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

Camille Francois, Manuel Pubellier, Christian Robert, Cédric Bulois, Siti Nur Fathiyah Jamaludin, Roland Oberhänsli, Michel Faure, Marc R. St-Onge

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

Camille François (1,), Manuel Pubellier (1,2,*), Christian Robert (2), Cédric Bulois (2,3), Siti Nur Fathiyah Jamaludin (4), Roland Oberhänsli (5), Michel Faure (6), Marc R. St-Onge (7) and the IGCP 667 Team (8)*

[1] *Commission for the Geological Map of the World – France*
[2] *Laboratoire de Géologie - CNRS UMR 8538, École Normale Supérieure - PSL University – France*
[3] *Université Côte d'Azur, IRD, CNRS, Géoazur – France*
[4] *Universiti Teknologi Petronas – Malaysia*
[5] *Potsdam University – Germany*
[6] *Univ. Orléans, CNRS, BRGM, ISTO, UMR 7327, F-45071 Orléans – France*
[7] *Geological Survey of Canada, 601 Booth Street, Ottawa, Ontario K1A 0E8 – Canada.*
[8] *https://ccgm.org/en/content/11-team*
[*] *Corresponding authors, email addresses: c.francois.geology@orange.fr (C. François), manupub.pubellier@gmail.com (M. Pubellier)*

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

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

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

The present overview of the world's orogens is part of the ongoing IGCP 667 Project "World Map of Orogens", in which the main driver is to design a map at scale 1:35 000 000 documenting the distribution of ancient to modern orogens throughout Earth history. The compilation of a "World Map of Orogens" necessarily requires a careful consideration of both the various types of global-scale compressional tectonic regimes and the best means of portraying the associated tectonic elements or attributes on such a map. Consideration of pre-plate tectonic, Early Earth crustal deformation processes and features also becomes necessary. 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

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

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

158 Type II (**Fig.2b**) involves micro- or ribbon-continents and slivers of the continental
159 margin rifted during the opening of back-arc basins. The slivers may also include oceanic
160 plateaus (e.g. Otong-Java in the Solomon Islands), exotic island arc terranes (e.g. North
161 American Cordillera) or continental accreted terranes (e.g. Cascadia). The convergent process
162 is often known as terrane docking or accretion and it results in a tectonic collage of pericratonic
163 and suspect terranes (*Coney et al., 1980; Howell, 1989).* The accretion of pericratonic, suspect,
164 or exotic terranes represents a milestone in the history of an oceanic basin. The orogen forms
165 when a crustal domain is rifted away from a continental margin and eventually reaches an active
166 continental margin. As subduction of the now-accreted crustal domain is not possible,
167 subduction jumps to the opposite side of the docked terrane (e.g. Anatolia, Iran; *Barrier and*
168 *Vrielynck, 2008*, or SE Asia; *Pubellier and Meresse, 2013*). Where observations are more
169 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 (*Courtillot, 1982*). Notable examples include the Woodlark Basin (*Weissel and Watts, 1979; Taylor et al., 1999;*), the Gulf of Aden (*Courtillot, 1982; d'Acremont et al., 2006; Fournier et al., 2010*) and the South China Sea (**Fig.3**; *Briais 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 serpentinized 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-trondjhemite-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

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

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

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

397 *5.2.2. Caledonian orogeny and closure of the Iapetus Ocean*

398 On the northern British Isles, during the Early-Middle Ordovician (c. 470-460 Ma), a
399 volcanic island arc (the Midland Valley arc terrane) that had formed along the margins of the
400 Iapetus Ocean, collided with Laurentia (**Fig.6**). This marked the onset of the Grampian phase
401 of the Caledonian orogeny (*Strachan & Woodcock, 2021*). Sedimentary and igneous rocks
402 trapped between the colliding landmasses were intensely folded and faulted and subject to
403 varying degrees of regional metamorphism (*Smooth, 2012*).
404 The subsequent (or continuing) Scandian phase of the Caledonian orogeny, which occurred
405 from the Ordovician to the Middle Silurian (450-430 Ma), resulted in full closure of the Iapetus
406 Ocean and amalgamtion of Laurentia, Baltica and Avalonia to form Laurussia (**Fig.6**). Early
407 phases of deformation and/or metamorphism are recognized in the Scandinavian Caledonides
408 and resulted from eastern Avalonia moving independently towards Baltica during the
409 Ordovician. This motion was accommodated by the subduction of the southeastern Iapetus
410 Ocean (so called Tornquist Sea) beneath Avalonia. The continental collision between Avalonia
411 and Baltica began in the Late Ordovician, about 450 Myr ago (**Fig.6**). As a result, the Tornquist
412 Sea was consumed and the resulting suture became the Tornquist Line. This was followed by
413 another phase of deformation and low-grade metamorphism during the Early Devonian (c. 405-
414 395 Ma), known as the early Acadian phase in the present-day NW North American and NE
415 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 collisonal 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,

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

mountain building occurred from the Paleocene to the Eocene, and continues today (**Fig.8**; refs). The Alpine Orogen *sensu lato* includes all the Tethyan system from the Moroccan Atlas to the Himalaya.

The Alpine Tethys comprises the Piemont-Liguria Ocean and the Valais Ocean. These are extensions of the Atlantic Ocean continuing eastward into the area of the Circum-Mediterranean orogens during rifting and drifting. The Piemont-Liguria Ocean becomes an extension of the Central Atlantic by Middle Jurassic time. In turn, the Valais Ocean (northeast of the Briançonnais microcontinent) forms a Northern Atlantic extension by Middle Cretaceous time. The opening of the Piemont-Liguria Ocean led to separation of the Adria plate from the European continent in Callovian-Oxfordian time (165-156 Ma) and its closure forms the suture zone between the Austro-alpine nappes (derived from Adria) and the Briançonnais microcontinent. Contemporaneously, the Triassic Meliata-Maliac Ocean subducted southeastward beneath the Jurassic upper plate portions of the Neo-Tethys Ocean (ophiolitic units from Vardar that later formed the Dinarides). By Cretaceous times, the closure of Meliata-Vardar produced the Eastern Alps orogeny (*Schmid et al., 2008; Bousquet et al., 2008; Handy et al., 2014*).

In Santonian time the two branches of the Alpine Tethys (Valais and Piemont-Liguria) were connected *via* the Carpathian embayment with the younger part of Neo-Tethys (Cretaceous in age), leading to a back-arc ocean that formed the future Sava Zone of Dinarides, which is considered a remnant of the Vardar Ocean. By Tertiary time, closure of the Alpine Tethys produced the Western and Central Alps orogeny (*Bousquet, 2008, Handy et al., 2010, 2014; Schmid et al., 2019*). Further east the Caucasian-Arabian belt is part of the late Cenozoic Alpine-Himalayan belt. The Trans-European Suture Zone-Kopetdag-Caucasus megafault separates the alpine orogen and the Eurasian plate. The Caucasian-Arabian segment consists of 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

This paper is the result of collaborative discussions within the framework of IGCP project 667. The project is funded by IUGS via UNESCO and we thank these organizations for their support. We also thank the CGMW for the support provided by their staff and for the organization of meetings. The RMCA is acknowledged for the organization of the 2019 annual meeting in Belgium, Romain Bousquet (Kiel University) for his precious help and his many suggestions for the figures, Solen Le Gardien (SGF) for her help with the references and Benjamin Sautter for his benevolent proofreading. The IGCP667 team is a large group and the name of all the participants can be found with the following link: CGMW/IGCP667

(https://ccgm.org/en/content/11-team). We thank the reviewers for their comments, which improved the content and clarity of the paper greatly. This is Earth Sciences Sector contribution #xxxx.

9. Captions

595 ***Table 1. Relevant geodynamic components and attributes by orogenic zones***

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-alcaline)	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)-sedimentatry terrigenous rocks	Molassic deposit

596

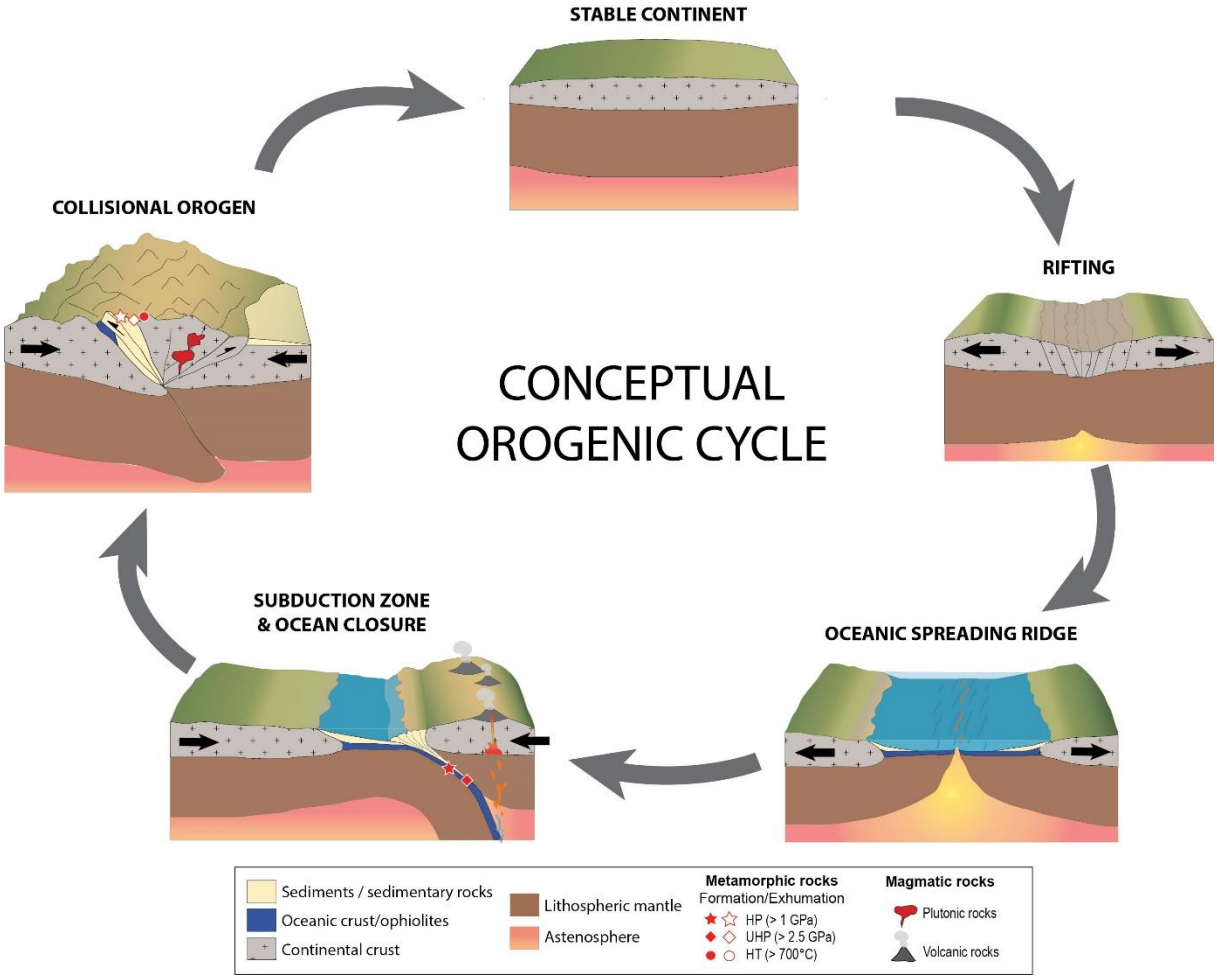
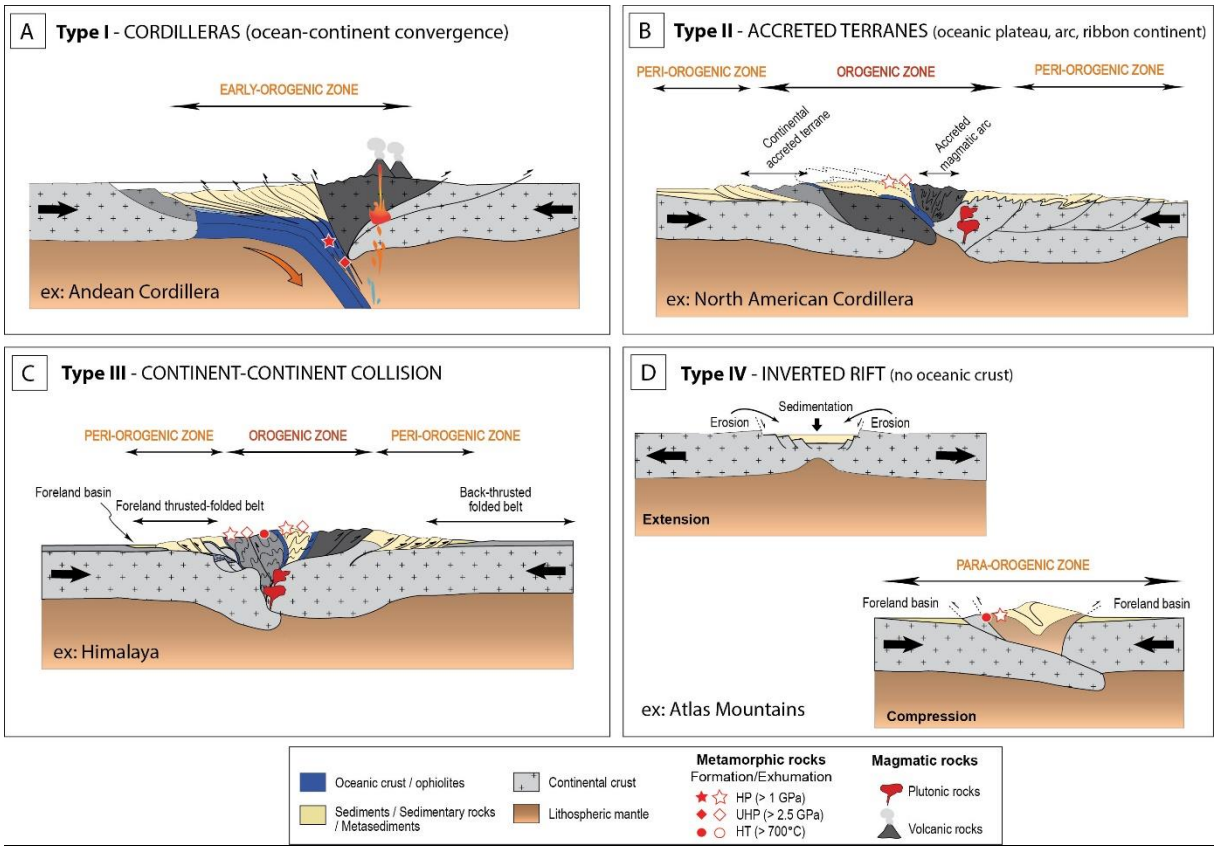


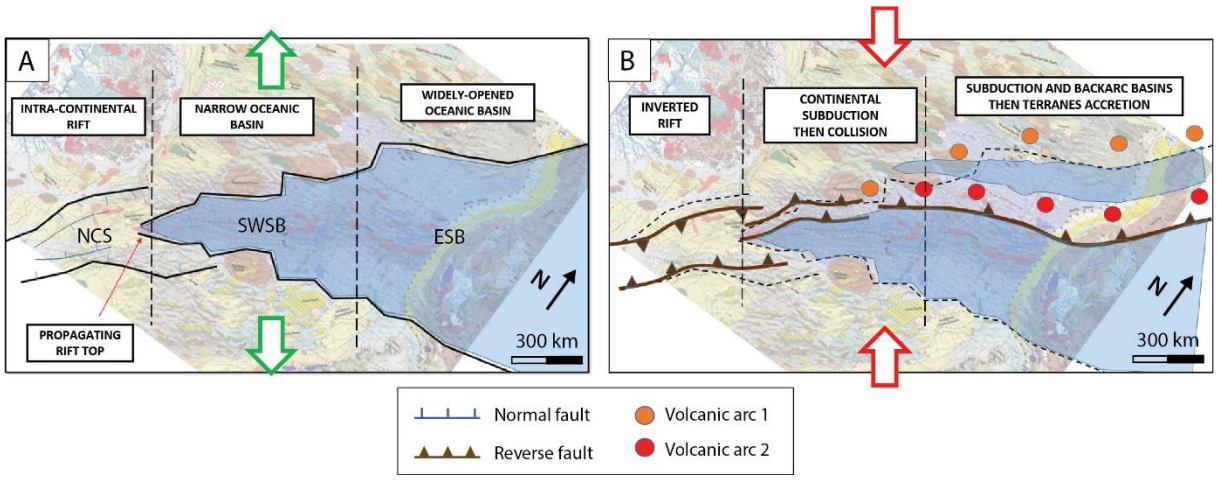
Fig.1. Cyclic view of the different stages of a theoretical orogenic cycle as an extension-compression continuum (Wilson Cycle). The uplift, erosion and unroofing processes including exhumation of metamorphosed rocks and crustal melting occur in the transition between the collisional orogen stage and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



604

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

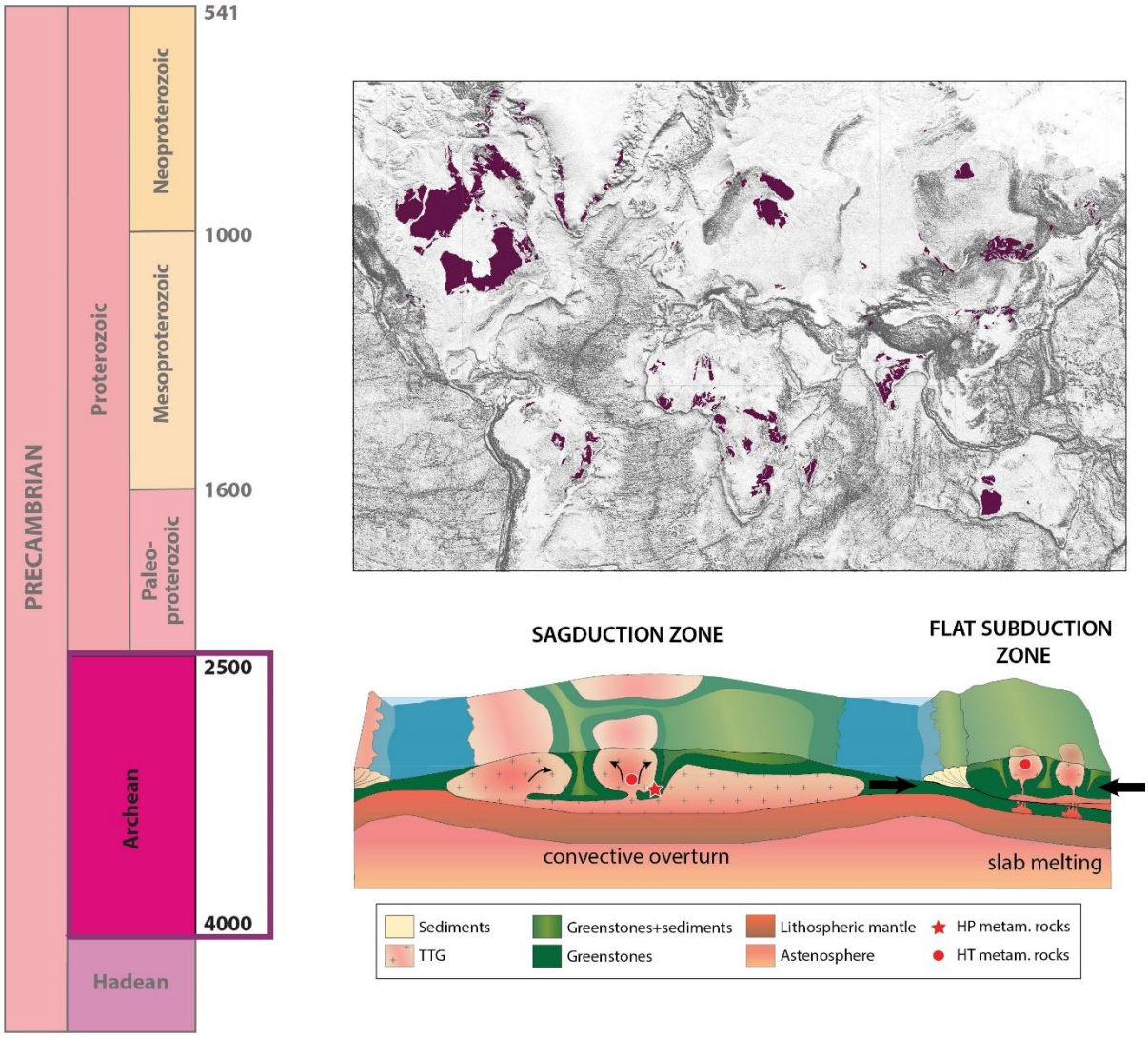
608



609

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

612 Con Son Basin, SWSB; SW sub-basin, ESB; Eastern Sub-basin. Figure (b) is a speculative
613 scenario for future basin closure.
614



615

616 **Fig.4.** Extent of Archean crust on the World Map and representative Archean dome and keel
617 structures formed by geodynamical processes that ended at about 2.5 Ga: flat subduction and
618 sagduction of dense greenstone lithologies into lighter felsic crystalline basement. The higher
619 mantle heat flux is envisioned as leading to the partial melting of the lower crust (flat
620 subduction) and the convective overturn of the Archean crust (sagduction).
621

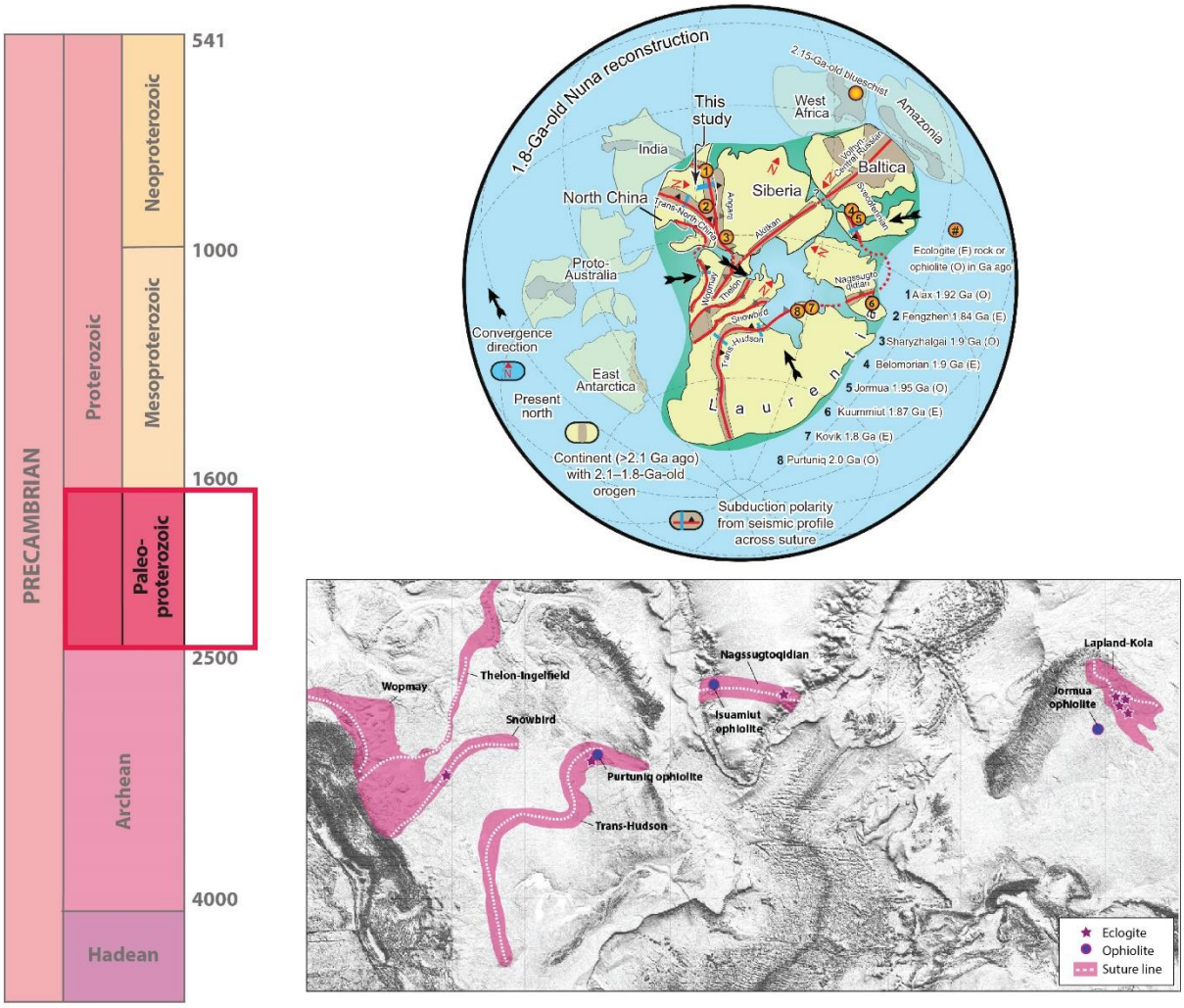


Fig.5. Present extent of some of the Paleoproterozoic orogens (may include reworked Archean crust) in the North Atlantic realm on the World Map, and in the past as shown on a Nuna reconstruction ~1.8 Ga ago (*Evans and Mitchell, 2011*), including the position of North China (*Wan et al., 2020*). Convergence 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*).

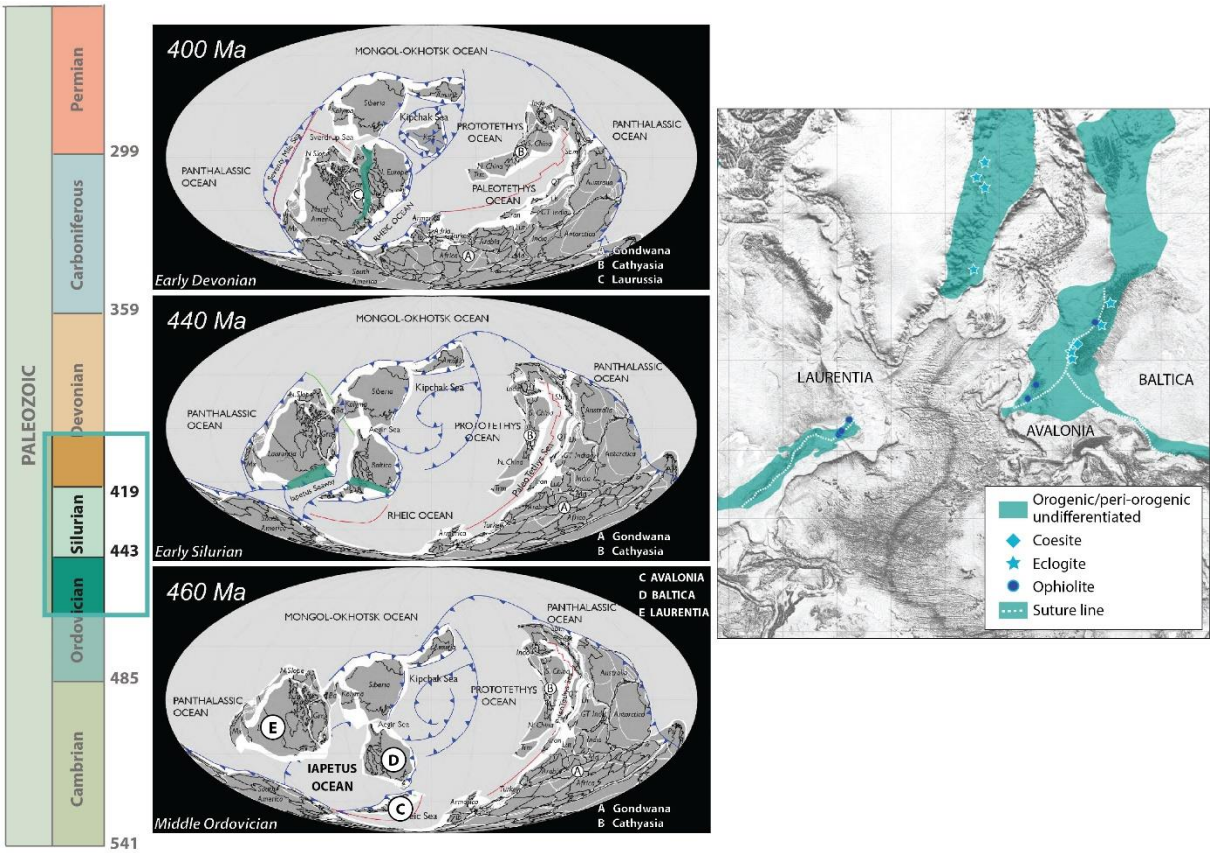
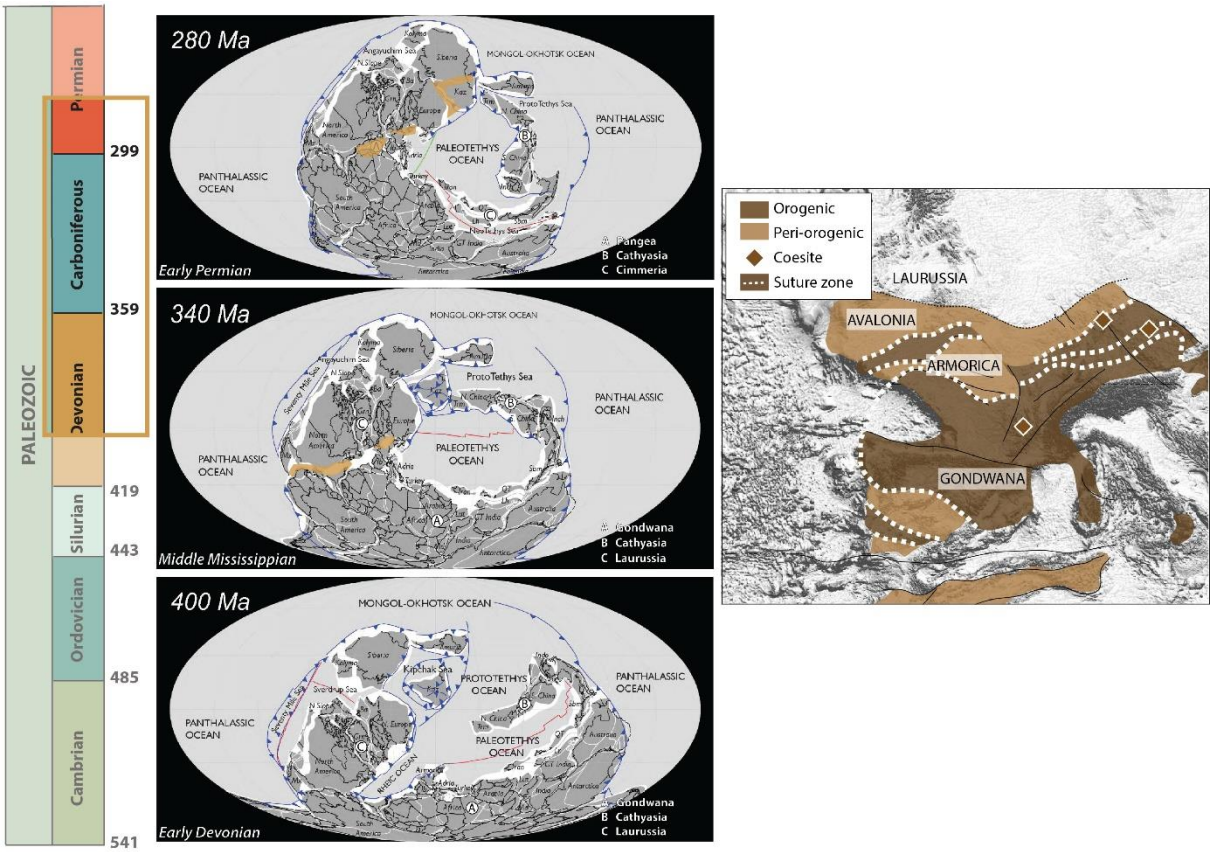


Fig.6. Present location of the Caledonian orogen resulting from the progressive closure of the Iapetus Ocean shown on the World Map and in the past as shown on paleogeographic reconstructions from Scotese, *PALEOMAP Project (www.scotese.com).*



643

644 **Fig.7.** Present location of the Variscan orogens shown on the World Map and in the past as

645 shown on paleogeographic reconstructions from Scotese, *PALEOMAP Project*

646 *(www.scotese.com).*

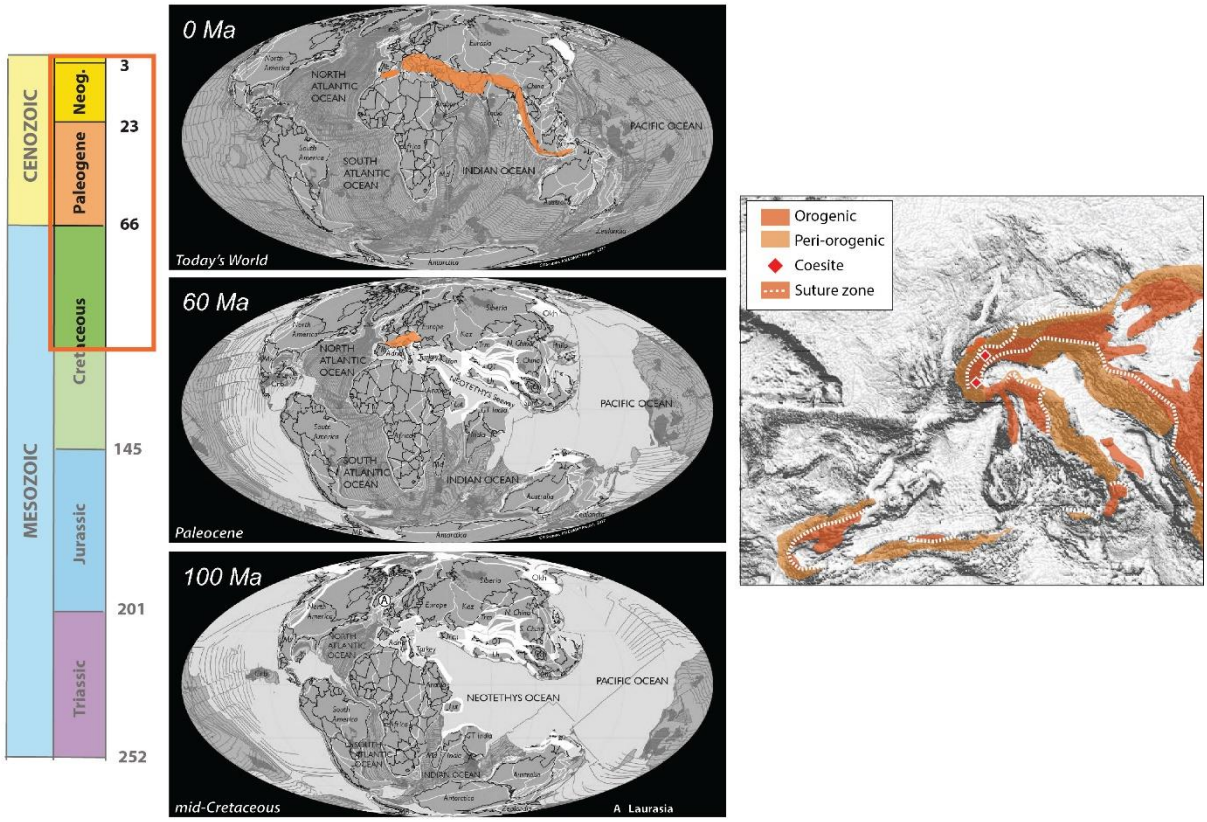


Fig.8. (a) Present location of the Alpine orogens (following closure of the Neo-Tethys and Alpine oceans) shown on the World Map and in the past as shown on paleogeographic reconstructions from Scotese, *PALEOMAP Project (www.scotese.com).*

10. References

Abbott, D. H., Hoffman, S. E., 1984. Archaean plate tectonics revisited 1. Heat flow, spreading rate, and the age of subducting oceanic lithosphere and their effects on the origin and evolution of continents. Tectonics 3 (4), 429–448.

Abbott, D., Burgess, L., Longhi, J., Smith, W. H. F. 1994. An empirical thermal history of the Earth's upper mantle. J. Geophys. Res. 99 (B7), 13835–13850.

Abu Sharib, A.S.A.A., Maurice, A.E., Abd El-Rahman, Y.M., Sanislav, I.V., Schulz, B., Bakhit, B.R., 2019. Neoproterozoic arc sedimentation, metamorphism and collision: Evidence from the northern tip of the Arabian-Nubian Shield and implication for the terminal collision between East and West Gondwana. Gondwana Research, 66, 13–42.

Arndt, N. T., Nisbet, E. G., 1982. Komatiites. Taylor and Francis.

Baldwin, J. A., Bowring, S. A., & Williams, M. L., 2003. Petrological and geochronological constraints on high pressure, high temperature metamorphism in the Snowbird tectonic zone, Canada. Journal of Metamorphic Geology, 21(1), 81-98.

Baldwin, J. A., Bowring, S. A., Williams, M. L. and Williams, I. S. 2004. Eclogites of the Snowbird tectonic zone: petrological and U-Pb geochronological evidence for Paleoproterozoic

high-pressure metamorphism in the western Canadian Shield. Contrib. Mineral. Petrol. 147(5), 528–548).

Ballèvre, M., Bosse, V., Ducassou, C., Pitra, P., 2009. Palaeozoic history of the Armorican Massif: Models for the tectonic evolution of the suture zones. C. R. Geoscience 341, 174-201.
Barrier, E. and Vrielynck, B. 2008. Atlas MEBE - Palaeotectonic Maps of the Middle East (Middle East Basins Evolution Programme). CCGM.org

Bastow, I.D., Thompson, D.A., Wookey, J., Kendall, J-M., Helffrich, G., Snyder, D.B., Eaton, D.W., and Darbyshire, F.A., 2011. Precambrian plate tectonics: seismic evidence from northern Hudson Bay, Canada. Geology. 39, 91–94.

Beauchamp W., Allmendinger R. W., Barazangi M., Demnati A., El Alji M., Dahmani M. 1999. Inversion tectonic and the evolution of the High Atlas mountains, Morocco, based on a geological-geophysical transect. Tectonics 18, 163-184.

Beaumont, C, Kamp, P. J. J., Hamilton J. and Fullsack P. 1996. The continental collision zone, South Island, New Zealand. Comparison of geodynamical models and observations. J. Geophys. Res. 101, B2, 3333-3359. doi.org/10.1029/95JB02401

Bédard JH, Harris LB, Thurston PC., 2013. The hunting of the snArc. Precambrian Res. 229:20–48

Bédard, JH., 2018. Stagnant lids and mantle overturns: implications for Archaean tectonics, magmagenesis, crustal growth, mantle evolution, and the start of plate tectonics. Geosci. Front. 9:19–49

Block: key observations and controversies. Gondwana Res. 23, 1273–1305.
Boniface, N., Schenk, V. and Appel, P. 2012. Paleoproterozoic eclogites of MORB-type chemistry and three Proterozoic orogenic cycles in the Ubendian Belt (Tanzania): Evidence from monazite and zircon geochronology, and geochemistry. Precambr. Res. 192, 16–33.

Bousquet R, Goffé. B, Wiederkehr M, Koller F, Schmid SM, Schuster R, Engi M, Martinotti G. 2008. Metamorphism of metasediments at the scale of an orogen: a key to the Tertiary geodynamic evolution of the Alps. Geol Soc Lond Spec Publ 298:393–411

Briais, A., Patriat, P. and Tapponnier, P., 1993. Updated interpretation of magnetic anomalies and seafloor spreading stages in the South China Sea: Implications for the Tertiary tectonics of Southeast Asia. Journal of Geophysical Research: Solid Earth, 98(B4), 6299-6328.

Brown, D. A., Tamblyn, R., Hand, M., & Morrissey, L. J., 2020. Thermobarometric constraints on burial and exhumation of 2-billion-year-old eclogites and their metapelitic hosts. Precambrian Research, 347, 105833.

Brown, M. 2008. Characteristic thermal regimes of plate tectonics and their metamorphic imprint throughout Earth history: When did Earth first adopt a plate tectonics mode of behavior? When Did Plate Tectonics Begin on Planet Earth? 440, 97.

Brown, M. and Johnson, T. E. 2018. Invited Centennial Article: Secular change in metamorphism and the onset of global plate tectonics. Am. Mineral. 103, 181-196.

Bulois, C., Pubellier, M., Chamot-Rooke, N., Delescluse, M. 2018. Successive rifting events in marginal basins: the example of the Coral Sea region (Papua New Guinea), 2017. Tectonics, Volume 37, Issue 1, 3-29. DOI10.1002/2017TC004783

Charvet J., Shu LS. Faure M., Choulet, F., Wang, B., Le Breton N., 2010. Structural development of the Lower paleozoic belt of South China : genesis of an intracontinental orogen. J. of Asian Earth Sci., 39, 309-330.

Chen K, Rudnick RL, Wang Z, Tang M, Gaschnig RM, et al., 2019. How mafic was the Archean upper continental crust? Insights from Cu and Ag in ancient glacial diamictites. Geochim. Cosmochim. Acta. https://doi.org/10.1016/j.gca.2019.08.002

Chen, L., Huang, B., Yi, Z. , Zhao, J., Yan, Y., 2013. Paleomagnetism of ca. 1.35 Ga sills in northern North China Craton and implications for paleogeographic reconstruction of the Mesoproterozoic supercontinent. Precambrian Res. 228, 36–47.

Choulet F., Faure M., Cluzel, Chen Y., Lin W., Wang B., Jahn B-M. 2012. Architecture and Evolution of Accretionary Orogens in the Altaids Collage: The Early Paleozoic West Junggar (NW China). American Journal of Science, 312, 1098-1145.

Clift, P., Lin, J., and Barckhausen, U. 2002, Evidence of low flexural rigidity and low viscosity lower continental crust during continental break-up in the South China Sea: Marine and Petroleum Geology, v. 19, no. 8, p. 951-970.

Cocks, L. R. M, and Torsvik, T. H. 2006. European geography in a global context from the Vendian to the end of the Palaeozoic. In Gee, D. G. and Stphenson, R. A. (eds). European Lithosphere Dynamics. Geological Society, London, Memoirs, 32, 83–95.

Collins, A. S., Reddy, S. M., Buchan, C. and Mruma, A. 2004. Temporal constraints on Palaeoproterozoic eclogite formation and exhumation (Usagaran Orogen, Tanzania). Earth Planet. Sci. Lett. 224(1), 175-192.

Collins, W. J., Van Kranendonk, M. J., Teyssier, C., 1998. Partial convective overturn of Archaean crust in the east Pilbara Craton, Western Australia: driving mechanisms and tectonic implications. J. Struct. Geol. 20 (9), 1405-1424.

Coltice, N., Bertrand, H., Rey, P., Jourdan, F., Phillips, B. R., Ricard, Y., 2009. Global warming of the mantle beneath continents back to the Archaean. Gondwana Res. 15 (3), 254-266.

Coney, P., Jones, D. and Monger, J. 1980. Cordilleran suspect terranes. Nature 288, 329-333 https://doi.org/10.1038/288329a0

Connelly, J.N., Thrane, K., Krawiec, A.W., and Garde, A.A., 2006. Linking the Palaeoproterozoic Nagssugtoqidian and Rinkian orogens through the Disko Bugt region of West Greenland. Journal of the Geological Society, London. 163, 319–335.

Corrigan, D., 2012. Paleoproterozoic crustal evolution and tectonic processes: Insight from the LITHOPROBE program in the Trans-Hudson orogeny, Canada: in Percival, J.A., Cook, F., Clowes, R., eds., Tectonic Styles in Canada: The LITHOPROBE Perspective. Geological Association of Canada, Special Paper 49, 237–284.

Corrigan, D., Pehrsson, S., Wodicka, N., and de Kemp, E., 2009. The Palaeoproterozoic Trans-Hudson orogen: a prototype of modern accretionary processes. Geological Society, London, Special Publication 327, 457–479.

Courtillot, V., 1982, Propagating rifts and continental breakup Tectonics, Volume 1, Issue 3. Doi: 10.1029/TC001i003p00239

d'Acremont, E., Leroy, S., Maia, M., Patriat, P., Beslier, M. O., et al. 2006. Structure and evolution of the eastern Gulf of Aden: insights from magnetic and gravity data (Encens-Sheba

MD117 cruise). Geophysical Journal International, Oxford University Press (OUP), 165 (3), pp.786-803. doi:10.1111/j.1365-246X.2006.02950.x

Daly, J.S., Balagansky. V.V., Timmerman, M.J., and Whitehouse, M.J., 2006. The Lapland-Kola orogen: Palaeoproterozoic collision and accretion of the northern Fennoscandian lithosphere. Geological Society, London, Memoirs, 32, 579–598.

Dana, J.D, 1873. On some results of the Earth's contraction from cooling, including a discussion of the origin of mountains and the nature of the Earth's interior. Part I Am. Jour. Sci., ser 3, v. 5, art 46. 423-443.

de Oliveira Chaves, A., and Porcher, C.C., 2020. Petrology, geochemistry and Sm-Nd systematics of the Paleoproterozoic Itaguara retroeclogite from São Francisco/Congo Craton: One of the oldest records of the modern-style plate tectonics. Gondwana Research, 87, 224–237.

de Oliveira Chaves, A., and Porcher, C. C. 2020. Petrology, geochemistry and Sm-Nd systematics of the Paleoproterozoic Itaguara retroeclogite from São Francisco/Congo Craton: One of the oldest records of the modern-style plate tectonics. Gondwana Research, 87, 224-237.

Ding, W. and Li, J., 2016. Propagated rifting in the Southwest Sub-basin, South China Sea: Insights from analogue modelling. Journal of Geodynamics, 100, pp.71-86.

Dumond, G., 2020. Tibetan dichotomy exposed in the Canadian Shield: A lower crustal perspective. Earth and Planetary Science Letters, 544, 116375.

Eaton, D.W., and Darbyshire, F., 2010. Lithospheric architecture and tectonic evolution of the Hudson Bay region. Tectonophysics, 480, 1–22.

Elie de Beaumont, M.L., 1852. Notice sur les systèmes des Montagnes. 3 vol. Paris. 1543 pp.

Evans, D. A. D., Mitchell, R. N., 2011. Assembly and breakup of the core of Paleoproterozoic–Mesoproterozoic supercontinent Nuna. Geology 39, 443–446.
Evans, D.A.D., and Mitchell, R.N., 2011. Assembly and breakup of the core of Paleoproterozoic-Mesoproterozoic supercontinent Nuna. Geology, 39, 443–446.

Evans, D.A.D., Veselovsky, R.V., Petrov, P.Y, Shatsillo, A.V., and Pavlov, V.E., 2016. Paleomagnetism of Mesoproterozoic margins of the Anabar Shield: A hypothesized billion-year partnership of Siberia and northern Laurentia. Precambrian Research, 281, 639–655.

Faill, R. T. (1997). A geologic history of the north-central Appalachians; Part 1, Orogenesis from the Mesoproterozoic through the Taconic Orogeny. American Journal of Science, 297(6), 551-619.

Faure M., Bé Mézème E., Cocherie A., Rossi P., Chemenda A., Boutelier D. 2008, Devonian geodynamic evolution of the Variscan Belt, insights from the French Massif Central and Massif Armoricain, Tectonics, 27, TC2008, http://dx.doi.org/10.1029/2007TC002115.

Faure M., Be Mézème E., Duguet M., Cartier C., Talbot J-Y. 2005. Paleozoic tectonic evolution of medio-Europa from the example of the French Massif Central and Massif Armoricain. In: R. Carosi et al. (eds). Journal of the Virtual Explorer, ISSN 1441-8142, Volume 19, Paper 5. doi:10.3809/jvirtex. 2005. 00120

Faure M., Shu LS., Wang,B., Charvet J., Choulet F., Monié P. 2009. Intracontinental subduction: a possible mechanism for the Early Palaeozoic Orogen of SE China. Terra Nova, 21, 360-368.

Faure, M., Nguyen V.V., Hoai, L.T.T., Lepvrier, C., 2018. Early Paleozoic or Early-Middle Triassic collision between the South China and Indochina Blocks: the controversy resolved? Structural insights from the Kon Tum massif (Central Vietnam). J.A.E.S. 166, 162-180. doi.org/10.1016/j.jseaes.2018.07.015.

Fournier, M., Chamot-Rooke, N., Petit, C., Huchon P., Al-Kathiri, A., Audin, L., Beslier, M.-O., d'Acremont, E., Fabbri, O., Fleury, J.-M., Khanbari, K., Lepvrier, C., Leroy, S., Maillot, B., and Merkouriev, S. 2010, Arabia-Somalia plate kinematics, evolution of the Aden-Owen-Carlsberg triple junction, and opening of the Gulf of Aden, J. Geophys. Res., 115, B04102, DOI: 10.1029/2008JB006257

François, C., Debaille, V., Paquette, J. L., Baudet, D., & Javaux, E. J., 2018. The earliest evidence for modern-style plate tectonics recorded by HP–LT metamorphism in the Paleoproterozoic of the Democratic Republic of the Congo. Scientific reports, 8(1), 1-10.

François, C., Philippot, P., Rey, P. and Rubatto, D. 2014. Burial and exhumation during Archean sagduction in the east Pilbara granite greenstone terrane. Earth Planet. Sci. Lett. 396, 235–251.

Frizon de Lamotte, D., Saint Bézar, B., Bracène, R., Mercier, E. 2000. The two main steps of the Atlas building and geodynamics of the western Mediterranean, Tectonics 19,740-761.

Ganne, J., De Andrade, V., Weinberg, R. F., Vidal, O., Dubacq, B., Kagambega, N., Naba, S., Baratoux, L., Jessell, M., Allibon, J., 2012. Modern-style plate subduction preserved in the Palaeoproterozoic West African craton. Nat. Geosci. 5, 60–65.

Gapais D, Cagnard F, Gueydan F, Barbey P, Ballèvre M., 2009. Mountain building and exhumation processes through time: inferences from nature and models. Terra Nova 21:188–94

Garde, A.A.,and Hollis, J.A., 2010. A buried Palaeoproterozoic spreading ridge in the northern Nagssugtoqidian orogen, West Greenland. Geological Society, London, Special Publications, 338, 213–234.

Grabau, A., 1924. Stratigraphy of China, part I, Paleozoic and older, Geol. Surv. of Agric. and Commerce. Peking.

Gressly, A. 1840. Observations géologiques sur le Jura Soleurois: Allgem. Schweiz. Ges. f.d. ges. Naturw., Neue Denkschr, 4, l l5-241.

Handy, M., R., Schmid, S. M, Bousquet, R., Kissling, E., Bernoulli, D. 2010. Reconciling plate-tectonic reconstructions of Alpine Tethys with the geological–geophysical record of spreading and subduction in the Alps. Earth-Science Reviews 102, 121-158. doi: 10.1016/j.earscirev.2010.06.002.

Handy, M., R., Ustaszewski, K., and Kissling, E. 2014. Reconstructing the Alps–Carpathians-Dinarides as a key to understanding switches in subduction polarity, slab gaps and surface motion. Int J Earth Sci (Geol Rundsch), 104 (1), 1-26. doi; 10.1007/s00531-014-1060-3.

Hernández-Montenegro, J. D., Palin, R. M., Zuluaga, C. A., & Hernández-Uribe, D. 2021. Archean continental crust formed by magma hybridization and voluminous partial melting. Scientific Reports, 11, Art-5263.

Hoffman, P.F., 1988. United Plates of America, the birth of a craton: Early Proterozoic assembly and growth of Laurentia. Annual Review of Earth and Planetary Sciences, 16, 543–603.

Hoffman, P.F., 1997. Tectonic genealogy of North America, in van der Pluijm, B.A., Marshak, S., eds., Earth Structure. An Introduction to Structural Geology and Tectonics: McGraw-Hill, New York, 459–464.

Howell, David G., 1989, Tectonics of Suspect Terranes: Mountain Building and Continental Growth, Eds : Chapman and Hall, 1989, ISBN 10 : 0412303701 ISBN 13 : 9780412303708Hutchison, C. 2005. Geology of North-West Borneo: Sarawak, Brunei and Sabah: Elsevier, Amsterdam, 448 pp

Jammes, S., Manatschal, G., Lavier, L., Masini, E. 2009 Tectonosedimentary evolution related to extreme crustal thinning ahead of a propagating ocean: Example of the western Pyrenees, Tectonics, Vol 28, 4, 2009

Jaupart, C., Labrosse, S., Mareschal, J. C. 2007. Temperatures, heat and energy in the mantle of the Earth. In: Treatise on Geophysics, vol. 7, pp. 253-303.

Kearey, P. 1993. Orogeny in: Kearey, P. The Encyclopedia of the Solid Earth, p 443-444. Blackwell, 713 pp.

Kolb, J., 2014. Structure of the Palaeoproterozoic Nagssugtoqidian Orogen,South-East Greenland: Model for the tectonic evolution. Precambrian Research, 255, 809–822.

Korja, A., Lahtinen, R., and Nironen, M., 2006. The Svecofennian orogen: a collage of microcontinents and island arcs. Geological Society, London, Memoirs, 32, 561–578.

Lagabrielle, Y., and Bodinier, J. L. 2008. Submarine reworking of exhumed subcontinental mantle rocks: field evidence from the Lherz peridotites, French Pyrenees. Terra Nova, 20(1), 11-21.

Lahtinen, R., Garde, A.A., and Melezhik, V.A., 2008. Paleoproterozoic evolution of Fennoscandia and Greenland. Episodes, 31, 1–9.

Lahtinen, R., and Huhma, H. 2019. A revised geodynamic model for the Lapland-Kola orogen. Precambrian Research, 330, 1-19.

Lardeaux JM., Schulmann K., Faure M., Janousek V., Lexa O., Skrzypek E., Edel J-B., Stipska P., 2014. The Moldanubian Zone in French Massif Central, Vosges/Schwarzwald and Bohemian Massif revisited: Differences and similarities. In « The Variscan Orogeny: Extent, Timescale and the Formation of the European Crust ». Schulmann, K. et al. (eds). Geological Society, London, Special Publications, 405, http://dx.doi.org/10.1144/ SP405.14

Lesley, J. P. cited in Chamberlin, T. C., & Salisbury, R. D., 1906. Geology, Vol. II. Chicago, p.125.

Lewry, J., and Stauffer, M., 1990. The Early Proterozoic Trans-Hudson Orogen of North America. Geological Association of Canada, Special Paper, 37.

Li, L., Clift, P. D., Stephenson, R., and Nguyen, H. T., 2014. Non-Uniform Hyper-Extension in Advance of Seafloor Spreading on the Vietnam Continental Margin and the SW South China Sea: Basin Research, no. 26, p. 106-134. doi: 10.1111/bre.12045.

Li, Z. X., Li, X. H., Wartho, J. A., Clark, C., Li, W. X., Zhang, C. L., & Bao, C. (2010). Magmatic and metamorphic events during the early Paleozoic Wuyi–Yunkai orogeny, southeastern South China: New age constraints and pressure- temperature conditions. Geological Society of America Bulletin, 122, 772–793.

Loose, D. and Schenk, V. 2018. 2.09 Ga old eclogites in the Eburnian-Transamazonian orogen of southern Cameroon: Significance for Palaeoproterozoic plate tectonics. Precambr. Res. 304, 1-11.

Macgregor, A. M. 1951. "Some milestones in the Precambrian of Southern Rhodesia." The Transactions and Proceedings of the Geological Society of South Africa 54: 39-50.

Martínez Catalán, J.R., Arenas, R., Abati, J., Sánchez Martínez, S., Díaz García, F., Fernández Suárez, J., González Cuadra, P., Castiñeiras, P., Gómez Barreiro, J., Díez Montes, A., González Clavijo, E., Rubio Pascual, F.J., Andonaegui, P., Jeffries, T.E., Alcock, J.E., Díez Fernández, R., López Carmona, A., 2009. A rootless suture and the loss of the roots of a mountain chain: the Variscan belt of NW Iberia. C.R. Geosci. 341, 114–126.

Mattauer, M., Tapponier, P., Proust, F. 1977. Sur les mécanismes de formation des chaînes intracontinentales. L'exemple des chaînes atlasiques du Maroc., Bull. Soc. Géol. Fr. (7) 19, 521-526.

Matte, P., 1986. Tectonics and plate tectonics model for the Variscan belt of Europe, Tectonophysics, 126, 329-374.

Matte, P., 2001. The Variscan collage and orogeny (480-290 Ma) and the tectonic definition of the Armorica microplate: a review. Terra Nova 13, 122–128.

Meert, J.G., 2002, Paleomagnetic evidence for a Paleo-Mesoproterozoic supercontinent Columbia. Gondwana Research, 5, 207–215.

Meert, J.G., 2012. What's in a name? The Columbia (Palaeopangea/Nuna) Supercontinent. Gondwana Research, 21, 987–993.

Meert, J.G., 2014, Strange attractors, spiritual interlopers and lonely wanderers: the search for pre-Pangæan supercontinents. Geoscience Frontiers, 5, 155–166.

Meert, J.G., and Santosh, M., 2017. The Columbia supercontinent revisited. Gondwana Research, 50, 67–83.

Michard, A., Soulaimani, A., Hoepffner, C., Ouanaimi, H., Baidder, L., Rjimati, E.C., Saddiqi, O. 2020. The South-Western Branch of the Variscan Belt: Evidence from Morocco. Tectonophysics, 492, 1–24.

Missenard Y., Zeyen H., Frizon de Lamotte D., Leturmy P., Petit C., Sébrier M., Saddiqi O. 2006. Crustal versus asthenospheric origin of the relief of the Atlas mountains of Morocco, J. Geophys. Res. 111 (B03401) doi:10.1029/2005JB003708

Möller, A., Appel, P., Mezger, K. and Schenk, V. 1995. Evidence for a 2 Ga subduction zone: eclogites in the Usagaran belt of Tanzania. Geology 23(12), 1067-1070.

Moyen, J.-F., Stevens, G. and Kisters, A. 2006. "Record of mid-Archaean subduction from metamorphism in the Barberton terrain, South Africa." Nature 442(7102): 559-562.

Müller, S., Dziggel, A., Kolb, J., and Sindern, S., 2018, Mineral textural evolution and PT-path of relict eclogite-facies rocks in the Paleoproterozoic Nagssugtoqidian Orogen, South-East Greenland. Lithos, 296–299, 212–232.

Najman, Y., Jenks, D., Godin, L., Boudagher-FadelI, M., Millar, I., Garzanti, E., Horstwood, M., Braccial, L., 2017. The Tethyan Himalayan detrital record shows that India–Asia terminal collision occurred by 54 Ma in the Western Himalaya. Earth and Planetary Science Letters, 459, 301–310.

Nance, R., D., Gutiérrez-Alonso, G. Keppie, J., D., Linnemann, U., Murphy, J., B. Quesada, C., Strachan, R., A. and Woodcock, N., A. 2010. Evolution of the Rheic Ocean. Gondwana Research, 17, 194-222.

Nutman, A. Phillip., Kalsbeek, F. & Friend, C.RL., 2008b. The Nagssugtoqidian orogen in South-East Greenland: evidence for Paleoproterozoic collision and plate assembly. American Journal of Science, 308, 529–572.

Nutman, A.P., Dawes, P.R., Kalsbeek, F., and Hamilton, M.A., 2008a. Palaeoproterozoic and Archaean gneiss complexes in northern Greenland: Palaeoproterozoic terrane assembly in the High Arctic. Precambrian Research, 161, 419–451.

Palin, R. M., Santosh, M., Cao, W., Li, S. S., Hernández-Uribe, D., & Parsons, A., 2020. Secular metamorphic change and the onset of plate tectonics. Earth-Science Reviews, 103172.

Park, R. G. 2000. Mountain-building (orogenesis) in: The Oxford Companion of the Earth, p 707. Oxford University Press, 1174 pp.

Park, R.G., 1995. Paleoproterozoic Laurentia-Baltic relationships: a view from the Lewisian, in Coward M.P., and Ries, A.C., eds., Early Precambrian processes. Geological Society, London, Special Publication 95, 211–224.

Perchuk, A. L. and Morgunova, A. A. (2014). "Variable P-T paths and HP-UHP metamorphism in a Precambrian terrane, Gridino, Russia: Petrological evidence and geodynamic implications." Gondwana Research 25(2): 614-629.

Pesonen, L.J., Elming, S.-Å., Mertanen, S., Pisarevsky, S.A., D'Agrella-Filho, M.S., Meert, J.G., Schmidt, P.W., Abrahamsen, N., and Bylund, G., 2003. Palaeomagnetic configuration of continents during the Proterozoic. Tectonophysics, 375, 289–324.

Petrov, O., Shokalsky, S., Pospelov, I., Kashubin, S., Morozov, A., Sobolev, N., Petrov, E., Baluev, A., Sokolov, S., Grikurov, G., Vernikovsky, V., St-Onge, M., Harrison, C., Ernst, R., Guarnieri, P., Labrousse, L., Lemmonier, N., Pubellier, M., Piepjohn, K., Smelror, M., Brekke, H., Faleide, J., Stephens, M., Moore, T., Box, S., Grantz, A., and Orndorff, R., 2018. Tectonic Map of the Arctic: CGMW-VSEGEI, doi:10.14683/2018TEMAR10M.

Pin, C. 1990. Variscan oceans: ages, origins and geodynamic implications inferred from geochemical and radiometric data, Tectonophysics, 177, 215-227.

948 Piper, J.D.A., 1976, Paleomagnetic evidence for a Proterozoic supercontinent. Philosophical
949 Transactions of the Royal Society of London, A280, 469–490.

950 Pourteau, A., Candan, O. and Oberhänsli, R. 2010. High-Pressure metasediments in central
951 Turkey: constraints on the Neotethyan closure history. Tectonics, doi:10.1029/2009TC002650.

952 Pubellier, M., Meresse, F. 2013, Phanerozoic Growth of Asia; Geodynamic Processes and
953 Evolution. Journal of Asian Earth Sciences, 72, 118-128, doi: 10.1016/j.jseaes.2012.06.01.

954 Pubellier, M., Savva, D, Sapin, F., Aurelio, M., 2016. Structural Map of the South China Sea;
955 scale 1:3M, Commission for the Geological Map of the World,
956 doi:10.14682/2017STRUCTUSCS.

957 Puchkov, V. N., 2009. The evolution of the Uralian orogen. Geological Society, London,
958 Special Publications, 327(1), 161-195.

959 Robert, B., Domeier, M., and Jakob, J. 2020. Iapetan Oceans: An analog of Tethys? Geology.

960 Robert, C. and Bousquet, R. 2018. Geowissenschaften: Die Dynamik des Systems Erde.
961 Springer-Spektrum, 1016 pp

962 Roman A, Arndt N., 2019. Differentiated Archean oceanic crust: its thermal structure,
963 mechanical stability and a test of the sagduction hypothesis. Geochim. Cosmochim. Acta.
964 https://doi.org/10.1016/j.gca.2019.07.009

965 Schmid S. M., Bernouilli, D, Fügenschuh, B., Mattenco, L., Schefer, S., Schuster, R., Tischler,
966 M. and Ustaszewski, K., 2008. The Alpine-Carpathian-Dinaridic orogenic system; correlation
967 and evolution of tectonic units. Swiss J. Geosci. 101, 139-183. doi: 10.1007/s00015-008-1247-
968 3.

969 Schmid, S. M., Fügenschuh, B., Kounov, A. et al., 2019. Tectonic units of the Alpine collision
970 zone between Eastern Alps and western Turkey, Gondwana Research.
971 https://doi.org/10.1016/j.gr.2019.07.005

972 Scotese, C. R. 2016. PALEOMAP Paleo Atlas for GPlates and the Paleo Data Plotter Program,
973 PALEOMAP Project, http://www.earthbyte.org/paleomap-paleoatlas-for-gplates/

974 Scotese, C. R., 1991. Jurassic and Cretaceous plate tectonic reconstructions. Palaeogeography,
975 Palaeoclimatology, Palaeoecology, 87(1-4), 493-501.

976 Şengör, A. M. C., 1979. Mid-Mesozoic closure of Permo-Triassic Tethys and its implications.
977 Nature 279, 590-593.

978 Sengör, A. M. C., Natal'in, B. A., 1996. Paleotectonics of Asia: fragments of a synthesis. In:
979 Yin, A., Harrison, M. (Eds.), The Tectonic Evolution of Asia. Cambridge University Press,
980 Cambridge, pp. 486-640.

981 Sengör, A. M. C., Natal'in, B. A., Burtman, V. S., 1993. Evolution of the Altaid tectonic collage
982 and Palaeozoic crustal growth in Eurasia. Nature 364, 299-307.

983 Şengör, A. M. C., Yılmaz Y., 1981. Tethyan evolution of Turkey: A Plate Tectonics approach.
984 Tectonophysics 75, 181-241.

985 Sharkov, E., Lebedev, V., Chugaev, A., Zabarinskaya, L., Rodnikov, A., Sergeeva, N. and
986 Safonova, I. 2015. The Caucasian-Arabian segment of the Alpine-Himalayan collisional belt:
987 Geology, volcanism and neotectonics. Geoscience Frontiers, 503-522.

988 Shirey, S. B. and Richardson, S. H. 2011. Start of the Wilson cycle at 3 Ga shown by diamonds
989 from subcontinental mantle. Science 333(6041), 434-436.

Skublov, S. G. et al., 2011. New data on the age of eclogites from the Belomorian mobile belt at Gridino settlement area. In Doklady Earth Sci. 439, 1163-1170.

Slabunov, A. I., Volodichev, O. I., Skublov, S. G. and Berezin, A. V. 2011. Main stages of the formation of paleoproterozoic eclogitized gabbronorite: Evidence from U-Pb (SHRIMP) dating of zircons and study of their genesis. In Doklady Earth Sci. 437, 396 ().

Smithies, R., Champion, D. and Cassidy, K. 2003. "Formation of Earth's early Archaean continental crust." Precambrian Research 127(1): 89-101.

Smoot C. 2012. Tectonic Globaloney: Closing Arguments. Author House. 158 pp.

Snoeyenbos, D. R., Williams, M. L., & Hanmer, S., 1995. Archean high-pressure metamorphism in the western Canadian Shield. European Journal of Mineralogy-Ohne Beihefte, 7(6), 1251-1272.

Stampfli, G. M., & Borel, G. D., 2002. A plate tectonic model for the Paleozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrons. Earth and Planetary Science Letters, 196(1-2), 17-33.

Stampfli, G. M., 2000. Tethyan oceans. Geological Society of London Special Publications 173, 1-23.

Stampfli, G. M., and Borel, G. D., 2004. The TRANSMED Transects in Space and Time: Constraints on the Paleotectonic Evolution of the Mediterranean Domain. In: Cavazza W. et al. The TRANSMED Atlas. Springer, p. 53-80.

Stampfli, G. M., von Raumer, J. and Borel, G., D. 2002.Paleozoic evolution of pre-Variscan terranes: from Gondwana to the Variscan collision. Geological Society of America, Special Paper 364, 263-280.

Stampfli, G., M. and Borel, G., D. 2002. A plate tectonic model for the Paleozoic and Mesozoic constrained by dynamic plate boundaries and restored synthetic oceanic isochrons. Earth and Planetary Science Letters 196, 17-33.

Stille, H. 1920. Die Begriffe Orogenese und Epiorogenese: Deutsch. Ceo. Gesell. Zeitsch, 71, 164-240.

Stille, H. 1940. Einfiihrnng in den Bau Amerikas: Berlin, Borntraeger, 717.

St-Onge, M.R., Scott, D.J., Rayner, N., Sanborn-Barrie, M., Skipton, D.R., Saumur, B.M., Wodicka, N., and Weller, O.M., 2020. Archean and Paleoproterozoic Cratonic Rocks of Baffin Island: In “Baffin Island and the Labrador-Baffin Seaway Geological Synthesis”. (ed.) L. T. Dafoe & N. Bingham-Koslowski. Geological Survey of Canada Bulletin, 608, 01–29.

St-Onge, M.R., Searle, M.P., and Wodicka, N., 2006. Trans-Hudson orogen of North America and Himalaya-Karakoram-Tibetan orogen of Asia: structural and thermal characteristics of the lower and upper plates. Tectonics, 25, 1–22.

St-Onge, M.R., Van Gool, J.A.M., Garde, A.A., and Scott, D.J., 2009. Correlation of Archaean and Palaeoproterozoic units between northeastern Canada and western Greenland: constraining the pre-collisional upper plate accretionary history of the Trans-Hudson orogen. Geological Society, London, Special Publication 318, 193–235.

St-Onge, M.R., Wodicka, N., and Ijewliw, O, 2007. Polymetamorphic evolution of the Trans-Hudson Orogen, Baffin Island, Canada: Integration of petrological, structural and geochronological data. Journal of Petrology, 48, 271–302.

Strachan R.A., and Woodcock N.H., 2021. The Tectonic Pattern of Britain and Ireland. In: Alderton, David; Elias, Scott A. (eds.) Encyclopedia of Geology, 2nd edition. 4, 328-337. United Kingdom: Academic Press.

Suess, E., 1893. Are Great Ocean Depth Permanent ? Natural Science, 2123, 180-187.

Tang M, LeeC-TA, Rudnick RL, Condie KC., 2019. Rapid mantle convection drove massive crustal thickening in the late Archean. Geochim. Cosmochim. Acta. https://doi.org/10.1016/j.gca.2019.03.039

Tardy, M., Blanchet, R., Zimmermann, M. 1989. The Texas and Caltam Lineaments from the American Cordillera to the Mexican Sierra Madres, nature, origin and structural evolution, p. 219-226. In; Open-File Report 385, Geological study of the North American Cordillera, Bull. Centres Rech. Explor.-prod. Elf-Aquitaine, 13, 2., ISSN 0396-2687 – BCREDP

Taylor, B., Goodliffe, A. M. and Martinez, F., 1999. How continents break up: insights from Papua New Guinea. Journal of Geophysical Research: Solid Earth, 104(B4), pp.7497-7512.

Taylor, S. R., McLennan, S. M., 1985. The Continental Crust: Its Composition and Evolution. Blackwell Scientific Publication, Carlton, 312 pp. OSTI ID: 6582885.

Thébaud, N., & Rey, P. F., 2013. Archean gravity-driven tectonics on hot and flooded continents: controls on long-lived mineralised hydrothermal systems away from continental margins. Precambrian Research, 229, 93-104.

Thompson, D.A., Bastow, I.D., Helffrich, G., Kendall, J-M., Wookey, J., Snyder, D.B., and Eaton, D.W., 2010. Precambrian crustal evolution: seismic constraints from the Canadian Shield. Earth and Planetary Science Letters, 297, 655–666.

Thurmann, M. J. 1854. Résumé des lois orographiques générales du système des monts Jura : Soc. Geol. France Bull., 2, 11, 41-57.

Tugend, J., Manatschal G., Kusznir, N. J. Masini E., Mohn, G., et al. 2014. Formation and deformation of hyperextended rift systems: Insights from rift domain mapping in the Bay of Biscay-Pyrenees. Tectonics, American Geophysical Union (AGU), 33, pp.1239-1276. ⟨10.1002/2014TC003529⟩

Upham, W. 1894. Wave-like progress of an epeirogenic uplift: Jour. Geology, 2, 380-395.

Van Gool, J.A.M., Connelly, J.N., Marker, M., and Mengel, F.C., 2002. The Nagssugtoqidian Orogenof West Greenland: Tectonic evolution and regional,correlations from a West Greenland perspective. Canadian Journal of Earth Sciences, 39, 665–686.

Van Kranendonk, M. J., Collins, W. J., Hickman, A., Pawley, M. J., 2004. Critical tests of vertical vs. horizontal tectonic models for the Archaean East Pilbara granite-greenstone terrane, Pilbara craton, Western Australia. Precambrian Res. 131 (3), 173–211.

Volodichev, O. I., Slabunov, A. I., Bibikova, E. V., Konilov, A. N. and Kuzenko, T. I. 2004. Archean eclogites in the Belomorian mobile belt, Baltic Shield. Petrology 12(6), 540-560.

Wan, B., Windley, B. F., Xiao, W., Feng, J., Zhang, J., 2015. Paleoproterozoic high-pressure metamorphism in the northern North China Craton and implications for the Nuna supercontinent. Nat. Commun. 6, 8344.

Wan, B., Yang, X., Tian, X., Yuan, H., Kirscher, U., and Mitchell, R.N., 2020. Seismological evidence for the earliest global subduction network at 2 Ga ago. Science Advances, 6. eabc5491.

Wang, Y., Fan, W., Zhang, G., & Zhang, Y. 2013. Phanerozoic tectonics of the South China Block: key observations and controversies. Gondwana Research, 23(4), 1273-1305.

Weissel, J. K., & Watts, A. B. 1979. Tectonic evolution of the Coral Sea basin. Journal of Geophysical Research: Solid Earth, 84(B9), 4572-4582.

Weller, O.M., and St-Onge, M.R., 2017. Record of modern-style plate tectonics in the Palaeoproterozoic Trans-Hudson orogeny. Nature Geoscience, 10, 305–313.

Weller, O.M., Jackson, S., Miller, W.G.R., St-Onge, M.R., and Rayner, N., 2020. Quantitative elemental mapping of granulite- facies monazite: Textural insights and implications for petrochronology. Journal of Metamorphic Geology, 38, 853–880.

Wilson, R. W., Houseman, G. A., Buiter, S. J. H., McCaffrey, K. J. W. and Doré, A. G. 2019. Fifty years of the Wilson Cycle Concept in Plate Tectonics: an Overview. Geological Society, London, Special Publications. DOI: https://doi.org/10.1144/SP470-2019-58

Xiao, W. J., Han, C. M., Yuan, C., Sun, M., Lin, S. F., Chen, H. L., Li, Z. L., Li, J. L., and Sun, S., 2008. Middle Cambrian to Permian subduction-related accretionary orogenesis of North Xinjiang, NW China: implications for the tectonic evolution of Central Asia. Journal of Asian Earth Sciences, 32, 102-117.

Xin Y., Li J., Ratschbacher L., Zhao G., Zhang Y., Dong S., Xia X-P., Yu Y. 2020. Early Devonian (415 -400 Ma) A-type granitoids and diabases in the Wuyishan, eastern Cathaysia: A signal of crustal extension coeval with the separation of South China from Gondwana. Geol. Soc. Am. Bull.

Xu, C., Kynický, J., Song, W., Tao, R., Lü, Z., Li, Y. and Fei, Y. 2018. Cold deep subduction recorded by remnants of a Paleoproterozoic carbonated slab. Nature communications, 9(1), 2790.

Yao W., Li ZX., Li WX. 2014. Was there a Cambrian ocean in South China? – Insight from detrital provenance analyses. Geol. Mag. 1-8 Cambridge University Press doi:10.1017/S0016756814000338.

Yu, H.L., Zhang, L.F., Wei, C.J., Li, X.L., and Guo, J.H., 2017. Age and P–T conditions of the Gridino-type eclogite in the Belomorian Province, Russia. Journal of Metamorphic Geology, 35, 855–869.

Zhao, G. and Cawood, P., 1999. Tectonothermal evolution of the Mayuan assemblage in the Cathaysia Block: implications for Neoproterozoic collision-related assembly of the South China Craton. Am. J. Sci., 299, 309–339.

Zhao, G., Cawood, P.A., Wilde, S.A., and Sun, M., 2002. Review of global 2.1–1.8 Ga orogens: implications for a pre-Rodinia supercontinent. Earth-Science Reviews, 59, 125–162.

1109 Zhao, G., Li, S., Sun. M., and Wilde, S.A., 2011. Assembly, accretion, and break-up of the
1110 Palaeo-Mesoproterozoic Columbia supercontinent: record in the North China Craton revisited.
1111 International Geology Review, 53, 1331–1356.

1112 Zhao, G., Sun, M., Wilde, S.A., and Li, S., 2004. A Paleo-Mesoproterozoic supercontinent:
1113 assembly, growth and breakup. Earth-Science Reviews, 67, 91–123.