or both. The study focuses primarily on the effect of the maturity of the
58 margin, i.e. the residual rift thermal anomaly, on the structure of the nascent mountain belt.

59 **2. Natural constraints for inverted rifts from the Pyrenees**

60 The Pyrenean belt is probably one of the best documented examples of an inverted rift
61 system. The shortening of the former extensional domains with different rift maturities/width
62 along strike resulted in the preservation of different stages of orogeny along strike. The Bay
63 of Biscay fossilized a nascent subduction zone involving an oceanic domain with a classical
64 oceanic crust. Eastward, the convergence in the Pyrenees led to a fully developed orogenic
65 wedge (Figure 1b). The role of rift segmentation is reflected across the Pyrenees as in the
66 Mauléon basin where a former rift structure is interpreted as an inverted hyper-extended
67 margin with an exhumed mantle domain (Jammes et al., 2009; Lagabriele et al., 2010;
68 Lagabriele and Bodinier, 2008; Masini et al., 2014; Teixell et al., 2016; Tugend et al., 2014;
69 Wang et al., 2016), while the eastern part of the belt shows a more complex pattern of

70 discontinuous exhumed mantle zones (Figure 1b; Grool et al., 2018; Vacherat et al., 2014).
71 According to paleogeographic reconstructions (Jammes et al., 2009; Tugend et al., 2014;
72 Nirrengarten et al., 2018), a major transtensional rifting phase started in the Late Jurassic-
73 Early Cretaceous (~150-120 Ma). Extension between Europe and Iberia was first distributed
74 between the Central Iberian Belt and the Pyrenees. Then, rifting localized in the Pyrenean
75 domain in the Late Aptian-Albian (~115 Ma) and resulted in mantle exhumation in the North
76 Pyrenean zone (Jammes et al., 2009; Lagabriele et al., 2010; Lagabriele and Bodinier, 2008;
77 Masini et al., 2014; Teixell et al., 2016; Tugend et al., 2014). The late stages of rifting are
78 marked by a HT-LP metamorphism (Albarède and Michard-Vitrac, 1978; Montigny et al.,
79 1986). It corresponds to a strong thermal anomaly affecting both pre- and syn-rift sediments
80 that locally exceed 600°C in sedimentary basins close to exhumed mantle domains (Clerc et
81 al., 2012; Ducoux, 2017; Vacherat et al., 2014). Vacherat et al. (2014) showed that the rift-
82 inherited thermal anomaly prevailed as convergence started in Late Cretaceous, which
83 supports a fast plate boundary inversion with no evidence for thermal relaxation.

84 Pyrenean orogeny is often considered as two-folds (Dumont et al., 2015; Mouthereau et al.,
85 2014). A first early orogenic phase consists of the inversion of former north pyrenean rifts
86 (Dumont et al., 2015; Mouthereau et al., 2014; Tugend et al., 2014; Vacherat et al., 2016) as
87 early as the Santonian (~85 Ma). Incipient orogenic reliefs are also recorded during this phase
88 as evidenced by thermochronology (Vacherat et al., 2016). The second orogenic phase
89 consists in the Eocene-Oligocene crustal nappe-stacking of the Axial zone (Pyrenean high
90 range) corresponding to the main stage of crustal thickening, relief and topographic growth
91 until the early Miocene post-orogenic time (Mouthereau et al., 2014). Although the timing
92 and history of the Pyrenean rifting and orogeny seems to be well constrained, the way the
93 former influence the latter is often postulated but poorly evaluated physically. As a result, first
94 order questions such as whether the actual pre-orogenic width of the exhumed mantle/oceanic

95 domain between Iberia and Europe in the Pyrenees was 10 km or 200 km remain open,
96 leaving large uncertainty in regional plate kinematic scenario.

97 **3. Dynamic modelling of tectonically-inverted hyper-extended rifts**

98 Numerous numerical modelling studies have been performed to understand the role of the
99 rheology and of the properties and architecture of continental lithosphere during collision
100 (Beaumont et al., 2009, 2001; Burg and Gerya, 2005; Burov et al., 2014, 2011, 2001; Duretz
101 et al., 2012; Duretz and Gerya, 2013; Dymkova et al., 2015; Jourdon et al., 2017; Warren et
102 al., 2008; Yamato et al., 2008). However, the majority of these models did not account for the
103 complexity of rifted margins. Based on the Pyrenean study case, studies (Erdös et al., 2014;
104 Jammes et al., 2014; Jammes and Huisman, 2012) showed that the extension phase preceding
105 the collision can explain the observed presence of mantle bodies at shallow depth and that
106 inherited structures facilitate the propagation of the deformation in the orogenic wedge. These
107 models assume narrow rifts with limited amount of extension (thinning factors of about 2).
108 The lack of a large domain with exhumed mantle is at odd with commonly accepted
109 reconstructions of the Pyrenean rifts where extension led to a mature and wide hyper-
110 extended domain but failed to record seafloor spreading. Offshore mapping of rift domains in
111 the Bay of Biscay indeed argued for the occurrence of a 100 km large widespread area of
112 hyper-thinned continental crust, exhumed domains and even oceanic crust (e.g. Tugend et al.,
113 2014). This result accords with geophysical imaging of magma-poor rifted margins showing a
114 70 to 200 km wide domain of hyper-extended crust, typically thinned below 10 km,
115 sandwiched between the necking zone and a domain of exhumed mantle (e.g. Chenin et al.,
116 2017).

117 Numerical models which are able to produce such magma-poor hyper-extended margins
118 require a weak crustal channel to decouple crust and mantle which in turn produces a delay in

the necking of the mantle. In 2D, they require slow extension rates ($< 1\text{cm/yr}$) (Brune et al., 2017, 2014; Manatschal et al., 2015; Naliboff et al., 2017; Tetreault and Buiter, 2017) or a mechanically layered lithosphere (Duretz et al., 2016) in which pre-existing softened zones permit to keep the deformation sufficiently distributed during lithospheric thinning to delay the break-up. During the final mantle exhumation phase, the 1200°C (LAB) isotherm typically lies at ~ 20 km or more beneath a magma-poor type Mid Oceanic ridge axis involving a “cold” rift geotherm (Dick, 1989; Lagabrielle and Lemoine, 1997; Sauter et al., 2013, 2004).

Among these 2D models, the crustal thinning is firstly accommodated by simultaneous faulting, and then followed by a sequential faulting (Brune et al., 2017, 2014; Manatschal et al., 2015; Naliboff et al., 2017; Tetreault and Buiter, 2017) rooting along a main detachment shear zone called “exhumation channel” (Brune et al., 2014) or “viscous channel” (Lavie and Manatschal, 2006) which allows rift migration and the formation of asymmetric rift systems. This process of strain localization along shallow dipping shear zones and rift asymmetry is interpreted as resulting from mechanical softening with ongoing deformation (Huisman et al., 2005; Lavie and Manatschal, 2006).

3.1. Rheology

Our models setup took benefit of former numerical models that were successful in reproducing the formation of magma-poor hyper-extended margins. The initial layering is designed to represent a simplified pre-rift lithosphere. The continental crust is divided in three layers (Figure 2a): (1) the upper crust from the surface to 10 km, (2) the middle crust from 10 km to 20 km and (3) the lower crust from 20 km to 35 km. Upper and middle crust rheology is approximated with a quartz flow law (Table 1, Ranalli and Murphy, 1987) according to the granitic composition of the upper crust (Denèle et al., 2012; Wickham, 1987) while the lower crust rheology is simulated with an anorthite flow law (Table 1, Rybacki and Dresen, 2000)

representing a more felsic composition (Wickham, 1987). This layering allows a decoupling between the middle and lower crust (Figure 2b).

For all runs, the mantle is divided into two phases: (1) the lithospheric mantle from 35 km to 120 km and (2) the asthenosphere from 120 km to 200 km. The rheology of these two phases is modelled with a dry olivine dislocation creep law (Table 1, Hirth and Kohlstedt, 2003).

In addition, the brittle parts of the lithosphere are simulated using the Drucker-Prager yield criterion adapted to continuum mechanics (see Supplementary Material). We implement at the beginning of the simulation a random Gaussian distribution (see supplementary material for details) of a small plastic strain amount (0 to 0.4) to simulate an already damaged lithosphere. This initial plastic strain will decrease the friction angle following a linear plastic strain softening (Figure 2c). The plastic strain softening is used to roughly approximate the behaviour of large faults that record long-term slip and localization due to phyllosilicates crystallisation in crustal rocks or serpentinization in mantle rocks.

3.2. Thermal setup

The initial thermal state is computed with the analytical solution provided in Burov & Diament (1995), with a radiogenic heat production in the crust of $8.91 \mu\text{W.m}^{-3}$. A thermal perturbation is added at the centre of the domain (Figure 2a). The 1300°C isotherm representing the lithosphere-asthenosphere boundary is located at 120 km and the Moho temperature is 603°C (Müntener et al., 2000) (Figure 2b). The thermal boundary conditions to solve energy conservation are: 0°C at the surface, 1340°C at the bottom of the domain (200 km) and null heat flux on the left and right sides of the model.

3.3. Velocity boundary conditions

In order to investigate the influence of extensional structures during the collision we simulate an extension phase followed by compression. During the first 30 Myr we impose a constant

extension velocity (Figure 2d and Supplementary material Table S1, given velocity values are the absolute velocity difference between the two model sides). We adopt a 5 Myr window between extension and compression to simulate a smooth plate boundary inversion (Figure 2d) rather than a sharp extension to compression inversion. In detail, model velocity decreases linearly from 30 Ma to reach zero at 32.5 Ma. Then, from 33 to 35 Ma the compression velocity linearly increases to reach 2.5 mm.yr^{-1} and remains constant until the end of the simulation. The extension velocity varies from a model to another in order to investigate the role of along-strike variations of margin width and maturity during mountain building for an extensive system which undergoes the same period of extension.

Our models are not aimed to represent orogenies which went through a long convergence history like the India-Asia hyper-collision in which colliding lithospheres have very different thermo-mechanical properties (Molnar and Tapponnier, 1981, 1975). However, these models represent alpine-type collisional belts, where the upper plate is weakly deformed and structures of the former passive margin are well preserved.

Models presented in this study are performed with pTatin (May et al., 2015, 2014) a parallel implementation of the finite element method, which employs an Arbitrary Lagrangian Eulerian discretization, together with the material point method for the rheologies described in supplementary material. It is coupled to a diffusive surface processes model (see details in Jourdon et al. 2018).

4. Results

4.1. *Structure of hyper-extended domains with variable amount and rates of extension*

All our experiments show the systematic development of lithospheric-scale detachment during rifted margin formation, which number, direction of slip and lateral extent are shown to depend on the amount and rates of extension.

In runs 1 to 3 deformation first localizes at the middle-lower crust boundary. As extension increases the shear zones exhumed through the crust and progressively reach the surface. The decoupling between middle and lower crust is accommodated by a shallow dipping shear zone in both proximal and necking domain during the whole rifting phase. A second decoupling level develops between the lower crust and the mantle. The oceanward sense of shear eventually leads to lower crust extraction in the distal domain resulting in mantle exhumation to upper crustal levels. Depending on the extension velocity, which is equivalent to the total amount of stretching in our model, the final rift structure differs and the geothermal gradient becomes greater with increasing extension velocity. At the end of extension, Run 1 displays an aborted rift system characterized by four antithetic normal fault zones (Figure 3). The most external fault zones are synchronous conjugate shear zones which developed during the first 15 Myrs and constitute the necking domain. Then two sequential shear zones are formed in the centre of the rift. At the end of the rifting phase the mantle is not exhumed and a 10-20 km thick crust remains.

Runs 2 and 3 display more mature passive margin structures and the mantle is exhumed to the surface. During crustal thinning the major part of the deformation is accommodated by crustal normal shear zones in the distal domain connected to a single low-angle detachment dipping oceanward at the Moho and rooting into the mantle in steeply dipping shear zone. This

lithospheric scale detachment represents the upper-lower plate boundary. It is remarkable that once the crust is thinned enough (~5 km thick) the upper-lower plate relationship switches and a second steep mantle detachment of opposite vergence initiates (Figure 4). Run 3 (Figure 5) characterized by more extension, shows a succession of sequentially formed detachments exhuming the mantle asymmetrically along mantle core complex structures. During mantle exhumation, the shear zone flattens until it becomes mechanically easier to create a new detachment rather than continuing to exhume along the previous one. At higher extension velocities, these structures vanish due to a warmer thermal regime which can be assimilated to a mid-oceanic ridge.

4.2. Rift inversion and mountain building

The onset of inversion is accommodated by reactivating former weak crustal extensional shear zones driven by first strain localization in the mantle where the rift thermal anomaly was maximal. The first back-thrust develops in the upper plate 10 to 15 Myrs after the polarity change. Following the onset of continental accretion, the crustal strain propagates into the down going plate along the reactivated middle-lower crust décollement. As the lowermost crust is underthrust, the middle/upper crust is shortened and thickened to form a typical thick-skinned orogenic wedge. Shortening ratios of each former passive margin domains remain close between the different runs. Indeed, the lower plate hyper-extended domain shortens from 40% to 60%. The lower plate necking domain shortens from 35% to 50% while the upper plate records no shortening up to 30%. Internal deformation within the wedge results in tilting of former thrusts and their passive transport toward the rigid upper plate along the retro-wedge thrust. The mantle exhumed in the upper crust during the rifting acts as a buttress during inversion. The back-thrust plays an important role in the deformation process as it localizes almost all the upper plate shortening. Coupled with orogenic wedge thickening and erosion, it allows exhumation of the deepest units.

235 In the slowest rift case (run 1; Figure 3) the mantle is not exhumed. However, the lithospheric
236 mantle remains incorporated to the belt at the end of collision. This mantle unit is bordered by
237 a reactivated shear zone, which is the first shortening structure in the crust. The former
238 extensional shear zones which linked the two decoupling levels at middle-lower crust
239 boundary and along the Moho are passively transported in the hanging wall of the retro-
240 wedge thrust, and the extensional relationships between the crust and mantle are preserved.
241 Finally, a second back-thrust is activated during the last steps of the collision and reactivates a
242 former extensional shear zone.

243 In runs 2 and 3, mantle exhumation during rifting is observed. Larger exhumed mantle
244 domain leads to larger mantle units outcropping in the mountain belt. Run 2 (Figure 4) shows
245 a first back-thrust structure located in the mantle, which is abandoned 20 Myr after the onset
246 of tectonic inversion. A new retro-wedge thrust forms closer to the surface. This thrust
247 nucleates in the mantle and then connects upward into a former extensional shear zone in the
248 crust. This structure accommodates all the upper plate shortening, letting the most distant
249 normal shear zones intact. At the end of the simulation the upper plate preserves most of the
250 architecture inherited from the rifting stage. Once again, the mantle and crust units carried by
251 the retro-wedge thrust exhibit rifting structures which are tilted but not deformed during the
252 collision.

253 Run 3 (Figure 5) displays a wide exhumed mantle domain (~70 km), which formation was
254 accommodated by extension across 5 detachments. The inversion process starts as in the other
255 runs along a shear zone connecting the thinnest part of the rifted margin with the thermal
256 anomaly in the mantle. As the shortening increases, the margin underthrusts the exhumed
257 mantle assisted by the reactivation of crustal normal faults and mantle detachments. At the
258 end of the experiment, the orogenic wedge is shaped by the underthrust continental passive

margin and a wide suture zone of obducted exhumed mantle while the upper continental plate is poorly deformed.

5. Discussion

5.1. Rift architecture

All our experiments show the systematic development of lithospheric-scale detachment during rifted margin formation, which number, direction of slip and lateral extent are shown to depend on the amount and rates of extension. Detachment structures develop sequentially in the mantle, always landward-dipping regarding the position of the plate boundary. A first detachment is created when the continental crust reaches a thickness less than 5 km. Then, detachments develop regularly every 1 Myr or 2 Myrs and result in regularly spaced structures. This deformation and mantle exhumation sequence is a characteristics of slow-spreading ridges as the southwest Indian ridge (Reston and Mcdermott, 2011; Sauter et al., 2013) and has been imaged in magma-poor conjugate margins of the Newfoundland (Hopper and Funck, 2004) and Iberia (Krawczyk et al., 1996; Reston et al., 2007) margins. Moreover, these detachments can sometimes be found in exhumed mantle or ophiolites as in the Tasna nappe in the Swiss Alps (Froitzheim and Rubatto, 1994; Manatschal, 2014; Reston and Mcdermott, 2011). In the model displaying this detachment sequence (run 3, Figure 5), the obducted mantle exhibits inverted detachment faults.

5.2. Thermal structure

The temperature peak of the rifted margin in the Pyrenees is evaluated between 200°C and 600°C in the pre- and syn-rift sediments (Clerc et al., 2012; Ducoux, 2017; Vacherat et al., 2014). However, the most part of metamorphosed pre- and syn-rift sediments reached 300°C to 350°C (Clerc et al., 2012; Vacherat et al., 2014). In our study, syn-rift sediments undergo

200°C to 330°C, i.e., the average peak temperature recorded in syn-rift Pyrenean sediments. This temperature is limited by the fact that in 2D, wide margins form at small rate of extension promoting cold geotherm or in 3D, the presence of an out of plane propagator also permits to distribute extension at larger rate than in 2D enabling warmer regime to occur at similar crustal thickness and margin width (Le Pourhiet et al., 2018). Alternatively, the presence of highly conductive salt layer in the pre-rift sediment could also have affected the temperature recorded in the sediments and none of the model above account for possible hydrothermal heating.

5.3. *Mantle exhumation*

Mantle bodies exposed in the Pyrenees are interpreted to be remnants of the subcontinental mantle exhumed during the rifting phase (Lagabrielle et al., 2010; Lagabrielle and Bodinier, 2008). We note, however, that in the case of a small amount of extension as in run 2, the mantle may be exhumed while being passively transported along the retro-wedge. Moreover, either in compression or in extension, the exhumed mantle still displays at the end of the collision the structural patterns acquired during the rifting phase. For instance, the extensional shear zones bounding the exhumed mantle unit are not reactivated. Natural examples of exhumed mantle units incorporated into mountain belts display preserved extensional structures such as the Beni Bousera peridotite in the Rif (Frets et al., 2014) or the Ronda peridotite in the Betics (Précigout et al., 2013). The finding of extensional structures associated with exhumed mantle bodies cannot therefore be considered as a definitive explanation of the main process of mantle exhumation.

5.4. *Inversion process*

In all runs of this study, inversion is localized on the rift thermal anomaly. Indeed, the upward advection of the isotherms leads to strong temperature-dependent weakening of the

lithospheric mantle through diffusion or dislocation creep. Therefore, when the inversion starts, strain localizes on the weakest part of the domain. One important result is that strain localization controlled by the rifting thermal anomaly triggers the underthrusting of one of the rifted margins beneath the exhumed mantle domain. This result implies that the width of the exhumed mantle domain cannot exceed a certain size, beyond which obduction occurs (Figure 6). In the Pyrenees, the time laps between extension and compression is short and occurs within a high temperature gradient (Masini et al., 2014; Vacherat et al., 2014). Our study shows that such boundary conditions can reproduce a case like the Pyrenean orogeny only for a small (< 50 km) exhumed mantle domain developed between the Iberian and European plates (Figure 6), which is consistent with estimates from Mouthereau et al. (2014).

5.5. *Preserved rift structures in pro- and retro-wedge sides of the orogen*

The growth of an asymmetric collisional orogenic wedge in our experiments leads to the formation of classical pro- and retro-wedge structures. Such pop-up like orogenic structure is generally well-reproduced in models assuming a kinematic singularity (Jamieson and Beaumont, 2013) and more recently in models accounting for thermal perturbations at the base of the lithosphere (Jaquet et al., 2018). Deformation sequence in the lower plate propagates from the thinnest part of the former hyper-extended domain to the proximal margin. The former hyper-extended domain shortens from 40% to 60%. The deformation is mainly localized on former normal faults. Then, it propagates to the necking domain which shortens around 40% before reaching the proximal domain. Such values are in the range of crustal strains estimated from balanced cross-sections in external orogens (Mouthereau et al., 2013). Our experiments may therefore help geologists estimating shortening values in orogens from the knowledge of the shortening rifted segment. In our models that explicitly account for rift inheritance, the retro-wedge side of the orogen is a direct consequence of extension. Indeed, the upper plate that hosts the retro-wedge is characterized by limited shortening.

Former normal faults are well preserved and the crust is poorly or even not thickened. This behaviour is due to: (i) the small amount of underthrust mantle limiting heat transfer (or melt) in the upper plate and therefore thermal weakening; (ii) the large amount of exhumed mantle at upper crustal levels acts as a buttress where the retro-wedge structures root; (iii) the ability of the lower plate to deform easily during its underthrusting. Moreover, the former rifting phase strongly influences the deep structure of the belt, particularly beneath the suture zone and the upper plate. Indeed, large mantle bodies exhumed beneath hyper thinned margins are preserved beneath the upper plate and above the crustal root. Otherwise, when the crust remains as thick as in the run 1, the deep structure of the belt exhibits less mantle and a sharp vertical transition from mantle to crust in the suture zone.

6. Conclusions

Through thermo-mechanical modelling we are able to provide new constraints and interpretations about the Pyrenean rifting and its influence on the following mountain building stage:

- The rapid extension to compression switch of stress regime accompanied by a hot thermal regime implies that the width of the mantle exhumed domain could not exceed 50 km to 70 km, otherwise obduction of large mantle bodies would have occurred.
- Mantle exhumation during extension is driven by continentward dipping detachment faults developing sequentially at the plate boundary.
- Extensional shear zones that originally developed in the lower plate are reactivated as thrusts while extensional structures of the upper plate are mainly preserved due to strain localization in the lower plate and along the retro-wedge thrust.
- The exhumed mantle incorporated in the mountain belt does not display reactivated normal shear zones. The relationships between the crust and exhumed mantle acquired during the rifted phase are conserved.

- The presence of the mantle in the mountain belt is not necessarily due to mantle exhumation during rifting. Mantle can also exhume during shortening stages. However, models displaying mantle exhumation during extension end with a higher proportion of exhumed mantle in the mountain belt.
- All simulations showed that the shortening amounts are very close regarding the former passive margin domains regardless of the size of those former domains. Once the former passive margin domains identified, this result can help to evaluate the shortening amount in mountain belts when no markers are available.

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376 **Table**

	Units	Quartz	Anorthite	Olivine
Reference		Ranalli and	Rybacki and	Hirth and
		Murphy	Dresen 2000	Kohlstedt
		1987		2003
A	MPa ⁿ .s ⁻¹	6.3x10 ⁻⁶	13.4637	1.1x10 ⁵
n		2.4	3	3.5
Q	KJ.mol ⁻¹	156x10 ³	345x10 ³	530x10 ³
V	m ³ .mol ⁻¹	0	3.8x10 ⁻⁵	18x10 ⁻⁶

377 Table 1

378

379 **Figures captions**

380

381 Figure 1: a) Structural map of the Pyrenees (modified from Mouthereau et al., 2014). Thick
382 black dashed lines indicate cross-sections c) and d). Coloured dots indicate the former
383 extensional domains (modified from Tugend et al., 2014). b) Structural sketch map of the
384 Pyrenean rift during the Santonian (~85 Ma, modified from Tugend et al., 2014). c)
385 Geological cross-section of the western Pyrenees interpreted from tomographic model
386 (modified after Wang et al., 2016). d) Geological cross-section of the central Pyrenees
387 (modified from Mouthereau et al., 2014).

388 Figure 2: a) Numerical phases geometry, thermal structure of the models and boundary
389 conditions. b) Yield-stress envelopes computed for a strain rate of 10⁻¹⁵ s⁻¹ for the different
390 materials used to simulate the lithosphere. The Yield-stress envelopes represent the high stress

cutoff used in the simulations (400 MPa). This cutoff is set to roughly approximate the strength limitation for high confinement pressures (Watremez et al., 2013). The red line indicates the initial undisturbed geotherm. c) Plastic strain softening, the friction angle decreases from 30° to 3° with the increase of plastic strain. d) Variation during the runs of the absolute velocity difference between the two model sides. Ext: extension, Comp: compression.

Figure 3: Run 1 evolution. Extension phase velocity of 4 mm/a over 30 Myr. Inversion starts at 35 Myr. Colours of the left panel represent the cumulative plastic and viscous strain. Colours of the right panel represent the numerical phases of the run. Sediments are indicated by 1 Myr alternating colour bands. Black lines represent the extensive shear zones, red lines represent the compressive shear zones and dashed black and red lines represent the reactivated shear zones. The frictional shear zones are drawn for a corresponding plastic strain superior or equal to 1. The viscous shear zones are drawn for a corresponding viscous strain superior or equal to $\sim 1.8-2$. Abbreviations represent the passive margin domains, N.D.: Necking domain. Colours are the same than in the horizontal shortening legend.

Figure 4: Run 2 evolution. Extension phase velocity of 5 mm/a over 30 Myr. Inversion starts at 35 Myr. Colours of the left panel represent the cumulative plastic and viscous strain. Colours of the right panel represent the numerical phases of the run. Sediments are indicated by 1 Myr alternating colour bands. Black lines represent the extensive shear zones, red lines represent the compressive shear zones and dashed black and red lines represent the reactivated shear zones. The frictional shear zones are drawn for a corresponding plastic strain superior or equal to 1. The viscous shear zones are drawn for a corresponding viscous strain superior or equal to $\sim 1.8-2$. Abbreviations represent the passive margin domains, N.D.: Necking domain, H.E.: Hyper-Extended domain. Colours are the same than in the horizontal shortening legend.

Figure 5: Run 3 evolution. Extension phase velocity of 6 mm/a over 30 Myr. Inversion starts at 35 Myr. Colours of the left panel represent the cumulative plastic and viscous strain. Colours of the right panel represent the numerical phases of the run. Sediments are indicated by 1 Myr alternating colour bands. Black lines represent the extensive shear zones, red lines represent the compressive shear zones and dashed black and red lines represent the reactivated shear zones. The frictional shear zones are drawn for a corresponding plastic strain superior or equal to 1. The viscous shear zones are drawn for a corresponding viscous strain superior or equal to ~1.8-2. Abbreviations represent the passive margin domains, N.D.: Necking domain, H.E.: Hyper-Extended domain. Colours are the same than in the horizontal shortening legend.

Figure 6: Interpretative sketches based on models results. Narrow mantle exhumed domain rifts result in continental collision when inverted during hot geotherm. Wide mantle exhumed domain rifts result in obduction during inversion process with hot geotherms.

Table 1: Parameters used for Arrhenius flow laws.

References

- Albarède, F., Michard-Vitrac, A., 1978. Age and significance of the North Pyrenean metamorphism. *Earth Planet. Sci. Lett.* 40, 327–332. [https://doi.org/10.1016/0012-821X\(78\)90157-7](https://doi.org/10.1016/0012-821X(78)90157-7)
- Audet, P., Bürgmann, R., 2011. Dominant role of tectonic inheritance in supercontinent cycles. *Nat. Geosci.* 4, 184–187. <https://doi.org/10.1038/ngeo1080>
- Ballato, P., Mulch, A., Landgraf, A., Strecker, M.R., Dalconi, M.C., Friedrich, A., Tabatabaei, S.H., 2010. Middle to late Miocene Middle Eastern climate from stable oxygen and carbon isotope data , southern Alborz mountains , N Iran. *Earth Planet. Sci. Lett.* 300, 125–138. <https://doi.org/10.1016/j.epsl.2010.09.043>

438 Beaumont, C., Jamieson, R.A., Butler, J.P., Warren, C.J., 2009. Crustal structure : A key
 439 constraint on the mechanism of ultra-high-pressure rock exhumation. *Earth Planet. Sci.*
 440 *Lett.* 287, 116–129. <https://doi.org/10.1016/j.epsl.2009.08.001>

441 Beaumont, C., Jamieson, R.A., Nguyen, M.H., Lee, B., 2001. Himalayan tectonics explained
 442 by extrusion of a low-viscosity crustal channel coupled to focused surface denudation.
 443 *Nature* 414, 738–742. <https://doi.org/10.1038/414738a>

444 Beltrando, M., Manatschal, G., Mohn, G., Dal Piaz, G.V., Vitale Brovarone, A., Masini, E.,
 445 2014. Recognizing remnants of magma-poor rifted margins in high-pressure orogenic
 446 belts: The Alpine case study. *Earth-Science Rev.* 131, 88–115.
 447 <https://doi.org/10.1016/j.earscirev.2014.01.001>

448 Brune, S., Heine, C., Clift, P.D., Pérez-Gussinyé, M., 2017. Rifted margin architecture and
 449 crustal rheology: Reviewing Iberia-Newfoundland, Central South Atlantic, and South
 450 China Sea. *Mar. Pet. Geol.* 79, 257–281.
 451 <https://doi.org/10.1016/j.marpetgeo.2016.10.018>

452 Brune, S., Heine, C., Pérez-Gussinyé, M., Sobolev, S. V., 2014. Rift migration explains
 453 continental margin asymmetry and crustal hyper-extension. *Nat. Commun.* 5, 1–9.
 454 <https://doi.org/10.1038/ncomms5014>

455 Burg, J., Gerya, T., 2005. The role of viscous heating in Barrovian metamorphism of
 456 collisional orogens : thermomechanical models and application to the Lepontine Dome in
 457 the Central Alps. *J. Metamorph. Geol.* 23, 75–95. [https://doi.org/10.1111/j.1525-](https://doi.org/10.1111/j.1525-1314.2005.00563.x)
 458 [1314.2005.00563.x](https://doi.org/10.1111/j.1525-1314.2005.00563.x)

459 Burov, E., 2011. Rheology and strength of the lithosphere. *Mar. Pet. Geol.* 28, 1402–1443.
 460 <https://doi.org/10.1016/j.marpetgeo.2011.05.008>

461 Burov, E., Diament, M., 1995. The effective elastic thickness (T_e) of continental lithosphere:
 462 What does it really mean? *J. Geophys. Res.* 100, 3905–3927.

463 Burov, E., Francois, T., Agard, P., Le, L., Meyer, B., Tirel, C., Lebedev, S., Yamato, P., Brun,
 464 J., 2014. Rheological and geodynamic controls on the mechanisms of subduction and HP
 465 / UHP exhumation of crustal rocks during continental collision : Insights from numerical
 466 models. *Tectonophysics* 631, 212–250. <https://doi.org/10.1016/j.tecto.2014.04.033>

467 Burov, E., Jolivet, L., Le Pourhiet, L., Poliakov, A., 2001. A thermomechanical model of
 468 exhumation of high pressure (HP) and ultra-high pressure (UHP) metamorphic rocks in
 469 Alpine-type collision belts. *Tectonophysics* 342, 113–136.
 470 [https://doi.org/10.1016/S0040-1951\(01\)00158-5](https://doi.org/10.1016/S0040-1951(01)00158-5)

471 Calignano, E., Sokoutis, D., Willingshofer, E., Gueydan, F., 2015a. Asymmetric vs .
 472 symmetric deep lithospheric architecture of intra-plate continental orogens. *Earth Planet.*
 473 *Sci. Lett.* 424, 38–50. <https://doi.org/10.1016/j.epsl.2015.05.022>

474 Calignano, E., Sokoutis, D., Willingshofer, E., Gueydan, F., Cloetingh, S., 2015b. Strain
 475 localization at the margins of strong lithospheric domains: Insights from analog models.
 476 *Tectonics* 1–17. <https://doi.org/10.1002/2014TC003756>.Received

477 Chenin, P., Manatschal, G., Picazo, S., Müntener, O., Karner, G., Johnson, C., Ulrich, M.,
 478 2017. Influence of the architecture of magma-poor hyperextended rifted margins on
 479 orogens produced by the closure of narrow versus wide oceans. *Geosphere* 13, 559–576.
 480 <https://doi.org/10.1130/GES01363.1>

481 Clerc, C., Lagabriele, Y., Neumaier, M., Reynaud, J.Y., De Saint Blanquat, M., 2012.
 482 Exhumation of subcontinental mantle rocks: Evidence from ultramafic-bearing clastic
 483 deposits nearby the Lherz peridotite body, French Pyrenees. *Bull. la Soc. Geol. Fr.* 183,

484 443–459. <https://doi.org/10.2113/gssgfbull.183.5.443>

485 Denèle, Y., Paquette, J.L., Olivier, P., Barbey, P., 2012. Permian granites in the Pyrenees: The
486 Aya pluton (Basque Country). *Terra Nov.* 24, 105–113. <https://doi.org/10.1111/j.1365->
487 3121.2011.01043.x

488 Dick, H.J.B., 1989. Abyssal peridotites , very slow spreading ridges and ocean ridge
489 magmatism Abyssal peridotites. *Geol. Soc. London, Spec. Publ.* 42, 71–105.

490 Ducoux, M., 2017. Structure, thermicité et évolution géodynamique de la Zone Interne
491 Métamorphique des Pyrénées. University of Orléans.

492 Dumont, T., Replumaz, A., Rouméjon, S., Briais, A., Rigo, A., Bouillin, J.P., 2015.
493 Microseismicity of the Béarn range: Reactivation of inversion and collision structures at
494 the northern edge of the Iberian plate. *Tectonics* 34, 934–950.
495 <https://doi.org/10.1002/2014TC003816>

496 Duretz, T., Gerya, T. V, 2013. Tectonophysics Slab detachment during continental collision :
497 In fl uence of crustal rheology and interaction with lithospheric delamination.
498 *Tectonophysics* 602, 124–140. <https://doi.org/10.1016/j.tecto.2012.12.024>

499 Duretz, T., Gerya, T. V, Kaus, B.J.P., Andersen, T.B., 2012. Thermomechanical modeling of
500 slab eduction 117, 1–17. <https://doi.org/10.1029/2012JB009137>

501 Duretz, T., Petri, B., Mohn, G., Schmalholz, S.M., Schenker, F.L., Müntener, O., 2016. The
502 importance of structural softening for the evolution and architecture of passive margins.
503 *Nat. Publ. Gr.* 1–7. <https://doi.org/10.1038/srep38704>

504 Dymkova, D., Gerya, T., Burg, J., 2015. 2D thermomechanical modelling of continent – arc –
505 continent collision. *Gondwana Res.* <https://doi.org/10.1016/j.gr.2015.02.012>

506 Erdős, Z., Huismans, R.S., Van Der Beek, P., Thieulot, C., 2014. Extensional inheritance and
507 surface processes as controlling factors of mountain belt structure. *J. Geophys. Res.*
508 *Solid Earth* 9042–9061. <https://doi.org/10.1002/2014JB011408>

509 Frets, E.C., Tommasi, A., Garrido, C., Vauchez, A., Mainprice, D., Kamaltarguisti, Amri, I.,
510 2014. The beni bousera peridotite (rif belt, morocco): An oblique-slip low-angle shear
511 zonethinning the subcontinental mantle lithosphere. *J. Petrol.* 55, 283–313.
512 <https://doi.org/10.1093/petrology/egt067>

513 Froitzheim, N., Rubatto, D., 1994. Continental breakup by detachment faulting : field
514 evidence and geochronological constraints (Tasna nappe, Switzerland). *Terra Nov.* 10,
515 171–176.

516 Grool, A.R., Ford, M., Vergés, J., Huismans, R.S., Christophoul, F., Dielforder, A., 2018.
517 Insights Into the Crustal-Scale Dynamics of a Doubly Vergent Orogen From a
518 Quantitative Analysis of Its Forelands: A Case Study of the Eastern Pyrenees. *Tectonics*
519 450–476. <https://doi.org/10.1002/2017TC004731>

520 Heron, P., Stephenson, R., Heron, P.J., Pysklywec, R.N., Stephenson, R., 2016. Identifying
521 mantle lithosphere inheritance in controlling intraplate orogenesis. *J. Geophys. Res.*
522 *Solid Earth* 121, 6966–6987. <https://doi.org/10.1002/2016JB013460>

523 Heron, P.J., Pysklywec, R.N., Stephenson, R., 2016. Lasting mantle scars lead to perennial
524 plate tectonics. *Nat. Commun.* 7, 1–7. <https://doi.org/10.1038/ncomms11834>

525 Hirth, G., Kohlstedt, D.L., 2003. Rheology of the Upper Mantle and the Mantle Wedge: A
526 View from the Experimentalists. *Geophys. Monogr.* 138, 83–105.

527 Hopper, J.R., Funck, T., 2004. Continental breakup and the onset of ultraslow seafloor
528 spreading off Flemish Cap on the Newfoundland rifted margin. *Geology* 32, 93–96.

529 <https://doi.org/10.1130/G19694.1>

530 Huismans, R.S., Buiter, S.J.H., Beaumont, C., 2005. Effect of plastic-viscous layering and
531 strain softening on mode selection during lithospheric extension. *J. Geophys. Res. B*
532 *Solid Earth* 110, 1–17. <https://doi.org/10.1029/2004JB003114>

533 Jamieson, R.A., Beaumont, C., 2013. On the origin of orogens. *Bull. Geol. Soc. Am.* 125,
534 1671–1702. <https://doi.org/10.1130/B30855.1>

535 Jammes, S., Huismans, R.S., 2012. Structural styles of mountain building: Controls of
536 lithospheric rheologic stratification and extensional inheritance. *J. Geophys. Res. B Solid*
537 *Earth* 117, 1–22. <https://doi.org/10.1029/2012JB009376>

538 Jammes, S., Huismans, R.S., Muñoz, J.A., 2014. Lateral variation in structural style of
539 mountain building: Controls of rheological and rift inheritance. *Terra Nov.* 26, 201–207.
540 <https://doi.org/10.1111/ter.12087>

541 Jammes, S., Manatschal, G., Lavier, L., Masini, E., 2009. Tectonosedimentary evolution
542 related to extreme crustal thinning ahead of a propagating ocean: Example of the western
543 Pyrenees. *Tectonics* 28, 1–24. <https://doi.org/10.1029/2008TC002406>

544 Jaquet, Y., Duretz, T., Grujic, D., Masson, H., Schmalholz, S.M., 2018. Formation of
545 orogenic wedges and crustal shear zones by thermal softening , associated topographic
546 evolution and application to natural orogens. *Tectonophysics* 746, 512–529.
547 <https://doi.org/10.1016/j.tecto.2017.07.021>

548 Jourdon, A., Le Pourhiet, L., Petit, C., Rolland, Y., 2017. The deep structure and reactivation
549 of the Kyrgyz Tien Shan: Modelling the past to better constrain the present.
550 *Tectonophysics*. <https://doi.org/10.1016/j.tecto.2017.07.019>

551 Krawczyk, C.M., Reston, T.J., Beslier, M., Boillot, G., 1996. Evidence for detachment
 552 tectonics on the Iberia Abyssal Plain rifted margin, in: Whitmarsh, R. et al. (Ed.),
 553 Proceedings of the Ocean Drilling Program, Scientific Results. Texas, pp. 603–615.

554 Lafosse, M., Boutoux, A., Bellahsen, N., Le Pourhiet, L., 2018. Role of tectonic burial and
 555 temperature on the inversion of inherited extensional basins during collision. *Geol. Mag.*
 556 153, 811–826. <https://doi.org/10.1017/S0016756816000510>

557 Lagabriele, Y., Bodinier, J.L., 2008. Submarine reworking of exhumed subcontinental mantle
 558 rocks: Field evidence from the Lherz peridotites, French Pyrenees. *Terra Nov.* 20, 11–21.
 559 <https://doi.org/10.1111/j.1365-3121.2007.00781.x>

560 Lagabriele, Y., Labaume, P., De Saint Blanquat, M., 2010. Mantle exhumation, crustal
 561 denudation, and gravity tectonics during Cretaceous rifting in the Pyrenean realm (SW
 562 Europe): Insights from the geological setting of the lherzolite bodies. *Tectonics* 29, 1–26.
 563 <https://doi.org/10.1029/2009TC002588>

564 Lagabriele, Y., Lemoine, M., 1997. Alpine, Corsican and Apennine ophiolites: the slow-
 565 spreading ridge model. *Comptes Rendus l'Academie Sci. - Sci. la Terre des Planetes*
 566 325, 909–920.

567 Lavier, L.L., Manatschal, G., 2006. A mechanism to thin the continental lithosphere at
 568 magma-poor margins. *Nature* 440, 324–328. <https://doi.org/10.1038/nature04608>

569 Le Pourhiet, L., Chamot-Rooke, N., Delescluse, M., May, D.A., Watremez, L., Pubellier, M.,
 570 2018. Continental break-up of the South China Sea stalled by far-field compression. *Nat.*
 571 *Geosci.* <https://doi.org/10.1038/s41561-018-0178-5>

572 Manatschal, G., 2014. New models for evolution of magma-poor rifted margins based on a
 573 review of data and concepts from West Iberia and the Alps. *Int. J. Earth Sci.* 93, 432–

574 466. <https://doi.org/10.1007/s00531-004-0394-7>

575 Manatschal, G., Lavier, L., Chenin, P., 2015. The role of inheritance in forming and
576 deforming hyper-extended rift systems and creating collisional orogens. *Gondwana Res.*
577 27, 140–164. <https://doi.org/10.1127/1864-5658/2015-69>

578 Masini, E., Manatschal, G., Tugend, J., Mohn, G., Flament, J.M., 2014. The tectono-
579 sedimentary evolution of a hyper-extended rift basin: The example of the Arzacq-
580 Mauléon rift system (Western Pyrenees, SW France). *Int. J. Earth Sci.* 103, 1569–1596.
581 <https://doi.org/10.1007/s00531-014-1023-8>

582 May, D.A., Brown, J., Le Pourhiet, L., 2015. A scalable , matrix-free multigrid preconditioner
583 for finite element discretizations of heterogeneous Stokes flow. *Comput. Methods Appl.*
584 *Mech. Eng.* 290, 496–523. <https://doi.org/10.1016/j.cma.2015.03.014>

585 May, D.A., Brown, J., Le Pourhiet, L., 2014. pTatin3D : High-Performance Methods for
586 Long-Term Lithospheric Dynamics. *Proceeding SC'14 Proc. Int. Conf. High Perform.*
587 *Comput. Networking, Storage Anal.* 274–284.

588 Mesalles, L., Mouthereau, F., Bernet, M., Chang, C., Lin, A.T., Fillon, C., Sengelen, X.,
589 2014. From submarine continental accretion to arc-continent orogenic evolution : The
590 thermal record in southern Taiwan. *Geology* 42, 907–910.
591 <https://doi.org/10.1130/G35854.1>

592 Molnar, P., Tapponnier, P., 1981. A possible dependence of tectonic strength on the age of the
593 crust in Asia. *Earth Planet. Sci. Lett.* 52, 107–114. [https://doi.org/10.1016/0012-](https://doi.org/10.1016/0012-821X(81)90213-2)
594 [821X\(81\)90213-2](https://doi.org/10.1016/0012-821X(81)90213-2)

595 Molnar, P., Tapponnier, P., 1975. Cenozoic Tectonics of Asia : Effects of a continental
596 collision. *Science* (80-.). 189, 419–426. <https://doi.org/10.1126/science.189.4201.419>

597 Montigny, R., Azambre, B., Rossy, M., Thuizat, R., 1986. K-Ar Study of cretaceous
 598 magmatism and metamorphism in the pyrenees: Age and length of rotation of the
 599 Iberian Peninsula. *Tectonophysics* 129, 257–273. [https://doi.org/10.1016/0040-](https://doi.org/10.1016/0040-1951(86)90255-6)
 600 1951(86)90255-6

601 Mouthereau, F., Filleaudeau, P.Y., Vacherat, A., Pik, R., Lacombe, O., Fellin, M.G.,
 602 Castelltort, S., Christophoul, F., Masini, E., 2014. Placing limits to shortening evolution
 603 in the Pyrenees: Role of margin architecture and implications for the Iberia/Europe
 604 convergence. *Tectonics* 33, 2283–2314. <https://doi.org/10.1002/2014TC003663>

605 Mouthereau, F., Lacombe, O., Vergés, J., 2012. Building the Zagros collisional orogen :
 606 Timing , strain distribution and the dynamics of Arabia / Eurasia plate convergence.
 607 *Tectonophysics* 532–535, 27–60. <https://doi.org/10.1016/j.tecto.2012.01.022>

608 Mouthereau, F., Watts, A.B., Burov, E., 2013. Structure of orogenic belts controlled by
 609 lithosphere age. *Nat. Geosci.* 6, 1–5. <https://doi.org/10.1038/ngeo1902>

610 Müntener, O., Hermann, J., Trommsdorff, V., 2000. Cooling History and Exhumation of
 611 Lower- Crustal Granulite and Upper Mantle (Malenco , Eastern Central Alps). *J. Petrol.*
 612 41, 175–200.

613 Müntener, O., Manatschal, G., Desmurs, L., Pettke, T., 2009. Plagioclase peridotites in ocean-
 614 continent transitions: Refertilized mantle domains generated by melt stagnation in the
 615 shallow mantle lithosphere. *J. Petrol.* 51, 255–294.
 616 <https://doi.org/10.1093/petrology/egp087>

617 Naliboff, J.B., Buiter, S.J.H., Péron-Pinvidic, G., Osmundsen, P.T., Tetreault, J., 2017.
 618 Complex fault interaction controls continental rifting. *Nat. Commun.* 1–9.
 619 <https://doi.org/10.1038/s41467-017-00904-x>

620 Précigout, J., Gueydan, F., Garrido, C.J., Cogné, N., Booth-Rea, G., 2013. Deformation and
621 exhumation of the Ronda peridotite (Spain). *Tectonics* 32, 1011–1025.
622 <https://doi.org/10.1002/tect.20062>

623 Ranalli, G., Murphy, D.C., 1987. Rheological stratification of the lithosphere. *Tectonophysics*
624 132, 281–295.

625 Reston, T.J., Leythaeuser, T., Booth-Rea, G., Sawyer, D., Klaeschen, D., Long, C., 2007.
626 Movement along a low-angle normal fault: The S reflector west of Spain. *Geochemistry,*
627 *Geophys. Geosystems* 8, 1–14. <https://doi.org/10.1029/2006GC001437>

628 Reston, T.J., Mcdermott, K.G., 2011. Successive detachment faults and mantle unroofing at
629 magma-poor rifted margins. *Geology* 39, 1071–1074. <https://doi.org/10.1130/G32428.1>

630 Rybacki, E., Dresen, G., 2000. Dislocation and diffusion creep of synthetic anorthite
631 aggregates. *J. Geophys. Res. Solid Earth* 105, 26017–26036.
632 <https://doi.org/10.1029/2000JB900223>

633 Sauter, D., Cannat, M., Rouméjon, S., Andreani, M., Birot, D., Bronner, A., Brunelli, D.,
634 Carlut, J., Delacour, A., Guyader, V., Macleod, C.J., Manatschal, G., Mendel, V.,
635 Ménez, B., Pasini, V., Ruellan, E., Searle, R., 2013. Continuous exhumation of mantle-
636 derived rocks at the Southwest Indian Ridge for 11 million years. *Nat. Geosci.* 6, 314–
637 320. <https://doi.org/10.1038/ngeo1771>

638 Sauter, D., Rommevaux-Jestin, C., Parson, L.M., Fujimoto, H., Mével, C., Cannat, M.,
639 Tamaki, K., 2004. Focused magmatism versus amagmatic spreading along the ultra-slow
640 spreading Southwest Indian Ridge: Evidence from TOBI side scan sonar imagery.
641 *Geochemistry, Geophys. Geosystems* 5, 1–20. <https://doi.org/10.1029/2004GC000738>

642 Teixell, A., Labaume, P., Lagabrielle, Y., 2016. The crustal evolution of the west-central

643 Pyrenees revisited: Inferences from a new kinematic scenario. *Comptes Rendus - Geosci.*
644 348, 257–267. <https://doi.org/10.1016/j.crte.2015.10.010>

645 Tetreault, J.L., Buiter, S.J.H., 2017. The influence of extension rate and crustal rheology on
646 the evolution of passive margins from rifting to break-up. *Tectonophysics* 0–1.
647 <https://doi.org/10.1016/j.tecto.2017.08.029>

648 Tugend, J., Manatschal, G., Kuszniir, N.J., Masini, E., Mohn, G., Thinon, I., 2014. Formation
649 and deformation of hyperextended rift systems: Insights from rift domain mapping in the
650 Bay of Biscay-Pyrenees. *Tectonics* 1239–1276.
651 <https://doi.org/10.1002/2014TC003529>.Received

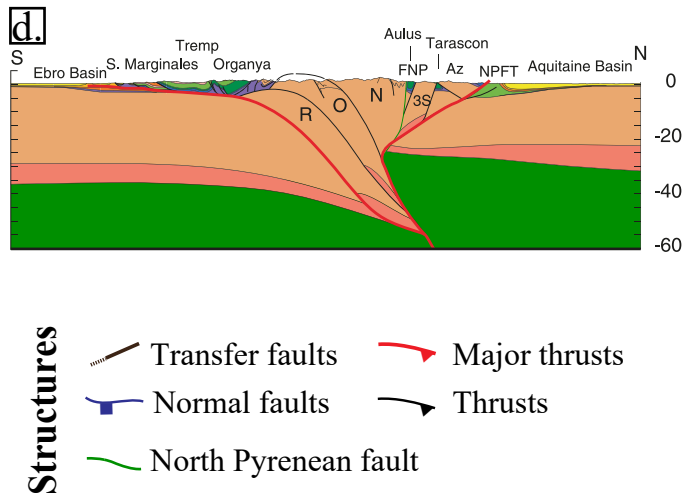
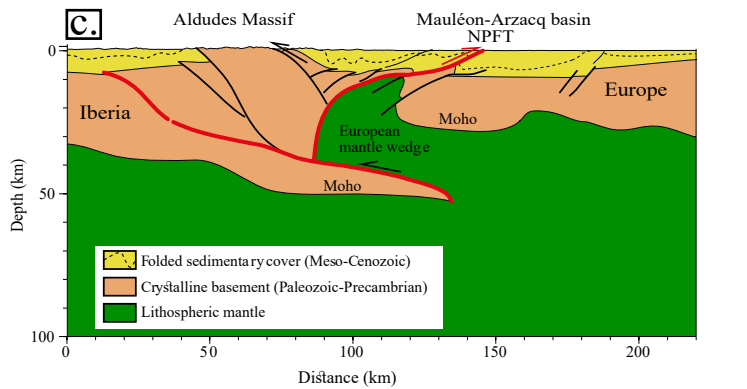
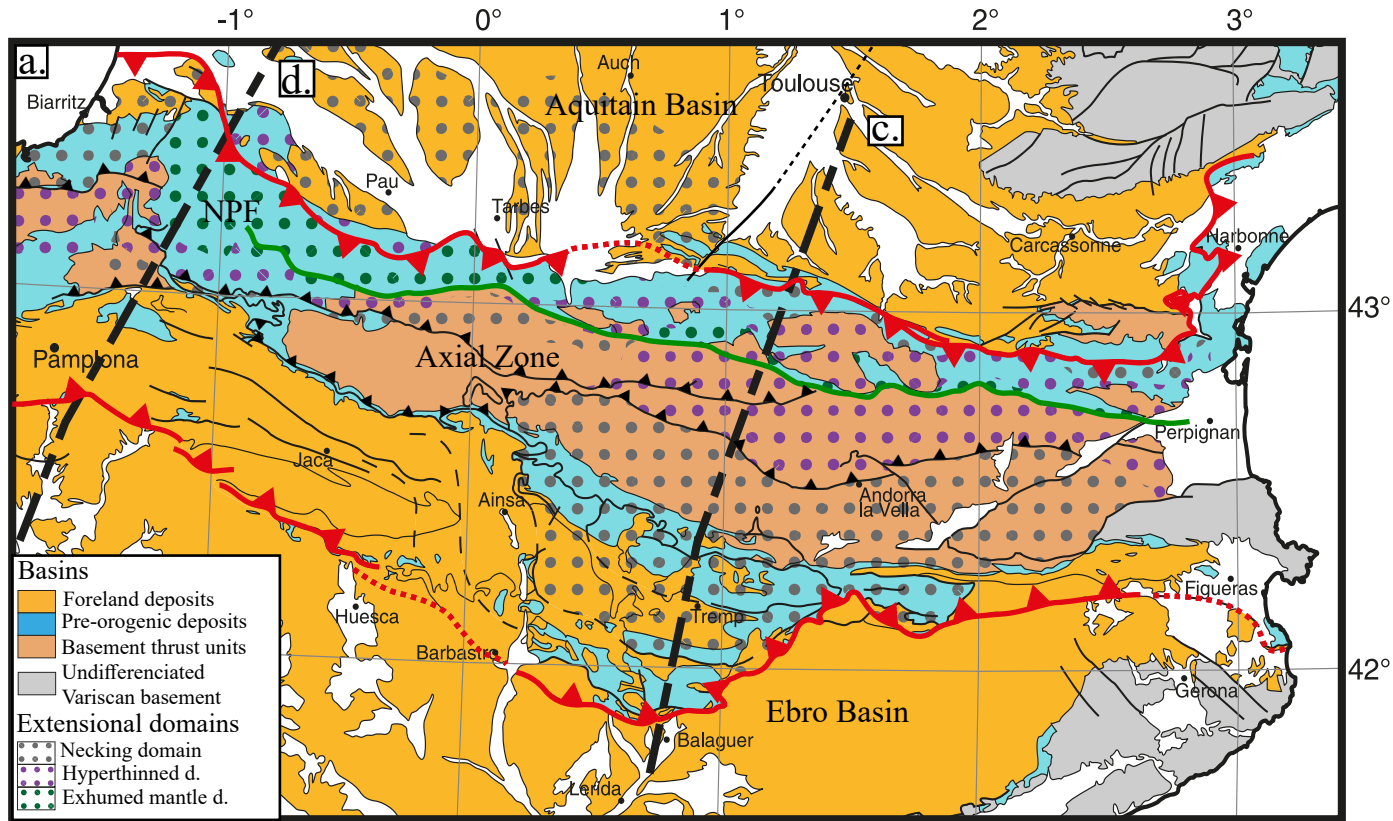
652 Vacherat, A., Mouthereau, F., Pik, R., Bellahsen, N., Gautheron, C., Bernet, M., Daudet, M.,
653 Balansa, J., Tibari, B., Jamme, R.P., Radal, J., 2016. Rift-to-collision transition recorded
654 by tectonothermal evolution of the northern Pyrenees. *Tectonics* 35, 907–933.
655 <https://doi.org/10.1002/2015TC004016>

656 Vacherat, A., Mouthereau, F., Pik, R., Bernet, M., Gautheron, C., Masini, E., Pourhiet, L. Le,
657 Tibari, B., Lahfid, A., 2014. Thermal imprint of rift-related processes in orogens as
658 recorded in the Pyrenees. *Earth Planet. Sci. Lett.* 408, 296–306.
659 <https://doi.org/10.1016/j.epsl.2014.10.014>

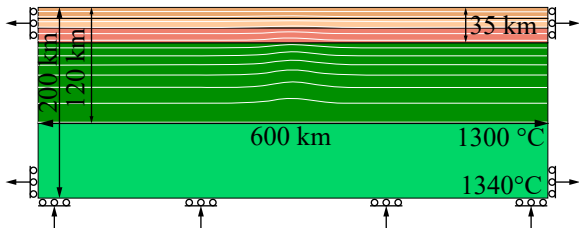
660 Velasque, P.C., Ducasse, L., Muller, J., Scholten, R., 1989. The influence of inherited
661 extensional structures on the tectonic evolution of an intracratonic chain : the example of
662 the Western Pyrenees. *Tectonophysics* 162, 243–264.

663 Wang, Y., Chevrot, S., Monteiller, V., Komatitsch, D., Mouthereau, F., Manatschal, G., 2016.
664 The deep roots of the western Pyrenees revealed by full waveform inversion of
665 teleseismic P waves. *Geology* 44, 475–478. <https://doi.org/10.1130/G37812.1>

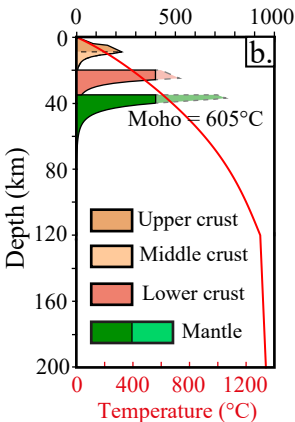
- Warren, C.J., Beaumont, C., Jamieson, R.A., 2008. Modelling tectonic styles and ultra-high pressure (UHP) rock exhumation during the transition from oceanic subduction to continental collision. *Earth Planet. Sci. Lett.* 267, 129–145.
<https://doi.org/10.1016/j.epsl.2007.11.025>
- Watremez, L., Burov, E., D'Acremont, E., Leroy, S., Huet, B., Le Pourhiet, L., Bellahsen, N., 2013. Buoyancy and localizing properties of continental mantle lithosphere: Insights from thermomechanical models of the eastern Gulf of Aden. *Geochemistry, Geophys. Geosystems* 14, 2800–2817. <https://doi.org/10.1002/ggge.20179>
- Wickham, S.M., 1987. Crustal anatexis and granite petrogenesis during low-pressure regional metamorphism: The trois seigneurs massif, pyrenees, France. *J. Petrol.* 28, 127–169.
<https://doi.org/10.1093/petrology/28.1.127>
- Yamato, P., Burov, E., Agard, P., Le Pourhiet, L., Jolivet, L., 2008. HP-UHP exhumation during slow continental subduction: Self-consistent thermodynamically and thermomechanically coupled model with application to the Western Alps. *Earth Planet. Sci. Lett.* 271, 63–74. <https://doi.org/10.1016/j.epsl.2008.03.049>



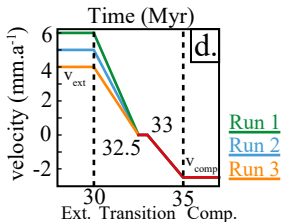
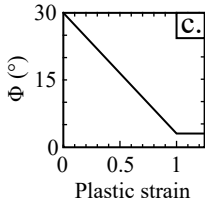
a. Free Surface

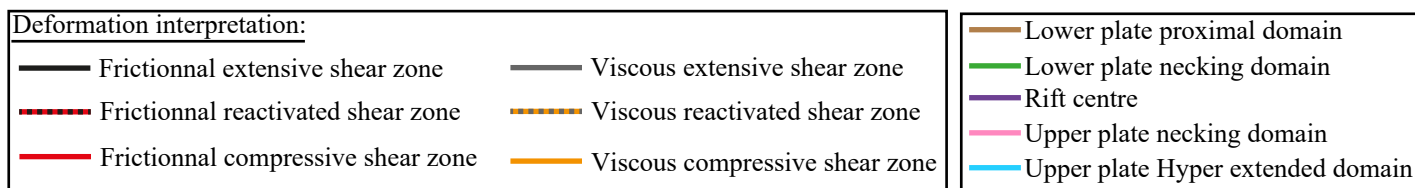
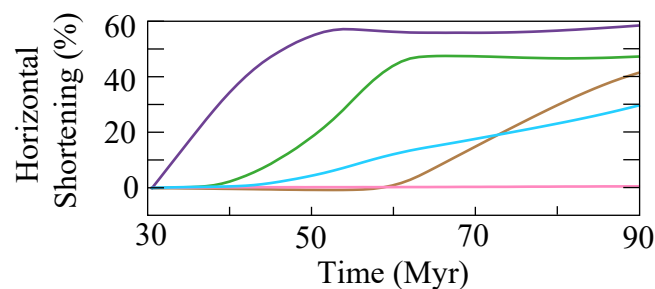
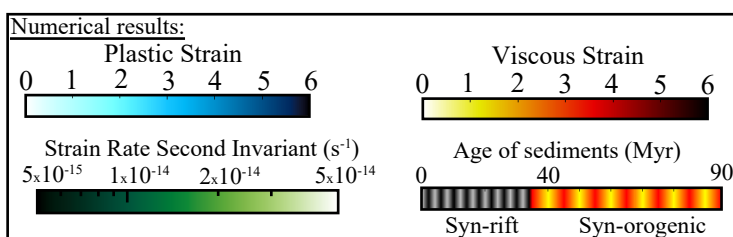
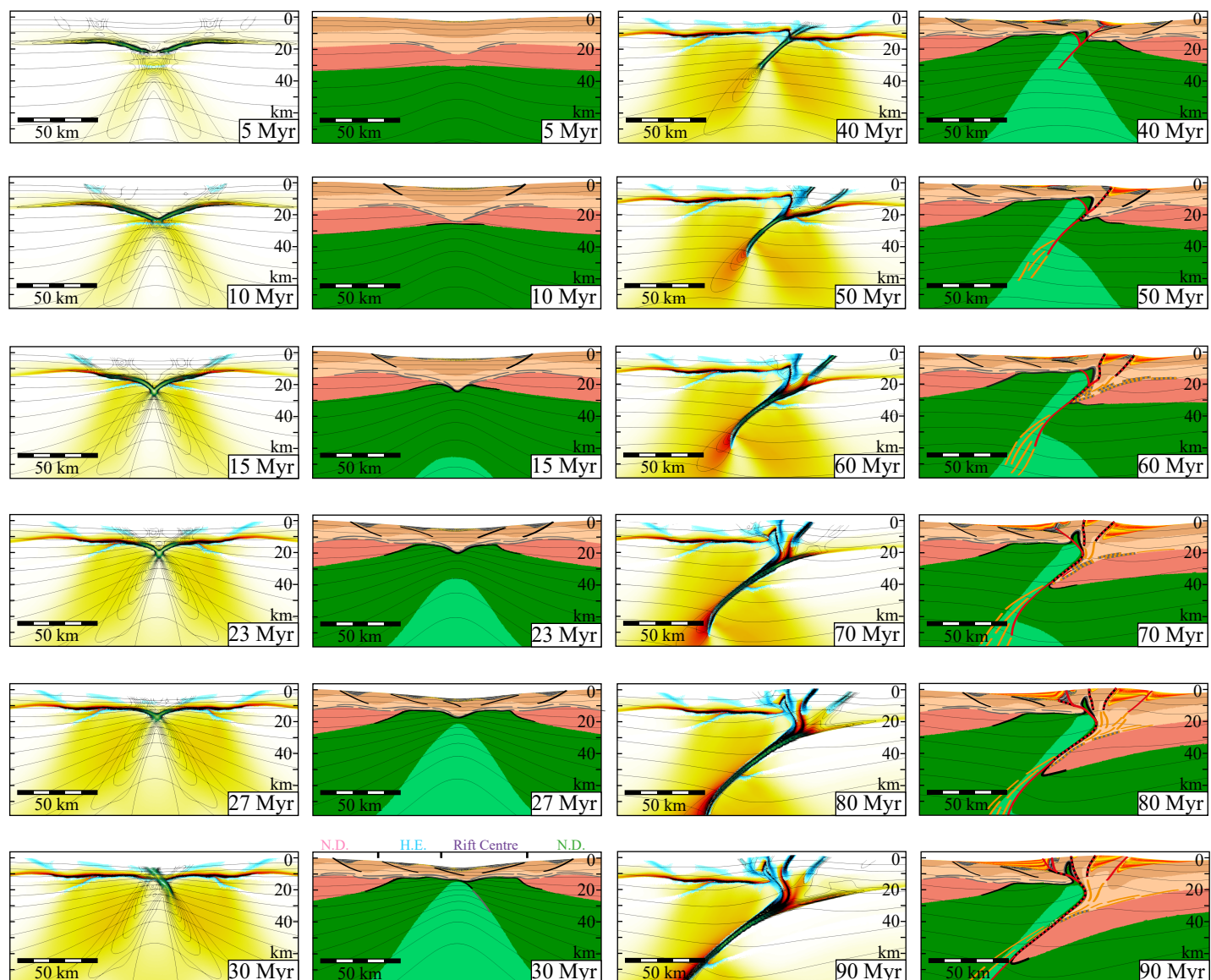


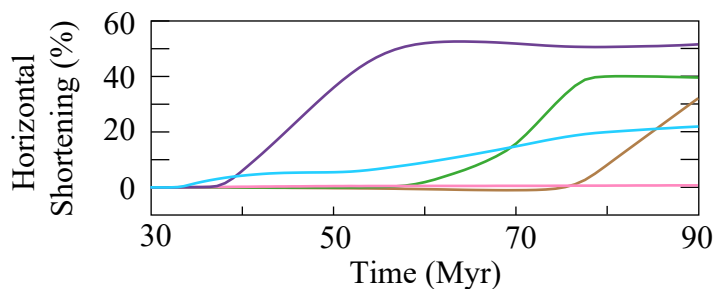
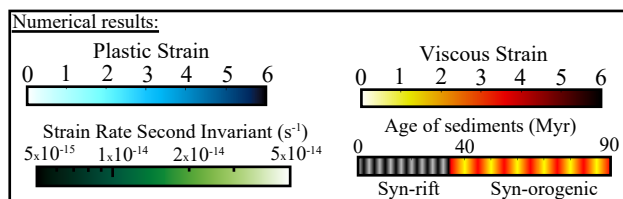
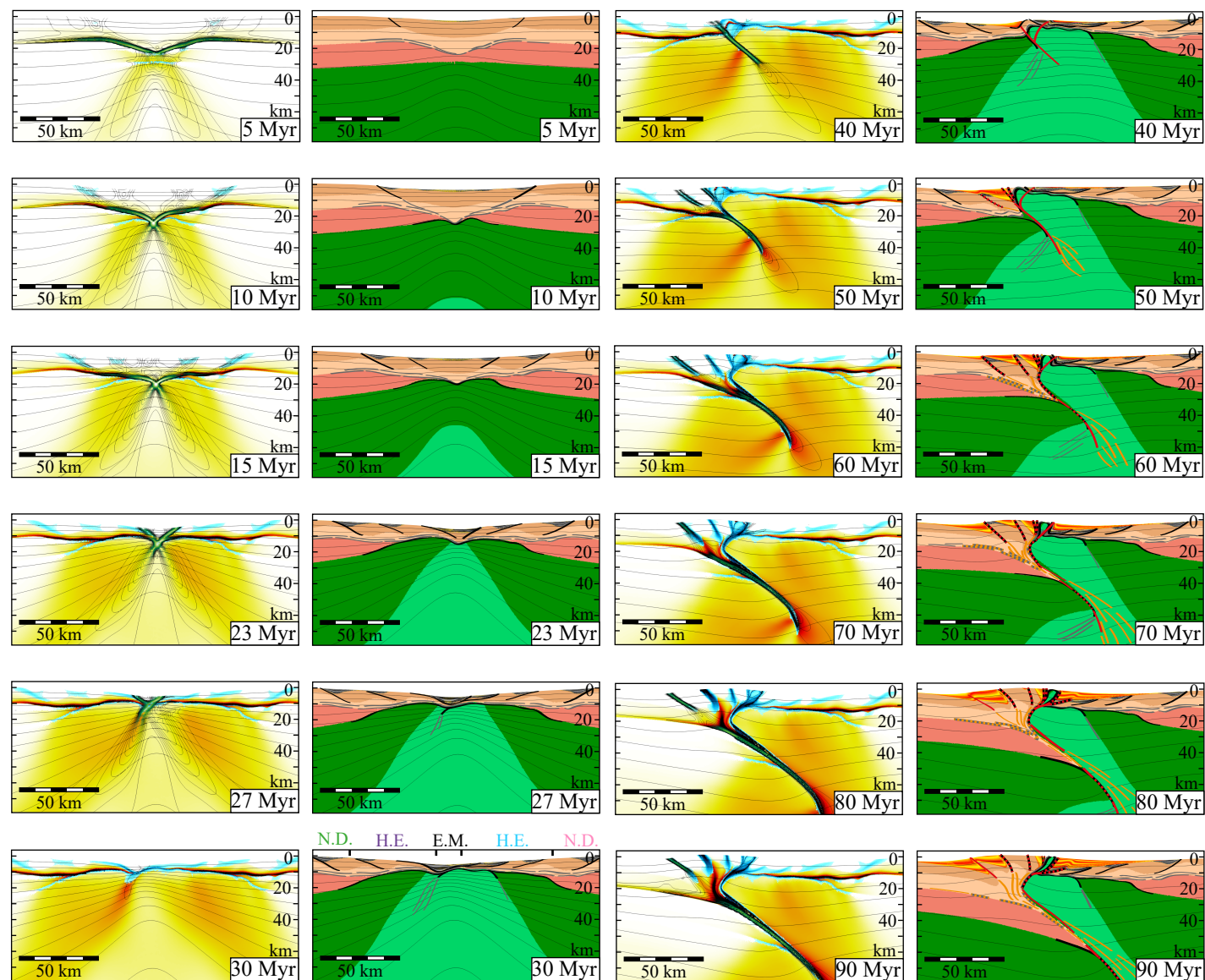
Differential stress (MPa)



Plastic strain softening



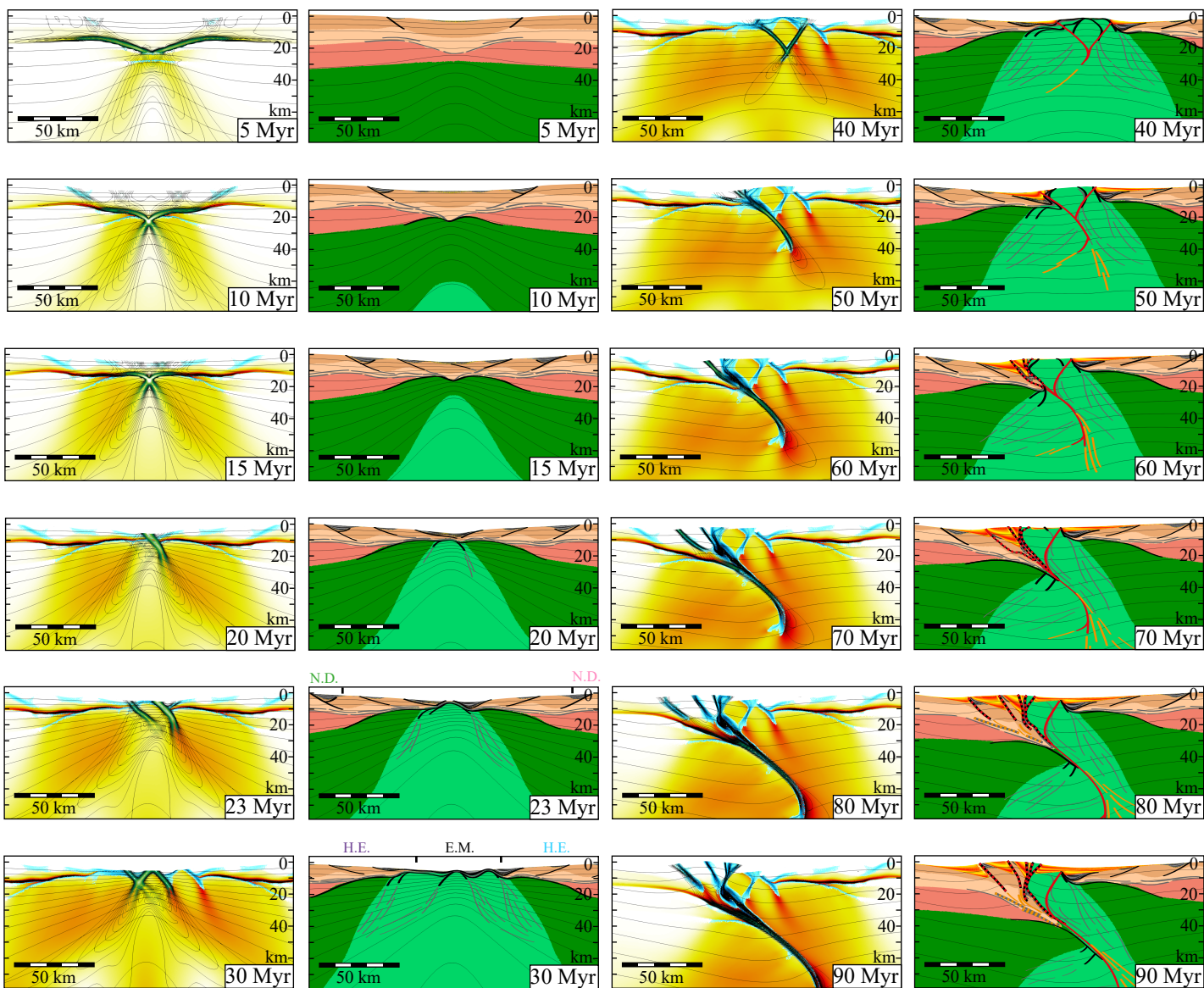




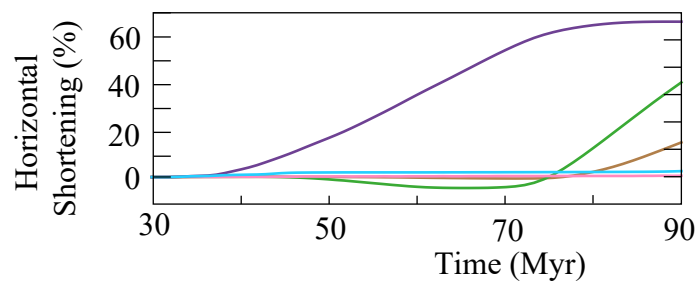
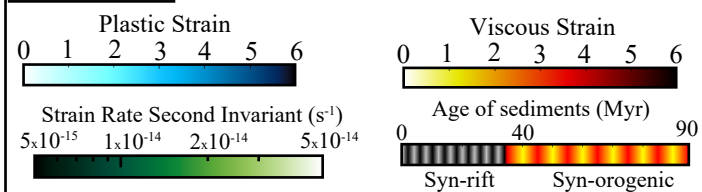
Deformation interpretation:

- | | |
|---|--------------------------------------|
| — Frictional extensive shear zone | — Viscous extensive shear zone |
| - - - Frictional reactivated shear zone | - - - Viscous reactivated shear zone |
| — Frictional compressive shear zone | — Viscous compressive shear zone |

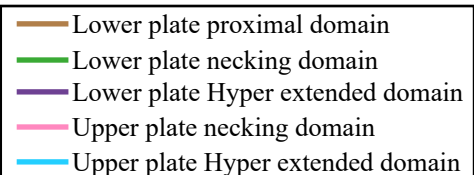
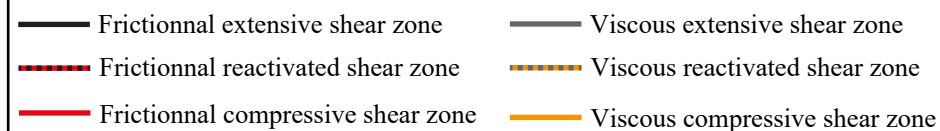
- | |
|-------------------------------------|
| — Lower plate proximal domain |
| — Lower plate necking domain |
| — Lower plate Hyper extended domain |
| — Upper plate necking domain |
| — Upper plate Hyper extended domain |



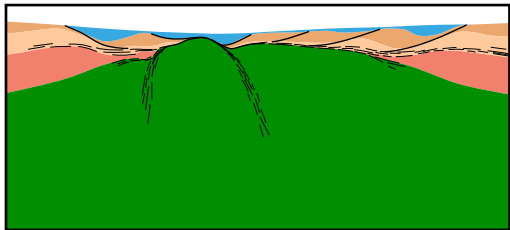
Numerical results:



Deformation interpretation:



Narrow exhumed domain

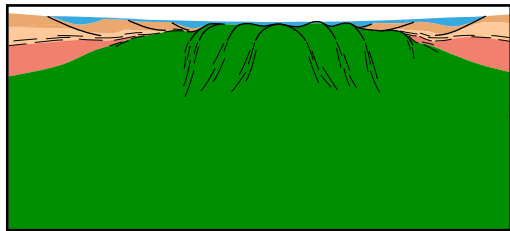


Hot geotherm
during inversion

Continental collision



Wide exhumed domain



Hot geotherm
during inversion

Obduction

