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A CASE OF AMPFERER-TYPE SUBDUCTION AND CONSEQUENCES FOR THE ALPS AND THE PYRENEES

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ABSTRACT. In this contribution, we review aspects of the petrological, metamorphic and sedimentological characteristics of Neotethyan and Pacific subduction zones and compare them to the Western and Central Alps and the Pyrenees. We argue that the current models of formation of the Western and Central Alps, which invoke the subduction of significant volumes of oceanic lithosphere and spontaneous subduction initiation are unable to explain fundamental characteristics unique to the Pyrenees and European Alps orogens. Their characteristic geodynamic features are surprisingly distinct from features of Wadati-Benioff-type subduction, as the latter typically have large subducted oceanic slabs, near-continuous magmatism and specific subduction initiation characteristics. The Pyrenees and the Western to Central Alps however, are characterized by subduction initiation without magmatism at rifted margins and inefficient subduction of hydrated lithologies into the convective upper mantle. The pre-collisional lithosphere from the Pyrenees to the Central Alps, or Western Tethys *sensu lato* comprised several sub-basins characterized by extremely thinned continental crust, exhumed subcontinental mantle and, locally, minor volumes of embryonic ultra-slow spreading ocean crust. This allows us to distinguish Benioff-type oceanic subduction resulting from the efficient subduction of hydrated oceanic lithosphere from Ampferer-type continental subduction. The latter records the closure of hyper-extended continental basins with minor volumes of oceanic crust, and subduction of predominantly dry lithosphere into the convective upper mantle.

Keywords: Ampferer-type subduction, Benioff-type subduction, Alps, Pyrenees, rift basins, amagmatic subduction initiation

INTRODUCTION

The primary driving force of subduction zones and present-day plate tectonic movements is thought to be a slab-pull mechanism whereby oceanic lithosphere, as it ages and moves away from the mid-ocean ridge, becomes colder, thicker and negatively buoyant, thereby dragging the plate down in the convective upper mantle (Forsyth and Uyeda, 1975; Carlson and others, 1983; Cloos, 1993). Subduction is a key mechanism of plate tectonics, driving the formation of new oceanic lithosphere at spreading ridges as well as growth of continental crust at convergent margins as a consequence of the

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recycling of oceanic crust, production of arc magmatism (for example: Gill, 1981; Grove and Kinzler, 1986) and accretion of terranes to the continental margin (Jones and others, 1982). The geological archives at active margins preserve distinct evidence of such dynamic conditions, from the initiation of subduction to the formation of long-lived subduction zones, development of arc magmatism, growth of accretionary orogens, terrane accretion and continental collision (Shervais, 2001; Stern, 2002; Ernst, 2005). Key geological evidence include: 1) the voluminous production of hydrous low-FeO, or "calc-alkaline", magmatism (Arculus, 2003) as a result of volatile fluxing of the mantle-wedge (Gill, 1981; Arculus, 2004); 2) the formation of subduction mélanges within accretionary wedges (Coleman and Lanphere, 1971); 3) high-pressure and low-temperature metamorphism (Miyashiro, 1961; Maekawa and others, 1993) as well as 4) seismic and tomographic imaging of downgoing slabs (Benioff, 1949). Such geological evidence has been used to infer the presence of ancient paleo-subduction zones and to constrain the timing of the initiation of a modern plate tectonic regime (Stern, 2005; Weller and St-Onge, 2017).

The Western to Central Alps have long played a key role in testing and developing new ideas concerning orogenesis and subduction zone processes (compare Trümpy, 2001). In particular, the structural and stratigraphic study of the Alps led to the observation that coherent structural units, or "*nappes*", were transported over long distances as a result of shallow tectonic processes (Schardt, 1893; Lugeon, 1902; Termier, 1904; Argand, 1916; Trümpy, 2001; Schmid and others, 2004; Pfiffner, 2014), also coined 'flake tectonics' (Oxburgh, 1972). Austrian geologist Otto Ampferer presented some of the first conceptual ideas regarding continental drifting and plate tectonics (compare Dullo and Pfaffl, 2019). He showed that the reconstructions of the Alpine orogen implied that a significant amount of basement rocks were missing and had been underthrust to great depth, a process he defined as "*Verschluckung*" in Ampferer and Hammer (1911) and later defined as "*subduction*" by Amstutz (1955). Following the geophysical imaging of oceanic crust below continental lithosphere within the circum-Pacific (*Wadati-Benioff Zone,* Benioff, 1949), the definition of subduction was further separated into a *Benioff-type* subduction (*"B-type")* driven by the subduction of oceanic lithosphere and an *Ampferer-type* subduction (*"A-type")* resulting from the subduction of continental lithosphere (Bally, 1975). This "*A-type*" subduction gradually fell into disuse as it became an accepted paradigm that the subduction of oceanic lithosphere ("*B-type*") was the key underlying mechanism driving plate tectonics and dragging continental fragments to considerable depth during the final stages of continental collision (Stampfli and others, 1998; Rosenbaum and Lister, 2005; Handy and others, 2010; Handy and others, 2014).

These paradigms have underpinned conceptual models of the closure of certain segments of the Western Tethys along the Western to Central Alps. The Western Tethys encompasses the hyper-extended rift basins and embryonic oceans that formed in the Jurassic-Cretaceous between Europe, Adria and Iberia which remnants are now found in the Apennines, Western and Central Alps and Pyrenees. The subduction and closure of the wider segment of the Western Tethys, or Piemonte-Liguria basin/ocean, along the Western to Central Alps has long been thought to result from the foundering of cold oceanic lithosphere along a passive margin prior to continental collision (Stampfli and others, 1998; Handy and others, 2010; Handy and others, 2014; Kissling and Schlunegger, 2018). The Western to Central Alps have thus been viewed as an analogue of Benioff-type oceanic subduction followed by subduction of continental crust and continental collision (Stampfli and Marthaler, 1990). The tectonic and structural evolution of the Alps during continental collision has been extensively studied, leading to a comprehensive understanding of thick-skinned and thin-skinned tectonics and constraints on the kinematics of nappe emplacement (Escher and others, 1993; Pfiffner, 2016; Schmid and others, 2017). However, the significance of the sparse magmatic record, from subduction initiation until continental collision, in the history of the Western Tethys as a whole has been largely set aside. Indeed, most contributions focusing on continental collision, paleogeographic reconstructions or on subduction-related metamorphism (in)directly imply the subduction of a large slab of oceanic lithosphere. Such an assumption is at odds with the narrow width of the Western Tethys and the lack of significant subduction-related magmatism (Stampfli and Marthaler, 1990; Stampfli and others, 1998; Babist and others, 2006; Herwartz and others, 2008; Groppo and others, 2009; Handy and others, 2010; Handy and others, 2014; Weber and others, 2015). However, because of the presumably small width of the Western Tethys some authors have suggested a forced subduction during the closure of the Western Tethys (De Graciansky and others, 2011; Kiss and others, 2019).

The sparse evidence of arc magmatism coeval with oceanic subduction was already highlighted by Stampfli and Marthaler (1990). More recently, McCarthy and others (2018) further illustrated this problem of invoking a classical Benioff-type subduction for the Alpine orogeny. The compilation of detrital zircon ages and ages of magmatism showed that the Alpine orogen is characterized by amagmatic subduction initiation at a passive margin followed by amagmatic subduction until continental collision. Such amagmatic characteristics over 40 to 50 Myr of subduction preclude a simple explanation of inhibition of magmatism through flat-slab subduction, slow- and oblique subduction or 'delays' in magmatism (Stampfli and Marthaler, 1990; Zanchetta and others, 2012; Bergomi and others, 2015). McCarthy and others (2018) suggested that the pre-subduction lithosphere acts as a key control on the mechanisms of subduction initiation, proposing that the Alpine orogen might be an example of Ampferer-type continental subduction. This initial definition implied the subduction of continental crust (Ampferer and Hammer, 1911). We update this term to imply as well the subduction of exhumed subcontinental lithospheric mantle without overlying continental crust. This was previously envisioned by Trümpy (1975) based on the observation that upper Cretaceous flysches lack the debris of volcanic activity. This contribution therefore aims to fill this "subduction gap", by discussing the magmatic timeline from rifting and mantle exhumation to subduction initiation up until continental collision. We aim to highlight the importance of the petrological characteristics of the subducted lithosphere prior to continental collision in the formation of the Alpine orogen.

In addition, similar amagmatic characteristics, pre-collision architecture of rift-basins and accommodation of convergence along passive margins are known from the Pyrenean orogeny as well (Jammes and others, 2009; Lagabrielle and others, 2010; Roca and others, 2011; Tugend and others, 2014). Although the mechanisms of rifting and mantle exhumation in the Pyrenees and in the Western and Central Alps are broadly comparable (Picazo and others, 2016), their similarities are largely ignored when focusing on the mechanisms of convergence. We will therefore present a compilation of available data regarding the latest 300 Myr of geological history of the Western and Central Alps and Pyrenees. We will focus in particular on the magmatism, metamorphism, sedimentation and regional tectonic behavior of the European, Iberian and Adria margins from rifting and oceanization to subduction and collision. We will also discuss key petrological characteristics of Benioff-type subduction zones, primarily with examples along the western Pacific and Neotethyan ophiolites. We show that the Western and Central Alps and the Pyrenees share similar characteristics that set them apart from Neotethyan ophiolites and *Benioff-type* subduction zones. These characteristics include the coherent formation of nappes and amagmatic subduction initiation at passive margins. We argue that subduction models in the Western Tethys

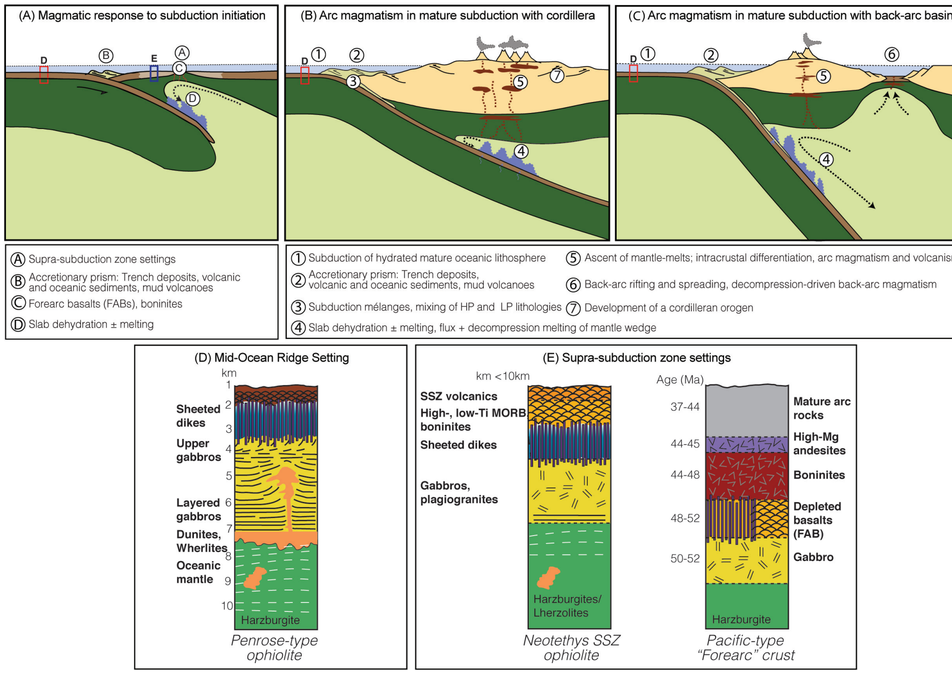


Fig. 1. (A,B,C) Snapshots in time of the main characteristics of Benioff-type subduction zones during subduction initiation (A), and during well-developped stages of subduction, leading to Cordilleran-type orogens at a convergent margin (B) (for example: Central Andes, after Jones and others, 2016) or to upper-plate extension and formation of back-arck basin (C); (D,E) Variations in the nature of oceanic lithosphere from "theoretical" oceanic crust from at fast-spreading mid-ocean ridges (Penrose-type), modified after Nicolas (1995) and oceanic lithosphere formed upon subduction initiation illustrated by simplified logs: Neotethyan ophiolites and Izu-Bonin "forearc" crust, after Maffione and others (2018) and Ishizuka and others (2011a) respectively.

based on Benioff-type ocean closure and subduction of significant volumes of oceanic lithosphere are not able to account for key geological characteristics.

CHARACTERISTICS OF BENIOFF-TYPE SUBDUCTIONS

In this section, we review the geological record that characterizes Benioff-type subduction zones at current active margins (fig. 1). By this term, we imply the subduction of oceanic lithosphere formed at a mid-ocean ridge. This oceanic lithosphere can be formed at a fast-spreading (Penrose-type ophiolite, Anonymous, 1972) or slow spreading ridge (Dick and others, 2003). This type of subduction implies the presence of one or two planar zone(s) of seismicity in the downgoing slab (Wadati-Benioff plane) (Benioff, 1949; Hasegawa and others, 1978) reflecting the active deformation of sinking oceanic lithosphere. Although in part enigmatic, these seismic planar structures are likely controlled by dehydration reactions involving the breakdown of hydrous minerals from the dehydrating oceanic lithosphere (Yamasaki and Seno, 2003; Hacker and others, 2003; Brudzinski and others, 2007). In particular, we will discuss the magmatic response to subduction and changes in subduction conditions, as well as the geophysical and sedimentary characteristics of active subduction zones.

Structure and Geometry of B-type Subduction

Over the last 50 to 60 years, a combination of seismic, velocity and attenuation tomography has allowed the imaging of both current and ancient subducting slabs in

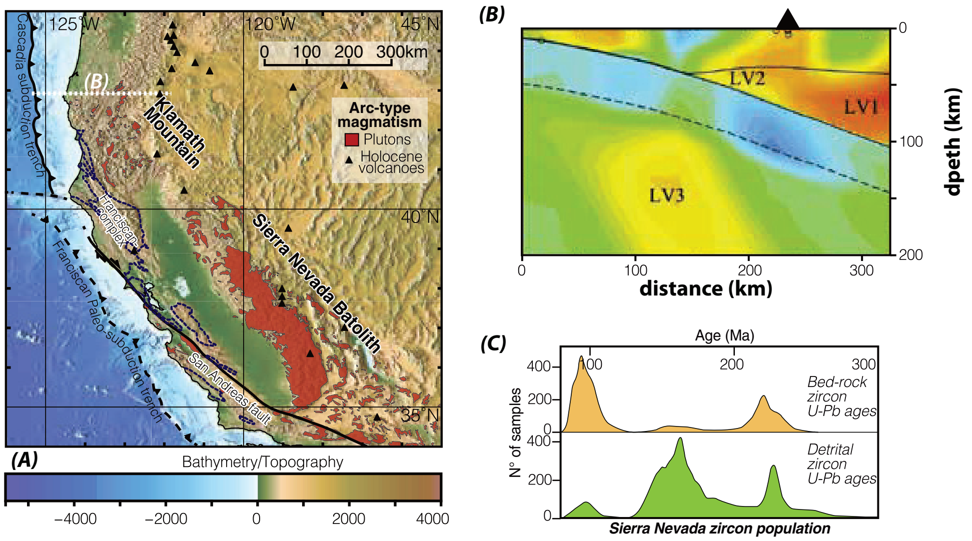
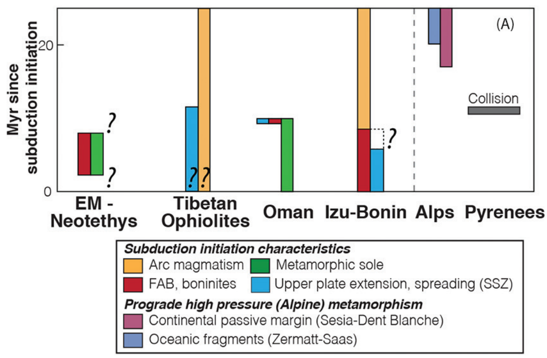


Fig. 2. (A,B) Geophysical imaging of the present-day Cascadian subduction zone and of the fossil accretionary Franciscan subduction complex; tomographic model after Chen and others (2015); black triangles mark the locations of active Holocene volcanoes from the Smithsonian Global Volcanism database. For a broader look at the distribution of arc plutons along the entire West coast of North, Central and South America, the reader is referred to Paterson and others (2011) and Ducea and others (2015); (C) distribution of U-Pb ages of Sierra Nevada detrital and bed-rock zircons, redrawn after Paterson and Ducea (2015), illustrating how magmatism, from the initiation of mesozoic magmatism along the Sierra Nevada batholith until the cessation of magmatism at 80 Ma (Barth and others, 2011; Paterson and Ducea, 2015) is well recorded in detrital zircon ages.

the convective mantle (Spakman and others, 1989; Spakman, 1990; Romanowicz, 1991). Earthquakes located in subduction zones are related at shallow depths to the coupling of overriding and subducting plates whereas at greater depths, earthquakes result from internal stresses and changes in thermal conditions of the downgoing plate, leading to the formation of deep, inclined seismic zones within subducting oceanic lithosphere (Wadati-Benioff zone, Benioff, 1949) (Hacker and others, 2003). This has allowed the identification of key characteristics and depths of subducting oceanic lithosphere in current convergent margins, from the circum-Pacific (Cascadia subduction, fig. 2, Chen and others, 2015) to smaller oceans such as the Tyrrhenian and Aegean subduction zones (Doglioni and others, 1999; Faccenna and others, 2003). In order to decipher aseismic sections of the slab, in particular remnant slabs from past collisions and ocean closures, tomography imaging is used to illustrate changes in seismic velocities within the mantle relative to a reference model (fig. 2). Areas faster than the average may be interpreted as cold oceanic crustal fragments (for example, Lippitsch and others, 2003). A variety of three-dimensional tomographic imaging techniques has established the heterogeneity of the convecting mantle and assigned specific anomalies to particular geological events and the tracking of slabs from ancient subductions (for example, Neotethyan slabs, Farralon/Kula plate, Alpine slab) (Pavlis and others, 2012; Wu and others, 2016; van der Meer and others, 2018).

Magmatism

There are distinct geodynamic environments that will produce magmatism during convergence (fig. 1). These include subduction initiation, mature subduction, back-arc magmatism and subduction of a mid-ocean ridge (figs. 1, 2, and 3). Based on key



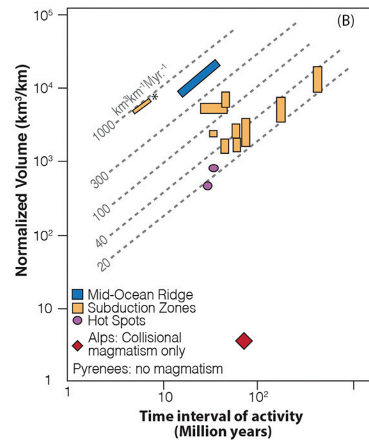


Fig. 3. (A) Time-lapse between subduction initiation and subduction-related magmatism and metamorphism. Subduction initiation for the Alps estimated at *ca.* 85–100 Ma (Zanchetta and others, 2012). Data from the Izu-Bonin come from Ishizuka and others (2011), Reagan and others (2013), and Ishizuka and others (2018). Oman data from Macleod and others (2013), Searle and others (2014), and Rollinson (2015). Age of subduction initiation of Oman is based on Guilmette and others (2018). Tibetan ophiolite forearc hyper-extension and estimations for subduction initiation after Maffione and others (2015b). Neotethys ophiolites after Maffione and others (2017). Prograde high-pressure metamorphism of the southern Piemonte-Liguria basin is related to the subduction of the thinned continental crust fragment of Sesia-Dent Blanche, with ages derived from Rubatto and others (2011) and Regis and others (2014). Prograde high-pressure metamorphism of oceanic fragments (Zermatt-Saas) after Skora and others (2009); **(B)** magma addition rates in subduction zones, Mid-Ocean Ridges, Hot-Spots and Alpine-type orogens. Data for subduction zones, Mid-Ocean Ridges and Hot-Spots from Arculus (2004) and modified with data from Jicha and others (2006) for the Aleutians. *= Estimated magmatism within the Izu-Bonin during the first *ca.* 5–7 Myr of subduction initiation, after Arculus and others (2015). Note that subduction zones include both slow and fast subduction zones (Izu-Bonin, Marianas, Aleutians and Lesser Antilles). Volume of Alpine magmatism is based on the Adamello pluton (2000 km^3 based on Schaltegger and others, 2019), Miagliano pluton (Berger and others, 2012: 20 km^2 and *ca.* 2 km thickness), Valle del Cervo pluton (Kapferer and others, 2012: *ca.* 50 km^2, and estimated pluton thickness: 4 km), Rieserferner pluton (Borsi and others, 1979; Steenken and others, 2000; Cesare and others, 2004: 40 km long, 1–10 km wide and 2.5 km thickness), and Rensen pluton (Krenn and others, 2003) (7 km long, 1 km wide and an unknown thickness which we assume as 2 km thick:). Estimated time interval of subduction activity in the Alps is estimated at 70–80 Ma for the Alps (that is, subduction initiation at *ca.* 85–100 Ma to 20 Ma). Length of active subduction zone for the Western Tethys between the Western and Central Alps is estimated at *ca.* 1000 km (Handy and others, 2014). The timescale of subduction activity for the Pyrenees is estimated to between the Santonien (convergence initiation) and Early Miocene (Aquitanien).

examples, we show that, though suppression of magmatism might occur during subduction, they remain short-lived.

Magmatic response to subduction initiation.—The mechanisms inducing subduction initiation are still poorly understood (McKenzie, 1977; Toth and Gurnis, 1998; Maffione and others, 2015a; Crameri and others, 2018; Stern and Gerya, 2018). In part this is because subduction initiation remains a short-lived, transient event where initial geological products are rapidly obscured by the growth of nascent arcs or destroyed upon subduction. However, a coherent picture of subduction initiation along the Izu-Bonin Mariana arc and Neotethyan ophiolites has been developed as a result of field-based studies of ophiolites and manned submersibles, dredging and coring expeditions in arc systems of the Western Pacific (Stern, 2004; Reagan and others, 2010; Arculus and others, 2015).

Pacific "forearc" crust.—The lithostratigraphy of western Pacific-type early arc crust has been well characterized in recent decades (Ishizuka and others, 2006; Ishizuka and others, 2011a; Stern and others, 2012; Arculus and others, 2015) (fig. 1E). Over >1000 km along the Izu-Bonin-Mariana arc, these studies have identified an oceanic lithosphere with petrological characteristics distinct from oceanic crust formed at an active fast-spreading mid-ocean ridge ("Penrose"-type ophiolitic sequences, Anonymous, 1972) (fig. 1A). In the Bonin Ridge forearc, a lower section of depleted harzburgites is overlain by gabbroic rocks (Ishizuka and others, 2011a). Lava flows, pillow basalts and occasional sheeted dikes of low K-Ti tholeiites, termed forearc basalts (FAB) on the basis of their present-day location, cover these gabbros (48–52 Ma). The youngest FABs overlap in age with, and are overlain by boninites and high-Mg# andesites (44–50 Ma) (Ishizuka and others, 2011a; Reagan and others, 2019). Compositional characteristics of FABS do not show any clear evidence of involvement of slab-derived, fluid-mobile trace elements and isotopic evidence of a slab component, characteristic of typical arc tholeiites (Reagan and others, 2010; Arculus and others, 2015, Yogodzinksi and others, 2018; Hickey-Vargas and others, 2018; Shervais and others, 2019), indicating that they formed from shallow decompression melting (fig. 1A). The signature of a subducting slab is only manifest in subsequent hydrous boninitic magmatism and high-Mg# andesites (Ishizuka and others, 2011a; Hickey-Vargas and others, 2018; Li and others, 2019) indicating the presence of a maturing arc system (figs. 1C and 1E).

The petrological and compositional characteristics as well as emplacement ages of FABs have been identified throughout much of the Izu-Bonin-Mariana arc and remnant back-arcs (Yogodzinski and others, 2018; Hickey-Vargas and others, 2018), implying that a 200 to 300 km wide domain along the IBM-arc was spreading synchronously with subduction initiation (Arculus and others, 2015; Ishizuka and others, 2018; Reagan and others, 2019). These petrological characteristics and ages of magmatism indicate that the oceanic crust underlying the Izu-Bonin-Mariana arc formed upon subduction initiation at *ca.* 52 Ma (Ishizuka and others, 2018; Reagan and others, 2019) and was rapidly followed by arc magmatism in about 5 to 10 Ma (Brandl and others, 2017; Ishizuka and others, 2018; Reagan and others, 2019). This implies that intra-oceanic subduction initiation leads to coeval upper plate extension and spatially widespread-, compositionally distinct-, and volumetrically important magmatism forming new oceanic lithosphere on which an incipient arc will subsequently develop (figs. 1A, 3A and 3B).

Although less well studied, additional evidence of magmatism upon subduction initiation comes from the incipient development of subduction zones in the Southwest Pacific. The Matthew and Hunter subduction zone initiated along a pre-existing arc <2 Ma (Patriat and others, 2015; Patriat and others, 2019) and at the Puysegur Trench along stretched continental crust (south of New Zealand, late Miocene–Pliocene) where a Wadati-Benioff zone is well imaged to 150 km depth (Eberhart-Phillips and Reyners, 2001; Gurnis and others, 2019).

Neotethyan ophiolites.—Ophiolites, ranging from the Dinarides-Hellenides through Turkey, Iran, Oman and Tibet are diverse and vary in thickness and in the presence or absence of individual ophiolitic sections. Neotethyan ophiolites have been emplaced over the continental margin through obduction processes (Ricou, 1971; Coleman, 1981; Agard and others, 2014). They generally show variably thick and continuous units of layered to isotropic gabbros overlying a mantle domain composed dominantly of harzburgites and minor lherzolites as well as dunites and wherlites (fig. 1E). Overlying these gabbros is a variably thick sequence of sheeted dike complexes, covered or intruded by compositionally diverse boninites, and MORB-like to arc tholeiitic magmatism with occasional pillow lava structures. Underlying these ophiolites are metamorphic soles, which are 10s to 100s of meter thick tectonic slices of metasediments and mafic rocks welded to the base of the ophiolites (Williams and Smyth, 1973; Hacker and others, 1996; van Hinsbergen and others, 2015; Agard and others, 2016; Soret and others, 2017). The tectonic environment of the ophiolitic belt of the Neotethys has long been debated as either deriving from "traditional" mid-ocean ridge spreading or from supra-subduction zone (SSZ) settings (Pearce and others, 1981; Pearce and others, 1984; Metcalf and Shervais, 2008). This is shown by the Albanian Mirdita ophiolite, which is separated into an eastern and western ophiolitic belt showing mid-ocean ridge and subduction zone affinities (Nicolas and others, 1999; Bortolotti and others, 2002; Maffione and others, 2018). Key petrological and geochemical indicators show that magmatism in Neotethyan ophiolites is related to a developing subduction zone (Miyashiro, 1973; Pearce and others, 1984; Shervais, 2001; Metcalf and Shervais, 2008) (fig. 1A). A contribution from a hydrous slab component is implied by hydrous MORB (Macleod and others, 2013), enrichment in large ion lithophile elements (K, Rb, Ba, Cs) compared to MORB and abundance of clinopyroxene over plagioclase in cumulates (Pearce and others, 1984). In part, these ophiolites are overlain and intruded by arc magmatic products, including trondjemites and andesitic to rhyolitic magmatism (Kusano and others, 2014). High degrees of partial melting of the mantle wedge by fluid-fluxing is recorded by low-FeO fractionation trends, abundance of depleted magmas (high-Mg# andesites and boninites) as well as by the abundance of refractory mantle domains (harzburgite and dunite) within the preserved mantle sections (Pearce and others, 1984; Shervais, 2001). Magmatism with geochemical and isotopic signatures reflecting an important sediment component also indicates the presence of a downgoing slab (Rollinson, 2015; Kusano and others, 2017). Metamorphic soles show inverted metamorphic gradients (Hacker and others, 1996) and tracers of fluid flow indicate that fluids pervasively affected the overlying mantle during high-temperature metamorphism (Prigent and others, 2018). Available ages of subduction-initiation magmatism overlap with the formation and cooling of the metamorphic sole, indicating that the formation of SSZ ophiolites, subduction related magmatism, obduction and cooling is contemporaneous with subduction initiation (Roberts and others, 2016) (fig. 3A). The majority of Jurassic-Cretaceous Neotethyan ophiolites are therefore interpreted as forming in a SSZ–setting at intra-oceanic subduction zones, either through spontaneous or induced subduction initiation near a mid-ocean ridge along a transform fault (Zhou and others, 2018) or, alternatively, along reactivated oceanic detachments (Maffione and others, 2015a; Maffione and others, 2018).

Arc magmatism in subduction zones.—In continental- and island arcs, magmatism is mostly driven by the release of mobile components, stored in hydrated minerals, into the convective upper mantle (Syracuse and Abers, 2006; Grove and others, 2012) (figs. 1B, 1C and fig. 2). Upon subduction, dehydration of the oceanic crust will flux the mantle wedge, lowering the peridotite solidus and produce primitive mantle-derived melts (Ulmer, 2001; Grove and others, 2012). A component of decompression melting is also commonly involved, as depicted in figure 1C (Sisson and Bronto, 1998). The volatile content of subduction-zone magmas will act as a primary control on the evolution of magmas, leading to the distinctive low-FeO magmatism characteristic of subduction zones (Miyashiro, 1974; Kay and Kay, 1985; Grove and Kinzler, 1986; Sisson and Grove, 1993; Zimmer and others, 2010; Nandedkar and others, 2014). Records of magmatism suggest significant variations in arc magma production rates over several tens of million years (Paterson and others, 2011; Ducea and others, 2015) indicating that arc systems are not steady state environments. Changes in subduction zone dynamics, such as age of the subducting lithosphere, obliquity, rates of convergence and subduction of buoyant oceanic plateaus are thought to play an important role on the composition and production of arc magmas (Stern, 2002; Kelemen and others, 2004; Ducea and others, 2015). Nevertheless, estimated magma production rates in slow subduction systems such as the Aleutians and Antilles island arcs (Reymer and Schubert, 1984; Jicha and others, 2006), remain within the range of other arc systems (Jicha and Jagoutz, 2015) (fig. 3B). Overall, large batholiths in subduction zone settings are incrementally grown over several tens of millions of years as exemplified by the Triassic to Cretaceous Sierra Nevada Batholith, with arc systems from diverse subduction zone settings generating a volume of arc crust on the scale of 20 to 300 km^3 per km of arc strike per million years ($km^3 km^{-1} Myr^{-1}$) (Arculus, 2004), with no clear evidence that magmatic contributions to crustal growth is *significantly* affected by changes in subduction dynamics over tens of millions of years (figs. 2A, 2C, and fig. 3).

Magmatic response to mid-ocean ridge subduction.—Prior to continental collision, recycling of an oceanic crust will lead to the subduction of a mid-ocean ridge. Subduction of an active ridge system leads to the intersection of buoyant, young, warm oceanic crust with an accretionary wedge, leading to changes in plate and subduction dynamics (Farrar and Dixon, 1993). In the circum-Pacific, the magmatic response to the intersection of an active mid-ocean ridge with the continental margin is suggested to have formed widespread near-trench magmatism in the Paleogene along the southern Alaska margin (Sanak-Baranof forearc magmatic belt; Marshak and Karig, 1977; Farris and Paterson, 2009). Migration of the triple junction is thought to have caused high-heat flow, an increase in fluid flow and hydrothermal circulation within the trench (Haeussler and others, 1995) as well as low-pressure–high-temperature amphibolite facies metamorphism of the accretionary complex (Zumsteg and others, 2003). The magmatic response to the formation of a triple junction is recorded along the southwestern margin of Chile. The recent subduction of the mid-ocean ridge has resulted in compositionally heterogeneous forearc magmatism within an active accretionary prism (Forsythe and others, 1986; Kaeding and others, 1990; Lagabrielle and others, 1994; Guivel and others, 1999), elevated heat-flow within the forearc (Behrmann and others, 1994) and the obduction of ophiolites on the forearc (Taitao Ophiolite, Nelson and others, 1993).

Back-arc rifting and spreading.—Backarc basins and ocean-spreading can form as a consequence of extension of a magmatic arc (for example: Izu-Bonin Mariana arc, Lau Basin, Scotia Sea, Calabrian arc) (Saunders and Tarney, 1984; Faccenna and others, 2001a, 2001b; Stern and others, 2004; Ducea and others, 2015) (fig. 1C). As a result of the near-vertical structure of the downgoing Pacific plate and the rapid movement of the Philippine Sea plate away from the Pacific trench (Stern and others, 2004 and references therein), the Izu-Bonin Mariana arc is characterized by the episodic growth of a series of back-arc rift/spreading centers following subduction initiation. This include the formation of remnant arcs such as the Kyushu-Palau ridge and west Mariana ridge and opening of the back-arc systems such as the Parece-Vela and Shikoku ocean basins and, more recently, the Mariana trough (Fryer, 1995; Stern and others, 2004; Ishizuka and others, 2011b). This allows for the formation of a variety of compositionally distinct magmatic rocks grading from N-MORB to arc tholeiites as a result of the imprint on the underlying mantle left behind by the subducting slab (back-arc basalts, Saunders and Tarney, 1984; Stern and others, 2004). On a smaller scale, back-arc rifting along the Calabrian arc in the Western Mediterranean has formed the Liguro-Provencal and Tyrrhenian back-arc basins (Faccenna and others, 2001a, 2001b).

Magmatic arc gaps.—Gaps in arc magmatism have been well documented in Benioff-type subduction zones, primarily along the eastern Pacific in Cordilleran-type orogens (Barazangi and Isacks, 1976; Livaccari and others, 1981; Ramos and Folguera, 2009; Pfiffner and Gonzales, 2013; Jones and others, 2016) (fig. 1B). Subduction of active mid-ocean ridges, buoyant oceanic plateaus and aseismic ridges are thought to be the primary causes of flat slab subductions (Livaccari and others, 1981; Nur and Ben-Avraham, 1983; Liu and others, 2010; Manea and others, 2017), with flat-slab geometries occuring in 10 percent of subduction zones worldwide (Gutscher and others, 2000). Numerical thermo-mechanical simulations show that a flat-slab geometry allow for higher interplate coupling and a removal of the asthenospheric wedge and subcontinental mantle of the overriding plate, thereby partially or completely supressing arc magmatism (Axen and others, 2018). Current flat-slab geometries on the order of 1000 to 1500 km wide segments include the Pampean–, Peruvian– and Bucaramanga flat-slabs along South America (Ramos and Folguera, 2009). These flat-slabs are characterized by a generalized suppression of arc magmatism over the last *ca.* 13 Myr and a possible migration of magmatism continentward. Along the western United States, the Late Cretaceous-Paleogene Laramide Orogeny resulted from a flat slab subduction related to the subduction of large oceanic plateaus (Livaccari and others, 1981). This orogen suppressed magmatism along the Sierra Nevada and pushed magmatism >600 km inland (Livaccari and others, 1981; Miller and others, 1992). Shorter magmatic arc gaps on the order of a few million years have also been recorded as resulting from the collision of oceanic plateaus within the Cretaceous-Paleogene forearc of Costa Rica, coeval with forearc uplift resulting from plateau accretion (Andjić and others, 2018). Though magmatic arc gaps occur in convergent settings, they result from short-term responses to changes in the subducting plate geometry and are mostly transient features.

Accretionary Complex and HP-LT Metamorphism

Subduction of oceanic crust followed by collision leads to distinct sedimentary deposits preserved in accretionary complexes (Underwood and Bachman, 1982) (figs. 1 and 2A) and ophiolitic sutures which contain high-P and low-T metamorphic rocks. Ophiolitic paleo-sutures, evidence of ancient subduction zones, are widespread throughout the circum-Pacific and Neotethys (Miyashiro, 1961; Robertson and others, 2004). Field observations show that such complex subduction zones contain predominantly cm to km-scale large exotic blocks and lenses that record high-P and low-T metamorphism imbedded in low-grade metamorphic matrix, as shown by mélanges in Costa Rica (Flores and others, 2015), Venezuela (Sisson and others, 1997), Guatemala (Harlow and others, 2004), New South Wales (Offler, 1982), Western United States (Hopson and Pessagno, 2005), Iran and Turkey (Robertson and others, 2004; Rad and others, 2005).

Such subduction mélanges are well illustrated by the Franciscan complex (Western United States), which is a preserved accretionary prism recording 80 Myr of Franciscan subduction (fig. 2A) (Ernst, 1965; Hamilton, 1969) and dominated by sandstones, mudstones with a variety of fragments of cherts, limestones, and mafic (oceanic) fragments. The exact mechanisms leading to the structures of the Franciscan complex are still debated. The proposed mechanisms center around a combination of return-flow in a subduction channel, underplating-accretion and thrusting of coherent slivers of sedimentary deposits, erosion and mingling of accreted sediments as well as deposition of debris flows and turbidite density currents in an active trench (Cloos, 1986; Cloos and Shreve, 1988a, 1988b; Wakabayashi, 2015; Raymond, 2017; Krohe, 2017). In places, coherent, low to high-grade metamorphic rocks (zeolite to epidote-blueschist facies) form large thrust sheets of metasedimentary and metavolcanic rocks (Wakabayashi and Dumitru, 2007; Wakabayshi, 2015). However, the Franciscan complex contains important subduction mélange zones formed by a low-metamorphic grade (zeolite facies) matrix of shales occasionally intermingled with serpentinites enclosing masses of scattered blocks over the length of the complex (>1000 km). These enclosed blocks are dominated by variably metamorphosed chert, conglomerate, gabbro, basalt, serpentinite, gneiss, and occasional amphibolite, blueschist- to eclogite facies rocks derived from the subducting oceanic crust, forearc lithosphere and magmatic arc (Raymond, 2017; Krohe, 2017 and references therein). High-pressure blocks are estimated to make up <1 percent of Franciscan metamorphic rocks (Coleman and Lanphere, 1971), and might reach peak blueschist- to eclogitic metamorphism at 2.4 GPa and *ca.* 600 °C, with P-T conditions varying significantly between individual blocks whereas the high-pressure paragenesis of the matrix is <0.8 GPa and <300 °C (Tsujimori and others, 2006; Wakabayashi, 2015; Raymond, 2017).

The structural complexity of mélanges within the trench is further enhanced by the formation of 10 to 50 km in diameter and up to 2.5 km high serpentinite mud-volcanoes (seamounts) forming on the overriding plate within the forearc, as shown in the Western Pacific, Mediterranean, Carribbean and Makkran accretionary complex (Higgins and Saunders, 1974; Robertson and Kopf, 1998; Fryer and others, 2000; Wiedicke and others, 2001; Fryer, 2012). These form as a result of the transfer of sediments and fluids along fractures from the downgoing slab to the overriding wedge and plate. Minerals and clasts within the serpentinite matrix of such mud-volcanoes are derived from the subducting plate, supra-subduction zone mantle and crustal fragments from the upper plate, and can include dunites, harzburgites, basalts, cherts as well as blueschist-facies metamorphic rocks (Maekawa and others, 1992; Maekawa and others, 1993; Fryer, 2012 and references therein).

Therefore, subduction leads to significant heterogeneity in sedimentary deposits within an active forearc trench (figs. 1B and 1C) with distinct sharp changes in P-T conditions at the outcrop- and regional scale, with high-pressure rocks from a variety of sources found as isolated blocks within a low-pressure sedimentary matrix.

FINGERPRINTS OF SUBDUCTION IN ALPINE-TYPE OROGENS

In this section, we review the same geological characteristics described in the previous section for Benioff-type subductions but in the context of the Western and Central Alps and Pyrenees identified as Alpine-type orogens (fig. 4). This orogenic system is formed of imbricated sequences of nappes resulting from the closure of narrow rift basins between two continental blocks (Sengör, 1991; Ernst, 2005; Mohn and others, 2014). As shown in this contribution, they are characterized by lack of significant evidence of magmatism during closure. These orogens are distinct from "Pacific-type" orogens formed over long-lived subduction systems (Ernst, 2005). This will allow us to compare the similarities and distinctive characteristics of Alpine-type orogens and Benioff-type subduction in the discussion.

The Western and Central Alps

Prior to Late Cretaceous convergence and subduction, the Alpine domain recorded a major Jurassic rifting phase between Europe and Adria (fig. 4) (Stampfli and others, 1998). Jurassic rifting leading to the formation of the Piemonte-Liguria and Valaisan basins is constrained by its stratigraphic record (Lemoine and others, 1987; Lemoine and Trümpy, 1987; Steinmann, 1994; Masini and others, 2013) combined with P-T-t conditions of rift-related thinning and exhumation of the continental lithosphere (Müntener and others, 2000; Mohn and others, 2012). Extreme thinning of continental lithosphere leading to the separation of continental crust was associated with exhumation of subcontinental mantle in relation with top-basement detachment faults (Manatschal and Müntener, 2009). Exhumed subcontinental mantle can be overlain by extensional crustal allochthons, late syn-rift to post-rift sedimentary successions, while magmatic additions remain scarce (Decandia and Elter, 1972; Steinmann and Stille, 1999; Manatschal and Müntener, 2009; Beltrando and others, 2012). All together these complex lithostratigraphic associations are characteristic of Western Tethyan ophiolites. The opening of the Piemonte-Liguria basin is determined by U-Pb ages of zircons in MOR type gabbros (*ca.* 155–165 Ma) (fig. 4) and the mid- to late Oxfordian – Bajocian age of the first overlying pelagic sediments (radiolarian chert, De Wever and Baumgartner, 1995; Cordey and Bailly, 2007; Cordey and others, 2012) (fig. 4). In contrast, the timing of rifting of the Valaisan basin is controversial. It has been related to (a) a renewed rifting in the Cretaceous after an initial Jurassic event

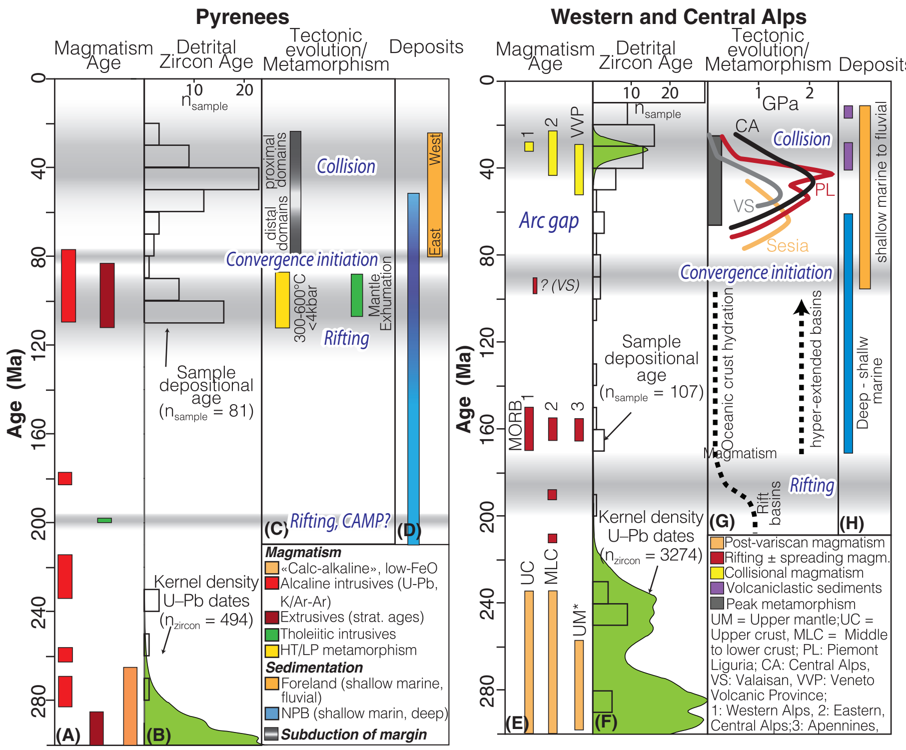


Fig. 4. Timescales of the main magmatic, metamorphic and tectonic events in the Pyrenees (A,B,C,D) and Alps (E,F,G,H). Pre-rift magmatism is from Briqueu and Innocent (1993), Martí (1996), Solé and others (2002), Rossi and others (2003), Rodríguez-Méndez and others (2014), Lago and others (2004), Denèle and others (2012), Pereira and others (2014), Druguet and others (2014). Ages and composition of syn-rift magmatism is from Schoeffler and others (1964), Montigny and others (1986), Fernández-Mendiola and García-Mondéjar (2003), Solé and others (2003), Castañres and Robbles (2004), Carracedo-Sánchez and others (2012), and Ubide and others (2014). A Cretaceous high-temperature-low pressure (HT-LP) metamorphism is recorded in the Pyrenean sedimentary cover (~105–85 Ma; see compilation in Clerc and others, 2015) and commonly attributed to crustal thinning and mantle exhumation during rifting (Golberg and Leyreloup, 1990; Vacherat and others, 2014; Clerc and others, 2015 and references therein). Timing of rifting and progressive margin inversion during collision is based on Tugend and others (2014) and Tugend and others (2015) and may slightly differ with the timing deduced from the eastern Pyrenees only (Mouthereau and others, 2014). Onset of convergence is deduced from plate kinematic models and field observations (respectively Olivet, 1996 and Garrido-Megias and Rios Aragues, 1972). Onset of rifting is recorded by the deposition of deep-water sediments (marls followed by siliciclastic turbidites, Masini and others, 2014) then in narrow rift basins progressively widening to reach up to 100–150 km in a north south direction (Tugend and others, 2015; Nirrengarten and others, 2018). Syn-collisional sedimentation is distinct between the Northern and Southern Pyrenees due to variations in localization of extension and rates of exhumation affecting the transport and sedimentation of eroded products. From 70 Ma and younger, a coarsening upwards of sedimentation is observed leading to the deposition of sandstones and conglomerates either in foreland basins (Northern Pyrenees) or in shallow, fluvial-dominated environments (Southern Pyrenees) (Vacherat and others, 2017 and references therein). Detritial zircon populations are from Whitchurch and others (2011), Filleaudeau and others (2012), Mouthereau and others (2014), Hart and others (2016), Thomson and others (2017), and Vacherat and others (2017). Data from the Alps is modified from McCarthy and others (2018). Additional alkaline magmatism at 197–192 Ma along the Ivrea zone during rifting is from Galli and others (2019). It should be noted that a few zircons from amphibolites of the European margin (VS, Valaisan) have been dated to *ca.* 93 Ma (Liati and others, 2003; Liati and Froitzheim, 2006) as is shown in (E). The origin of these metabasic rocks, as well as their age, remains dubious, particularly as a result of the significant inheritance and variation in zircon ages found in individual samples (40 Ma–2.4 Ga). Nevertheless, these metabasic rocks are found on the northern margin of the Piemonte-Liguria basin and are unrelated to subduction initiation along the southern margin of the Piemonte-Liguria basin. * = Sm-Nd depletion model age of Alpine depleted spinel-peridotites from McCarthy and Müntener (2015).

(Loprieno and others, 2011), (b) the rifting and separation of the "exotic" terrane of the Briançonnais from the European margin during the Late Jurassic to Early Cretaceous (Stampfli, 1993, Stampfli and others, 2002) or (c) a portion of a Jurassic hyper-extended margin (Manatschal and others, 2006). However, U-Pb dating of amphibolites along the Valaisan basin has never given an unequivocal Cretaceous age (Liati and others, 2003). Moreover, pre-rift sediment deposits within this basin link the Briançonnais and European margin together (Galster and others, 2012). In addition, Ribes and others (2019) consistently show that the Valais basin most likely formed more or less simultaneously with the Piemonte-Liguria basin, based on stratigraphic and pateontological observations. Thus, the geological evidence therefore favors an origin of the Valaisan basin as part of protracted late Jurassic rifting.

Initiation of convergence is well established on the basis of paleomagnetic data showing the northward movement of Iberia and North Africa at *ca.* 100 to 85 Ma (Rosenbaum and Lister, 2005). This is consistent with dating of accretionary thrusting and initiation of sedimentary flysch deposits within the Piemonte-Liguria basin (Stampfli and others, 1998; Handy and others, 2010, Zanchetta and others, 2012) as well as with prograde high-pressure metamorphism starting at *ca.* 85 Ma along the southern passive margin of the Piemonte-Liguria basin (Skora and others, 2009; Manzotti and others, 2014) (fig. 4).

Geophysical imaging of an Alpine slab.—The structure of the Alpine orogen has been extensively imaged along the entire belt and includes the ECORS-CROP profile (Roure and others, 1990; Roure and others, 1996) and NFP20 along the Western and Central Alps (Schmid and others 1996; Escher and others, 1997; Pfiffner and others, 1997) as well as TRANSALP along the Eastern Alps (Transalp Working Group 2002). The structure of the Alpine orogen remains debated and susceptible to diverse interpretations (Lacassin and others, 1990; Nicolas and others, 1990; Roure and others, 1996; Schmid and Kissiling, 2000; Schmid and others, 2004). The scientific consensus has long been that the underlying reflectors at shallow (<50 km) depth below the Alps represent the subducted continental crust, either with or without a large lower crustal duplex (Schmid and Kissling, 2000; Schmid and others, 2017). This is required in order to balance the restorations of pre-collisional Alpine transects as a result of an assumed pre-collisional equilibrated continental crust thickness (that is the missing continental crust, Ampferer and Hammer, 1911). However, Mohn and others (2014) suggested, in light of the formation of hyper-extended and hyperthinned continental crust, that the geophysical images are representing hyper-extended margins of the Piemonte-Liguria basin. Thus, the interpretation of current geophysical imaging in the Alps, and tectonic features in the Alps in general, are strongly affected by our understanding of the pre-collisional architecture of the Alpine orogen (Butler, 2013; Mohn and others, 2014; Beltrando and others, 2014).

At greater depth, the Wadati-Benioff zone, representing the subducting oceanic slab, remains conspicuously absent beneath the Alpine orogeny (Doglioni and others, 1999; Malusà and others, 2017). Moreover, geophysical imaging of the convective mantle underlying central Europe shows a large anomaly at 600 km depth (Spakman, 1990) which has been interpreted as being the remnant Piemonte-Liguria oceanic slab (Spakman and Wortel, 2004; Handy and others, 2010; Handy and others, 2014). However, geophysical imaging of the Alpine orogen has produced inconsistent results (Garzanti and others, 2018; Kästle and others, 2020). In part, the problems of identifying subducted slabs in the mantle beneath the Alps result from difficulties in constraining the vertical continuity of tomographic anomalies, different inversion parameters and in the assumption that detected anomalies are related to thermal differences (Foulger and others, 2013). These include different models indicating a detached slab in the Western Alps and an imaged ~250 km deep slab in the Central

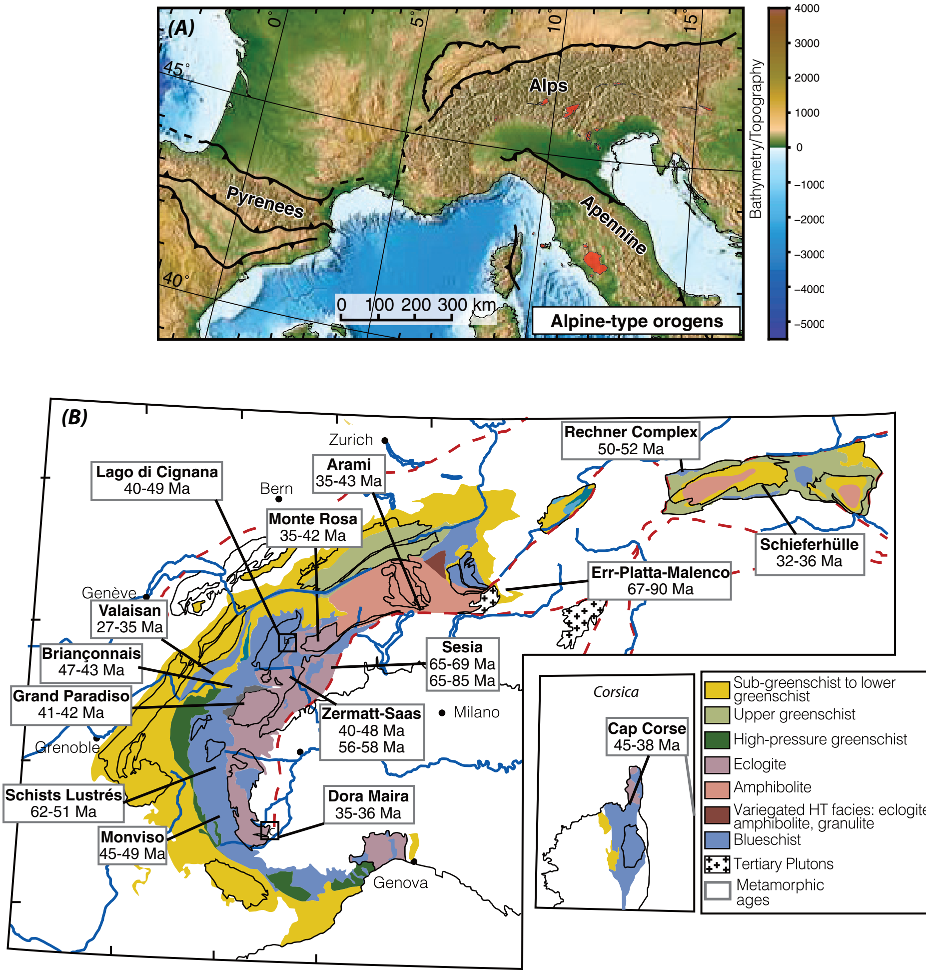


Fig. 5. Characteristics of subduction in Alpine-type orogens: (A) Distribution of magmatism in the Pyrenees and European Alps during convergence; (B) map of metamorphic facies in the European Alps. Redrawn after Berger and Bousquet (2008). Ages of metamorphism after Berger and Bousquet (2008) and references therein. Additional ages for Zermatt-Saas are from Weber and others (2015), for Sesia from Rubatto and others (2011) and Regis and others (2014), for the Gran Paradiso Massif (Manzotti and others, 2018) and for Dora-Maira from Gauthiez-Putallaz and others (2016).

Alps (Lippitsch and others, 2003), or a more continuous slab in the Western Alps and a detached slab in the Eastern Alps (Piromallo and Faccenna, 2004). More recently, Zhao and others (2016) postulated that a large slab is still attached along the entire Western and Central Alps (compare Kästle and others, 2020). Discontinuous slabs along strike have inspired slab breakoff models, which have in part been interpreted as resulting in syn-collisional magmatism in the Alps (Davies and von Blackenburg, 1995). However, there is no consensus about the mechanism and timing of slab breakoff, which brings into question its role for the evolution of the Alps (Garzanti and others, 2018). In addition, the predominant model of ocean closure in the Western Alps suggests continuous southward-dipping subduction between the Cretaceous and Paleogene (Handy and others, 2010; Handy and others, 2014). Other models suggest that subduction was jammed and migrated northwards once buoyant continental frag-

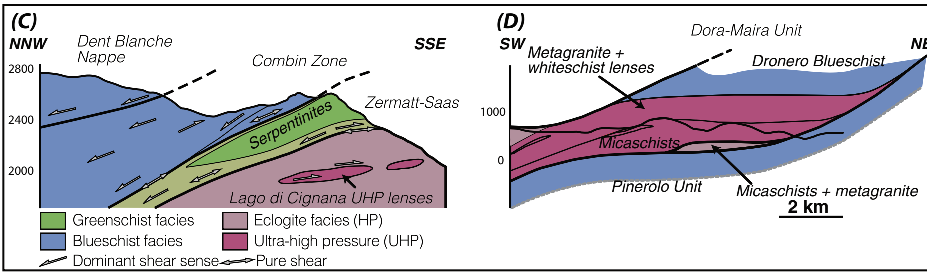


Fig. 5—Continued
(C) Cross-section of Zermatt-Saas Jurassic ophiolites, showing the structural consistency of lenses of hundreds of meters (Angiboust and others, 2009), redrawn from Kirst and Leis (2017). The Combin zone (oceanic derived Tsaté and continental derived Cime Blanche units) reached upper-greenschist facies to blueschist facies metamorphism (Kirst and Leis, 2017 and references therein). Note that the Combin zone is pervasively overprinted by retrograd greenschist metamorphism (Dal Piaz and Ernst, 1978). The Dent Blanche nappe is interpreted as a continental allochthon rifted from the Adriatic margin (Froitzheim and others, 1996; Manzotti and others, 2014); (D) cross-section of the Dora-Maira Massif, which is composed of three tectonically imbricated continental fragments from the European margins. Redrawn after Avigad and others (2003) and Schenker and others (2015). These cross-sections, derived from the oceanic and continental parts of the Piemonte-Liguria basin illustrate the coherence of metamorphism overs hundreds of meters or kilometers. The ultra-high pressure unit of the Dora Maira massif is estimated 15 km wide and 1–2 km thick (Avigad and others, 2003). Note that the Pyrenees show no magmatism nor metamorphism from initiation of convergence until collision.

ments (or "micro-continents", Babist and others, 2006; Weber and others, 2015) reached the trench, allowing for two (Stampfli and others, 1998; Babist and others, 2006; Herwartz and others, 2008) or three (Weber and others, 2015) subducting slabs to form.

Magmatism.—The Alps preserve three distinct phases of magmatism in the last 300 Ma (fig. 4) (McCarthy and others, 2018). Permo-Triassic magmatism is related to the transtentional movements and thinning of the continental lithosphere during post-Variscan extension and is recorded by intrusive rocks within the lower- to upper crust as well as volcaniclastic deposits and lava flows (Schaltegger and Corfu, 1995; Petri and others, 2017; Manzotti and others, 2017; Kunz and others, 2018). A few depleted spinel-peridotites from Western Tethyan ophiolites have Permian depletion ages interpreted as decompression mantle melting in Permian times (Rampone and others, 1998; Müntener and others, 2004; McCarthy and Müntener, 2015).

Magmatism during rifting was sparse, forming local alkaline and carbonatite dikes along the Ivrea Zone (Schaltegger and others, 2015; Galli and others, 2019) whereas magmatism upon mantle exhumation to the seafloor allowed for localized gabbroic intrusions (Costa and Caby, 2001; Desmurs and others, 2002; Kaczmarek and others, 2008), lava flows and pillow lavas (Chalot-Prat, 2005) as well as melt-percolation in exhumed mantle rocks (Müntener and Piccardo, 2003) (figs. 4 and fig. 6). Jurassic-age MORB-type magmatism in Alpine-Apennine ophiolites is widespread, but remains short-lived (Rampone and others, 1998; Schaltegger and others, 2002; Tribuzio and others, 2016) (fig. 4).

A third distinct phase of magmatism is related to Alpine collision at 42 to 25 Ma (fig. 4) mostly recorded by plutons and dikes along the Periadriatic (Insubric) line (Del Moro and others, 1983; Steenken and others, 2000; Bergomi and others, 2015; Samperton and others, 2015) (fig. 5A) but also by protracted crustal melting in the central Alps (Rubatto and others, 2009). The alkaline Veneto Volcanic Province (VVP) in northern Italy is slightly older (45–40 Ma) and its relation to Alpine magmatism is unknown (Visonà and others, 2007).

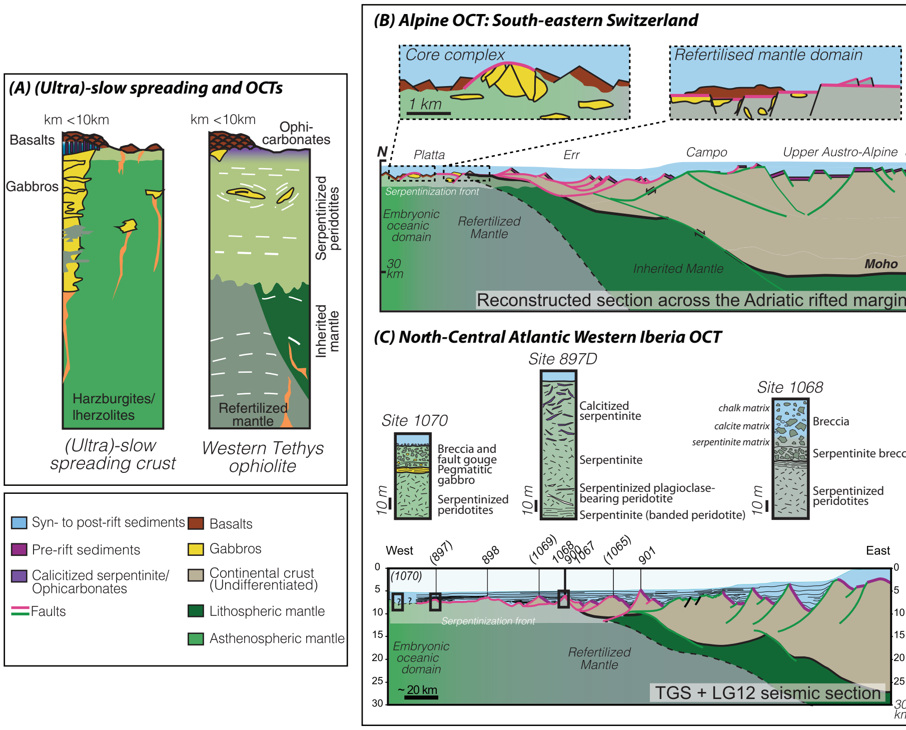


Fig. 6. Setting and lithostratigraphic associations of Western Tethyan ophiolites and comparison with (ultra-)slow spreading systems and OCTs along the Iberia passive margin. (A) Representative cross-section of (ultra-)slow spreading systems (Atlantic ocean) (modified after Dick and others, 2003; Dick and others, 2006) and Western Tethyan ophiolites, dominated by exhumed mantle, serpentinized peridotites and minor MORB-type magmatism; (B) representative fossil passive margins from the Tasna-Err Ocean-continent transition (OCT), Western Alps (Mohn and others, 2010; Picazo and others, 2016); note that for clarity, the sediments are not shown; (C) North-Central Atlantic Western Iberia OCT characterized by hyper-extended continental crust, local exhumation of lower crust, serpentinized mantle covered by carbonated tectonic breccias, and sparse magmatism, after Sutra and others (2013). IODP drill Sites 1070, 897D and 1068 are from Shipboard Scientific Party (1994, 1998a, 1998b).

Overall, syn- to post-collisional magmatism remains volumetrically limited in the Alps (figs. 3A, 3B and 5A). The two main phases of magmatism, Tertiary- and Permo-Triassic magmatism are well recorded both in terms of intrusive and erupted magmatic products and also in terms of detritial zircons found in sedimentary deposits (fig. 4). However, Alpine Paleogene collisional crustal melting and formation of migmatites and pegmatitic to aplitic dikes remains spatially and volumetrically limited (Southern Steep Belt, Central Alps, Burri and others, 2005) whereas most granulite-facies metamorphism and crustal melting found within the European Alps involved pre-Alpine (Variscan orogeny and older) events (Blattner, 1965; Schnetger, 1994; Del Moro and others, 1999; Müntener and others, 2000; Burri and others, 2005; Genier and others, 2008; Schuster and Stüwe, 2008; Manzotti and others, 2017; Kunz and others, 2018).

The most remarkable feature is, however, the complete lack of evidence of arc-like magmatism from the initiation of subduction between 100 and 85 Ma until continental collision at *ca.* 42 Ma (fig. 4) (McCarthy and others, 2018). This age-gap contrasts sharply with the continuous record of detrital zircons for the preceeding 1 Ga (see Appendix) and of the continuous record of detritial zircon ages in Benioff-type subduction zones (fig. 2C).

HP-LT metamorphism in the Alps.—The internal part (that is, axial belt) of the Alps recorded a complex and polyphase orogenic related deformation/metamorphic history ranging from greenschist to UHP eclogite facies conditions (fig. 5, see Berger and Bousquet, 2008; Beltrando and others, 2010a; Pfiffner, 2014 for reviews). It samples distinct paleogeographic units deriving from continental (European/Briançonnais and Adriatic margins) and oceanic (Piemonte-Liguria and Valaisan basins) units. Extensive studies in both continental and oceanic units permit a general characterization of their P-T-t evolution combined with an estimation of the peak pressure of each of these units (Berger and Bousquet, 2008; Agard and others, 2009). A general younging for the HP-LT metamorphism is recorded from internal to external units, from the Adriatic to the European margins. The timing of high-pressure metamorphism (from ~80 Ma to ~35 Ma) of individual paleogeographic units (for example: Margna, Sesia-Dent Blanche, Briançonnais, Internal Crystalline Massif) separated by exhumed mantle domains (Piemonte-Liguria and Valaisan basins) (Froitzheim and others, 1996) has led to a variety of subduction models. In detail, an evolution in the age of high-pressure metamorphism has been identified corresponding to the progressive closure and subduction of the Piemonte-Liguria and Valaisan basins from Late Cretaceous to early Oligocene (80 – 35 Ma). The oldest ages for HP-LT metamorphism (Late Cretaceous to early Paleocene) are reported from distal continental units (Sesia zone) (figs. 3A and 4). These ages are interpreted as the initial subduction of the Adriatic margin representing possibly former extensional allochthons (Babist and others, 2006; Manzotti and others, 2014). The Paleocene to early Oligocene period corresponds to the successive subduction of the Piemonte-Liguria basin, Briançonnais block, Valaisan basin and eventually to the European distal margin (fig. 4). Evidence for ultra-high pressure metamorphism is volumetrically minor in the Alps compared to high-pressure lithologies (Berger and Bousquet, 2008) (fig. 5) and has been reported from both continental-derived units (small Brossasco-Isasca unit in the Dora-Maira Internal Crystalline Massif, Chopin, 1984) and oceanic units (small Lago di Cignana unit in the Zermaat-Saas, Reinecke, 1991) (figs. 5B, 5C and 5D).

The Pyrenees

There are a number of uncertainties regarding the formation of the Pyrenees prior to the magnetic oceanic anomaly A34 (83 Ma, Santonian; Roest and Srivastava, 1991). This uncertainty is mainly related to the pre-collisional tectonic setting that led to the formation of the Cretaceous Pyrenees Basins (within the so-called North Pyrenean Zone) and westward Bay of Biscay. The traditional interpretation of Cretaceous Pyrenees Basins suggests that they correspond to pull-apart basins formed during left-lateral strike-slip to transtensional deformation (Le Pichon and others, 1971; Mattauer and Séguret, 1971; Choukroune and Mattauer, 1978; Peybernès and Souquet, 1984; Lagabrielle and Bodinier, 2008). More recently, a two-step model aiming to account for both marine and field observations was proposed by Jammes and others (2009, 2010). They propose that Aptian to Cenomanian north-south extension was preceded by a Late Jurassic to early Aptian transtension (fig. 4). This geological scenario was subsequently integrated in a global plate kinematic model (Nirrengarten and others, 2018) accounting for the partitioning of deformation, which occurred at the scale of the Iberian-European plate boundary (Tugend and others, 2015). A more controversial interpretation proposed by Vissers and Meijer (2012) and Vissers and others (2016) suggests that these basins formed during Aptian subduction of a Jurassic–early Cretaceous Neotethyan lithosphere. This subduction would be followed by slab breakoff in late Albian time that would explain, according to the authors, the current absence of a velocity anomaly beneath the Pyrenees that could be interpreted as former subduction (Vissers and Meijer, 2012; Vissers and others, 2016). The kinematic setting leading to the formation of the Cretaceous Pyrenean rift basins remains debated. However, most authors agree that they formed subsequently to extreme thinning (Peybernes and Souquet, 1984; Velasque and others, 1989; Saspiturry and others, 2019) and local mantle exhumation during Albo-Cenomanian time (Lagabrielle and Bodinier, 2008; Jammes and others, 2009; Lagabrielle and others, 2010; Masini and others, 2014; Tugend and others, 2014) (fig. 4). This was followed by compression after *ca.* 85 Ma, leading to the closure of these hyper-extended basins and a northward dipping "subduction".

Interpretation of geophysical data.—Our understanding of the present-day deep structure beneath the Pyrenean orogen relies on several generations of tomographic models (Souriau and Granet, 1995; Vacher and Souriau, 2001; Souriau and others, 2008; Chevrot and others, 2014). The first tomographic models imaged a poorly resolved low velocity anomaly down to *ca.* 100 km that was interpreted as the subducted Iberian crust (Souriau and Granet, 1995; Vacher and Souriau, 2001). However, these models were derived from a highly heterogeneous and poor distribution of seismic stations and these results were partly invalidated by subsequent tomographic studies including an improved array coverage (Souriau and others, 2008). The new models reveal the strong dependency on crustal corrections, a key parameter that can influence the resolution of the data up to 200 km depth. Therefore, the absence of a velocity anomaly at depth is highly uncertain (Souriau and others, 2008). The deployment of a large network of seismic stations during the PYROPE and IBERRAY experiments in addition to higher resolution crustal corrections recently confirmed the absence of an anomaly that could be attributed to a subduction beneath the Pyrenees (Chevrot and others, 2014). Current images of the deep structure of the Pyrenees are derived from 2D tomographic models (Chevrot and others, 2015) and full-waveform inversions of P waves (Wang and others, 2016). Instead of a lithospheric-scale slab, only a northward dipping crustal root is imaged down to 50 to 60 km depth.

Magmatism.—As in the Alps, several magmatic episodes preceding the main Cretaceous rift event are documented in the Pyrenees (fig. 4). A Permian magmatic event is characterized by the presence of granitic sills, dikes, and plutons whose ages range between ~267 and ~278 Ma (fig. 4; see synthesis in Vacherat and others, 2017) and volcanic tuffs in several basins (Barnolas and Chiron, 1996; Lago and others, 2004). During the Triassic and Jurassic two types of magmatic events are distinguished and interpreted as the effect of a rift episode (Azambre and Fabries, 1989). Alkaline intrusives, mostly Norian in age (Montigny and others, 1982), are interpreted as part of a multi-episodic and scattered igneous province in eastern Iberia (fig. 4; Lago and others, 1996; Sanz and others, 2013; Ubide and others, 2014). A distinct short-lived but widespread tholeiitic (ophitic) magmatic event appears restricted to the Triassic-Jurassic boundary (fig. 4; Montigny and others, 1982; Marzoli and others, 1999; Rossi and others, 2003) and is commonly related to the Central Atlantic Magmatic Province (CAMP).

Widespread magmatism, alkaline in composition, also occurred between *ca.* 110 to 80 Ma after the onset of crustal extension (Montigny and others, 1986; Golberg and others, 1986; Ubide and others, 2014; Clerc and others, 2015) and persisted during the post-rift stage (fig. 4) (Choukroune and Mattauer, 1978). Alkaline magmatic rocks are distributed across the Pyrenees as small bodies corresponding to submarine flows, dikes and sills and ranging from basalts and gabbros to syenites and lamprophyres (Azambre and Rossy, 1976; Ubide and others, 2014; Clerc and others, 2015). The age of this volcanic activity is often constrained by stratigraphic correlations from intercalated sediments indicating an Albian to Turonian activity (Dubois and Seguin, 1978). Absolute ages from K–Ar methods (Montigny and others, 1986) constrain this magmatic activity to ~113 to ~85 Ma (Aptian–Santonian) also corroborated by recent $^{40}Ar/^{39}Ar$ dating (Ubide and others, 2014), with different stages of magmatism being recognized (Montigny and others, 1986). No arc–related magmatism is reported across the Pyrenees during convergence, an absence noted by Vissers and others (2016) and a record consistent with the lack of detrital zircons with magmatic ages younger than *ca.* 260 Ma (fig. 4).

Metamorphism.—No evidence for subduction-related or Barrovian metamorphism resulting from basin closure has been reported from the Pyrenees (Muñoz, 1992). In contrast, a pre-orogenic early Cretaceous thermal event is recognized in the North-Pyrenean basins (Azambre and Rossi, 1976; Albarède and Michard-Vitrac, 1978; Montigny and others, 1986; Goldberg and others, 1988; Thiébaut and others, 1988). This event reaches HT-LP metamorphic conditions, characterized by temperatures of 550 to 650 °C and pressures of 0.3 to 0.4 GPa (fig. 4; Golberg and Leyreloup, 1990) and affects the Mesozoic cover of the North Pyrenean Zone forming a narrow band along the Pyrenees (Clerc and others, 2015 and references therein). This metamorphic event is generally attributed to the Albo-Cenomanian rifting (Golberg and Leyreloup, 1990; Lagabrielle and Bodinier, 2008; Clerc and Lagabrielle, 2014; Clerc and others, 2015), most likely correlated to the combined effect of the local exhumation of sub-continental mantle rocks and syn to post-rift sedimentation (Lescoutre and others, 2019).

The Nature of Western Tethys Ophiolites

Key petrological characteristics.—Western Tethys ophiolites of the Apennines (Elter, 1972; Decandia and Elter, 1972; Rampone and others, 1995; Rampone and others, 1996), the Western- and Central Alps (Lagabrielle and Cannat, 1990; Picazo and others, 2016) and in the Pyrenees (Albarède and Michard-Vitrac, 1978; Sautter and Fabriès, 1990), share characteristics making them different to traditional "Penrose"-type ophiolites, Neotethyan SSZ-ophiolites and Pacific "forearc" crust (figs. 1 and 6). Western Tethys ophiolites (compare Steinmann, 1905) are dismembered slivers dominated by a heterogeneous serpentinized mantle with only minor basaltic dikes and basaltic flows as well as local gabbro intrusions. The serpentinized mantle is overlain locally by ophicarbonates, radiolarian chert or siliciclastic turbidites and other oceanic sedimentary deposits (fig. 6) (Lagabrielle and Cannat, 1990; Lagabrielle and others, 2015). Ultramafic rocks are heterogeneous and can be subdivided into two major groups (compare Picazo and others, 2016). *Inherited mantle* is predominantly formed of cold (<950 °C), heterogeneous spinel peridotites containing an array of compositionally distinct spinel and garnet pyroxenites and dunites (Rampone and others, 1995; Müntener and others, 2010; Guarnieri and others, 2012; Borghini and others, 2013; Basch and others, 2019). This mantle is, in certain cases, associated with thinned continental crust or extensional continental allochthons (Manatschal and Nievergelt, 1997; Müntener and others, 2000) (fig. 6). This group includes the majority of mantle peridotite bodies found across the Pyrenees (Verschure and others, 1967; Bodinier and others, 1988; Fabriès and others, 1998; Henry and others, 1998; Le Roux and others, 2007; Riches and Rogers, 2011). The petrological and chemical characteristics of *inherited mantle* domains are interpreted as representing ancient subcontinental lithospheric mantle exhumed to the sea-floor following rifting (Picazo and others, 2016) (fig. 4). *Refertilized mantle* in the Alps on the other hand is formed of plagioclase-peridotites characterized by elevated temperature (1000–1200 °C), fertile compositions and isotopic compositions in equilibrium with Jurassic-age MORB (Müntener and Piccardo, 2003; Rampone and others, 2008; Kaczmarek and Müntener, 2008; Rampone and others, 2009; Müntener and others, 2010; Kaczmarek and Müntener, 2010). Refertilized peridotites represent mainly inherited subcontinental lithospheric mantle that has been impregnated with MORB-melts during rifting and exhumation at the seafloor (Picazo and others, 2016) (fig. 6). Associated with these refertilized peridotites are residual spinel-peridotites affected by near-fractional melting and which are not in isotopic equilibrium with Jurassic MORB magmatism (Rampone and others, 1998; Müntener and others, 2004; McCarthy and Müntener, 2015; McCarthy and Müntener, 2019). Widespread Permian magmatism and partial melting of the underlying asthenosphere is proposed to have caused the formation of these peridotites, indicating that they are depleted post-orogenic mantle domains accreted to the subcontinental mantle prior to rifting and not mantle domains formed along a Jurassic mid-ocean ridge (McCarthy and Müntener, 2015; Picazo and others, 2016).

Similarities with ultra-slow spreading ocean crust and magma-poor OCTs.—Some Western Tethyan ophiolites show similarities with magma-poor to (ultra-)slow spreading systems such as the present-day Arctic ridges, Southwest Indian Ridge and Mid-Atlantic Ridge (Lagabrielle and Cannat, 1990; Cochran and others, 2003; Dick and others, 2003; Sauter and others, 2013; Lutz and others, 2018; Cannat and others, 2019; Bickert and others, 2020) (fig. 6). (Ultra)-slow spreading systems have important lithological, bathymetric and geophysical characteristics (Dick, 1989; Cannat, 1993; Dick and others, 2003; Cannat and others, 2019). They preserve sparse and localized magmatism and volcanism, minor gabbroic intrusions, and minor- to nonexistent dike-complexes, heterogeneous and serpentinized mantle rocks. These characteristics lead to the interpretation that Western Tethys ophiolites are remnants of an ultra-slow spreading (<2 cm/yr full spreading rate) system (Lagabrielle and Cannat, 1990; Tricart and Lemoine, 1991). Other similarities include the evidence of long-lived oceanic detachment faults accommodating extension. These have been characterized by geophysical imaging in the central Atlantic and Southwest Indian Ridge (Tucholke and others, 1998; Macleod and others, 2002; Sauter and others, 2013; Cannat and others, 2019) and hypothesized to occur in the Alps based on field evidence (Manatschal and others, 2011; Lagabrielle and others, 2015; Festa and others, 2015).

Other Western Tethyan ophiolites have been interpreted as remnants of magma-poor Ocean-Continent Transition zones (OCT) similar to the Iberia-Newfoundland margin (Boillot and others, 1988; Manatschal, 2004; Manatschal and Müntener, 2009; Mohn and others, 2010; Mohn and others, 2012; Beltrando and others, 2014) (figs. 6B and 6C). Key characteristics of the emplacement of ultramafic rocks along an OCT or hyper-extended domains in the Pyrenees (Arzacq-Mauléon basin) and Alps (for example: Central Alps, Tasna OCT) are the complex lateral variation and termination of crustal- and sedimentary lithologies, subcontinental mantle fragments remobilized in breccias indicating their exhumation at the ocean-floor (Lagabrielle and Bodinier, 2008), the identification of tectono-sedimentary breccias containing continental clasts overlying the subcontinental mantle (Manatschal and others, 2006; Jammes and others, 2009; Debroas and others, 2010), the presence of continental derived allochthons separating exhumed mantle domains (Manatschal, 2004), thinned mid- to lower crustal rocks overlapped and remobilized by synextensional sediments (Claude, ms, 1990; Jammes and others, 2009) and preservation of pre-Alpine contacts between subcontinental mantle and continental crust (Müntener and Hermann, 1996). Moreover, the exhumed mantle is heterogenous, grading from lherzolite to harzburgite and dunite, and locally containing garnet-pyroxenite dikes, supporting a subcontinental mantle origin (Trommsdorf and others, 1993; Rampone and others, 1995; Müntener and Hermann, 1996; Montanini and others, 2006). Such characteristics are consistent with geophysical imaging and results of ocean drilling expeditions of magma-poor rifted margins along the Iberia-Newfoundland (Manatschal and others, 2001; Sutra and others, 2013) (fig. 6C), Bay of Biscay (Thinon and others, 2003; Tugend and others, 2014) and Australia-Antarctica conjugate margins (Gillard and others, 2015) indicating thinned continental crust and exhumed mantle rocks over large areas (>100 km).

DISCUSSION

Basin Closure in Alpine-Type Orogens: Similarities with B-Type Subduction

Although the Pyrenees remain an exception regarding the lack of an inferred subducting slab during convergence and collision (Chevrot and others, 2014; Chevrot and others, 2015), there are apparent similarities between Benioff-type subductions and the Alpine orogen. These include geophysical imaging identifying south-verging tectonic structures, the European Moho dipping towards the subduction direction at 40 to 80 km depth as well as the imaging of a 200 to 250 km anomaly connected to the European margin which is interpreted as a downgoing slab (Lippisch and others, 2003; Spakman and Wortel, 2004; Schmid and others, 2004; Kissling and others, 2006; Zhao and others, 2016). Although other similarities commonly attributed to Benioff-type subduction in the Alps includes the formation of blueschist to eclogitic facies rocks (Ernst, 1971; Ernst, 1973; Agard and others, 2009) (fig. 5), circum-Pacific subduction zones lack the presence of ultra-high-pressure continental fragments reaching up to 4 GPa which are found in the Alpine orogen (Chopin, 1984). However, at present the only reported continental crustal rocks exhibiting peak metamorphic pressure of *ca.* 4 GPa in the Western and Central Alps are restricted to minor volumes in the Brossasco-Isasca unit within the Dora Maira massif, whereby peak pressures in all other Dora Maira units are 2.0 to 2.4 GPa (Groppo and others, 2019). The preservation of ophiolites combined with paleomagnetic data showing that Adria migrated north starting at 80 to 100 Ma (Schmid and others, 1996; Rosenbaum and Lister, 2005), and well–characterized changes in peak metamorphic ages for distinct paleogeographic units during convergence (Berger and Bousquet, 2008, fig. 5) are all key points which have been used to develop a "*classical plate-tectonic*" model of Benioff-type oceanic subduction leading to continental collision.

Basin Closure in Alpine-Type Orogens: Key Differences with B-type Subductions

Amagmatic subduction initiation and basin closure.—The lack of magmatic activity related to subduction initiation and closure of basins in either the Pyrenees or the Western and Central Alps is surprising (figs. 3, 4, and 6). The geological record of subduction initiation in Alpine-orogens is distinct from the record of subduction initiation in the western Pacific and Neotethys, as shown in figure 3A. These differences are underscored by the lack of SSZ-ophiolites in both the Alps and Pyrenees.

In the first ~20 Myr following subduction initiation, Alpine orogens do not show any magmatic response to subduction. Within the Alpine domain, convergence is recorded by the formation of flysch sediments, and within 10 to 20 Myr by a few prograde to peak HP-LT metamorphism of continental- and oceanic fragments from the Adria margin (figs. 3 and 4). As argued in McCarthy and others (2018), the absence of magmatism during subduction is not related to a lack of preservation of plutonic and volcanic products, as detrital zircons deposited since the Cretaceous do not show any evidence of magmatic activity at that particular time interval (fig. 4). Upper Cretaceous sediments do not show evidence of remobilized volcanoclastic material (Trümpy, 1975; Caron and others, 1989; Vacherat and others, 2017). These amagmatic characteristics are illustrated in figure 2, figure 3B and figure 5A, which show a striking contrast in magmatic activity associated with oceanic subduction (for example: Sierra Nevada batholith, Western U.S.) versus Alpine-type orogens. Magmatism during convergence remains nonexistent within the Pyrenees (figs. 4 and 5A). In the Alps, magmatism is only found as dikes or small plutons along the peri-Adriatic lineament as well as in sparse volcanic debris in Paleogene flyschs. The volume of magmatism is anomalously low compared to other arc systems (fig. 3B). In the Alps, multiple-saturation experiments on primitive dikes indicate that the source of magmatism of syncollisional plutonism, at 2.7 GPa, is unusually deep, at least for the Adamello massif (Ulmer, 1988; Hürlimann and others, 2016). This is significantly deeper than most primitive magmas from circum-Pacific subduction zones, which generally form at depths corresponding to pressures <2 GPa (Ulmer, 2001; Grove and others, 2012). The lack of arc magmatism during subduction in Alpine-type orogens is a peculiarity that has been discussed in few studies (Zanchetta and others, 2012; Bergomi and others, 2015). These contributions suggest that either subduction was slow and oblique or subduction of crustal- and metamorphic lithologies inhibited magmatism. Convergence in the Alps was likely oblique and slow (Beltrando and others, 2010b; Pfiffner, 2016). However, conditions of slow, shallow subduction zones lead to increasing dehydration and/or melting of subducting slabs and significant fluxing of the mantle wedge, producing characteristic hydrous basalts, high-Mg# andesites and high-Mg# dacites in young, hot subduction zones (Kay, 1978; Wood and Turner, 2009; Yogodzinski and others, 2015; Mitchell and Grove, 2015; Yogodzinski and others, 2017). Thus, slow- and oblique subduction cannot readily account for the lack of magmatism. Since the Piemonte-Liguria basin was partly sediment–starved (Masini and others, 2013), it could be argued that subduction did not lead to flux-melting of the convective upper mantle due to the lack of subducted fluid-rich sediments. However, in sediment-poor, slow subducting environments, such as the Western Aleutian Arc, fluids are released from the hydrated mantle section of the subducting slab due to the breakdown of serpentinite (Yogodzinksi and others, 2017). In addition, magmatism in the Lesser Antilles Arc is initiated by the breakdown of subducting hydrated slow-spreading oceanic crust at 60 to 100 km depth beneath the volcanic arc (Paulatto and others, 2017). Thus, the lack of magmatism in Alpine-type orogens has to be related to the inefficient subduction of hydrated sediments and serpentinites to mantle depth and their preservation in the orogenic wedge instead.

Subduction initiation at magma-poor passive margins and hyper-extended basins.—In the Western Alps, the age of peak HP-LT metamorphism becomes younger from the internal (Adriatic) to external (European) units (figs. 4G and 5B). Continental allochthons, separated from the Adriatic continental margin by exhumed mantle, reach peak HP-LT metamorphism prior to fragments of embryonic oceanic crust (Manzotti and others, 2014 and references therein). In the Pyrenees, geophysical and field data suggest that deformation during closure of hyper-extended basins focused along former rift domain boundaries, notably at the edge of the continental crust at the location where mantle is exhumed (Tugend and others, 2014). Therefore, most geodynamic interpretations in the Alps suggest that subduction is initiated at the edge of passive margins of rift basins (Stampfli and others, 1998; Manzotti and others, 2014; Marroni and others, 2017). The mechanism behind subduction initiation at passive margins is debated (Stern and Gerya, 2018) but is classically suggested to be gravitational instability due to the contrast between the negative buoyancy of oceanic lithosphere established after ageing for ≥10 Ma (Cloos, 1993) and positively buoyant continental crust. This has, in part, been supported by analogue experiments and numerical modeling (Mart and others, 2005; Goren and others, 2008; Nikolaeva and others, 2010). However, these results have been disputed by Leng and Gurnis (2015) who argue that spontaneous subduction initiation at passive margins is inhibited due to the amount of force required for plate bending and overcoming shear coupling. These results supported earlier numerical modeling on the mechanisms of spontaneous subduction initiation by Mueller and Phillips (1991). Spontaneous subduction initiation at passive margins is also disputed by well–characterized examples indicating that (a) subduction initiation occurs oceanward (Izu-Bonin-Mariana arc, Neotethys) within crustal weaknesses (Stern and Gerya, 2018); and (b) the age of (hyper)-extended margins in the Atlantic ocean (>100Ma), which remain stable even though they are significantly older than passive margins along the Piemonte-Liguria basin upon subduction initiation (Müller and others, 2008). Thus, current models of subduction in the Alps suffer from two weaknesses. Firstly, models of buoyancy-controlled subduction initiation indicate extension of the upper plate during the initial foundering of the down-going slab, which in turn should produce magma-rich back-arc basins and/or forearc extension (Nikolaeva and others, 2008; Nikolaeva and others, 2010; Leng and others, 2012; Maffione and others, 2015b; Leng and Gurnis, 2015; Arculus and others, 2015). This is in contradiction with the upper plate (Adria) being under compression during subduction initiation at 85 to 100 Ma (Froitzheim and others, 1996; Zanchetta and others, 2012). Secondly, models do not take into account the limited width (<700 km) (Rosenbaum and Lister, 2005; Vissers and others, 2013; Manzotti and others, 2014) and the nature of these types of hyper-extended basins. In addition, the "oceanic" lithosphere is unlike Penrose-type, but is formed of exhumed and serpentinized subcontinental mantle, refertilized mantle rocks as well as oceanic core complexes formed of gabbroic and mantle rocks (fig. 6). The proximal crustal domains are thinned from 30 km to 10 km across the so-called necking-zone located continentward of the hyper-extended domain (Mohn and others, 2012). Thus, these OCTs should be less negatively buoyant or even neutrally buoyant, which will inhibit spontaneous nucleation of subduction at passive margins (Nikolaeva and others, 2010; Leng and Gurnis, 2015). Traditional models of spontaneous subduction initiation are therefore not able to answer a key question of closure of the Western Tethys: *why was subduction initiated along hyper-extended and buoyant passive margins under amagmatic- and compressive conditions?* Recently, Kiss and others (2019) presented thermo-mechanical numerical simulations of forced, or induced, subduction initiation at a hyper-extended margin. Their results support the interpretation of forced subduction of the Piemonte-Luguria basin (De Graciansky and others, 2011), presumably caused by convergence and associated compressive stresses resulting from the northward motion of Adria and Africa.

The exhumation of (ultra-)high-pressure lithologies.—Most high-pressure rocks in the Western and Central Alps form large-scale, laterally continuous, coherent tectonic units (Escher and others, 1987; Schmid and others, 1996; Pfiffner, 2014), meaning that rocks were subducted and exhumed as nappes and were not dismembered or mechanically mixed together with pieces originating from other tectonic units during transport and deformation. As shown in figures 5B and 5C, a west-to-east transect across the Western and Central Alps shows a gradual pattern of generally increasing metamorphic gradients to eclogite facies as one moves from the European margin to the Piemonte-Liguria basin and the Adratic margin (Berger and Bousquet, 2008). Convergence lead to the imbrication of nappes of internal Alpine domains thrust onto external domains separated by crustal-scale shear-zones (Avigad and others, 2003; Schenker and others, 2015; Kirst and Leis, 2017). Significant jumps in recorded pressure between individual nappes indicate that distinct nappes were juxtaposed at different intervals during exhumation (Avigad and others, 2003; Babist and others, 2006; Beltrando and others, 2010a; Brovarone and others, 2013). Subduction mélanges in accretionary channels have been inferred for the Central and Western Alps (Engi and others, 2001; Angiboust and others, 2014). However, this has been questioned by Schenker and others (2015) and Manzotti and others (2017), in part on the basis of the coherent exhumation of the Adula nappe from high-pressure (Schmid and others, 1996; Pleuger and Podladchikov, 2014) and new field observations and chemical analysis of magmatic rocks around the contact between Tsaté and Dent Blanche nappes (Manzotti and others, 2017). In most cases, the lithostratigraphic association of serpentinized mantle with gabbros, continental allochthons and post-rift basin sediments is inherited from the hyper-extended margin and not formed by a tectonic mélange during subduction (Beltrando and others, 2010a). The exhumation of (ultra-)high pressure rocks in the Alps and the preservation of rift-related hyper-extended margins is therefore at odds with subduction-mélange models in Benioff-type subduction zones (Brovarone and others, 2013; Schenker and others, 2015) precisely because they do not represent large-scale, incoherent tectonic mélanges implied by numerical models (Stöckert and Gerya, 2005) and field observations (Coleman and Lanphere, 1971).

One key assumption of most geodynamic interpretations is that peak metamorphic pressures in individual nappes are recording lithostatic pressures (Malusà and others, 2015; Weber and others, 2015; Fassmer and others, 2016; Gauthier-Putallaz and others, 2016). Because the density of lithospheric rocks varies by only *ca.* 10 percent, the lithostatic pressure is essentially a function of depth. Commonly, the highest reported metamorphic pressure of a tectonic nappe is used as geobarometric estimate of subduction depth, which is subsequently used for retrotranslations of tectonic plates (Handy and others, 2010). However, field-based and numerical model studies have shown that tectonic forces and reaction-induced volume changes could produce local stresses that can deviate significantly from the lithostatic pressure. (Mancktelow, 2008; Moulas and others, 2013; Schmalholz and Podladchikov, 2013; Moulas and others, 2014; Tajčmanová and others, 2014; Schmalholz and others, 2014; Jamtveit and others, 2018; Moulas and others, 2019; Luisier and others, 2019). These studies propose that significant jumps in recorded high-pressure conditions (typically 0.4–0.8 GPa) are recorded at the grain-scale, outcrop-scale and nappe-scale which are difficult to reconcile with traditional lithostatic models or by variably sluggish reaction-kinetics, retrogression and tectonic mixing (Tajčmanová and others, 2014; Schenker and others, 2015; Moore and others, 2019; Luisier and others, 2019). With regards to constraining exhumation rates or rapid decompression-burial cycles, deviations from lithostatic pressure could relate a part of the metamorphic pressure decrease during exhumation to mechanical decompression and hence the actual exhumation rates would be significantly decreased.

Retrotranslations and associated estimates of the width of oceans and basins, which result from lithostatic pressure and corresponding subduction depth, provide likely maximal estimates. The true subduction depth of coherent tectonic units could have been less than the lithostatic depth estimates based on minor rock volumes exhibiting (ultra)high-pressures that are not recorded by the majority of the rocks in the corresponding tectonic unit, such as the Dora Maira massif. As such, the P-T-t paths, and mechanisms of burial and exhumation of coherent (ultra)-high-pressure units within Alpine-type orogens remain contentious (Malusà and others, 2015; Schenker and others, 2015), and can therefore not be readily linked to *classical* P-T-t paths and mechanisms of burial and exhumation of sediments. Generally, deviations from the lithostatic pressure must exist during the evolution of orogens due to lateral variations in topography and crustal thickness, and due to crustal and lithospheric flexure (Schmalholz and others, 2014, 2019). However, the magnitude of these pressure variations, their impact on metamorphic reactions and, hence, their impact on lithostatic-based depth estimates is currently disputed (Moulas and others, 2013; Moulas and others, 2019). Several recent studies show how the metamorphic pressure record in tectonic units could be used to estimate pressure variations and deviations from the lithostatic pressure (Jamtveit and others, 2018, 2019; Luisier and others, 2019). A main aim of these studies is to quantify the paleo-stress state during peak metamorphic conditions and ultimately to reconstruct the burial-exhumation cycle of rocks based on tectono-metamorphic models considering dynamic stress and pressure.

Ampferer-Type Subduction Along the Western Tethys

Importance of the pre-collisional architecture of the Western Tethys.—Classical geodynamic models in the Western and Central Alps interpret a large geophysical anomaly at

~600 km depth as a remnant oceanic slab separated by slab-breakoff following continental collision, and interpret an original size of the subducted oceanic lithosphere between 550 and 1000 km in length (Stampfli and others, 1998; Handy and others, 2010). In this case, subduction would have followed intra-oceanic subduction initiation dynamics similar to intraoceanic Neotethyan subduction. Initial high-pressure metamorphism of continental fragments could then be explained by trench-related erosion of the upper plate (Polino and others, 1990; Dal Piaz, 1993). However, this would imply an intra-oceanic Benioff-type subduction zone with specific magmatic characteristics upon subduction initiation followed by arc magmatism, which is not the case (fig. 1). In addition, the Western Tethys was tectonically complex, narrow (<600–700 km wide, Vissers and others, 2013) and formed of a complex network of hyper-extended basins and domains of embryonic oceanic crust, with continental blocks (Margna-Sella and Sesia-Dent Blanche, Trümpy, 1975; Froitzheim and Eberli, 1990; Babist and others, 2006; Manzotti and others, 2014; Weber and others, 2015) forming large continental allochthons, or "extensional boudins" detached from the southern Adriatic margin and the Briançonnais continental ribbon detached from the European margin during protracted rifting (Steinmann, 1994; Galster and others, 2012) (fig. 6). These continental allochthons were separated from the distal margins by exhumed serpentinized mantle associated with variable magmatic additions (Froitzheim and Manatschal, 1996). Thus, the architecture of the Western Tethys could be described as a lithospheric-scale pinch-and-swell architecture (Trümpy, 1975) composed of continental crustal fragments forming the swells, or boudins, and exhumed mantle in the region of the pinch, or inter-boudin, structures where the continental crust has been separated. This architecture would not necessarily indicate a complex amalgamation of independent, short-lived, tectonic continental micro-plates in the Western Alps or in the Eastern Alps (Kurz, 2006; Handy and others, 2010; Hosseinpour and others, 2016). Similarly, the pre-convergence lithospheric structure of the Pyrenees was formed of smaller, disconnected rift-basins involving extreme crustal thinning and local subcontinental mantle exhumation at the seafloor that all together did not exceed *ca.* 100 to 200 km width (Lagabrielle and Bodinier, 2008; Roca and others, 2011; Tugend and others, 2015; Nirrengarten and others, 2018) (table 1).

The closure of these Pyrenean and Alpine basins was therefore likely not the result of spontaneous subduction initiation due to cooling oceanic lithosphere but rather resulted from a compressive environment, or so-called forced subduction, of hyper-thinned continental crust locally separated by embryonic, ultra-slow spreading domains. Such a compressive tectonic environment was likely related primarily to the northern movement of Africa and Adria between 100 to 80 Ma (Rosenbaum and others, 2002; Rosenbaum and Lister, 2005; Hosseinpour and others, 2016). Alternatively, or in addition to these large-scale tectonic forces, foundering of gravitationally unstable subcontinental lithosphere underlying Adria could have helped trigger a forced "intra-continental" subduction (Stüwe and Schuster, 2010). Nevertheless, in both scenarios, subduction in the Pyrenees and Alps was initiated by the forced subduction of inherited, subcontinental mantle and continental fragments, or Ampferer-type continental subduction (Trümpy, 1975; McCarthy and others, 2018).

Numerical simulations of Ampferer-type subductions: The importance of serpentinites.—Rift basins forming the Western Tethys offer an array of potential weaknesses enabling the initiation of subduction. For example, extensional detachment systems capping the exhumed mantle and rooted beneath hyperthinned crust may act as a guide to localize the deformation during convergence (Mohn and others, 2011; Tugend and others, 2014; Beltrando and others, 2014). In addition, numerical models indicate that ocean-continent transitions are weaker than the crustal necking zone of the passive margin (Burov and Poliakov, 2001; Leroy and others, 2008). This is also consistent with

TABLE 1

Summary of characteristics of Pyrenees, Alps (Alpine-type orogens) and Wadati-Benioff-type subductions

Characteristics	Pyrenees	Alps	Oceanic lithosphere subductions
Ocean/basin width (km)	50-200 km	< 600-700 km	>1000 km
Pre-subduction lithosphere	Rift basins	Rift basins + embryonic ultra-slow spreading ocean crust	Oceanic lithosphere
Magmatism during rifting/spreading	Exhumed SCLM, limited alkaline magmatism; no mid-ocean ridge	Exhumed SCLM, sparse MORB magmatism; short-lived mid-ocean ridge	Formation of mature oceanic crust, mature mid-ocean ridge
Initial 20 Myr markers of subduction initiation	Basin compression	Basin + upper plate compression	Distinct magmatism, mature arc, upper-plate extension, SSZ ophiolites, metamorphic soles
HP-LT metamorphism	No	Yes, coherent units (nappes)	Yes, subduction mélanges, suture zones
Magmatism	No	Only collisional	Yes
Magmatism volume (km3km-1Myr-1)	No	<0.1	20-1000
Back-arc basins	No	No	Yes possible, with spreading
Subduction initiation location	Passive margin	Passive margin	Intra-oceanic (e.g. detachment faults, transform faults, buoyant plateaus, remnant arcs)
Tomography imaging	<100 km slab	250 km slab ?	Well imaged, down to 600 km ?
What is subducted	Dry mantle lithosphere (subcontinental)	Dry mantle lithosphere (subcontinental ± oceanic)	Hydrated oceanic lithosphere, oceanic sediments
Type of subduction	Ampferer-type continental subduction	Ampferer-type continental subduction	Benioff-type oceanic subduction

HP-LT metamorphism = high-pressure and low-temperature metamorphism (blueschist-eclogite facies); MORB = mid-ocean ridge basalts.

the incipient stages of convergence observed along the Bay of Biscay-Pyrenean system (Thinon and others, 2001; Gallastegui and others, 2002; Tugend and others, 2014) where deformation is accommodated primarily in the exhumed mantle domain along the edge of the continental crust (Tugend and others, 2014; Tugend and others, 2015).

A crucial factor upon subduction in small basins is the lack of efficient subduction of hydrated and buoyant lithologies to great depth. Figures 7 and 8 illustrate a numerically modeled cycle of rifting (50 Myrs with absolute extension velocity of 1 cm/yr), cooling (60 Myrs with no far-field velocity) and subsequent convergence (1.5 cm/yr absolute convergence velocity) resulting in collision of Alpine-type orogens. Figures 7A–7D shows a model of this cycle without employing a serpentinization front at the top of the exhumed mantle at the end of cooling. Figures 8A–8C show the same model, except that a 7.5 km thick serpentinization front on top of the exhumed mantle is added at the end of cooling. The serpentinization front is 7.5 km thick because with the applied numerical resolution of 1 × 1 km the internal deformation of a layer with smaller thickness cannot be resolved. Embrittlement of stretched exhumed mantle leads to pervasive serpentinization of mantle rocks at the seafloor (Pérez-Gussinyé and Reston, 2001) with serpentinization reaching depths of 8 km (Andréani and others, 2007). This leads to a several km-thick layer of serpentinized peridotites (figs. 6 and 8). Such depths of serpentinization are consistent with geophysical imaging of the southern North Atlantic ocean-continent transition zones (Chian and others, 1999). (Ultra-)slow rifting at absolute extension velocities of 1 cm/yr leads to crustal thinning, mantle exhumation and the formation of *ca.* 360 km wide basin of exhumed mantle bounded by thinned continental margins, a size similar to the estimated width of the Piemonte-Liguria basin (Manzotti and others, 2014). Rifting is followed by a period of cooling, consistent with an observed lag between MORB-magmatism (*ca.* 155 Ma) and subduction initiation (*ca.* 100 Ma) in the Alpine domain (fig. 4). The cooling period allows for the formation of stable, thermally relaxed basins. After thermal cooling follows a general phase of compression, modeled with an absolute convergence velocity of 1.5 cm/yr, consistent with the slow and oblique subduction proposed for the Alps (Stampfli and others, 1998; Carry and others, 2011) and evidence of compressional upper-plate dynamics at 100 Ma (Zanchetta and others, 2012). Our numerical simulations show that an initial lithospheric structure with no initial weakness will, under compression, lead to subduction initiation along weak hyper-extended continental margins (see also Kiss and others, 2019). In our model, focusing of deformation and propagation of the subduction interface at depth is triggered by shear-heating (Burg and Schmalholz, 2008; Thielmann and Kaus, 2012; Jaquet and Schmalholz, 2018; Kiss and others, 2019) as illustrated by the perturbation of isotherms along the subduction interface. Strain localization and subduction initiation is controlled in our model by thermal softening, which is a consequence of energy conservation in combination with dissipative deformation. We focus on macroscale (>> mineral grain size) processes and do not consider microscale processes such as softening by grain size reduction. Additional microscale processes would generally intensify the strain localization (Thielmann and others, 2015).

In the model without serpentinization, a significant volume of sediments is accreted in the trench (fig. 7C) and incorporated in the orogenic wedge during continent-continent collision (fig. 7D). Although the model with serpentinization of the exhumed mantle evolved in the same way as the reference model during the rifting and cooling period, convergence leads to distinct outcomes. Serpentinites are significantly less viscous than peridotites, and even a small percentage of serpentinization (10–15%) of peridotites drastically lowers the brittle strength of the subcontinental mantle (Escartin and others, 2001). Serpentinites are thus sheared off to a large extent

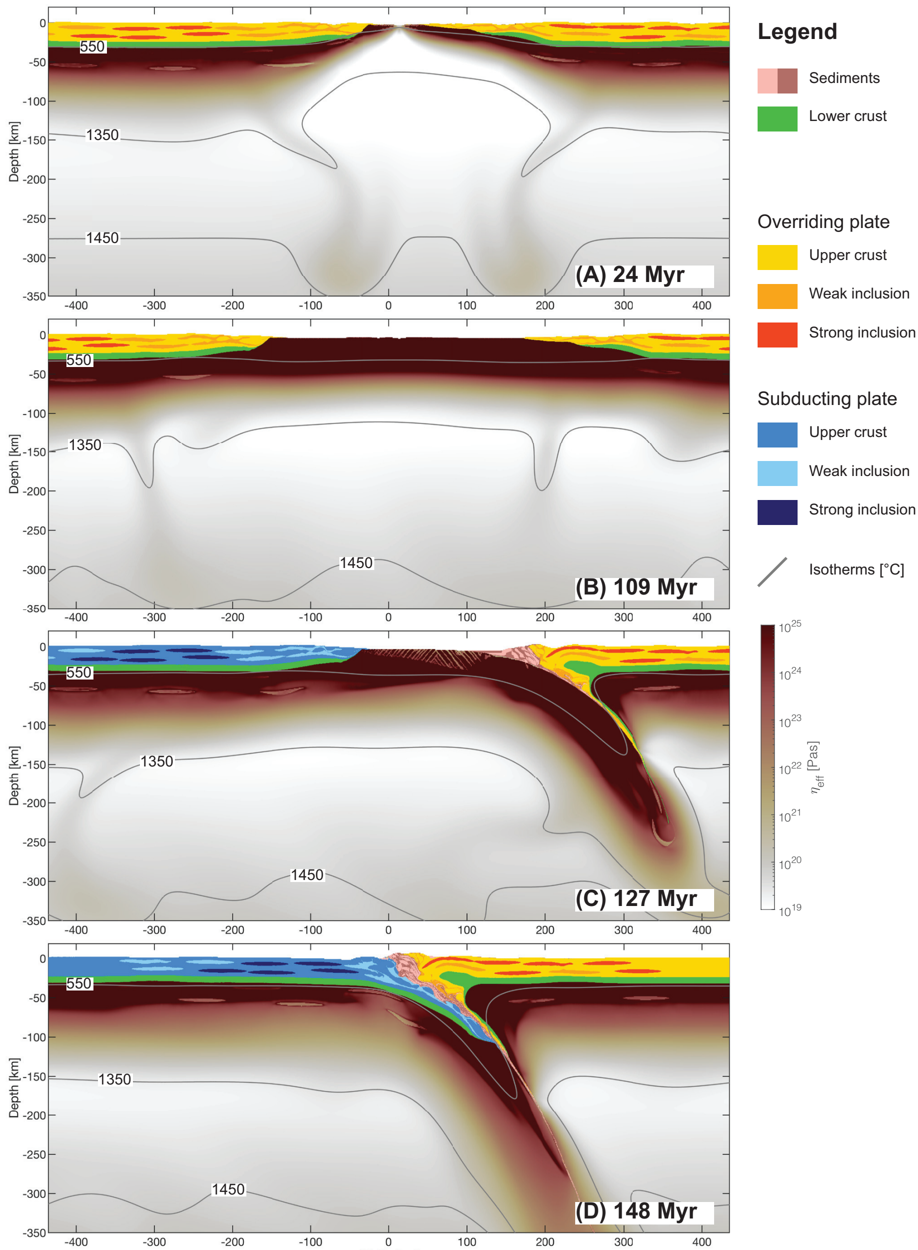


Fig. 7. Numerical modeling illustrating the extension and compression of a narrow rift basin without the presence of serpentinization following mantle exhumation. Crustal composition patch on top of the effective viscosity field of the mantle lithosphere and asthenosphere calculated by the applied numerical algorithm for a model without a serpentinization front (A) at the stage of rifting and crustal separation; (B) after thermal relaxation; (C) during subduction and (D) at the onset of continent-continent collision. Yellow colors represent the upper crust, dark yellow colors the weak and orange colors the strong inclusions, green colors the lower crust and brown and salmon colors the sediments. Note that the color coding of the upper crust and inclusions changes from (C) onward: light to dark blue colors indicate upper crustal phases of the subducting plate and yellow-orange colors indicate upper crustal phases of the overriding plate. White to red colors show values for effective viscosity on a logarithmic scale and the gray contour lines denote isotherms. Further details can be found in the Appendix.

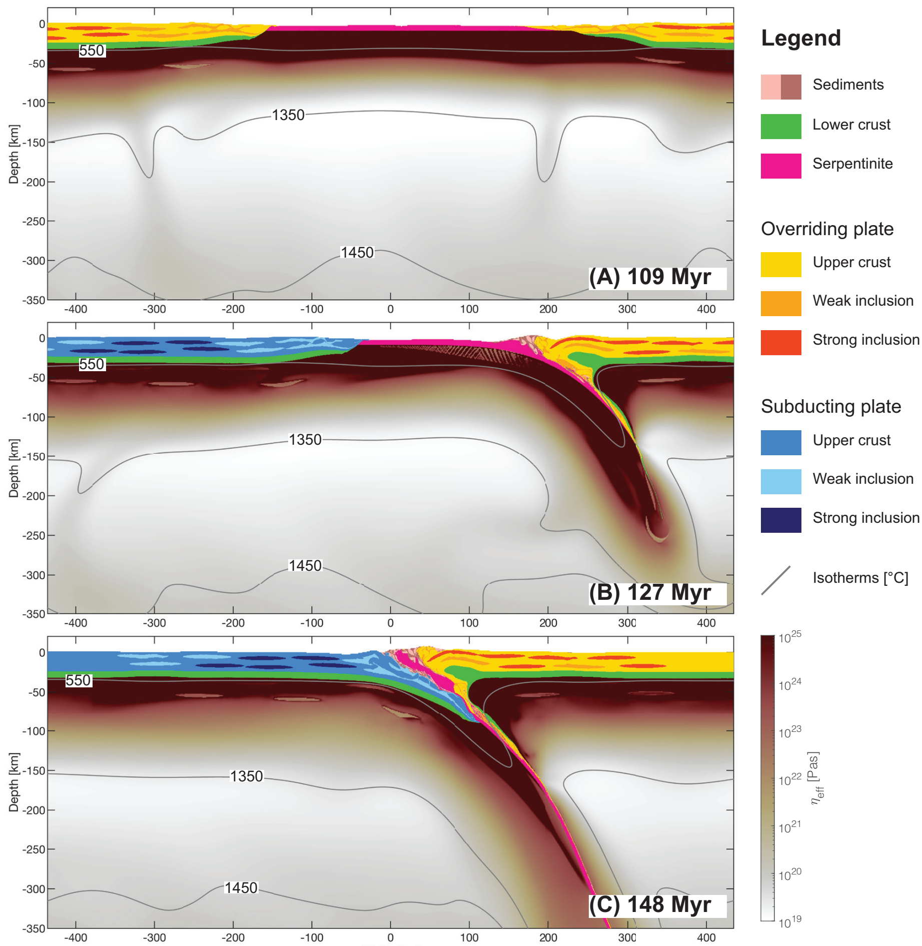


Fig. 8. Numerical modeling illustrating the extension and compression of a narrow rift basin with the presence of a thick serpentinite layer following mantle exhumation. Crustal composition patch on top of the effective viscosity field of the mantle calculated by the applied numerical algorithm for a model with a serpentinization front (A) at the end of thermal relaxation after the emplacement of the serpentinite layer, (B) during subduction and (C) at the onset of continent-continent collision. Magenta colors represent the serpentinite layer. Other colors as in fig. 7. Note that the color coding of the upper crust and inclusions changes from (B) onward: light to dark blue colors indicate upper crustal phases of the subducting plate and yellow-orange colors indicate upper crustal phases of the overriding plate.

and decoupled from the non-serpentinized mantle during subduction (figs. 8B and 8C), as proposed in McCarthy and others (2018). The shearing-off of the serpentinized mantle causes a shallower trench and, hence, less sediment accumulation in the trench (fig. 8B). Additionally, serpentinites are more buoyant than peridotitic mantle, which inhibit its burial to great depths. The majority of the serpentinized mantle is incorporated into the orogenic wedge and only a thin layer of serpentinized mantle remains coupled to the slab and is subducted to greater depths (fig. 8C), because with the numerical resolution of 1 × 1 km, the shear zone at the base of the serpentinites will have a thickness of a few kilometers. For increased numerical resolution, the thickness of the sheared serpentinite layer coupled to the non-serpentinized mantle is expected to decrease and significantly less volume of serpentinites is expected to be buried to great depths. In consequence, only the dry lithospheric mantle is subducted and the majority of serpentinites and hydrated sedimentary deposits are incorporated in the growing orogenic wedge (fig. 8C). Thus, our numerical models indicate that the majority of serpentinites will be sheared-off prior to reaching 80 km depth, consistent with numerical models showing tectonic slicing upon subduction of serpentinized Penrose-type oceanic crust (Vogt and Gerya, 2014; Ruh and others, 2015). At such depths, temperatures were probably not high enough to induce significant dehydration of serpentinites causing flux melting and arc magmatism (Ulmer and Trommsdorff, 1999) and are overall consistent with eclogite facies metamorphism recorded in the Western and Central Alps (Berger and Bousquet, 2008). With ongoing shortening, only minor amounts of crustal material are subducted to depths >80 km (fig. 8C). Incorporation of the majority of the serpentinite volume in the orogenic crustal wedge (fig. 8C), while mostly dry lithosphere is subducted to mantle depth, provides an explanation for the lack of arc magmatism and traditional characteristics of Benioff-type intra-oceanic subduction initiation events.

A Spectrum of Subduction Zone Systems: Why Size Matters

Two end-member scenarios for subduction processes in the Western and Central Alps emerge. One scenario envisages that Benioff-type oceanic subduction lead to the near-complete subduction and destruction of an oceanic domain, including >95 percent of oceanic lithosphere and 50 to 70 percent of sediments (Agard and others, 2009). Alternatively, a model of Ampferer-type subduction implies a near-complete preservation of the upper section of hyper-extended basin domains, including continental allochthons, km-thick layers of serpentinites and marine sediments. This model suggests that self-sustained subduction did not develop, with shortening accommodated by reactivation and inversion of rift basins and formation of thrusts leading to coherent imbrication of nappes (Tricart and Lemoine, 1986; Mohn and others, 2012; Brovarone and others, 2013; Schenker and others, 2015; Lagabrielle and others, 2015).

Benioff-type subduction models of the Alps are based on a classical plate-tectonic framework (Stampfli and Marthaler, 1990; Schmid and others, 1996). In such a model, the generation of an ocean basin is due to oceanic spreading during lateral drifting of passive continental fragments (Europe–Adria). The rigid oceanic lithosphere becomes a key driver of plate-tectonic movement upon subduction, magmatism and continental collision (Kissling and Schlunegger, 2018). However, contrary to the horizontal rigid body motion of oceanic lithosphere, continental interiors can be affected by significant diffuse deformation and rotation upon compression and/or extension, a feature defined as continental tectonics by Molnar (1988). The Western Tethys, from the Central Alps to the Pyrenees, was a product of extreme continental crustal extension, leading to the formation of 10s to 100s of km wide hyper-extended basins floored by hyper-thinned crustal allochthons, exhumed subcontinental mantle and minor oceanic domains similar to ultra-slow spreading seafloor. Convergence did not lead to (significant) destruction of the subducting plate, but to the imbrication of rifted margins within a collisional orogen (Beltrando and others, 2010a; Tugend and others, 2014; Mohn and others, 2014). Thus, Alpine-type orogens and Ampferer-type subduction might be the products of extreme intracontinental extension, ultra-slow plate separation and compression of mainly hyper-extended continental domains.

In table 1 and in figure 9, we summarize key parameters of continental Ampferer-type or oceanic Benioff-type subduction. Figure 9 highlights the variations in subduction processes from the Pacific or Neotethys to the Alpine and Pyrenean systems. In large oceanic systems, such as the Neotethyan and Western Pacific subduction zones, and smaller oceanic subductions within the Mediterranean (Faccenna and others, 2001a, 2001b; Faccenna and others, 2003; Royden and Faccenna, 2018), intra-oceanic

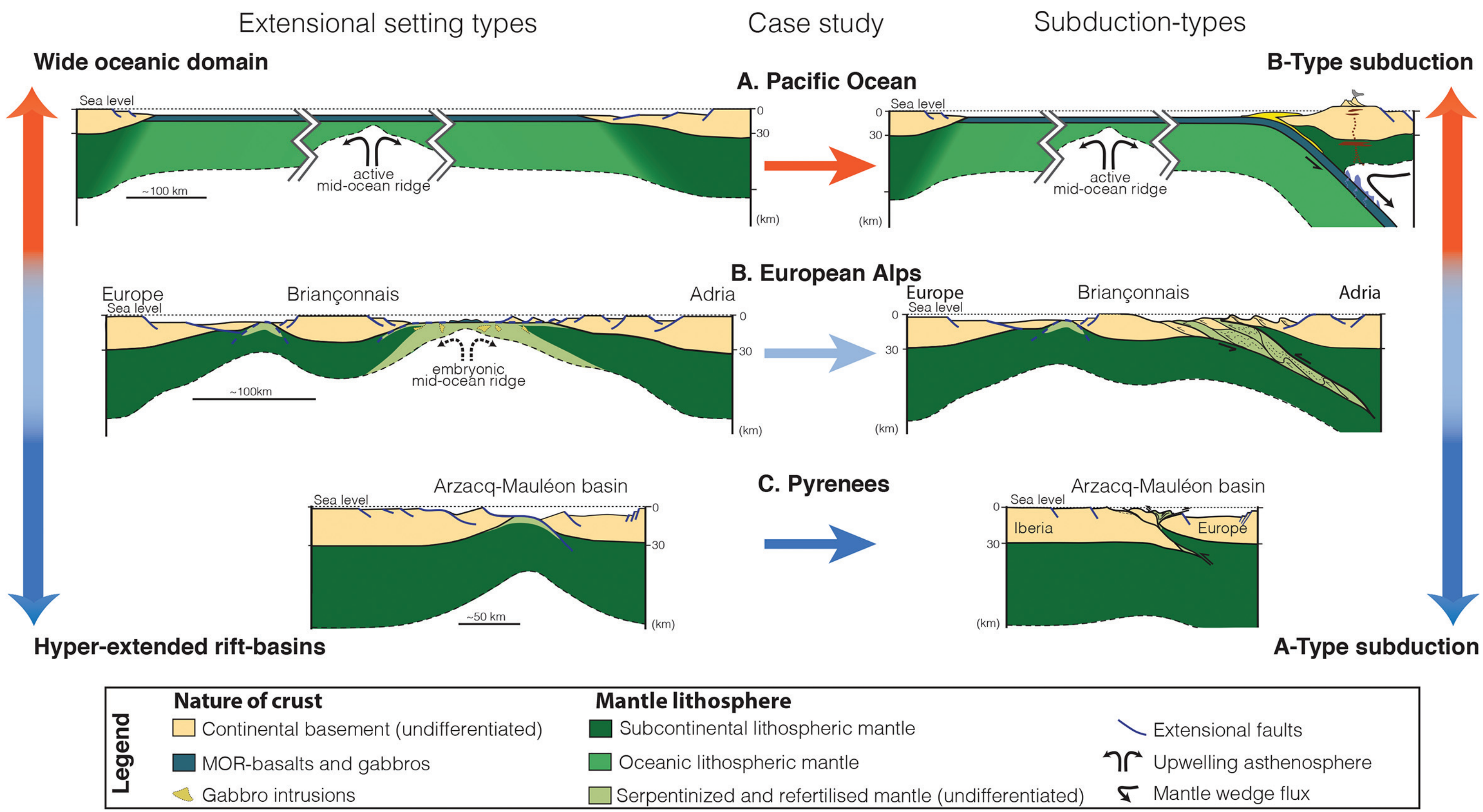


Fig. 9. Illustration of two end-members types of subduction. (A) Benioff-type subduction, related to the subduction of wide oceanic domains, efficient subduction of hydrated lithologies and voluminous arc magmatism; and (C) Ampferer-type subduction related to the closure of narrow hyper-extended rift basins leading to amagmatic basin closure, subduction initiation at passive margins and inefficient subduction of hydrated lithologies; (B) the Western Alps represent a transitional environment between both end-members, where wider hyper-extended basins and embryonic oceanic domains lead to high-pressure and low-temperature metamorphism and localized collisional magmatism, but convergence is still accommodated by amagmatic subduction initiation at passive margins as well as inefficient subduction of hydrated lithologies. The Western Tethys is modified from Mohn and others, 2012 and Schenker and others, 2015) and Pyrenees from Tugend and others (2014).

subduction initiation along weaknesses in oceanic lithosphere allows for rapid propagation of subduction along strike and initiation of magmatism (Zhou and others, 2018; Ishizuka and others, 2018; Reagan and others, 2019). Such Benioff-type subduction allows for the efficient subduction of hydrated lithologies into the convective upper mantle, allowing for significant arc magmatism. Prolonged oceanic subduction allows for upper-plate extension inducing back-arc rifting and spreading. On the other hand, Alpine-type orogens such as the Pyrenees and the Alps, are characterized by narrow, buoyant basins floored by serpentinized mantle, thinned continental crust and occasional large crustal fragments. Such extremely thinned domains along magma-poor OCTs and rift-basins are mechanically weak regions that are key for focusing deformation during compression. This is consistent with deformation and high-grade metamorphism along passive margins being focused at the hyper-thinned margin (Mohn and others, 2012; Beltrando and others, 2014; Mohn and others, 2014; Tugend and others, 2014). Overall, this implies that both hydrated, ultramafic lithologies and narrow oceanic basins or hyper-extended basins are a pre-requisite for the formation of Ampferer-type subduction, wherease larger oceanic domains will invariably lead to Benioff-type oceanic subduction zones.

CONCLUSIONS

We argue that many "traditional" markers of Benioff-type oceanic subduction are absent within Alpine-type orogens such as the Alps or Pyrenees. Subduction initiation and basin closure in Alpine-type orogens is amagmatic and occurs along passive margins, which is inconsistent with Benioff-type models of subduction. These peculiar features of ocean closure can be explained by the pre-collisional architecture of rift basins and the nature of the oceanic crust. The Western Tethys was formed of narrow, hyper-extended continental rift basins floored by serpentinized subcontinental mantle, isolated crustal allochthons and laterally heterogeneous ultra-slow spreading oceanic lithosphere. In such environments, subduction is unlikely to have resulted from spontaneous subduction initiation due to the buoyant architecture of passive margins and the small basin size. The closure of such basins was likely supported by the reactivation of inherited, rift-related structures during a general phase of compression induced by the large-scale movement of Iberia, North Africa and Adria. The formation of Alpine-type orogens was related to the closure of hyper-extended rift basins that resulted from significant internal extensional deformation of the continental lithosphere. Induced, or forced, subduction of "dry" subcontinental mantle and inefficient subduction of hydrated lithologies, which were accreted to the nascent orogenic wedge, inhibited magmatism during the closure of the Western Tethys. The Pyrenees as well as the Western and Central Alps might well define the type localities of the previously known "Ampferer-type"continental subduction. Hence, we argue that the geodynamic cycle of extension and subsequent convergence, which formed the Pyrenees and Western and Central Alps, was dominated by extreme deformation of the continental lithosphere and oceanic lithosphere played a minor role only.

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APPENDIX

1. Method and initial configuration of the numerical simulations

In the applied numerical algorithm (see also Schmalholz and others, 2019), the equations for continuity and force balance without inertial forces for an incompressible viscoelastoplastic fluid are discretized on a two-dimensional finite difference grid in the Cartesian coordinate system as

$$\frac{\partial v_i}{\partial x_i} = 0 \tag{A1}$$

$$\frac{\partial \sigma_{ij}}{\partial x_j} = -\rho \mathbf{a}_i, \tag{A2}$$

where v_i denotes velocity vector components x_i denotes the spatial coordinate vector component with {i,j} = 1 indicating horizontal and {i,j} = 2 indicating vertical direction, respectively, ρ is density and $\mathbf{a} = [0; g]$ is a vector with g being the gravitational acceleration. Constitutive relationships are formulated to relate stresses to strain rates as

$$\sigma_{ij} = -P + 2\left(\frac{1}{\eta} + \frac{1}{G\Delta t}\right)^{-1} \dot{\varepsilon}_{ij} + \left(1 + \frac{G\Delta t}{\eta}\right)^{-1} \sigma_{ij}^{o} + J_{ij}, \tag{A3}$$

where σ_{ij} are the components of the total stress tensor, P is pressure (negative mean stress), η is effective viscosity, G is the elastic shear modulus and Δt is the numerical time step, σ_{ij}^{o} are stress tensor components from the previous time step and J_{ij} comprises all terms resulting from the Jaumann stress rate. A viscoelastic Maxwell model is used to describe the rheology, implying that the components of the deviatoric strain rate tensor $\dot{\varepsilon}_{ij}$ are additively decomposed into contributions from the viscous (dislocation, diffusion, exponential creep), plastic and elastic deformation as

$$\dot{\varepsilon}_{ij} = \dot{\varepsilon}_{ij}^{dis} + \dot{\varepsilon}_{ij}^{dif} + \dot{\varepsilon}_{ij}^{pei} + \dot{\varepsilon}_{ij}^{pla} + \dot{\varepsilon}_{ij}^{ela}. \tag{A4}$$

The effective viscosity for the dislocation and exponential – or Peierls – creep law mechanisms is a function of the respective strain rate invariant $\dot{\varepsilon}_{II}^{dis,pei} = \tau_{II}/(2\eta^{dis,pei})$,

$$\eta^{dis} = f\frac{2^{\frac{1-n}{n}}}{3^{\frac{1+n}{2n}}} A \left(\dot{\varepsilon}_{II}^{dis}\right)^{\frac{1}{n}-1} \exp\left(\frac{Q+PV}{nRT}\right) (f_{H2O})^{-\frac{r}{n}}, \tag{A5a}$$

where the ratio in front of the strain rate invariant pre-factor A results from the conversion of the one-dimensional flow law, obtained by experimental rock deformation, to a flow law for tensor components. The Peierls viscosity is computed as

$$\eta^{pei} = \frac{2^{\frac{1-s}{s}}}{3^{\frac{1+s}{2s}}} \hat{A} \left(\dot{\varepsilon}_{II}^{pei}\right)^{\frac{1}{s}-1}, \tag{A5b}$$

where the effective stress exponent s is

$$s = 2\gamma \frac{Q}{RT}(1-\gamma) \tag{A6}$$

and the strain rate pre-factor $\hat{A}$ is defined as

$$\hat{A} = \left[A \exp\left(-\frac{Q(1-\gamma)^2}{RT} \right) \right]^{-\frac{1}{s}} \gamma \sigma^{\mathrm{pei}}, \tag{A7}$$

where γ is a fitting parameter. For the mantle lithosphere and asthenosphere, we also take diffusion creep of the material into account and the corresponding viscosity is formulated as

$$\eta^{\mathrm{dif}} = A d^m \exp\left(\frac{Q+PV}{RT} \right) (f_{\mathrm{H2O}})^{-\frac{\mathrm{r}}{\mathrm{n}}}, \tag{A8}$$

where d is a constant grain size and m is a grain size exponent.

Stresses are limited by a yield stress τ_y obtained by a Drucker-Prager criterion

$$\tau_y = C\cos(\varphi) + P\sin(\varphi) \tag{A9}$$

where C is cohesion and φ is the internal friction angle. Above the plastic limit we use an iterative procedure to reduce the effective viscosity until the invariant of the deviatoric stress tensor components equals the yield stress. Consequently, the plastic viscosity is only computed for $\tau_{II} \geq \tau_y$ and takes the following expression

$$\eta^{\mathrm{pla}} = \frac{\tau_y}{2\dot{\varepsilon}_{\mathrm{II}}^{\mathrm{pla}}} \{\tau_{\mathrm{II}} \geq \tau_y. \tag{A10}$$

When the iteration cycle is completed the effective viscosity in (A3) is computed as the inverse of the pseudo-harmonic average of the dissipative viscous and plastic contributions for the corresponding invariants of the strain rate tensor components of the distinct deformation mechanisms as

$$\eta = \left(\frac{1}{\eta^{\mathrm{dis}}(\dot{\varepsilon}_{\mathrm{II}}^{\mathrm{dis}})} + \frac{1}{\eta^{\mathrm{dif}}(\dot{\varepsilon}_{\mathrm{II}}^{\mathrm{dif}})} + \frac{1}{\eta^{\mathrm{pei}}(\dot{\varepsilon}_{\mathrm{II}}^{\mathrm{pei}})} + \frac{1}{\eta^{\mathrm{pla}}(\dot{\varepsilon}_{\mathrm{II}}^{\mathrm{pla}})} \right)^{-1}. \tag{A11}$$

In order to fully couple mechanical and thermal processes, the change of temperature with time is computed with the heat transport equation as

$$\rho c_V \frac{DT}{Dt} = \frac{\partial}{\partial x_i}\left(k \frac{\partial T}{\partial x_i} \right) + H_D + H_R + T\alpha\rho g v_z, \tag{A12}$$

with c_V being the specific heat capacity at constant volume V, DT/Dt is the material time derivative of temperature T, k is thermal conductivity and H_D and H_R are additional source terms for dissipative work and radiogenic heat production, α is the coefficient of thermal expansion and v_z is the vertical component of the velocity vector. A list of all physical parameters is given in table A1.

Effective density of the material phases is a function of pressure and temperature calculated via a linearized equation of state for the crustal phases, the sediments and the serpentinite phase as

$$\rho(P,T) = \rho_0(1 - \alpha[T - T_0])(1 + \beta[P - P_0]), \tag{A13}$$

where ρ_0 is the density at reference pressure P_0 and reference temperature T_0 for a given material, β is the coefficient of compressibility. In case of the mantle phase the effective density is precomputed using phase diagrams for given bulk rock compositions. These look-up tables are calculated using the software package PERPLE_X (Connolly, 2005, 2009). In the simulation, the model density is read in from these thermodynamic look-up tables for density according to the local pressure and temperature values at each grid cell. A marker-in-cell advection scheme is applied to move the

TABLE A1

Material parameters used in the numerical simulations

Parameter	Units	Upper crust[1]	Lower crust[1]	Strong inclusion[2]	Weak inclusion[3]	Calcite sediments[4]	Mica sediments[5]	Strong mantle[6,7]	Weak mantle[6,7]	Serpentinite[8]
ρ_0	Kgm^{-3}	2800	2900	2800	2800	2800	2800	-	-	2585
G	Pa	$2x10^{10}$	$2x10^{10}$	$2x10^{10}$	$2x10^{10}$	$2x10^{10}$	$2x10^{10}$	$2x10^{10}$	$2x10^{10}$	$1.81x10^{10}$
c_V	$JK^{-1}kg^{-1}$	1050	1050	1050	1050	1050	$2x10^{10}$	1050	1050	1050
k	$Wm^{-1}K^{-1}$	2.25	2.23	2.25	2.25	2.37	2.55	2.75	2.75	2.75
Q_R	Wm^{-3}	$1.02x10^{-6}$	$0.26x10^{-6}$	$1.02x10^{-6}$	$1.02x10^{-6}$	$0.56x10^{-6}$	$2.9x10^{-6}$	$2.1139x10^{-8}$	$2.1139x10^{-8}$	$2.1139x10^{-8}$
C	Pa	$1x10^{7}$	$1x10^{7}$	$1x10^{7}$	$1x10^{6}$	$1x10^{7}$	$1x10^{7}$	$1x10^{7}$	$1x10^{7}$	$1x10^{7}$
φ	°	30	30	30	5	30	15	30	30	25
α	K^{-1}	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$	$3x10^{-5}$	$4.7x10^{-5}$
β	Pa^{-1}	$1x10^{-11}$	$3x10^{-5}$	$1x10^{-11}$	$1x10^{-11}$	$1x10^{-11}$	$3x10^{-5}$	$1x10^{-11}$	$1x10^{-11}$	$1x10^{-11}$
d	m	$1x10^{-3}$	$3x10^{-5}$	$1x10^{-3}$	$1x10^{-3}$	$1x10^{-3}$	$1x10^{-3}$	$1x10^{-3}$	$1x10^{-3}$	$1x10^{-3}$
A^{dis}	$Pa^{-n}s^{-1}$	$3.9811x10^{-16}$	$3.9811x10^{-16}$	$5.0477x10^{-28}$	$5.0717x10^{-18}$	$1.5849x10^{-25}$	$1x10^{-138}$	$1.1x10^{-16}$	$5.6786x10^{-27}$	$4.4738x10^{-38}$
n^{dis}	-	3.0	3.0	4.7	2.3	4.7	18	3.5	3.5	3.8
V^{dis}	m^3mol^{-1}	0	0	0	0	0	0	$14x10^{-6}$	$11x10^{-6}$	$3.2x10^{-6}$
Q^{dis}	$kJmol^{-1}$	356	356	485	154	297	51	530	480	8.9
f^{dis}	-	0.3	1	1	1	1	1	1	1	30
r^{dis}	-	0	0	0	0	0	0	0	1.2	0
$f^{dis}_{H_2O}$	Pa	0	0	0	0	0	0	0	$1x10^{9}$	0
A^{dif}	$Pa^{-n}s^{-1}$	-	-	-	-	-	-	$1.5x10^{-15}$	$2.5x10^{-23}$	-
n^{dif}	-	-	-	-	-	-	-	1	1	
m^{dif}	-	-	-	-	-	-	-	3.0	3.0	-
V^{dif}	m^3mol^{-1}	-	-	-	-	-	-	$7.5x10^{-6}$	$9x10^{-6}$	-
Q^{dif}	$kJmol^{-1}$	-	-	-	-	-	-	370	375	-
d^{dif}	m	-	-	-	-	-	-	10^{-3}	10^{-3}	
r^{dif}	-	-	-	-	-	-	-	0	1	-
$f^{dif}_{H_2O}$	Pa	-	-	-	-	-	-	0	$1x10^{9}$	-
A^{pei}	s^{-1}	-	-	-	-	-	-	$5.7x10^{11}$	$5.7x10^{11}$	-
γ	-	-	-	-	-	-	-	0.1	0.1	-
σ_p	Pa	-	-	-	-	-	-	$8.5x10^{9}$	$8.5x10^{9}$	-
Q^{pei}	$kJmol^{-1}$	-	-	-	-	-	-	540	540	-

[1] (Rybacki and Dresen, 2004), [2] (Mackwell, and others, 1998), [3] (Ranalli, 1995), [4] (Schmid and others, 1977), [5] (Kronenberg and others, 1990), [6] (Hirth and Kohlstedt, 2004), [7] (Goetze and Evans, 1979) regularized after (Kameyama and others, 1999), [8] (Hilairet and others, 2007).

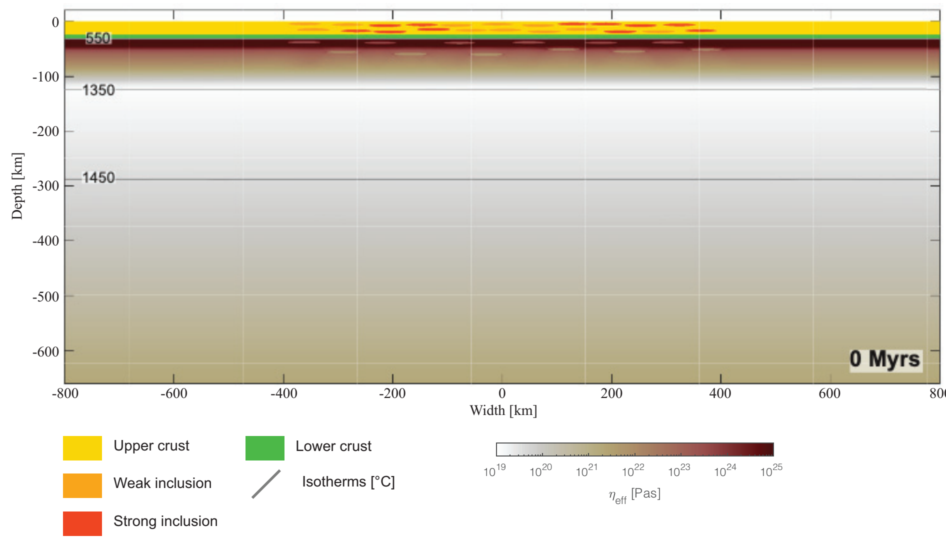

Fig. A1. Initial model configuration. Crustal phase patch on top of effective viscosity field computed for the lithosphere and upper mantle by the applied numerical algorithm. Yellow colors indicate the upper crust, dark yellow colors the weak and orange colors the strong inclusions and green colors the lower crust. White to red colors show values for effective viscosity on a logarithmic scale. Gray contour lines indicate isotherms. Elliptical inclusions are emplaced with random distance between each other. The strength of sequential inclusions is varied randomly (weak or strong). Mechanical heterogeneities in the mantle lithosphere are parameterized by elliptical inclusions of weak mantle rheology (light ellipses in viscosity field).

Lagrangian markers which carry the material properties through the Eulerian grid. The numerical algorithm has been used to model lithospheric extension (Duretz and others, 2016) and to model the stability of the Tibetan plateau estimating the strength of the lithosphere by calculating crustal and lithospheric stresses (Schmalholz and others, 2019).

In this study, we model the evolution of an Alpine-type cycle consisting of the following three deformation stages: First, we model the evolution of a rift basin bounded by hyper-extended passive margins in an extensional deformation phase over a total period of 50 Myrs. Second, the evolved basin and hyper-extended margin system is thermally relaxed over 60 Myrs. During this relaxation period, no far field deformation is applied to the system. In one of the two presented numerical simulations, the rheology of top-most layer of exhumed mantle (dry olivine) is replaced with an antigorite flow law at *ca.* 1 Myr before the onset of convergence. To ensure sufficient numerical resolution of the antigorite layer, its thickness is set to an initial value of 7.5 km. Third, the thermally relaxed passive margin system is used as a self-consistently developed initial model configuration to model convergence, subduction initiation via shear heating and orogenic wedge formation during continent-continent collision.

Figure A1 shows the initial model configuration. The upper crust is initially 25 km thick and contains mechanically strong and weak elliptical inclusions. An 8 km thick lower crust beneath the upper crust overlies a 87 km thick dry peridotitic lithospheric mantle containing weaker elliptical inclusions of wet peridotite. The 1350 °C isotherm marks the lithosphere-asthenosphere boundary and lies initially at a depth of 120 km. The domain extends over 1600 km in width and 680 km in depth and the equations are discretized on a regular rectangular grid consisting of 1601 nodal points in horizontal and 681 nodal points in vertical direction, where the initial surface level is set at 0 km height and 20 km space is left free above to allow for evolution of topography. Surface evolution is modeled using a free surface (stress-free) boundary condition projected onto a surface contour line. Surface processes (for example, sedimentation, erosion) are taken into account by correcting the topography with a constant rate of 0.5 mm/yr upward and downward in case the predicted topography lies below a level of 5 km or above a level of 2 km, respectively. In case of upward correction, the generated free space between old and new topographic level is filled with sediments, whereas in case of downward correction all material points above the corrected topography are deleted. Free slip boundary conditions with constant material flow are applied at the right and left model boundary. Inflow or outflow velocities are set to a constant value of 1 cm/yr during extension and 1.5 cm/yr during compression as absolute values at the upper half of the two vertical boundaries. To ensure conservation of volume flowing through the boundaries, the inflow or outflow velocity through the lower half of the boundaries is calculated accordingly. The mechanical boundary condition at the bottom is set to free slip, that is the boundary is assumed to be impermeable. Temperature at the Moho is set initially to 550 °C and the temperature throughout the upper mantle and transition zone increases adiabatically. The adiabatic gradient can be either calculated analytically (Trubitsyn and Trubitsyna, 2015), or extracted from the thermodynamical data calculated for the bulk rock composition of the mantle. During the first time step, 10 internal thermal diffusion cycles with a numerical time step of 100 Myrs are performed to reach thermal equilibrium for the initial temperature field. In order to mimick thermal convection in the upper mantle during this thermal equilibration procedure, the thermal conductivity in the asthenosphere is set to unrealistically high values. Once a thermal equilibrium is reached the conductivity in the asthenosphere is set back to its original realistic value and the extensional phase is modeled. Thermal boundary conditions on the right and left boundary are set to zero heat flux and the temperature is kept constant at the top (15 °C) and bottom (1612 °C) boundary.

2. U-Pb age distribution of detrital zircons in the European Alps and the Pyrenees

Figure A2 and figure A3 illustrate the U-Pb age distribution of compiled detrital zircons in the European Alps and Pyrenees.

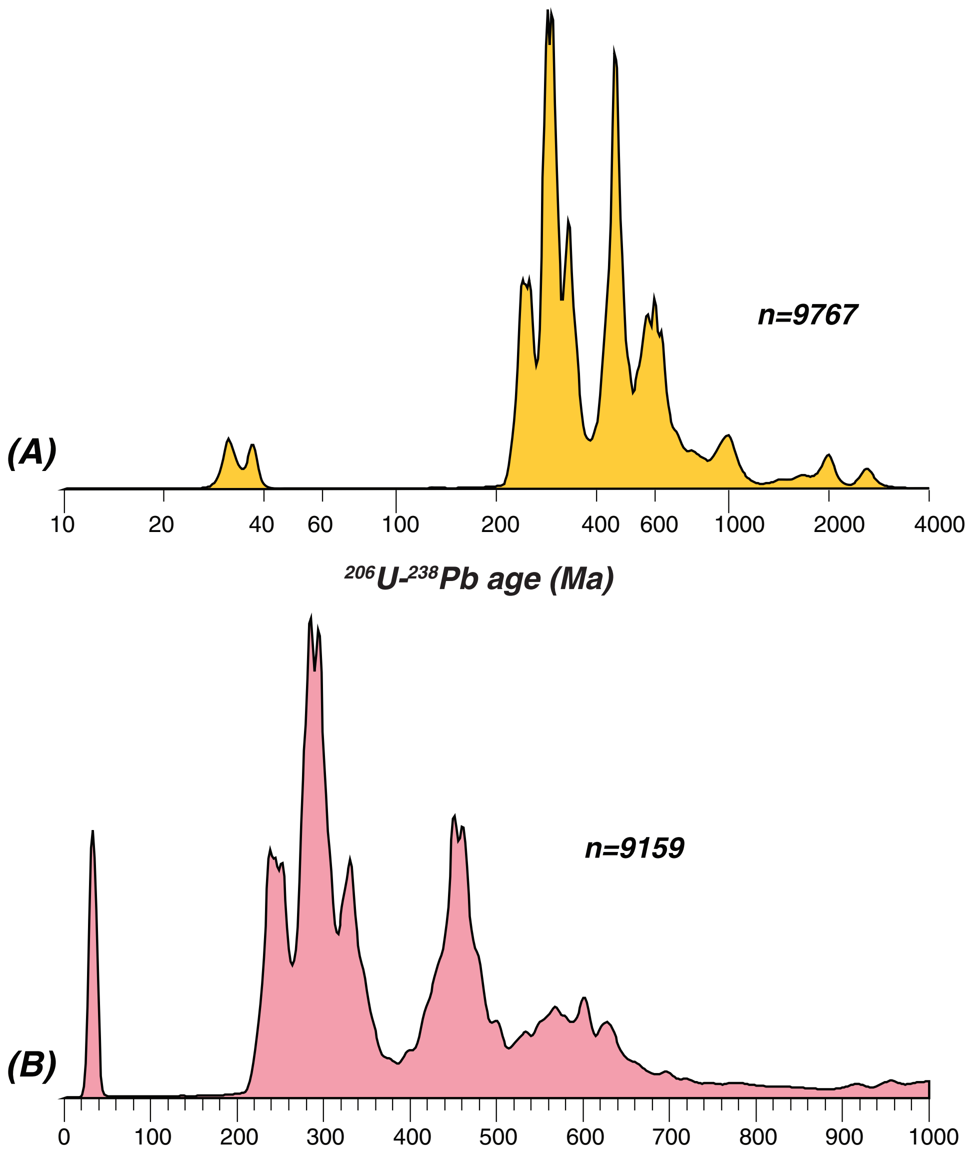

Fig. A2. U-Pb age distribution of detritial zircons in the European Alps. Data compilation from McCarthy and others (2018).

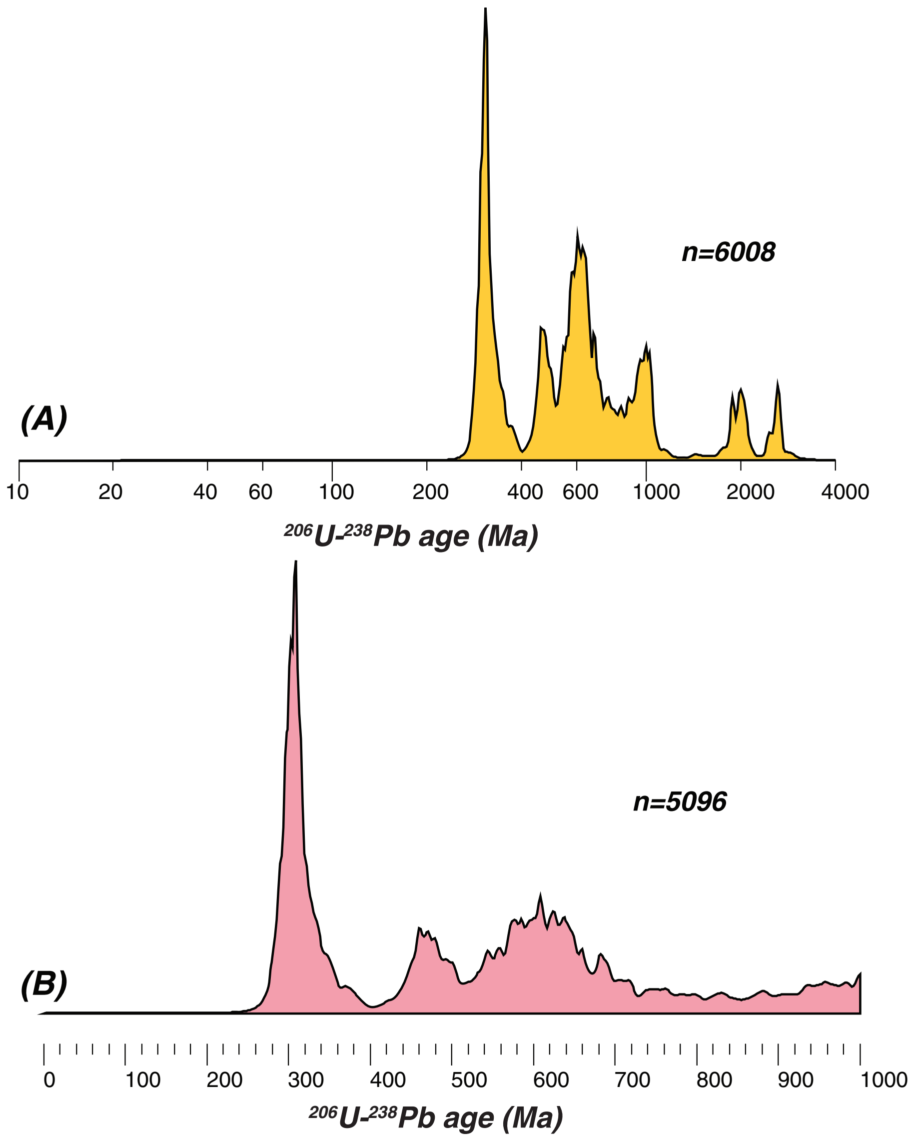

Fig. A3. U-Pb age distribution of detritial zircons in the Pyrenees. See main text for references.

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