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Temporal and spatial evolution of orogens: a guide for geological mapping

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Abstract

Orogens develop in convergent settings involving two or more continental and/or oceanic plates. They are traditionally defined as zones of crustal deformation associated with mountain building resulting from either the accretion of an arc, a continent-continent collision or the inversion of a rift basin. However, this somewhat narrow definition does not consider for example the genetic link between an oceanic domain and an intra-continental rift, even though extension associated with a scissor-shape opening can be demonstrated in many ocean-floored basins. Consequently, we propose a new concept of orogenic evolution based on the development of extensional margins subsequently subjected to crustal shortening. Thus orogens that develop as a result of the closure of wide basins, and involve the shortening of supra-subduction basins and arcs, are distinguished from mountain belts developed above subduction zones or that result from continental collision. Inverted intra-continental rifts (devoid of oceanic material) are also considered. Our review of a number of key orogens identifies apparent similarities and differences in geodynamic processes through geological time and emphasizes the lateral spatial and temporal changes that may occur from a subduction to a collisional setting, with such changes being intrinsic to the transition from intra-continental rifts to large basins floored by oceanic crust. We also review the geodynamic processes that operated prior to the onset of plate tectonics. As we go back in time, key tectonic attributes are apparently lost in many cases (but tellingly not all) due to a lack of data, a lack of understanding, or the deeper levels of erosion generally associated with older orogens.

We propose that mapping utilizing tools based on comparative tectonics to understand old terrains and fault zones is a good way to constrain such an evolution and that this can start with a global-scale map of past-to-modern orogens, aimed at re-exploring mountain building concepts spatially and temporarily. This is the primary objective of IGCP 667 project “World Map of Orogens”.

1. Introduction

Orogens form along (or in proximity to) lithospheric plate boundaries in a convergent setting (i.e. subduction, collision) and often involve two or more neighboring continental and/or oceanic plates. On the other hand, intracontinental orogens develop in an intraplate setting. The evolution of an orogen is a complex geodynamic process punctuated by several tectonic events that can be viewed as the “orogenic cycle”, and in which an “orogeny” *sensu stricto* is a fundamental step (**Fig.1**; *Wilson et al., 2019*). Orogenic belts may be defined as a combination of folding, faulting, regional metamorphism and igneous activity within contemporaneously uplifted terranes (*Kearey, 1993*). These orogenic components are found in belts resulting from either continent-continent collision or continental-island arc accretion leading to crustal thickening. These two tectonic settings control the final orogenic architecture through geological processes both at depth and at the surface, including for example: the inversion of a passive margin, the emplacement of tectonic nappes, strike-slip faulting, the accretion of volcanic arcs, granitic magmatism, suture zone deformation and regional metamorphism (*van der Pluijm and Marshak, 2004*). Nevertheless, it remains key to differentiate accretionary orogens (*Sengör and Okuurogullari, 1991*), where there is sequential amalgamation of terranes to a continental margin (e.g. the Cenozoic North American Cordillera or the Paleozoic Altaids Belt in Central Asia), from collisional orogens (*Windley, 1999*) where subduction leads to ocean-continent convergence (e.g. Andes of South America) and then continent-continent collision (e.g. Himalaya).

The definition of an orogen therefore becomes conceptually important and this obliges us to reconsider the identification of the key diagnostic attributes of an orogeny. It also requires an accurate portrayal of such orogenic attributes on a geological or tectonic map including their location relative to craton boundaries and any older orogen. Such an approach emphasizes the evolutionary steps responsible for continental growth and underline any inherent reactivation

process and/or history by documenting the superimposition of orogens through time and space. The resulting parsing of an orogenic belt into key attributes enables the construction of crustal cross-sections, along which overlapping or juxtaposed events can be considered. An important facet of IGCP 667 is thus the investigation of how basins and orogens interact with one another during continental growth, and how that interaction can be considered in terms of geological mapping and map compilations.

(Figure 1 is about here)

2. Historical overview

Tectonic displacement from a geological mapping point-of-view was poorly understood until the end of the 19th Century, although vertical movement intrinsic to mountain belt formation was already being considered (*e.g.* theory of mountain systems of de Beaumont (1852); geosyncline theory of Hall and Dana (1873); eustatic variations of Suess (1893)). The first use of the terms “orographic” as applied to a mountain structure in a descriptive sense and “orogenic” in a genetic sense was made by Gressly (1840) and Thurmann (1854) in the Jura Mountains of France. A fundamental step in the establishment of tectonics as an individual science occurred with the geometric description of folds and faults responsible for stratigraphic inversions in the Appalachians of eastern North America by Rogers (1858). Lesley (cited in Chamberlin and Salisbury, 1906) further advanced our understanding of mountain belt structures by imagining the horizontal forces required to form large-scale folds, without however defining their specific nature. De Beaumont’s theory of mountain systems somehow evokes the equivalent of an orogeny in the mid-19th Century, but the collage of stratigraphic terranes was not sufficient to explain the formation of mountain belts. Crustal movement as a result of deformation was first theorized by the end of 19th Century by Gilbert (1890) with the formulation of orogenic and epeiorogenic uplift concepts, and in particular the relationship between fold-and-thrust mountain belts and lateral-compressional tectonics. However, Gilbert

96 (1890) attached no specific time connotation to an orogeny. The concept of orogeny as “the
97 process of mountain-building” was first developed by Upham (1894) who defined orogeny as
98 “denoting the process of formation of mountains ranges by folds, faults, upthrusts and
99 overthrusts, affecting comparatively narrow belts and lifting them in great ridges”. Later, Haug
100 (1907) distinguished four major orogenies: Huronian (end of Precambrian), Caledonian (Early
101 Paleozoic), Hercynian (Late Paleozoic) and Alpine (Cretaceous). Stille (1920) reconsidered
102 orogenesis as an episodic change in the rock deformation fabric, thus equating the term to both
103 an event and a tectonic process. The definition of orogeny proposed by Stille (1940) involved
104 the pervasive deformation of pre-existing rock fabrics, limited in space and time and hence
105 implying a spatiotemporal event, was widely accepted.

106 Major orogens are commonly generated by continental collisions, but these can be
107 superimposed through time and space thus potentially masking key information on the
108 fundamental nature of the shortening processes involved. The understanding and classification
109 of such processes therefore becomes crucial and includes documenting the reactivation of older
110 tectonic features during mountain building, as well as in any subsequent ocean opening. It
111 therefore becomes important to revisit the concept of orogeny and to also identify potential
112 continental growth through mantle extraction processes through time and space.

113 The present overview of the world’s orogens is part of the ongoing IGCP 667 Project “World
114 Map of Orogens”, in which the main driver is to design a map at scale 1:35 000 000
115 documenting the distribution of ancient to modern orogens throughout Earth history. The
116 compilation of a “World Map of Orogens” necessarily requires a careful consideration of both
117 the various types of global-scale compressional tectonic regimes and the best means of
118 portraying the associated tectonic elements or attributes on such a map. Consideration of pre-
119 plate tectonic, Early Earth crustal deformation processes and features also becomes necessary.
120 As we go back in time, we propose a slight simplification of the main tectonic attributes

captured due to an increasing paucity of data. We conclude by presenting a number of orogens, from the Paleoproterozoic to the present, with a legend that includes the key elements of a “typical” orogen.

3. A four-fold geodynamic classification of orogens

Recent evolutionary models of continental margins and basins (*d’Acremont et al., 2006; Fournier et al., 2010; Franke et al., 2014; Pubellier et al., 2016; Bulois et al., 2018*), as well as comparisons between modern and ancient orogenic systems (e.g. *Weller and St-Onge, 2017; Brown and Johnson, 2018; Palin et al., 2020*), emphasize the notions of margin evolution, basin development, rift propagation and basement reactivation. In the documentation of most orogenic systems, a space-time continuum of deformation affecting the various geological elements of a mountain belt is reported. This continuity of tectonic events can be directly linked to the Wilson Cycle (**Fig.1**). Given the current understanding that geological processes are intimately linked to orogenic evolution, it seems timely to propose a different geodynamic classification of orogens in order to emphasize the overall tectonic processes occurring during lithospheric shortening.

Four main types of convergent orogenic systems can be identified within the present-day geodynamical settings encountered on Earth (**Fig.2**): (i) ocean-continent convergence accommodated by subduction, (ii) accretion/collision of crustal elements derived from subduction (island arc and/or back-arc basin), or occurring as a microcontinent, ribbon continent, oceanic plateau or ridge, (iii) continental collision, often following subduction ± accretion and (iv) shortening of an intracontinental sedimentary basin and continental crustal materials. Each of these types in turn may only represent one step in the development of an orogen in time and space, thus a transition between each stage can/should be considered. Indeed most of these orogens will result in construction of topography, particularly the continental collision, which is followed by orogenic collapse along large low angle normal faults, which

are generally located at the frontier between zones of strongly varying rheologies. This post orogenic stage aspect is not developed in the synthesis.

Type I (**Fig.2a**) involves ocean-continent convergence and leads to the formation of a mountain belt on the overriding upper plate, as a result of igneous material sourced from the underlying subduction zone. Compression may happen during oceanic subduction when there is a change in subduction attributes such as the dip or the relative velocity of the subducting slab, the obliquity of the convergence, or the amount of sedimentary material deposited in the trench, underplated or subducted with the downgoing plate. This results in a non-collisional system typical of Andean-type cordilleras (South America, Eastern Australia, and Antarctica). The Type I orogen is often considered as a precursor to continental collision (cf Type III), so that the resulting mountain range built on the upper plate can be referred to as the “early-orogenic zone”.

Type II (**Fig.2b**) involves micro- or ribbon-continentals and slivers of the continental margin rifted during the opening of back-arc basins. The slivers may also include oceanic plateaus (e.g. Ontong-Java in the Solomon Islands), exotic island arc terranes (e.g. North American Cordillera) or continental accreted terranes (e.g. Cascadia). The convergent process is often known as terrane docking or accretion and it results in a tectonic collage of pericratonic and suspect terranes (*Coney et al., 1980; Howell, 1989*). The accretion of pericratonic, suspect, or exotic terranes represents a milestone in the history of an oceanic basin. The orogen forms when a crustal domain is rifted away from a continental margin and eventually reaches an active continental margin. As subduction of the now-accreted crustal domain is not possible, subduction jumps to the opposite side of the docked terrane (e.g. Anatolia, Iran; *Barrier and Vrielynck, 2008*, or SE Asia; *Pubellier and Meresse, 2013*). Where observations are more complete (e.g. Cenozoic convergent settings), the docked crustal domains are often back-arc

basins floored by continental basement, or island arcs. This stage may therefore be considered as a prelude to the eventual formation of a continental collisional belt (Type III).

Type III (**Fig.2c**) is characterized by the collision of two or more continental landmasses. Continental collision, sometimes referred to as “terminal collision” in the literature (e.g. *Najman et al., 2017; Abu Sharib et al., 2019*), is characterized by a pre-collisional history that can involve a lengthy period of subduction and terrane accretion (cf Types I and II). Because of the crustal thickness of the subducting lithosphere and its positive buoyancy, most of the incoming material cannot be subducted, or remain subducted, leading to crustal shortening, accommodated by low-angle syn-metamorphic ductile shearing and orogeny. Therefore, a continent-continent collision follows the complete closure of an oceanic basin (e.g. Himalaya, European Alps) and the closure of the arc and/or back-arc system in the upper plate.

Type IV (**Fig.2d**) is an intra-continental orogen formed during the inversion of a rift system that originated from pre-orogenic rifting that never led to the formation of oceanic crust. The Atlas Mountains in NW Africa are a classic example of this type of orogen, in which only the upper structural levels and the metamorphic rocks they contain are exposed (e.g. *Mattauer et al., 1977; Beauchamp et al., 1999; Frizon de Lamotte et al., 2000; Missenard et al., 2006*). Another example from the South China Block comprises a Silurian intracontinental orogen, in which deep seated metamorphic zones are exposed. The Early Paleozoic Orogen (EPO) of South China, has long been recognized due to a Devonian unconformity (*Grabau, 1924*). This pre-Devonian belt resulted from the closure of a Neoproterozoic to Ordovician rift within continental crust and without involving oceanic crust (e.g. *Yao et al., 2014*). The rift closure was accommodated by a north-directed continental subduction as shown by south-verging folds and thrusts, and ductile kinematics in the southern part of the belt (*Faure et al., 2008; Charvet et al. 2010; Li et al., 2010; Wang et al., 2013; Xin et al. 2020 and enclosed references*). The inner zone of this orogen consists of mid temperature and mid pressure biotite-garnet-staurolite-

kyanite metapelites and garnet amphibolites that document P-T conditions at 1.0- 1.2 GPa and 500-600°C (*Zhao and Cawood 1999*). The driving force for the intracontinental orogen is interpreted as being the far-field effect of the Ordovician collisional orogeny recognized in central Vietnam, responsible for the welding of the Viet-Cambodia block and the Viet-Lao (or North Vietnam-South China) blocks (*Faure et al., 2018*).

Intracontinental orogens may also result from highly oblique continental collisions (e.g. Southern Alps of New Zealand; *Beaumont et al., 1996*). Transpressionnal components can also play important roles in the formation of intracontinental orogens. In spite of a unique initial pre-orogenic setting, the intracontinental orogens exhibit structural, metamorphic, plutonic and sedimentary features similar to those of type III orogens (cf. Table 1).

As a broad generalization, a Type I orogen can often represent an early stage of a Type II or Type III orogen. In this overview, we have defined an “early orogenic zone” (**Table 1**), which can be compared and contrasted to the three other (generally subsequent) orogenic stages and zones: peri-orogenic/orogenic zones (types II and III) and para-orogenic zones (Type IV).

The “orogenic zones” in types II and III are associated with the presence of ophiolites, (U)HP and (U)HT metamorphic rocks and plutonism (granites). The “peri-orogenic zones” correspond to areas of more moderate deformation, lower grade metamorphism, moderate topography, the presence of molassic foreland basins as well as fold-and-thrust belts developed above shallow detachment levels. The “para-orogenic zone” (**Table I**), is associated with Type IV orogens and is characterized by moderate deformation along inversed rifted features, occasionally low-grade metamorphism and the exhumation of mantellic material without the presence of oceanic crust as a result of the shortening of former propagator tips.

(Figure 2 is about here)

(Table 1 is about here)

4. The geometry of basins and orogenic implications

During crustal shortening, a previously extended sedimentary basin is often subjected to compression during the process of tectonic inversion. The precise end of the rifting (breakup) is contemporaneous with the first sedimentary units deposited on oceanic crust and can sometimes be confirmed with the presence of clear oceanic magnetic anomalies. The opening is rarely orthogonal, implying an along-strike migration of breakup with time (**Fig.3**). The newly formed oceanic crust is therefore wider at one end and narrower on the opposite end of the newly formed basin. It might even be connected to a continental rift inland, which is devoid of oceanic crust. The resulting V-shape geometry yields strong implications on the style of any subsequent shortening and timing of eventual basin closure.

4.1. Non-orthogonal opening of basins, and propagation of rift systems

Rift basin opening is rarely orthogonal. Most of the well-documented marginal basins have a scissor-shape geometry resulting from the propagation of a rift toward locked zones (Courtilot, 1982). Notable examples include the Woodlark Basin (Weissel and Watts, 1979; Taylor et al., 1999;), the Gulf of Aden (Courtilot, 1982; d'Acremont et al., 2006; Fournier et al., 2010) and the South China Sea (**Fig.3**; Briaud et al., 1993, Franke et al., 2014; Pubellier et al., 2016). In continental rift systems, such as the North Atlantic, a similar along-axis propagation of the deformation is also documented based on paleogeographic reconstructions (e.g. Ziegler, 1990) but the absence of clear oceanic crust in some cases (e.g. Gakkel ridge) remains problematic to fully understanding how deformation propagates spatially and temporarily. In general, the propagation of rifts is due to either a lateral variation in the timing of the initial breakup due to local weaknesses of the continental crust, or to the proximity of the Euler rotation pole with the tip of V-shape rifted basin. The breakup point migrates with continued extension as long as the basin is opening (e.g. Martin 1984; Taylor et al., 1999; Bellahsen et al. 2003; ArRajehi et al. 2010; Péron-Pinvidic and Manatschal 2010; Fletcher et

al., 2013), so that the breakup unconformity is diachronous along-strike, with break-up ages decreasing in the direction of propagation (Ding and Li, 2016, Franke et al., 2014; Gaina et al., 1999; Bulois et al., 2017). The integration of field and seismic studies documents that rift propagation slows down or stops when the rift tip abuts transversal structures of orogenic origin (e.g. Bulois et al., 2018). Numerical models have shown that tectonic loading in the propagation direction is crucial for slowing down or halting continental rifting and allowing the development of a wide triangular zone of stretching and thinning of the continental crust (Le Pourhiet et al., 2018).

One of the best-documented rift propagator is the South China Sea Basin (Taylor and Hayes, 1983), where all these characteristics are documented. The rift migration (Franke et al, 2014) produced a large ocean-floored basin in the NE (Eastern Sub-basin) and a narrow ocean floor in the central part (SW sub-basin) with the tip corresponding to a continental rift south of Vietnam - the Nam Con Son Basin (Fig. 3a; Li et al., 2014, Sibuet et al., 2016). The subdivisions of the basin reflect interaction with the surrounding continental and oceanic crust, for example in the form of depocenter locations. The opening of the South China Sea is related to neighbouring basins closure (Pubellier et al. 2013), implying a constant relocation of tectonic stresses within this portion of the continental crust. As such, it is important to consider that a scissor-shape opening is conditioned by the dynamics of the neighbouring subduction. In larger oceanic basins bounded by well-defined continental margins (e.g. Tasman Sea, Atlantic Ocean; e.g. Guiraud and Maurin, 1992; Seton et al., 2012a; b), similar opening mechanisms might be considered, recognizing that these may be masked by subsequent changes in basin boundary conditions.

(Figure 3 is about here)

4.2. Implications on the geodynamic classification of orogens

In oceanic rift systems, the oceanic crust is predominantly associated with exhumed mantle outboard of the continent-ocean transition (COT), and the divergent system transitions laterally to a younger continental extensional domain toward the propagator tip (Fig. 3a, *e.g.* Benes *et al.*, 1994; Taylor *et al.*, 1995; 1999; 2009; Goodliffe *et al.*, 1999; Martinez *et al.*, 2001; Clift *et al.*, 2002; Minshul, 2009; Franke *et al.*, 2014; Li *et al.*, 2014). This type of geological setting promotes a scissor-like opening of the extensional basin, with an oceanic basin forming away from the propagator tip and a continental rift or aulacogen forming at the propagator tip. This transition occurs in several steps controlled by transfer faults near the COT and larger transform faults in the oceanic domain. This is particularly well documented in recent rift propagators such as the South China Sea, the Woodlark Basin, the Gulf of Aden and the Gulf of Mexico. It is also envisaged for the Proto-Tethys Ocean in the Early Paleozoic (Scotese, 2016), as well as for the Paleo-Tethys in the Late Paleozoic (Stampfli and Borel, 2001). In the case of the Neo-Tethys Ocean, it appears that the continental blocks that rifted away from the main continent were larger in the East where the ocean was wider (Borel and Stampfli, 2002, Scotese, 1991, Pourteau *et al.*, 2009), resulting in wider Phanerozoic Tethysides (Sengör *et al.*, 1996).

The above extensional mechanisms imply along-strike time-variation of opening, which is reflected in the distance between the opposite conjugated margins. If this V-Shape geometry is spectacular in the South China Sea or the Woodlark Basin, it is also the case for the large but simple example of the Atlantic Ocean, which opened in a northward direction from the Early Cretaceous to the Eocene where it ends as a rift at the tip along the Gakkel Ridge.

In any theoretical subsequent plate convergence, the required time for the basin margins to close will produce a reverse migration of the locus of closure along strike, with early collision in the rift system to later subduction-accretion followed by continental collision toward the open end of the V (**Fig. 3b**). In other words the timing of collision should be diachronous along strike,

with resulting differences in the time of deformation of the accretionary wedges, as well as the formation of the post-orogenic molassic basins. Such a scenario also implies that the material engulfed in the collision zone evolves laterally from an accretionary wedge to a continental wedge as is documented from the Eastern Sunda wedge to the Burma-Bangladesh fold-and-thrust belt, or from the Muertos wedge to the Haïti fold-and-thrust Belt in Hispaniola. Intra-oceanic deformation prior to collision and ophiolite obduction would take place in that part of the oceanic realm that was formerly wide, while no oceanic or ultramafic material would be found in the intra-continental rift region. This type of along-strike variation is observed in the western Pyrenees where the oceanic-floored V-shaped Bay of Biscay is expressed inland as shortened packages of serpentized mantle within the orogenic belt (*Jammes et al. 2004; Lagabrielle and Bodinier, 2008; Tugend et al., 2014*).

5. Temporal evolution of orogens

5.1. Early Earth geodynamic processes (before 2.5 Ga)

Today, the global tectonic regime is dominated by plate tectonics or mobile-lid tectonics including deep and cold subduction and/or collision. However, the early Earth (> 2.5 Ga) was subject to a very different, higher mantle heat flux and a thermal regime characterized by the formation of unique lithomagmatic assemblages including the emplacement of tonalite-trondjemite-granodiorite (TTG) suites in the early continental lithosphere and the formation of granite-greenstone terrains (*e.g. Arndt and Nisbet, 1982; Abbott and Hoffman, 1984; Taylor and McLennan, 1985; Abbott et al., 1994; Jaupart et al., 2007; Coltice et al., 2009; Gapais et al., 2009; Bédard and Harris, 2013; Bédard, 2018; Tang et al., 2019; Chen et al., 2019; Roman and Arndt, 2019; Hernández-Montenegro et al., 2021*). Accordingly, the early Earth was dominated by different geodynamical processes (**Fig.4**), including flat or low angle subduction (mobile-lid tectonics; *Smithies et al., 2003*) and sagduction (stagnant-lid tectonics; *McGregor, 1951*). Sagduction is thought to have been driven by partial convective overturn due to the

318 density contrast between dense (ultra) mafic cover (greenstone belts) and the more felsic crustal
319 basement (TTG), coupled with partial melting in the lower crust (*Collins et al., 1998; Van*
320 *Kranendonk et al., 2004; Thébaud and Rey, 2013; François et al., 2014; Bédard, 2018; Roman*
321 *and Arndt, 2019*). This resulted in the typical Archean dome-and-keel structures with TTG
322 terrains forming ovoid domes surrounded by narrow and strongly pinched greenstone belts.
323 Such a configuration is not suggestive of large tangential movements of plates, and the complete
324 absence of ultra-high pressure (UHP) metamorphic assemblages also suggest a completely
325 different geodynamic regime (*Brown and Johnson, 2018*). In addition, the very local evidence
326 of Archean (flat) subduction is rare and controversial (*Volodichev et al., 2004; Moyen et al.,*
327 *2006 Perchuk and Morgunova, 2014*). It would seem that the thermal (i.e. a higher mantle heat
328 flux) and the rheological conditions prevalent during the Archean Eon probably made modern
329 steep subduction (burial of a lithospheric panel at great depth) impossible.

330 The Archean metamorphic rock record and implied pressure-temperature (PT) conditions
331 (apparent geothermal gradient) can be used to discriminate geodynamic processes in the early
332 Earth, and more particularly to infer whether or not deep and cold subduction similar to that of
333 modern plate tectonics was operational (*Brown, 2008; Brown and Johnson, 2018*). The oldest
334 evidences of high pressure (HP) rocks (> 15 kbar) are bracketed between 2.2 and 1.8 Ga (e.g.
335 *Möller et al., 1995; Baldwin et al., 2004; Collins et al., 2004; Volodichev et al., 2004; Boniface*
336 *et al., 2012; Weller and St-Onge, 2017; François et al., 2018; Loose and Schenk, 2018; Xu et*
337 *al., 2018; De Olivera Chavez, 2020*). Consequently, these authors suggest a beginning for
338 modern-style plate tectonics and therefore orogens (as per the types presented above) at or
339 following 2.5 Ga (*St-Onge et al., 2006; Shirey and Richardson, 2011; François et al., 2018;*
340 *Brown and Johnson, 2020*). It is for this reason that we decided to document orogens and their
341 attributes post 2.5 Ga and to represent old geodynamic features (i.e. Archean dome-and-keel
342 structures) in the craton map (**Fig.4**).

(Figure 4 is about here)

5.2. “Modern” orogens (after 2.5 Ga)

The convergent margin geodynamic processes that developed after the end of the Archean continued to undergo a progressive change as a result of the continued secular cooling of the Earth. Younger mountain belts show better-preserved components allowing a better definition of successive tectonic events with time. As a result, there may be a bias in the apparent extent of orogens with age. We choose to illustrate this aspect with examples from the Trans-Hudson, the Caledonian, the Variscan-Uralian and the Alpine systems, which show somewhat finer resolution with time. In all these examples, we restrict the definition of an “orogen” to spatial areas affected by tectonics events within and on the borders of former oceans that played a key role in the development of the crustal features related to ocean closure.

5.2.1. *Trans-Hudson orogeny and closure of the Manikewan Ocean*

The middle to late Paleoproterozoic, and specifically the aptly named Orisiran Period (2015-1800 Ma), is believed to have witnessed the assembly of Earth’s first genuine supercontinent variably referred to as either Nuna, Columbia, or Hudsonland (*Piper, 1976; Hoffman, 1988, 1997; Park, 1995; Meert, 2002, 2012, 2014; Pesonen et al., 2003*). The primary evidence for Nuna is based on paleomagnetic data and on comparative geology, and observed through alignment and broad synchronicity of geological features including mafic dyke swarms, rift and continental-margin supracrustal sequences, suture zones, oceanic and continental magmatic arcs, obducted ophiolites, thrust and fold belts, localities of high- and ultrahigh-pressure metamorphism, and in particular 2.00–1.74 Ga orogenic belts on a majority of the continents (**Fig. 5**; *Zhao et al., 2002, 2004, 2011; Evans and Mitchell, 2011; Evans et al., 2016; Meert and Santosh, 2017; de Oliveira Chaves et al., 2020; Wan et al., 2020*).

The largest of the Nuna-affiliated orogenic belts is the ca. 1.92–1.74 Ga Trans-Hudson orogen (Hoffman, 1988; Lewry and Stauffer, 1990), which comprises superbly exposed erosional remnants of a collisional mountain belt that extended ~3,000 km across North America into northeastern Arctic Canada (**Fig. 5**). In Greenland, the continuation of the Trans-Hudson orogen is the well-studied, but largely ice-covered, Nagssugtoqidian orogen (**Fig. 5**; van Gool *et al.* 2002; Connelly *et al.* 2006; Nutman *et al.*, 2008a; St-Onge *et al.*, 2009; Garde and Hollis, 2010; Kolb, 2014; and references therein). Further to the east, the Trans-Hudson orogen is manifest as the Lapland-Kola orogen in Baltica (**Fig. 5**; Daly *et al.*, 2006; Korja *et al.*, 2006; Lahtinen *et al.*, 2008; Petrov *et al.*, 2018; Lahtinen and Huhma, 2019). In Canada, numerous studies of the Trans-Hudson orogen have shown that it formed during progressive closure of the Manikewan Ocean and accretion of intervening suspect terranes, culminating with the collision of the lower-plate Archean Superior craton with an upper-plate collage of Archean and Paleoproterozoic crustal blocks (collectively known as the Churchill plate) at ca. 1.82 Ga (St-Onge *et al.*, 2006, 2007, 2009, 2020; Corrigan *et al.*, 2009; Corrigan, 2012; Weller and St-Onge, 2017; Dumond, 2020; Weller *et al.*, 2020; and references therein). Several parallels have been drawn between the Trans-Hudson orogen and the archetypical Himalayan orogen of central Asia as both orogens are of a similar scale and exhibit similar patterns of magmatism and metamorphism with respect to the timing of collision (St-Onge *et al.*, 2006; Eaton and Darbyshire, 2010; Thompson *et al.*, 2010; Bastow *et al.*, 2011; Corrigan, 2012; Weller and St-Onge, 2017).

Signature tectonic features that characterize Trans-Hudson orogen include mounting evidence for continental subduction during diachronous plate convergence as documented by high- to ultrahigh-pressure/low-temperature mineral assemblage localities, and closure of intervening oceanic domains as witnessed by obducted ophiolitic suites. Within the Trans-Hudson orogen of Laurentia and Baltica (**Fig. 5**) well-researched Paleoproterozoic eclogite localities include:

the 1.83 Ga Kovik eclogite (Weller and St-Onge, 2017), the 1.87 Ga Kuummiut eclogite (Nutman et al., 2008b; Müller et al., 2018) of SE Greenland, and the 1.9 Ga Belomorian eclogite (Yu et al., 2017) of western Russia. Taken as a set, the Trans-Hudsonian eclogites record a clear east-to-west diachronous trend in the age of continental subduction and orogenesis, and by inference, Manikewan Ocean closure. (Figure 5 is about here)

5.2.2. Caledonian orogeny and closure of the Iapetus Ocean

On the northern British Isles, during the Early-Middle Ordovician (c. 470-460 Ma), a volcanic island arc (the Midland Valley arc terrane) that had formed along the margins of the Iapetus Ocean, collided with Laurentia (**Fig.6**). This marked the onset of the Grampian phase of the Caledonian orogeny (Strachan & Woodcock, 2021). Sedimentary and igneous rocks trapped between the colliding landmasses were intensely folded and faulted and subject to varying degrees of regional metamorphism (Smooth, 2012).

The subsequent (or continuing) Scandian phase of the Caledonian orogeny, which occurred from the Ordovician to the Middle Silurian (450-430 Ma), resulted in full closure of the Iapetus Ocean and amalgamation of Laurentia, Baltica and Avalonia to form Laurussia (**Fig.6**). Early phases of deformation and/or metamorphism are recognized in the Scandinavian Caledonides and resulted from eastern Avalonia moving independently towards Baltica during the Ordovician. This motion was accommodated by the subduction of the southeastern Iapetus Ocean (so called Tornquist Sea) beneath Avalonia. The continental collision between Avalonia and Baltica began in the Late Ordovician, about 450 Myr ago (**Fig.6**). As a result, the Tornquist Sea was consumed and the resulting suture became the Tornquist Line. This was followed by another phase of deformation and low-grade metamorphism during the Early Devonian (c. 405-395 Ma), known as the early Acadian phase in the present-day NW North American and NE European Atlantic margins. Clearly then, the Iapetus Ocean first closed in the north (Robert et

al., 2020) with continued closure migrating south and resulting in a diachronous collisional orogeny. (Figure 6 is about here)

5.2.3. Variscan-Uralian orogenies and closure of the Rheic and Uralian oceans

The closure of the Paleozoic oceans that extended from Mexico to Central Asia gave rise to the Pangea supercontinent. To a certain extent, the resulting orogenies may also include the Acadian and Alleghanian belts in North America, but the main orogenic segment is usually referred to as the Variscan Orogen, which is well documented in Western Europe, from Iberia and Cornwall to Bohemia. The Variscan belt is also represented in basement domains to the Alpine orogen but its distribution is still poorly understood.

The late Acadian Orogeny (375-325 Ma) resulted from the collision of an island arc with Laurussia. It was followed by the Alleghenian Orogeny (325-260 Ma) driven by the collision between Laurussia and Gondwana. Together, both orogens formed the Ouachita and part of the Appalachian Mountains in North America, as well as the Moroccan Meseta and the Anti-Atlas in north-western Africa, a once continuous orogenic belt prior to the opening of the Atlantic Ocean starting in Triassic times (Michard et al., 2020).

The Variscan orogen of Southern Britain and Western Europe is related to the closure of several oceanic basins of unknown width between Laurussia in the north and Gondwana in the south (in the present coordinates). Some intervening microcontinents are also recognized between the two large convergent continents (e. g. Matte, 1986, 2001; Pin, 1990; Faure et al. 2005, 2008; Martinez-Catalan et al. 2009; Ballèvre et al., 2009, Lardeaux et al., 2014). In spite of several controversies on the spatial extent and age of the oceanic domains related to the Variscan orogeny, the Saxothuringian and Armorican microcontinents are interpreted as pericratonic to the Gondwanian margin, rifted in the Early Ordovician (ca 480-460 Ma), and containing evidence of the Neoproterozoic Cadomian orogeny. From the inner to outer parts of Gondwana,

441 three oceanic domains are recognized. i) The Medio-European (also called Galicia-Massif
442 Central) Ocean was probably a small basin similar to the present Red Sea, on the basis of similar
443 benthic faunas on both sides of the ocean. ii) The Tepla - Le Conquet Ocean separated the
444 Armorican and Saxothuringian blocks. iii) Lastly, the Rheic Ocean was the main oceanic basin
445 separating Laurussia and Saxothuringia, and the widest due to the presence of different faunal
446 assemblages observed on the Laurussia and Gondwana sides. The timing of closure for these
447 oceanic basins remains disputed. Some authors argue for a poly-orogenic evolution whereas
448 others argue for a monophased collision. In the monocyclic model, all three oceans closed in
449 the Late Devonian-Early Carboniferous, whereas, in the polycyclic model, the Medio-European
450 Ocean was closed by an early oceanic subduction below Armorica. The Gondwana-Armorica
451 Early Devonian collision gave rise to the eo-Variscan belt characterized by HP-UHP
452 metamorphic rocks (eclogites and granulites), SW-directed synmetamorphic nappes, and
453 crustal melting. In the Variscan orogeny, the main collisional event took place when both the
454 Rheic and Tepla-Le Conquet oceans closed after south-directed subductions in Late Devonian-
455 Early Carboniferous, at ca 360 Ma. Late collisional tectono-metamorphic events developed
456 from Viséan to Serpukhovian (ca 335-320 Ma). This deformation produced a doubly-vergent
457 orogen, i.e. north- and south-directed folds and thrusts on the north and south sides of the
458 orogenic belt, respectively. Due to the Permian Ibero-Armorican orocline, the northern foreland
459 developed from the Rheinische Schiefer Gebirge of eastern Germany, through the Ardennes,
460 SW England, and southern Portuguese zone of Iberia with various displacement directions. The
461 continuity of the southern foreland basin, well exposed in southern France (Massif Central,
462 Pyrenees), SE Spain, and SW Sardinia, was disrupted by the opening of the Provence-Algeria
463 basin in the Neogene. Of interest is the fact that the Paleotethys Ocean is not recognized to the
464 west of Turkey, and therefore did not play a significant role in the formation of the Variscan
465 orogen.

The Uralian orogen was the consequence of the Permian closure of the Uralian Ocean between Laurussia in the west and Siberia in the east. The former comprises the subducting plate, whereas the latter was the upper plate, upon which a magmatic arc was built (*Puchkov, 2009*). The southwestern margin of Baltica defined by the Trans European Suture Zone represents an Early Paleozoic suture (sometimes referred to as the “Polish Caledonides”), possibly reworked by strike-slip faulting in the Late Paleozoic. To the southeast of Baltica, the Uralian orogen formed when Laurussia and Kazakhstania collided with Siberia in the late Carboniferous to early Permian.

The eastern extension of the Variscan orogen in Central Asia remains poorly documented and is beyond the scope of the present contribution. The Central Asia Orogenic Belt (CAOB) is the consequence of multiple microcontinent collisions between the Siberia, Tarim, and North China Block (**Fig.7**; *Sengor et al., 1996*; *Sengor and Natal'in, 1993*; *Xiao et al. 2008*; *Choulet et al., 2012*). However, another Paleozoic ocean developed at the location of the Kunlun orogen between the Tarim and Qaidam blocks and was connected to South China and Indochina seas. (Figure 7 is about here)

5.2.4. *Alpine orogeny and closure of the Neo-Tethys and Alpine Tethys oceans*

The complex history of the Alpine cycle begins with the opening of the Triassic (Carnian, 220 Ma) Meliata-Maliac Ocean, followed by the opening of the Vardar Ocean in the Jurassic (Toarcian-Aalenian, 180 Ma), later referred to as Neo-Tethys (east) and Alpine Tethys oceans (*Stampfli, 2000*; *Bousquet et al., 2008*). In the Western Alps, the Liguro-Piemonte ocean opened in Middle Jurassic (Callovian-Oxfordian). The complexity of the Alpine orogeny comes from the fact that ~~the~~ its history is intrinsically related to the diachronous closure of two oceanic systems, the Neo-Tethys Ocean (eastern domain) and the Alpine Tethys Ocean (western domain ; *Stampfli, 2000* ; *Bousquet et al., 2008*) when Africa ~~and India~~ collided with Eurasia. Convergent plate movement began in the early Cretaceous, but the major phases of

491 mountain building occurred from the Paleocene to the Eocene, and continues today (**Fig.8**; refs).

492 The Alpine Orogen *sensu lato* includes all the Tethyan system from the Moroccan Atlas to the

493 Himalaya.

494 The Alpine Tethys comprises the Piemont-Liguria Ocean and the Valais Ocean. These are

495 extensions of the Atlantic Ocean continuing eastward into the area of the Circum-Mediterranean

496 orogens during rifting and drifting. The Piemont-Liguria Ocean becomes an extension of the

497 Central Atlantic by Middle Jurassic time. In turn, the Valais Ocean (northeast of the

498 Briançonnais microcontinent) forms a Northern Atlantic extension by Middle Cretaceous time.

499 The opening of the Piemont-Liguria Ocean led to separation of the Adria plate from the

500 European continent in Callovian-Oxfordian time (165-156 Ma) and its closure forms the suture

501 zone between the Austro-alpine nappes (derived from Adria) and the Briançonnais

502 microcontinent. Contemporaneously, the Triassic Meliata-Maliac Ocean subducted

503 southeastward beneath the Jurassic upper plate portions of the Neo-Tethys Ocean (ophiolitic

504 units from Vardar that later formed the Dinarides). By Cretaceous times, the closure of Meliata-

505 Vardar produced the Eastern Alps orogeny (*Schmid et al., 2008; Bousquet et al., 2008; Handy*

506 *et al., 2014*).

507 In Santonian time the two branches of the Alpine Tethys (Valais and Piemont-Liguria) were

508 connected *via* the Carpathian embayment with the younger part of Neo-Tethys (Cretaceous in

509 age), leading to a back-arc ocean that formed the future Sava Zone of Dinarides, which is

510 considered a remnant of the Vardar Ocean. By Tertiary time, closure of the Alpine Tethys

511 produced the Western and Central Alps orogeny (*Bousquet, 2008, Handy et al., 2010, 2014;*

512 *Schmid et al., 2019*). Further east the Caucasian-Arabian belt is part of the late Cenozoic

513 Alpine-Himalayan belt. The Trans-European Suture Zone-Kopetdag-Caucasus megafault

514 separates the alpine orogen and the Eurasian plate. The Caucasian-Arabian segment consists of

515 the Greater Caucasus to the north and the Caucasian-Arabian Syntaxis to the south. Greater

Caucasus represents the southern margin of the Eurasian plate uplifted over the Main Caucasian Fault (part of the Trans-European Suture Zone-Kopetdag-Caucasus megafault). The Caucasian-Arabian Syntaxis is a zone of indentation of the Arabian plate into the southern East European craton and the Alpine structure of the Caucasus is generated by the interaction of the Arabian indenter and the East European craton. By the late Cenozoic this interaction lead to a 200 to 400 km shortening of the Caucasian-Arabian Syntaxis (*Sharkov et al., 2015*). Therefore, in a tectonic sense the Alpine orogen (*lato sensu*) is the result of two orogenies associated with closure of two oceanic systems: Meliata (Neo-Tethys) by the end of Cretaceous time and Alpine Tethys (Atlantic) during Tertiary time. (Figure 8 is about here)

6. Key orogenic attributes and their use in a World Map of Orogens

Due to space and legibility issues, as well as the polycyclic nature of deformation and metamorphism, many orogenic features are difficult to represent on a World Map of Orogens. In order to overcome these constraints, it becomes necessary to identify deformation zones at a regional scale, and to organize them in terms of fundamental components of a “typical” mountain belt. In addition, another issue that requires careful consideration is the spatiotemporal overprint of orogenic events. As outlined in this paper, we have identified key elements of different types of orogens in order to define a number of “typical” orogens and map them on a World map. To do so, we propose the creation of several layers corresponding to time windows that encompass the development of a given orogen. This superposition of these layers allows us to highlight a range of continental growth processes and events through time. Thus, our approach hinges on the identification of a spatial zonation that is common to many (most?) mountain belts and uses geological concepts that are based on the global rock record. The proposed zonation comprises elements that characterize the architecture of many mountain belts and include in the internal zone: lower plate “crystalline” metamorphic complexes (often

but not always corresponding to older lithostratigraphic substratum) modified by thick-skin structures and (U)HP and (U)HT metamorphism, overlain by deformed sedimentary rocks deposited in former distal oceanic areas and possibly ophiolitic slivers of oceanic crust. In contrast, the orogenic upper plate usually contains relics of magmatic arcs and associated (U)HT metamorphism formed prior to collision and ophiolites corresponding to associated back-arc and fore-arc basins. The adjacent foreland and hinterland domains are devoid of high grade metamorphic assemblages and are generally dominated by thin-skin structures that deform sedimentary strata (**Table 1**).

In this broad view of orogenic architecture, there are of course many additional complexities that can originate with back-thrusting of lithostratigraphic units for example and that result in structurally-complex domains. Nevertheless, common features such as the direction and polarity of convergence as documented with folds, faults and fold-and-thrust belts can be documented globally. Thus, the proposed zonation (early orogenic, orogenic, peri-orogenic and para-orogenic zones) is useful and certainly reflects key geodynamic processes pertinent to the evolution of orogens through time.

The use of the four-fold classification of orogens is therefore expected to document the growth and modification of continents through time and space since 2.5 Ga. We propose to distinguish the various deformation zones (and their components) on the World Map of Orogens with different shades of the same color and thus improve our understanding of the evolution of orogens from early Earth geodynamic processes to current plate tectonic dynamics within a continuum of deformation.

7. Conclusion

In the classic Wilson Cycle model, there is a space-time continuum of deformation, during which the various geological elements of a mountain belt system are actively deformed. Since the Wilson Cycle nevertheless treats these elements independently, whereas modern

concepts consider the various geodynamic processes to be intimately linked during orogenic evolution, we propose that consideration of the various stages in mountain building may help in the classification of convergent orogens. To a certain extent, this implies considering inherent links between the various compressional stages, which can be either end-member type (i.e. well-defined spatially and temporarily) or transient (i.e. changing in space and time). We consider that orogenic zones, which are characterized by (U)HP metamorphism, (U)HT metamorphism, magmatism (aluminous granites) and/or ophiolites (= orogenic zone in types II and III), represent the core of the orogenic system and correspond to what is often referred to as the “internal zone”. Such zones can be readily documented in the older orogens even though erosion may have removed much of the higher structural level rock record. On the other hand, the peri-orogenic zones, in which deformation includes fold-thrust belts and low-grade metamorphism (= peri-orogenic zones in types II and III and early or para-orogenic zones in types I and IV), may have in part (but generally not completely) been eroded in ancient orogens. These are important elements to take in account to achieve a full understanding of the internal architecture of orogens, as well as their sequencing, history, and role in the growth and reorganization of continents with time.

8. Acknowledgments

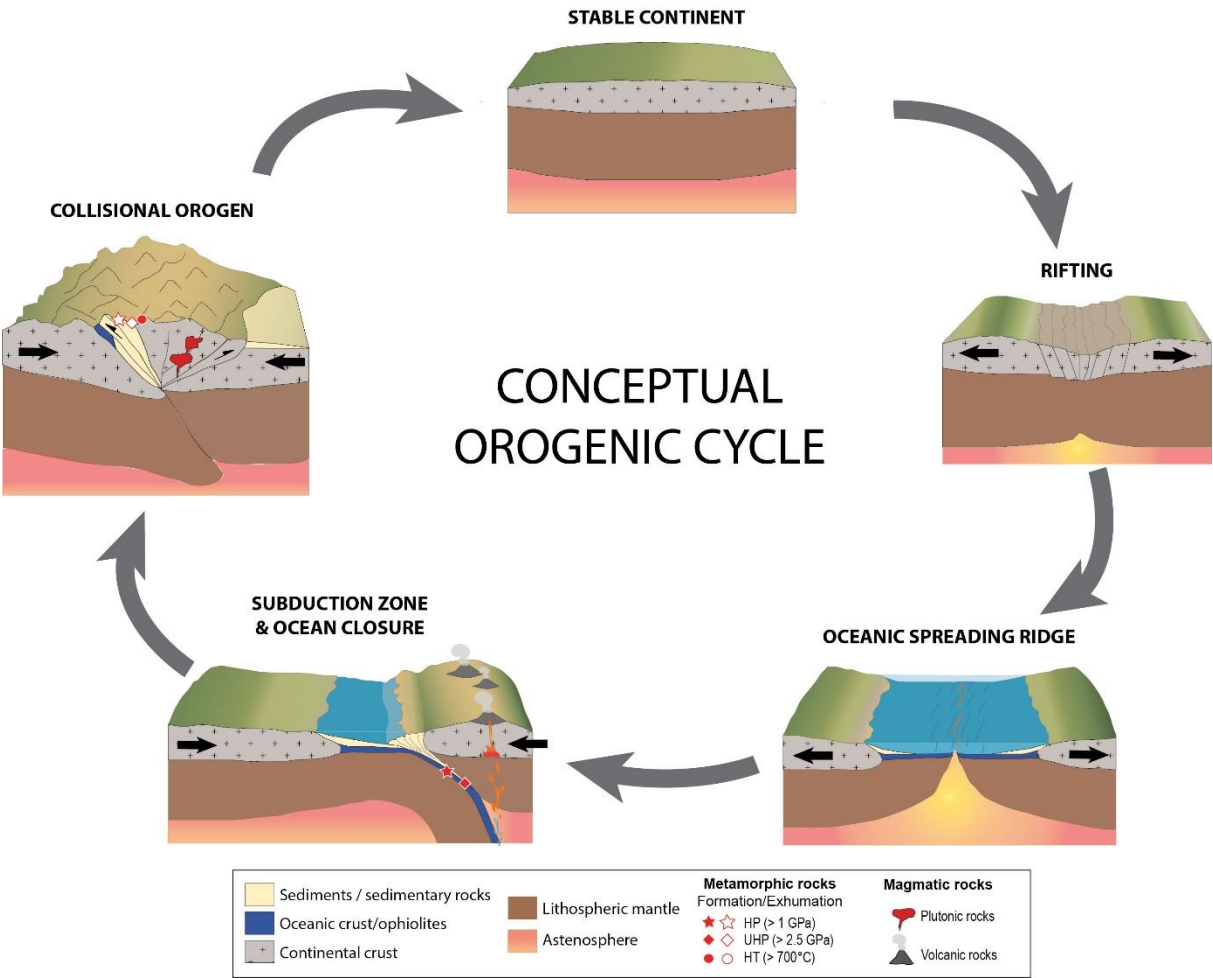
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591 (<https://ccgm.org/en/content/11-team>). We thank the reviewers for their comments, which
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593 #xxxx.

594 **9. Captions**

Types of orogens	Orogenic zones	Oceanic closure	Magmatism	Metamorphism	Fold and Thrust Belt	Pre- to syn-orogenic sediment deposit	Late- to post orogenic sediment deposit
Cordilleras (Type I)	EARLY-OROGENIC	in progress	Volcanic & plutonic rocks (Primary melts, dominantly calco-alkaline)	in lower part: HP/LT	Accretionay prism	Formation of a retro-arc foreland basin, Thick sedimentary sequence Flysch-Accretionary wedge	Absent but sediments deposited in the forearc basin may unconformably cover the FTB
Accreted terranes (Type II)	OROGENIC	YES (Multiple) ophiolites	Syn-collisional (Migmatites and aluminous granites)	(U)HP (U)HT	-	-	-
	PERI-OROGENIC	-	-	-	FTBs form where passive continental margins collide with a subduction zone	-	Molassic basin
Continent-Continent (Type III)	OROGENIC	YES ophiolites or ophiolitic mélange	Syn-collisional (Migmatites and (per)-aluminous granites)	HP/UHP (prograde) during continental subduction and MP-LP/(U)HT (retrograde) coeval with crustal melting	-	Passive continental margin deposits	-
	PERI-OROGENIC	-	-	-	Bilateral foreland/hinterland FTB	Passive continental margin deposits, (volcanic)-sedimentatry terrigenous rocks turbiditic deposts	Thick molassic deposits in foreland and hinterland basins. Intramontane basins

Inverted rift (Type IV)	INTRAPLATE / PARA- OROGENIC	NO real oceanic closure	No magmatic arc Migmatites and (per)aluminous granites	Prograde and retrograde metamorphism	FTB	Passive continental margin deposits, (volcanic)-sedimentary terrigenous rocks	Molassic deposit
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598

599 **Fig.1.** Cyclic view of the different stages of a theoretical orogenic cycle as an extension-
600 compression continuum (Wilson Cycle). The uplift, erosion and unroofing processes including
601 exhumation of metamorphosed rocks and crustal melting occur in the transition between the
602 collisional orogen stage and the stable continent stage.

603

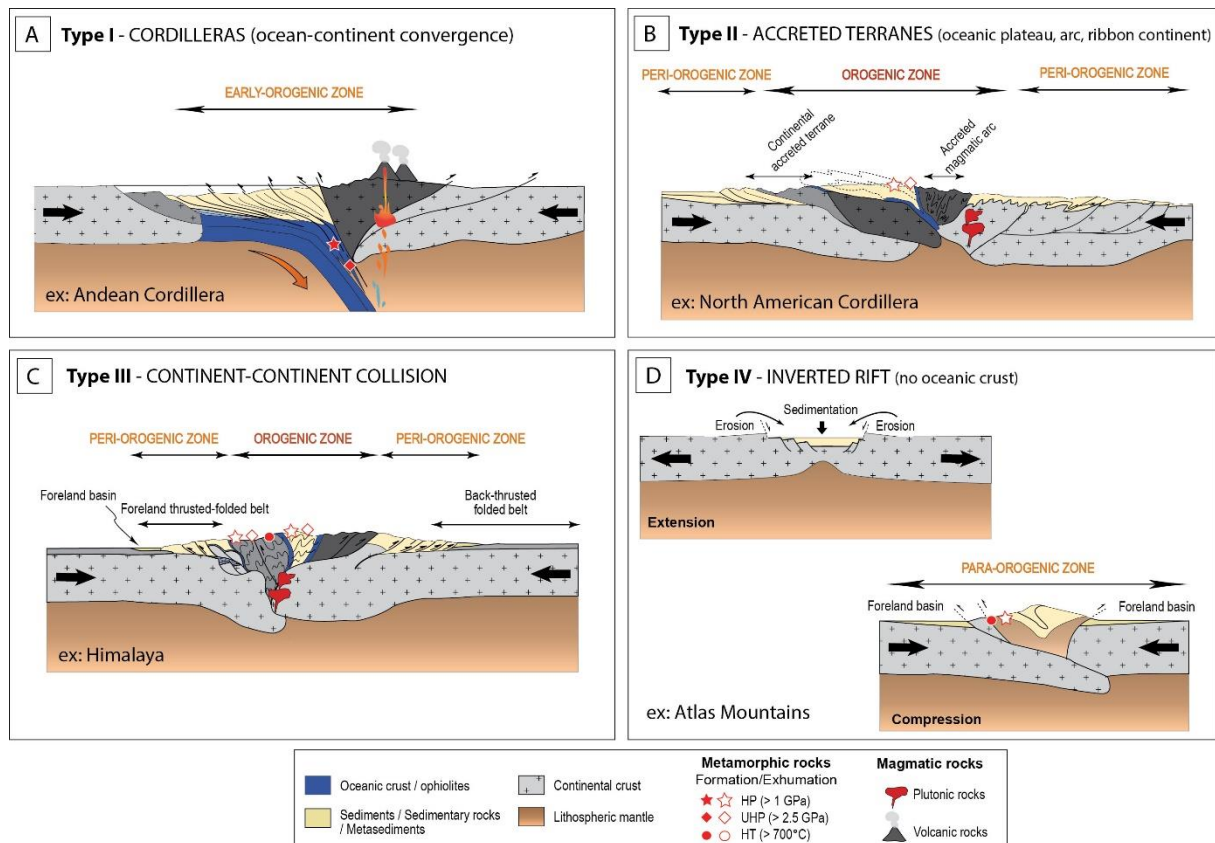


Fig.2. The different types of orogens. (a) Subduction type, (b) accreted terranes type, (c) continental collision type (d) intracontinental type. Figures modified from *Lagabriele and Bodinier, 2008; Robert and Bousquet, 2018.*

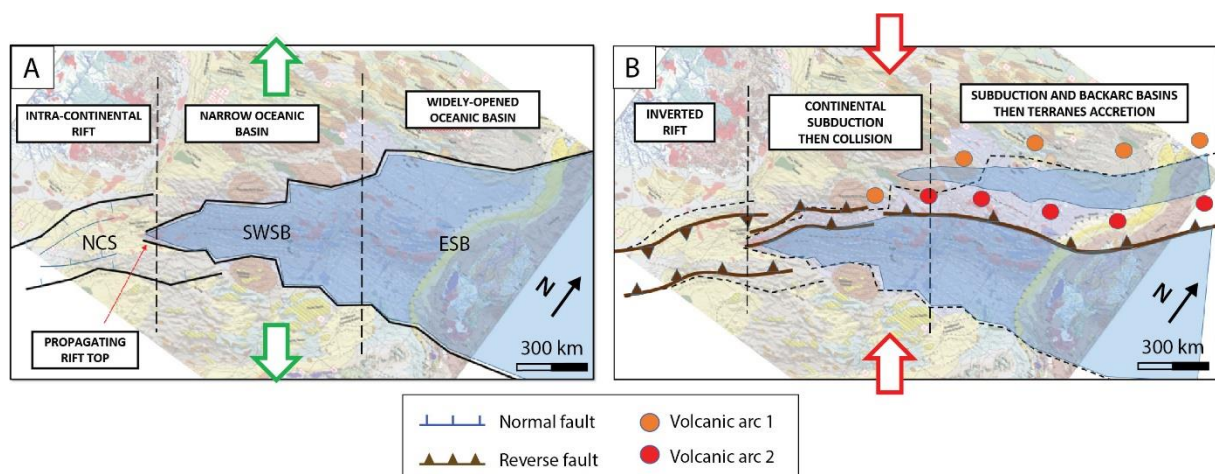


Fig.3. Theoretical “scissor shape”. (a) Active opening setting and (b) projected setting during shortening. The background is an example modified from *Pubellier et al., 2016; NCS; Nam*

Con Son Basin, SWSB; SW sub-basin, ESB; Eastern Sub-basin. Figure (b) is a speculative scenario for future basin closure.

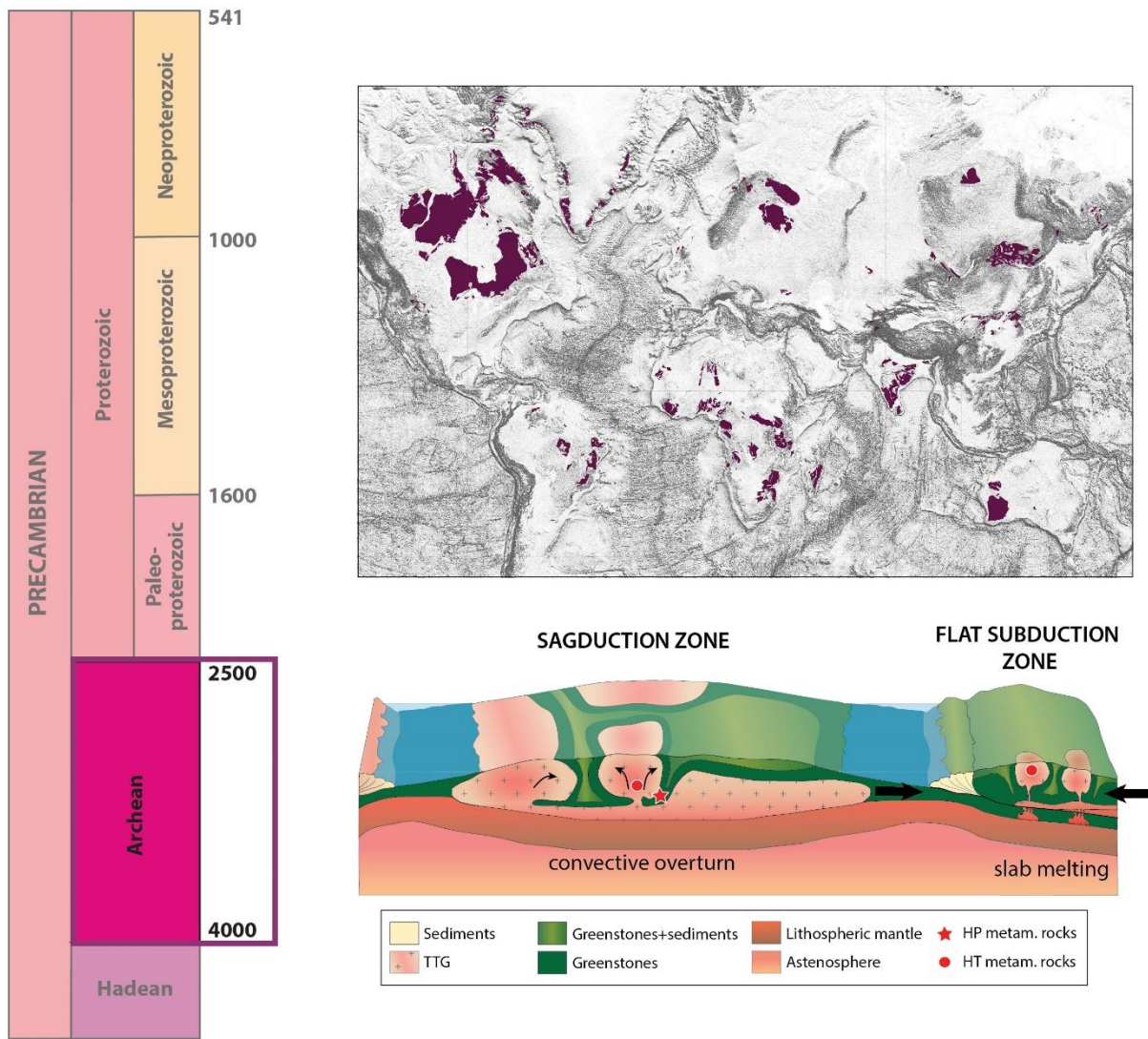


Fig.4. Extent of Archean crust on the World Map and representative Archean dome and keel structures formed by geodynamical processes that ended at about 2.5 Ga: flat subduction and sagduction of dense greenstone lithologies into lighter felsic crystalline basement. The higher mantle heat flux is envisioned as leading to the partial melting of the lower crust (flat subduction) and the convective overturn of the Archean crust (sagduction).

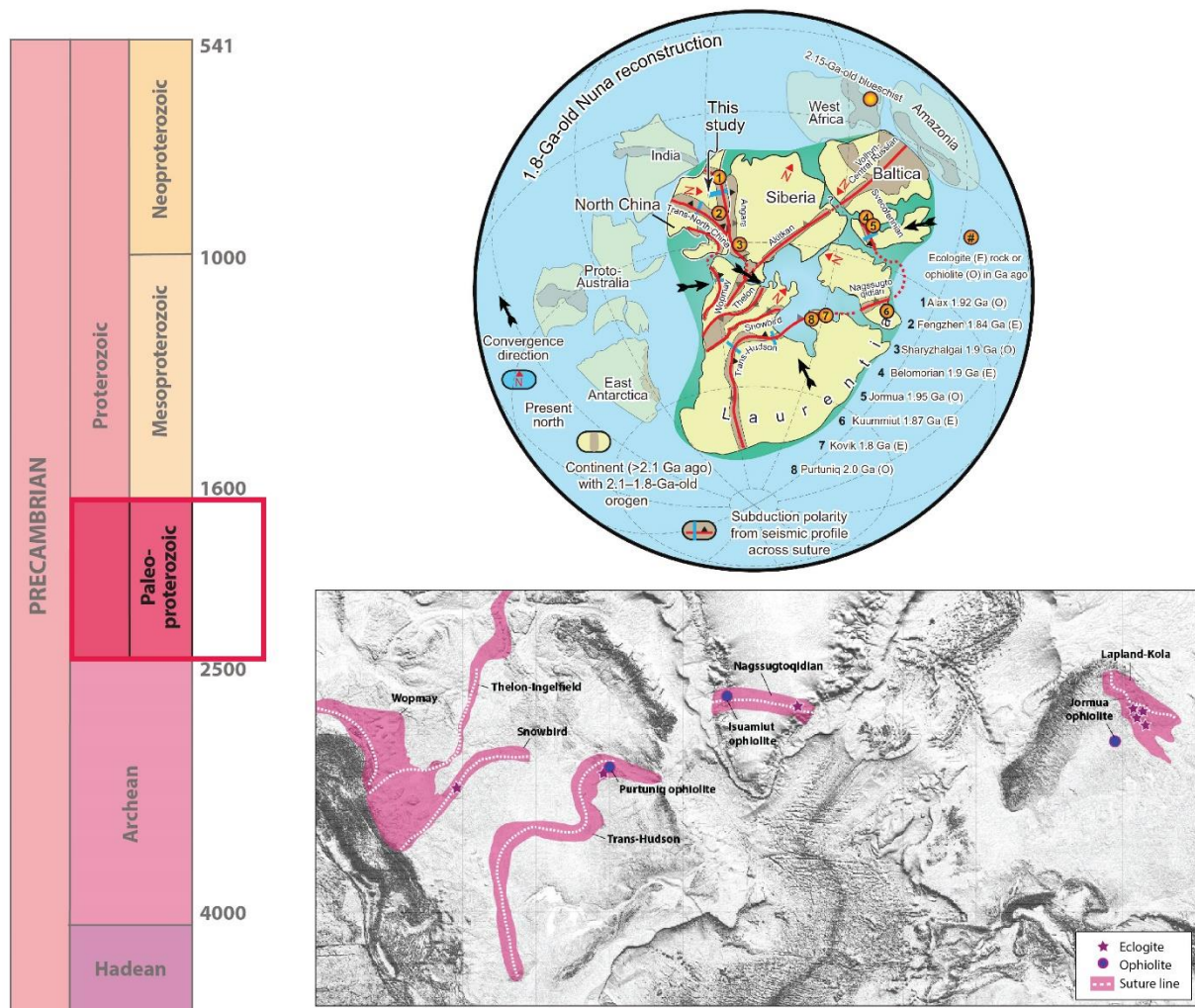


Fig.5. Present extent of some of the Paleoproterozoic orogens (may include reworked Archean crust) in the North Atlantic realm on the World Map, and in the past as shown on a Nuna reconstruction ~1.8 Ga ago (*Evans and Mitchell, 2011*), including the position of North China (*Wan et al., 2020*). Convergence directions (black arrows) are inferred from subduction polarity (black triangles) indicated at seismic profiles (blue). Speculated continuation of seismically confirmed sutures (red). A 2.15-Ga-old blueschist immediately preceding the Orosirian period is shown (*Ganne et al., 2012*). Green region marks configuration of core continents assembled through two to three major sutures. Putative positions in the larger supercontinent Nuna are suggested for other less studied continents (slightly transparent). Seismic evidence is consistent with previous paleomagnetic and geological configurations in the green region under the paradigm of a modern, global subduction-driven plate tectonic network. From *Wan et al.*

(2020). Location of eclogites and blueschist from: Kola-Lapland orogen, Belomorian Province, Russia (Volodichev *et al.*, 2004; Skublov *et al.*, 2011; Slabunov *et al.*, 2011); Nagssugtoqidian orogen, South-East Greenland (Nutman *et al.*, 2008; Müller *et al.*, 2018); Trans-Hudson orogen, North America (Weller & St-Onge, 2017); Snowbird Tectonic Zone, Canada (Snoeyenbos *et al.*, 1995; Baldwin *et al.*, 2003; 2004).

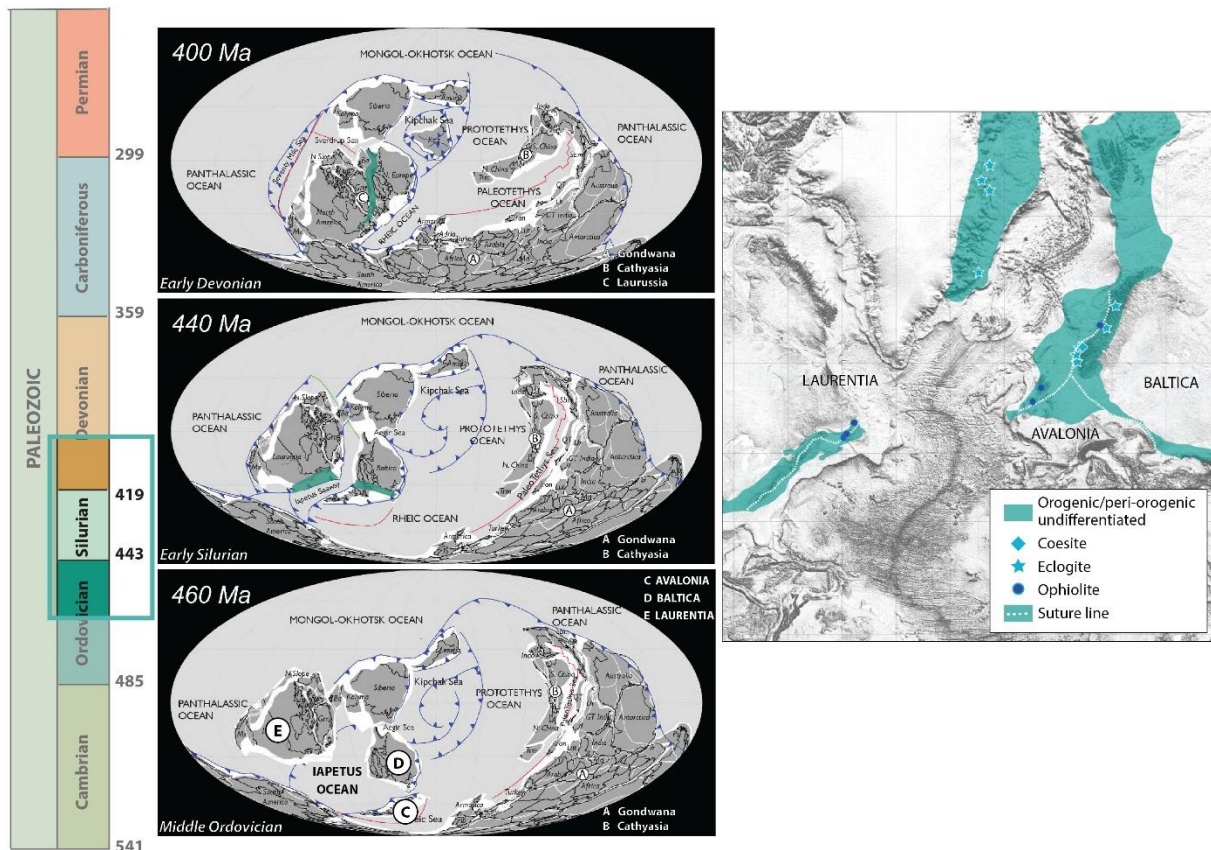


Fig.6. Present location of the Caledonian orogen resulting from the progressive closure of the Iapetus Ocean shown on the World Map and in the past as shown on paleogeographic reconstructions from Scotese, *PALEOMAP Project* (www.scotese.com).

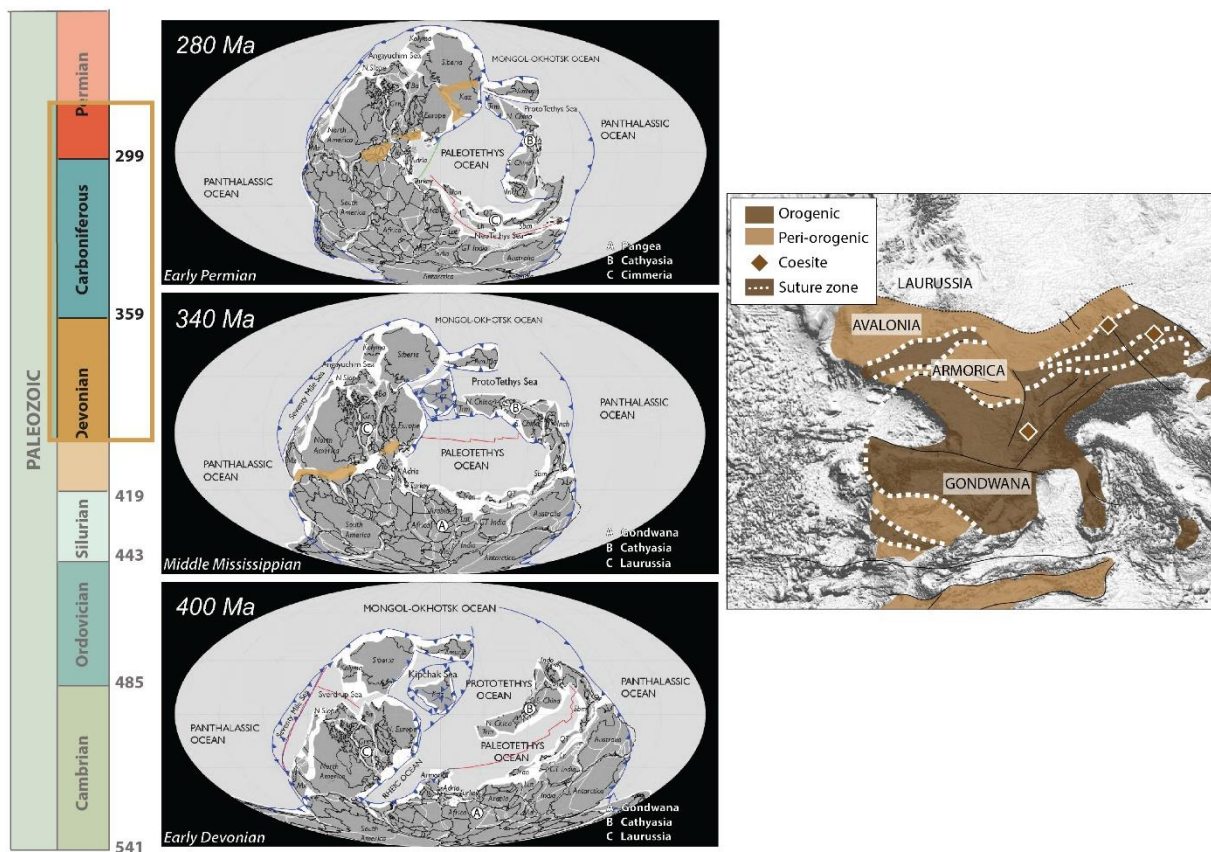


Fig.7. Present location of the Variscan orogens shown on the World Map and in the past as shown on paleogeographic reconstructions from Scotese, *PALEOMAP Project* (www.scotese.com).

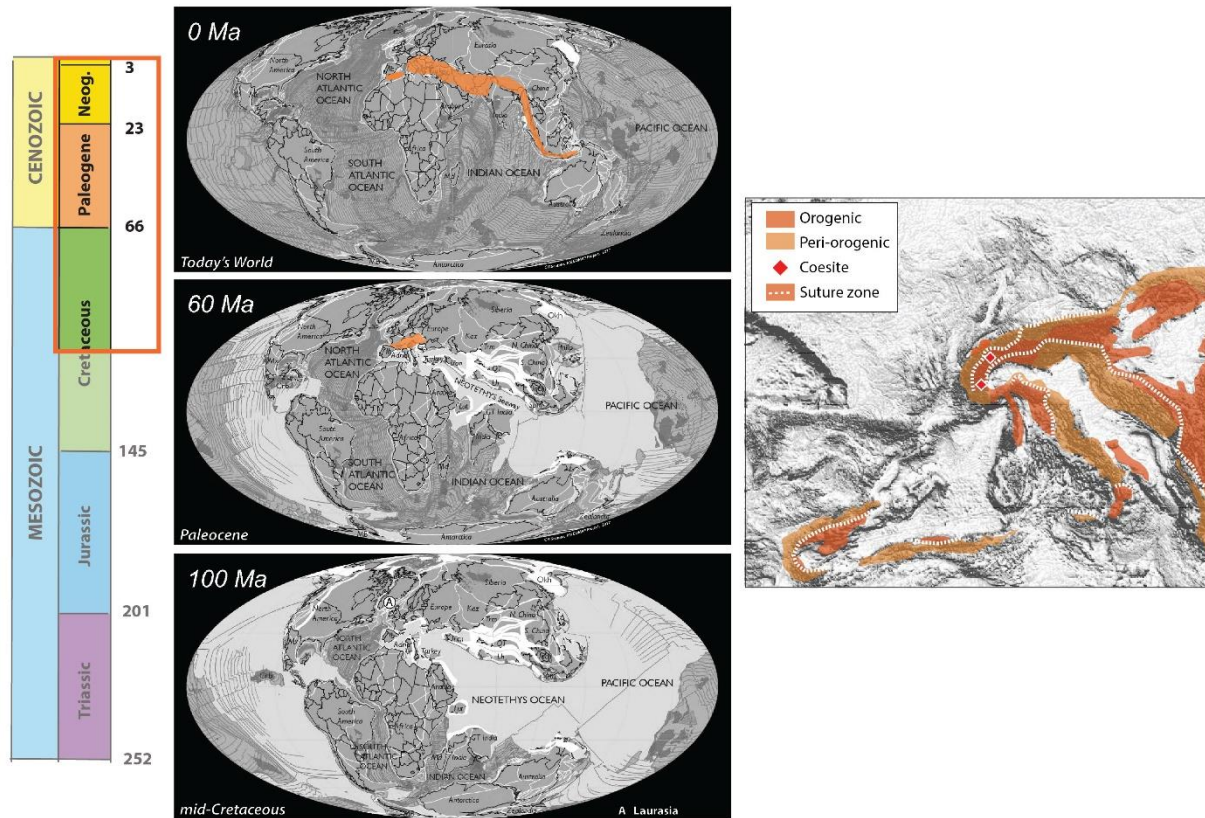


Fig.8. (a) Present location of the Alpine orogens (following closure of the Neo-Tethys and Alpine oceans) shown on the World Map and in the past as shown on paleogeographic reconstructions from Scotese, *PALEOMAP Project* (www.scotese.com).

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