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Review Article

Kinematic constraints on buckling a lithospheric-scale orocline along the northern margin of Gondwana: A geologic synthesis

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ABSTRACT

The Paleozoic Variscan orogeny was a large-scale collisional event involving amalgamation of multiple continents and micro-continents. Existing data, suggests oroclinal buckling of an originally near-linear convergent margin during the last stages of Variscan deformation in the late Paleozoic. Closure of the Rheic Ocean resulted in E–W shortening (present-day coordinates) in the Carboniferous, producing a near linear N–S trending, east-verging belt. Subsequent N–S shortening near the Carb-Permian boundary resulted in oroclinal buckling. This late-stage orogenic event remains an enigmatic part of final Pangea amalgamation.

The present-day arc curvature of the Variscan has inspired many tectonic models, with little agreement between them. While there is general consensus that two separate phases of deformation occurred, various models consider that curvature was caused by: dextral transpression around a Gondwana indentor; strike-slip wrench tectonics; or a change in tectonic transport direction due to changing stress fields. More recent models explain the curvature as an orocline, with potentially two opposite-facing bends, caused by secondary rotations. Deciphering the kinematic history of curved orogens is difficult, and requires establishment of two deformation phases: an initial compressive phase that forms a relatively linear belt, and a second phase that causes vertical-axis rotation of the orogenic limbs. Historically the most robust technique to accurately quantify vertical axis-rotation in curved orogens is paleomagnetic analysis, but recently other types of data, including fracture, geochemical, petrologic, paleo-current and calcite twin data, have been used to corroborate secondary buckling. A review of existing and new Variscan data from Iberia is presented that argues for secondary buckling of an originally linear orogenic system.

Together, these data constrain oroclinal buckling of the Cantabrian Orocline to have occurred in about 10 Ma during the latest Carboniferous, which agrees well with recent geodynamical models and structural data that relate oroclinal buckling with lithospheric delamination in the Variscan.

Contents

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1. Introduction

Folded and distorted strata have long fascinated artists and observers of the natural world. Since at least as far back as the time of Leonardo da Vinci in the late 1400s, natural scientists have questioned the meaning of warped and slanted rock layers (Jones, 1962; Rosenberg, 2001). In da Vinci's sketch of 'crumbled strata' his inference of dynamic processes that stressed horizontally layered rock into a distorted shape is clear, and is argued as a foundation to his postulations on geologic facies and relative time (Berger, 2005).

In the 1600s Nicholas Steno returned to the significance of folded strata for understanding the Earth when he used field observations of distorted strata to define his law of horizontality published in *Dissertationis Prodromus* (Brookfield, 2004; Rosenberg, 2006; Steno, 1669). Later in the 1700s, observations by fellow renaissance geologist Sir James Hall established that folded strata are a direct consequence of deformation in the Earth's crust (Carey, 1955). Finally, in the late 1800s and early 1900s it was geologists like Eduard Suess (1909), Émile Argand (1924), and William Hobbs (1914) who acknowledged folded and distorted layers in plan/map-view at the scale of entire orogenic systems, not just within, or between individual outcrops (Hobbs, 1914). This expansion of observations ultimately led Carey (1955, 1958), the true father of curved orogenic systems, to the realization that folds happen across at least 12 orders of magnitude — from the micro- to the crustal-scale. It is these crustal-scale folds that remain one of the largest and least understood structures on Earth, with one of the most fundamental remaining questions being — if the crust folds, how deep does deformation penetrate?

Unquestionably, one of the most spectacular crustal-scale folds on Earth today is found in the Western Europe Variscan Belt (Fig. 1). The Western Europe Variscan Belt is a complex continental-scale orogen (1000 km wide and 8000 km long) that formed through a series of protracted tectonic episodes from initial convergence at about 420 Ma to final collision at about 310 Ma (e.g., Franke et al., 2005; Martínez Catalán et al., 2007). Broadly speaking, the Variscan orogen represents the closing of at least two – and possibly four – oceans between Laurentia, Baltica, and Gondwana, and intervening microcontinents during the Paleozoic amalgamation of the Pangea supercontinent (e.g., Hatcher, 1989, 2002; Martínez Catalán et al., 1997, 2007; Matte, 2001; van Staal et al., 1998; Winchester et al., 2002). Late, to post-orogenic modification of the Western Europe Variscan Belt produced its characteristic sinuous shape that today traces at least one, and possibly four (Martínez-Catalán, 2011, 2012; Shaw et al., 2012a) complete arcs from Poland to Brittany, and then across the Bay of Biscay (Cantabrian Sea) into Iberia, where they are truncated by the Betic Alpine front in southern Spain (Fig. 1).

This review paper describes the collection of observations that led to the current orogenic model for the most thoroughly established orocline in the Western Europe Variscan Belt — the Cantabrian Orocline (Fig. 2). Herein, the Cantabrian Orocline is roughly equivalent to the classically defined Ibero-Armorican Arc (e.g., Brun and Burg, 1982; Lefort, 1979; Perroud and Bonhommet, 1981), and is characterized by a curved structural trend that traces an arc from Brittany across the Bay of Biscay into western Iberia. Recently published data and paleogeographic models from Iberia indicate that within the classic Ibero-Armorican Arc, whose original boundary was truncated in southern Iberia by the Alpine Front (Fig. 1), exists a coupled orocline system with two linked bends — the Cantabrian Orocline to the north and the Central Iberian Orocline to the south (Martínez-Catalán, 2011, 2012; Shaw et al., 2012a). This review focuses on the Cantabrian Orocline, which is better studied and exposed than the Central Iberian Orocline. Development of the Cantabrian Orocline requires the formation of a roughly linear orogenic belt during early Variscan closure of the Rheic Ocean, which was subsequently bent in plan-view into an orocline during late stages of Pangea amalgamation. Importantly, this model predicts that to accommodate oroclinal buckling at this scale, involvement of the entire lithosphere is required (Gutiérrez-Alonso et al., 2004, 2012). The resulting tectonic story is based on kinematic and geologic observations grounded in structural, strain, sedimentologic, geochronologic, geochemical, and paleomagnetic data from across Iberia.

2. Opening of the Rheic

The tectonic history of the Western Europe Variscan Belt begins with the development of the Late Cambrian to Early Ordovician northern Gondwanan margin (Fig. 3), which preserves passive margin sedimentary, intrusive and volcanic sequences that record the

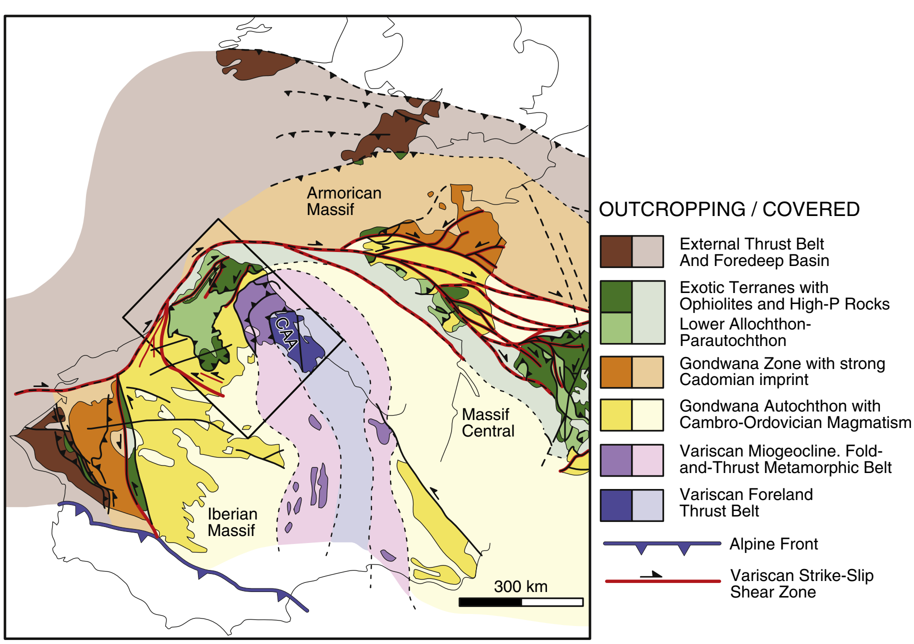


Fig. 1. Simplified tectonostratigraphic zonation of the Variscan orogen in southwestern Europe (modified from Franke, 1989 and Martínez Catalán et al., 2007) highlighting the location of the central Cantabrian–Asturian Arc (labeled as CAA). Inset box represents areal coverage of Fig. 2.

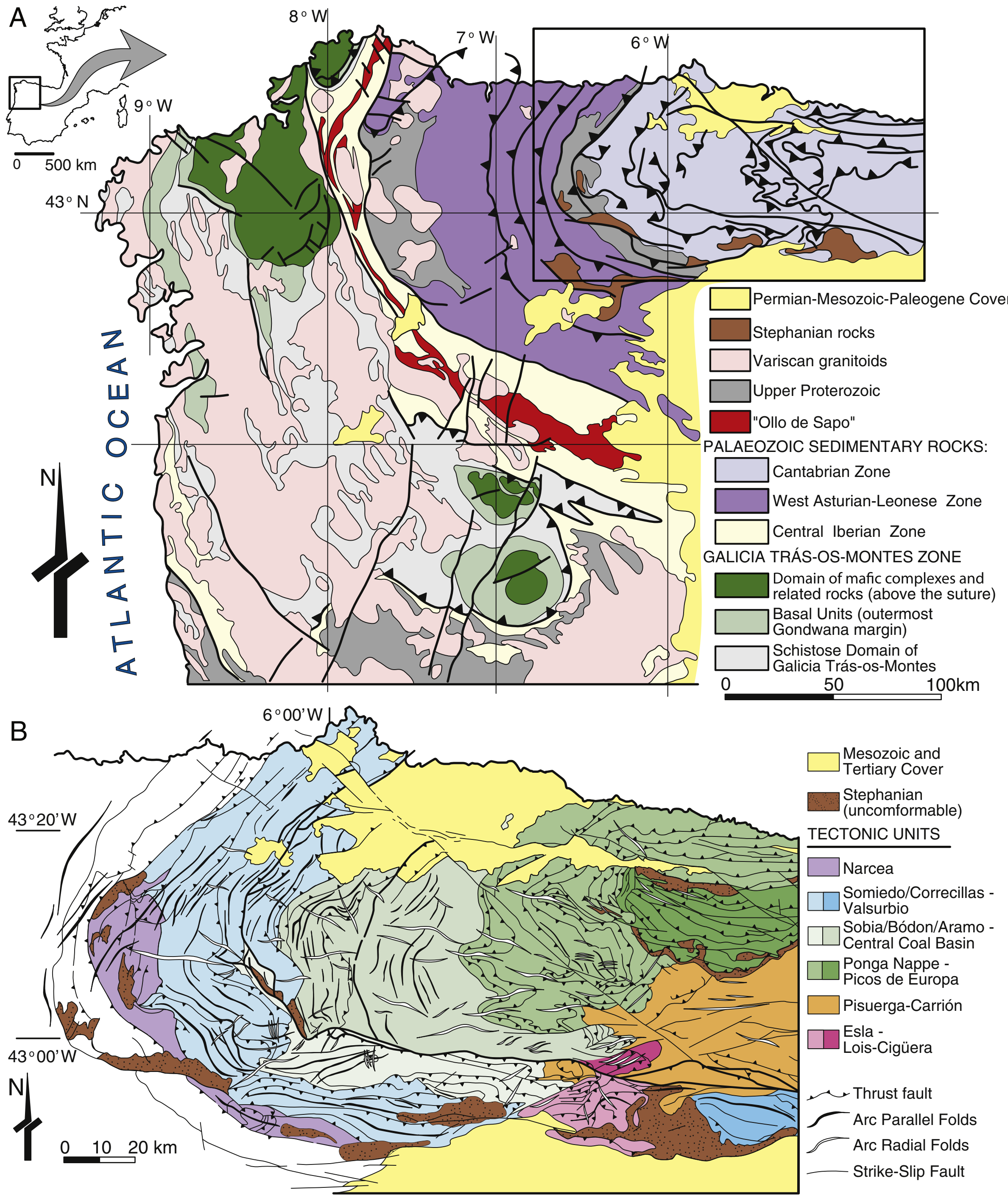


Fig. 2. (A) Simplified map of the major subdivisions of the Variscan Belt in NW Iberia that make up the Cantabrian Orocline. (B) Simplified structure/tectonic map of the Cantabrian–Asturian Arc, highlighting the geometry of major thrusts and the orientation of arc-parallel and arc-perpendicular folds.
Panel A is modified from Julivert et al. (1972) and Farias et al. (1987). Tectonic unit divisions in panel B are modified from Alonso et al. (2009).

opening of the Rheic Ocean (e.g. Cocks and Torsvik, 2002, 2005; McKerrow and Scotese, 1990). There is general agreement that the Rheic Ocean was the result of rifting and subsequent northward drift of a series of peri-Gondwanan terranes (e.g., Avalonia, Carolinia and Ganderia) from the northern margin of a southern hemisphere Gondwana (Fig. 3) (Cocks and Fortey, 1990; Cocks and Torsvik, 2002, 2005; Linnemann et al., 2012; McKerrow and Scotese, 1990; Stampfli and Borel, 2002; von Raumer et al., 2003); however, there remains uncertainty as to what initiated rifting (e.g., Crowley et al., 2000; Díez Fernández et al., 2012; Fuenlabrada et al., in press; Matte, 2001; Murphy et al., 2006; Stampfli and Borel, 2002; van Staal et al., 1998; von Raumer et al., 2002). In NW Iberia the sedimentary and igneous rock succession of the Late Cambrian to Early Ordovician is well preserved and provides one of the most complete records of the Rheic Ocean rift-to-drift history, as well as its closure during the peak of Variscan orogeny in the Late Devonian and Early Carboniferous (e.g., Aramburu et al., 2002; Gibbons and Moreno, 2002; Keller et al., 2007, 2008; Vera, 2004).

Prior to the opening of the Rheic Ocean, the peri-Gondwana terranes (e.g., Avalonia, Carolinia, Armorica and Iberia) record a Neoproterozoic history of protracted (ca. 750–550 Ma) low-grade tectonothermal evolution caused by the northern Gondwanan

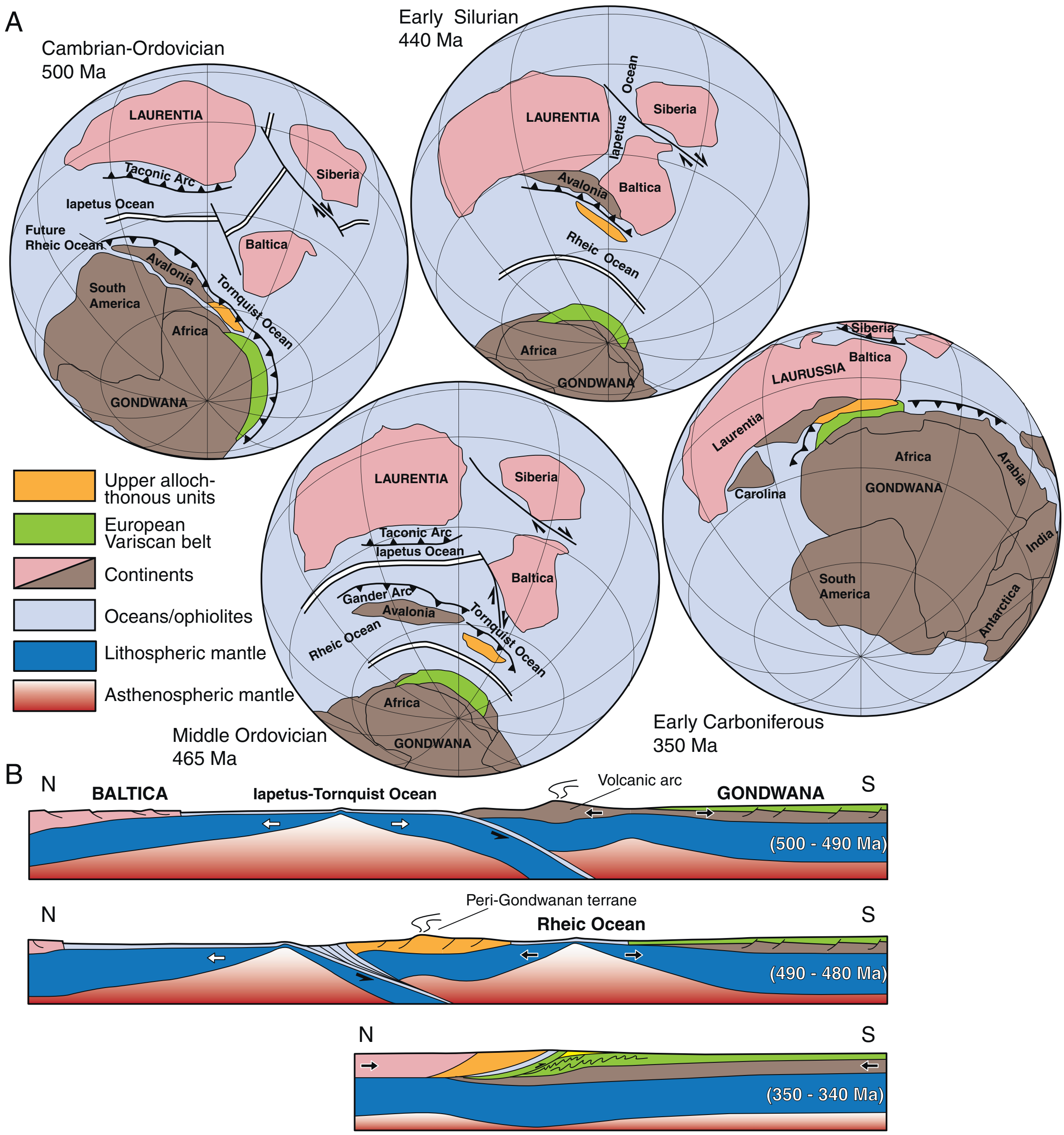


Fig. 3. (A) Paleogeographic reconstruction depicting opening and closing of the Rheic Ocean. Major tectonic domains relevant to the Western Europe Variscan Belt are color coded as in the cross-sections of (B). (B) Schematic tectonic cross-sections for the time slices depicted in the paleogeographic maps of (A).
Panel A reconstructions modified from Gómez Barreiro et al. (2007), Winchester et al. (2002) and Martínez Catalán et al. (2007). Panel B is modified from Martínez Catalán et al. (2007).

Cadomian subduction-related orogeny (Cuesta et al., 2004; Díaz García, 2006; Fernández-Suárez et al., 1998; Gutiérrez-Alonso, 1996; Gutiérrez-Alonso and Fernández-Suárez, 1996; Gutiérrez-Alonso, et al., 2004). The Cadomian orogeny in northern Gondwana is thought to have developed along a Cordilleran-style convergent plate boundary, characterized by a landward-dipping subduction zone and a continental magmatic arc (e.g., Cogné, 1990; Linnemann et al., 2007, 2012). The Cadomian orogen in north Gondwana was followed by lithospheric thinning during the Late Cambrian to Early Ordovician (Díaz García, 2002; Díez Fernández et al., 2011, 2012; Díez Montes, 2006; Martínez Catalán et al., 1992; Pérez-Estaún et al., 1991a; Valverde Vaquero et al., 2005), and ultimately to the formation of the Rheic Ocean.

Existing lithostratigraphic and paleontological data indicate that much of the Western Europe Variscan Belt was located adjacent to North Africa during the early Paleozoic evolution of the Rheic (e.g. Martínez-Catalán et al., 2004; Robardet, 2002). The age and origin of the basement to these sequences is poorly understood, and thus a definitive paleogeography has been elusive (Robardet, 2003). Early studies suggested that the basement of the Western Europe Variscan Belt was part of the Paleoproterozoic ca. 2.0 Ga West African craton (Guerrot et al., 1989; Samson and D'Lemos, 1998). However, more recent detrital zircon studies have pointed to other potential basement sources in NW Iberia based on age populations of Mesoproterozoic (ca. 1.1–1.4 Ga), and Archean age (e.g., Díez Fernández et al., 2010; Fernández-Suárez et al., 2000, 2002a,b;

Gutiérrez-Alonso et al., 2003; Pereira et al., 2012). Detrital zircon data from Neoproterozoic metasedimentary rocks of the Ossa–Morena zone contain zircon populations that are typical of the West African craton (Fernández-Suárez et al., 2002a; Gutiérrez-Alonso et al., 2003; Pereira et al., 2011) and directly correlate with those found in the North Armorican Massif (Samson et al., 2005). Conversely, the presence of ca. 1.0 Ga detrital zircons in Neoproterozoic clastic rocks suggested an Amazonian source for at least some of the detritus (Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2003; Pereira et al., 2012), although a recently discovered potential source of ca. 1.0 Ga zircons exists in the Central Sahara (de Wit et al., 2005; Meinhold et al., 2011, 2012). Additionally, Sm–Nd isotope data for Neoproterozoic sedimentary rocks (Fernández-Suárez et al., 1998; Ugidos et al., 2003) yield ε_{Nd} values between −2.2 and −0.4, which are consistent with a source outside the typical range of time-equivalent sedimentary rocks from the West African craton (e.g. Abati et al., 2010; Morag et al., 2011; Potrel et al., 1998). Recent findings of ca 1.0 Ga zircons and source rocks in the southern Sahara (Avigad et al., 2003, 2012; Meinhold et al., 2011) provide new possibilities for the reconstruction of the paleoposition of NW Iberia both in Neoproterozoic and early Paleozoic times, though this controversy is not yet resolved.

Moving into the early Paleozoic, the Cambrian to Devonian detrital zircons and stratigraphy in NW Iberia suggests a North African connection (Díez Fernández et al., 2010; Pastor-Galán et al., 2012c; Shaw et al., 2012b). These associations include the Armorican Quartzite, Late Ordovician glaciomarine diamictites, and the presence of shallow-marine Tethyan (African proximity) fossils (Robardet and Gutiérrez-Marco, 1990). Sm–Nd data for the Cambrian sedimentary rocks (Ugidos et al., 2003) yield more negative ε_{Nd} values, interpreted to reflect a peri-West African source; whereas detrital zircons in the Ordovician, Silurian and Devonian sedimentary rocks yield mixed populations (Fernández-Suárez et al., 2002b; Martínez-Catalán et al., 2004; Pastor-Galán et al., 2012a; Shaw et al., 2012b) with both West African and an increasing Mesoproterozoic signatures of ambiguous origin. These various data suggest that, in contrast to the Neoproterozoic sources, Palaeozoic detritus within NW Iberia have mixed Mesoproterozoic and West African components. This possible switch in source region has been linked to portions of the Western Europe Variscan Belt being transferred along the Gondwanan margin by the Precambrian–Cambrian boundary (Fernández-Suárez et al., 2002b; Gutiérrez-Alonso et al., 2003).

The protracted and complex tectonic evolution of the peri-Gondwana terranes likely controlled the site of initial rifting and subsequent development of the Rheic Ocean (Murphy et al., 2006). Rift initiation began with the ca. 500–490 Ma separation of Avalonia and Carolinia from Gondwana, which left Cadomian-type terranes (e.g., Iberia) along the northern Gondwanan margin (Fig. 3) (e.g., Cocks and Torsvik, 2002, 2005, 2011). By ca. 460 Ma Avalonia and Carolinia had drifted about 2000 km north of this margin (Fig. 3a) (Gómez Barreiro et al., 2010; Winchester et al., 2002). The timing of the Rheic rift-to-drift transition is constrained by Lower Ordovician (Tremadoc–Arenigian) volcanism present throughout NW Iberia (Castro et al., 1999, 2003; Díez Montes, 2006; Díez Montes et al., 2010; Gutiérrez-Alonso et al., 2007; Montero et al., 2007, 2009; Talavera et al., 2012; Valverde Vaquero et al., 2005; Valverde-Vaquero and Dunning, 2000). The northern portion of the Central Iberian Zone preserves the most volumetrically significant Early Ordovician volcanic sequence in NW Iberia, including the "Ollo de Sapo" belt where voluminous felsic volcanics and related intrusions outcrop along a continuous NW- to N-trending belt (Castro et al., 1999, 2003; Díez Montes, 2006; Valverde Vaquero et al., 2005; Valverde-Vaquero and Dunning, 2000). Coeval with Ollo de Sapo magmatism, ca. 4500 m of strata accumulated in sub-basins and troughs parallel to the northern Gondwanan margin, marking an increase in subsidence related to tectonic extension and the rift-drift transition stage of Rheic Ocean development (Aramburu et al., 1992; Martínez Catalán et al., 1992; Pérez-Estaún et al., 1990). This extensional event is coeval with genesis of Lower Ordovician granitoid and volcanic rocks interpreted as intra-crustal melts generated in response to steep geothermal gradients associated with rifting (Díez Montes, 2006; Gallastegui et al., 1987; Pin et al., 1992; Ribeiro and Floor, 1987; Rubio-Ordoñez et al., 2012; Valverde Vaquero et al., 2005). Although volcanic activity continued into the Upper Ordovician, the resulting volcanic rocks are scarce and only locally represented (Corretgé and Suárez, 1990; Gallastegui et al., 1992; Heinz et al., 1985).

Some models for the rift-to-drift transition (e.g., Díez Fernández et al., 2012; van Staal et al., 1998) imply that the Rheic Ocean initiated as a backarc basin, but evidence for arc-related rocks coeval with rifting along the northern Gondwanan margin is equivocal. Alternatively, since the opening of the Rheic Ocean is coeval with a polarity flip along the northern Iapetus margin and the onset of northwesterly directed subduction and ridge–trench collision (e.g., Stampfli and Borel, 2002; van Staal et al., 1998), the portion of the Avalonian–Carolinian microplate captured from Gondwana during the Early Cambrian may have been pulled from Gondwana by slab pull forces (Murphy et al., 2006) in a manner analogous to the opening of the Neotethys in the Mesozoic (Stampfli and Borel, 2002). The slab pull model requires the absence of a spreading ridge between Avalonia–Carolinia and the Laurentian margin and, given the moderately rapid northerly component of motion of Avalonia between 480 and 460 Ma (8 cm/yr; Hamilton and Murphy, 2004), the presence of an east–west striking spreading ridge in the Rheic Ocean (Fig. 3a).

The Schistose Galicia–Trás-os-Montes Domain (Fig. 2a) (Farias et al., 1987; Marcos et al., 2002; Martínez-Catalán et al., 1997) rests tectonically above the Central Iberian Zone and consists of a thick siliciclastic sequence with interbedded volcanic rocks. Some of the volcanic rocks yield Lower Ordovician (475 ± 2, Valverde Vaquero et al., 2005) ages that are interpreted as the most outboard parts of the passive margin sedimentary wedge of Gondwana. Structurally above the sedimentary sequence, the Schistose Galicia–Trás-os-Montes Domain contains an allochthonous complex, which has a lowermost unit of continental, Gondwana basement affinity (the Basal Units) (Martínez-Catalán et al., 1997). This unit is structurally overlain by two ophiolitic units of Early Ordovician and Devonian age that are interpreted to be the remnants of the Rheic Ocean or subsidiary oceans closed during the Variscan orogeny (Martínez-Catalán et al., 1997; Arenas et al., 2007a and b; Sánchez Martínez et al., 2007, 2011). These units were affected by high-pressure and low- to intermediate temperature metamorphism (Arenas et al., 1995; López Carmona et al., 2009; Rodríguez et al., 2003), and yield Lower Ordovician igneous protolith ages (ca. 480 Ma; Santos Zalduegui et al., 1995). The uppermost units, resting on top of the ophiolites, are interpreted to be rocks of the northern (Laurussian) margin of the Rheic Ocean (Fig. 3b) (Arenas et al., 2007a,b; Martínez-Catalán et al., 1997).

The Paleozoic passive margin rocks of Iberia are divided into several zones based on differences in their Lower Paleozoic sedimentary successions, which likely reflect variations in relative proximity to the Gondwana margin (Fig. 2). The Cantabrian Zone preserves a coastal environment, whereas the West Asturian–Leonese, Central Iberian, Galicia–Trás-os-Montes (Lower Schistose Domain) and/or Ossa-Morena zones preserve a more outboard tectonostratigraphic succession (Fig. 2a) (Aramburu et al., 2002; Gutiérrez-Marco et al., 1999; Julivert et al., 1972; Marcos and Farias, 1999; Martínez-Catalán et al., 1997, 1999; Pérez-Estaún et al., 1990; Quesada, 1990; Quesada et al., 1991; Ribeiro et al., 1990; Robardet, 2002, 2003). Today the boundaries that separate the major tectonostratigraphic zones in Iberia are defined by major Variscan fault systems that in some cases were reactivated as extensional and/or strike-slip structures in the aftermath of the Variscan orogeny (e.g., Martínez Catalán et al., 1992, 2003).

3. Closing of the Rheic and the Variscan orogeny

The Variscan orogeny, named for a fabled Germanic tribe in NE Bavaria (Suess, 1888), represents a protracted deformation phase that amalgamated major and minor continental blocks and island arcs into what is now Western Europe (e.g. Matte, 2001) (Figs. 1 and 3). Rocks deformed during Variscan orogeny can be found today from Portugal east to Poland, north to the British Isles and Germany to south along the northern margin of the present-day Mediterranean. The Variscan is interpreted to result from the closure of the Rheic Ocean, which was subducted northward beneath the southern margin of the drifted peri-Gondwanan terranes that closed the Iapetus Ocean (Fig. 3) (e.g. Díez Fernández et al., 2012). Rheic subduction included consumption of the Rheic mid-ocean ridge by 395 Ma (Gutiérrez-Alonso et al., 2008; Woodcock et al., 2007), which resulted in an increase in convergence rate between opposing margins. The continuing debate about the number of basins that existed during the Variscan has caused confusion as to the geodynamic evolution of Variscan collisions and the paleogeography of the various crustal blocks.

Most models of Variscan dynamics suggest that the putative major and minor basins that existed between the peri-Gondwana terranes and Gondwana were closed in the Late Devonian (e.g., Azor et al., 2008; Dallmeyer et al., 1997; Franke et al., 2005; Martínez Catalán et al., 2007; Pin and Paquette, 1997; Rodríguez et al., 2003). Structural relationships and their timing indicate that deformation, metamorphism and exhumation of oceanic and continental rocks found today within interpreted sutures, represent collision and the progressive transport of a deformation front towards the foreland (Dallmeyer et al., 1997; Pérez-Estaún et al., 1991b). The foreland is mainly represented by Paleozoic passive margin sedimentary rocks that today are mostly preserved in NW Iberia (Fig. 3b). Collision initially produced recumbent folds that verged and migrated from the suture towards the foreland. Continued shortening then led to the extensional collapse of the thickened orogenic hinterland (Arenas and Martínez-Catalán, 2003; Escuder Viruete et al., 1994) at ca. 320 Ma (Martínez-Catalán et al., 2009), an event that was coeval with the development of a non-metamorphic foreland fold-thrust belt within the Cantabrian Zone (Pérez-Estaún et al., 1994). Final deformation associated with the closure of the Rheic Ocean produced large-wavelength upright folds and strike-slip ductile shear zones (Martínez-Catalán et al., 2009), along with regional metamorphism that diminishes towards the internal zones to the foreland. Metamorphic grade ranges from high-middle grade rocks in the hinterland (Arenas and Martínez-Catalán, 2003), to low and very low grade and even diagenetic in the foreland (Abad et al., 2003; Gutiérrez-Alonso and Nieto, 1996).

Today the structural trace of the Western Europe Variscan Belt in northern Iberia (Bard et al., 1971; Ribeiro et al., 1995) outlines one of the most dramatic curved orogenic systems on Earth (the Cantabrian Orocline), with an arc-trace that spans 180° of curvature (Fig. 1). Ries and Shackleton (1976) divide the Cantabrian Orocline into three structural zones based on observed longitudinal tangential strain: the outer arc (tangential extension), the inner arc (tangential compression), and a small (ca. 10 km wide) dividing neutral zone (no arc-parallel strain). In their model, arc-parallel stretching in the outer arc increases away from the core (Ries and Shackleton, 1976), and shortening in the inner arc increases towards the core (Julivert and Marcos, 1973). The outer arc extension was primarily accommodated by dextral strike-slip faulting in the upper crust, and ductile elongation in the lower crust (Gutiérrez-Alonso et al., 2004). At the core of the Cantabrian Orocline is the Cantabrian–Asturian Arc (Fig. 2b), which is considered by many to consist of strata deposited along the south margin of the Rheic Ocean (Martínez Catalán et al., 1992; Murphy et al., 2006; Robardet, 2002, 2003) (Fig. 2). Curvature of the Cantabrian Orocline is most extreme within the Cantabrian–Asturian Arc (Fig. 2).

The Cantabrian Orocline has been the object of numerous paleomagnetic and structural studies ever since early studies by Schultz (1858), Barrois (1882), Suess (1909) and Staub (1926) first documented the dramatic curvature; likewise, many varying explanations have been given for its origin and development. Following Carey's (1955) original orocline hypothesis for the Cantabrian Orocline, a number of models have been suggested for its formation (Fig. 4). Matte and Ribeiro (1975) and Matte (1986) suggested that the curvature is due to displacement of a 'Cantabria microplate' westward, resulting in a primary non-rotational arc (Fig. 4a); whereas Ries and Shackleton (1976) proposed that the curvature formed by late Variscan north–south compression that produced a counterclockwise rotation of central and southern Iberia relative to Brittany (Fig. 4b). Lorenz (1976), Lorenz and Nicholls (1984), Lefort (1979) and Dias and Ribeiro (1995) suggested that irregular coastlines of opposing margins, or promontory–salient pairs, caused the curvature (Fig. 4c). Later, Ries et al. (1980) used strain and limited paleomagnetic data to argue that at least a portion of the curvature in the Cantabrian Orocline is secondary, or post-orogenic (Fig. 4d). Brun and Burg (1982) offered what they considered to be a progressive model, with the observed structural curvature being caused by sinistral wrenching during a single collision event, which produced at least some of the observed curvature seen today (Fig. 4e). Pérez-Estaún et al. (1988) proposed a thin-skinned model, explaining the arcuate shape of the Cantabrian–Asturian Arc as the result of progressive clockwise rotational emplacement of a series of nappes (the so called photographic iris model), with a simultaneous initiation of radial and longitudinal structures during thrust emplacement (Fig. 4f). In a review of the geodynamics of the SW Europe Variscides, Ribeiro et al. (2007) (Fig. 4g) argued for a "soft plate tectonics" model in which some of the present-day curvature in the Cantabrian Orocline is primary, with secondary tightening due to the impingement of the 'Cantabrian indentor' in the Devonian. Most recently, Martínez-Catalán (2011) and Martínez-García (2012) argue that the oroclines of the Variscan are related to late strike-slip tectonics along a broad dextral intracontinental shear zone (Fig. 5h).

Since Carey's seminal work on oroclines, reconstructing the kinematic evolution and mechanics of curved mountain systems has been a fundamental component of understanding the paleogeographic and tectonic evolution of continents (e.g. Marshak, 2004; Sussman and Weil, 2004; Van der Voo, 2004). The keys to testing various models for curvature formation in the Cantabrian Orocline, is establishing a robust kinematic model that constrains the timing of rotations relative to the larger tectonic events that occurred during and subsequent to the main phase of Variscan convergence in this region. Deciphering this kinematic history is difficult due to the significant vertical-axis rotations involved. For oroclinal buckling to be a viable model, two deformation phases need to be differentiated: an initial compressive phase that forms a relatively linear belt with little to no rotation of developing structures, and a second phase that causes vertical-axis rotation of existing orogenic limbs. Below is a review of data collected over the years that support an orocline model of secondary vertical-axis rotation – as first envisioned by Carey (1955) – for formation of the Cantabrian Orocline.

4. Observations used for establishing secondary rotation of the Cantabrian Orocline

4.1. Paleomagnetic data

The most robust method for quantifying vertical axis-rotation in curved orogens is paleomagnetic analysis (e.g., Irving and Opdyke, 1965; Lowrie and Hirt, 1986; Muttoni et al., 1998; Van der Voo and Channell, 1980; Weil and Sussman, 2004), and even then, such analysis needs to be accompanied by geologic and structural observations that help constrain the rotation, strain and translation components of the three-dimensional deformation field (e.g., Hindle and Burkhard,

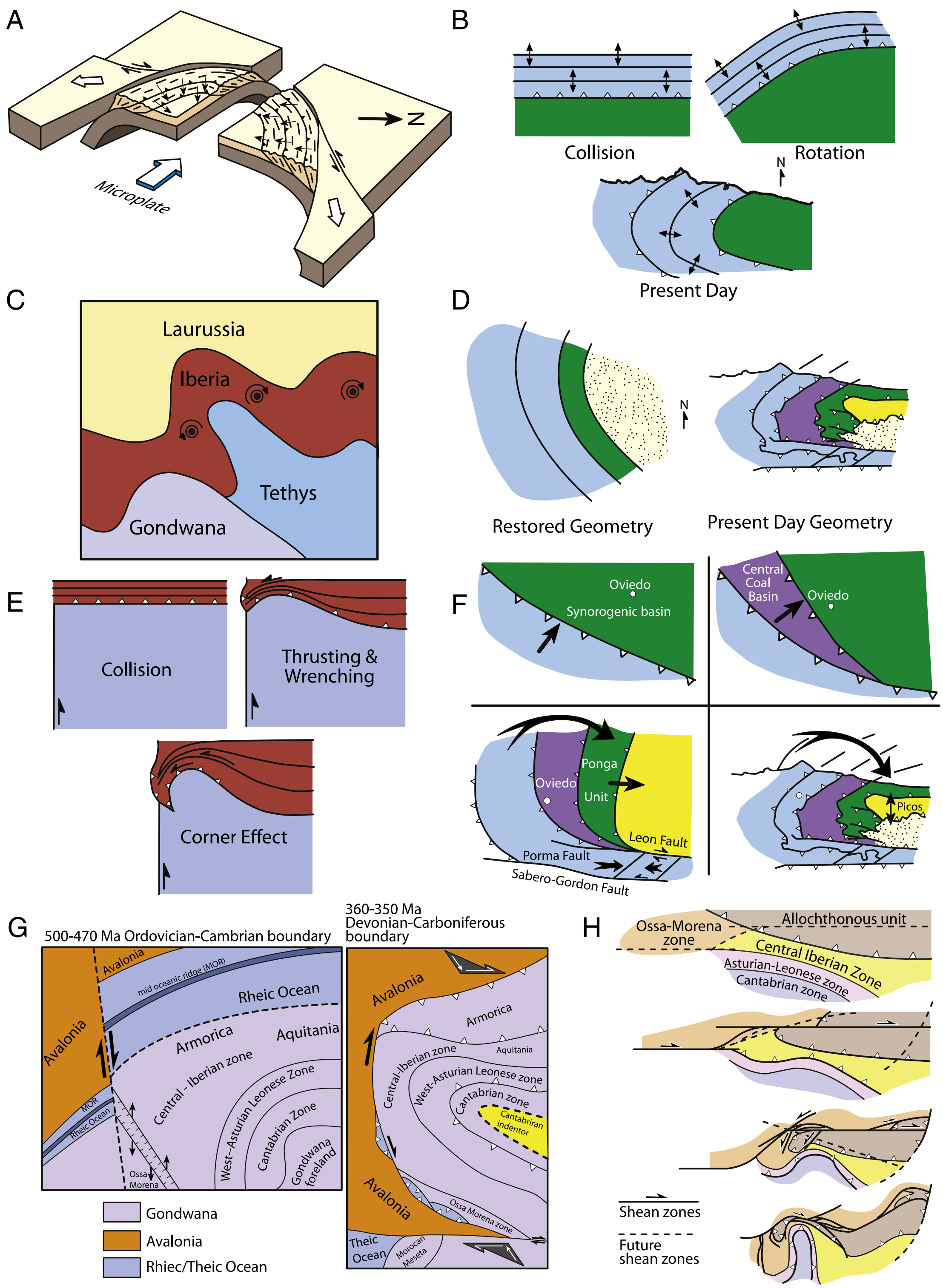


Fig. 4. Schematic sketches of various models proposed for development of the Cantabrian Orocline. (A) Displacement of a 'Cantabria microplate' westward, resulting in a primary non-rotational arc (after Matte and Ribeiro, 1975, and Matte, 1986). (B) Curvature formed by late Variscan north–south compression that produced counterclockwise rotation of central and southern Iberia relative to Brittany (after Ries and Shackleton, 1976). (C) Curvature formed by impinging irregular coastlines during tectonic convergence (after Lorenz, 1976, and Lefort, 1979). (D) Progressive arc model based on early strain and paleomagnetic data (after Ries et al., 1980). (E) A progressive arc model based on sinistral wrenching (after Brun and Burg, 1982). (F) A thin-skinned model that explains the Cantabrian–Asturian Arc's arcuate shape by progressive clockwise emplacement of nappes (after Pérez-Estaún et al., 1988). (G) A "soft plate tectonic" model in which some of the present-day curvature is primary, with secondary tightening due to impingement of the 'Cantabrian indentor' in the Devonian (after Ribeiro et al., 2007). (H) Orocline development by strike-slip tectonics (after Martínez-Catalán, 2011).

1999; Kwon and Mitra, 2004; Pueyo et al., 2003, 2004; Weil and Sussman, 2004; Yonkee and Weil, 2010a). If deformation occurs subsequent to magnetization acquisition, then the magnetization will record all ensuing rotations. Consequently, if magnetizations are found that vary systematically as a function of strike around a curved orogen, then secondary oroclinal buckling, regardless of the mechanism, is likely the most viable kinematic model for curvature formation (Yonkee and Weil, 2010b).

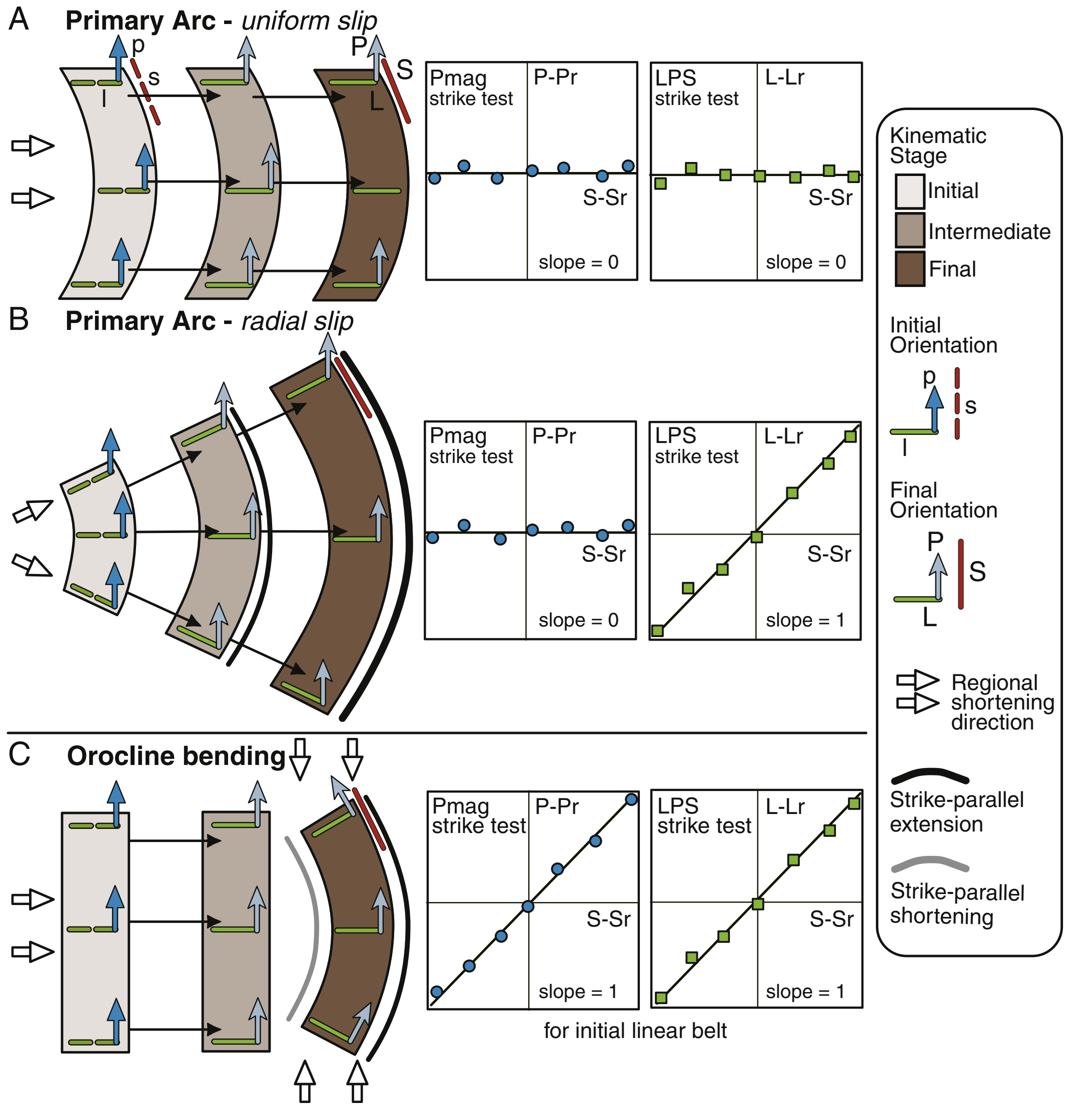


Fig. 5. Schematic sketches of kinematic models proposed to explain orogenic curvature and their predicted strike tests for paleomagnetic and structural data: A) primary arc with uniform slip, B) primary arc with radial slip, and C) secondary orocline with superimposed buckling of an older fold-thrust belt. Each of these models predicts rotation patterns that can be tested with available paleomagnetic and structural data.
Modified from Yonkee and Weil (2010b).

Correlations between changes in regional structural trend (relative to a reference trend), and rotations estimated from paleomagnetic or deformation fabric directions (relative to a reference direction) are evaluated using a strike test (Eldredge et al., 1985; Lowrie and Hirt, 1986; Schwartz and Van der Voo, 1983; Yonkee and Weil, 2010b). In this way, end member kinematic models can be statistically tested by collecting sufficient sites that are distributed evenly around a given curved structural trace (Yonkee and Weil, 2010b). For the purposes of simplicity, only end member models are described to provide perspective for the datasets and figures described in this review (Fig. 5). In the "Primary arc" end-member model structures initiate with curvature (primary arcs) and do not experience additional rotation during subsequent deformation (Fig. 5a and b) (Weil and Sussman, 2004). In this case, paleomagnetic directions remain unchanged, regardless of structural trend, resulting in a strike test with a slope of 0. Within the family of primary arcs there can be cases were thrust transport has a uniform slip direction, or alternatively a radial slip pattern. For a uniform slip model, both paleomagnetic and shortening directions remain unchanged, and both strike tests have a slope of 0 (Fig. 5a); whereas for radial slip, paleomagnetic directions are not rotated and define a slope of 0, while initial radial shortening directions define a slope of 1 (Fig. 5b). The other end member is an orocline (Fig. 5c), a belt with initially linear thrusts and consistent shortening fabrics that undergoes 100% secondary rotation during subsequent deformation. In this case strike tests would yield slopes of 1 for both paleomagnetic and shortening directions. In summary, a strike test slope of 0 for paleomagnetic data indicates a primary arc, and a slope of 1 indicates an orocline. However, a strike test for deformation fabric data, or any geologic fabric, can only be uniquely interpreted if constraints can be placed on initial fabric orientation.

Most geologic, structural and paleomagnetic studies done to test the orocline hypothesis for the Cantabrian Orocline have been done in its central core — the Cantabrian–Asturian Arc (Fig. 3b). Within the Cantabrian–Asturian Arc there are two well-defined fold sets that outline the present-day curvature: an arc-parallel set and an arc-perpendicular radial set (Fig. 3b) (established by Aller and Gallastegui, 1995 Julivert and Marcos, 1973;). In general, major thrusts of the Cantabrian–Asturian Arc are concave towards the foreland and propagate towards the east (the present-day core) with a common decollement surface in the Cambrian Láncara Formation (Julivert, 1971).

Early paleomagnetic studies in the Cantabrian–Asturian Arc were used to test the various proposed rotational models, and demonstrated that at least some of the arc's curvature is of a secondary nature (e.g., Bonhommet et al., 1981; Hirt et al., 1992; Perroud, 1986; Perroud and Bonhommet, 1981; Stewart, 1995). However, the ability to establish robust evidence for oroclinal buckling was limited due to a lack of sufficient data around the entire arc, and the failure to recognize and distinguish between secondary syn-tectonic and post-tectonic remagnetizations. Building on these early studies, more recent paleomagnetic investigations from around the arc have established the Cantabrian–Asturian Arc as an orocline that tracked the entire progression from an early linear orogen to a secondarily folded arc (Fig. 6a and b) (Parés et al., 1994; Van der Voo et al., 1997; Weil et al., 2000, 2001; Weil, 2006; Weil et al., 2010; Weil et al., 2012).

Studying the tectonic units located in the inner arc (Fig. 2b) Weil et al. (2000, 2001) identified three separate magnetization components carried by Paleozoic carbonates within the Cantabrian–Asturian Arc. Two of the magnetizations are Carboniferous in age, and a third is Permian–Triassic. Local and regional fold tests for the two Carboniferous components indicate that these magnetizations were secondary overprints acquired either during (the C component) or after (the B component) early longitudinal folding that occurred during the Bashkirian in the West Asturian–Leonese Zone and during the Moscovian in the Cantabrian Zone, but prior to secondary vertical-axis rotation in the late Moscovian, Kasimovian and Gzhelian (Weil et al., 2000, 2010). When all available B and C component paleomagnetic site means are compared to deviations in structural trend around the outer provinces of the Cantabrian–Asturian Arc, a strike test slope of $0.98 \pm .06$ is established, indicating that the orocline end-member model is the best kinematic model for the Cantabrian Orocline (Fig. 6c). The later Permo-Triassic (P–T) component has seen little to no distortion since the time of magnetization acquisition, and is within error of reference P–T paleomagnetic poles for stable Iberia (Weil et al., 2001). From these data a loose P–T upper age limit was first placed on the final phase of oroclinal buckling in the Cantabrian–Asturian Arc (Weil et al., 2001). More recently, Tohver et al. (2008) used $^{40}Ar/^{39}Ar$ analysis of the smectite–illite transition in Cantabrian–Asturian Arc carbonates to date the thermal-fluid event that is interpreted to have potentially altered these rocks during Variscan deformation. The age data collected from different clay grain size fractions, coupled with quantitative polytype modeling, indicated an authigenic age that is coeval with the established late Paleozoic syn-orocline remagnetization (the B and C components) age of Cantabrian–Asturian Arc carbonates. The fluids caused transformation of Fe-rich smectite to Fe-poor illite and created a population of authigenic magnetite that was then responsible for carbonate remagnetization (Weil and der Voo, 2002). Weil (2006) and Weil et al. (2012) found a similar magnetic history for portions of the innermost tectonic units of the Cantabrian–Asturian Arc. Finally, Weil et al. (2010) presented new paleomagnetic data from Early Permian samples from the northern Cantabrian Orocline, and the southern Central Iberian Orocline. After minor structural correction to the analyzed directions, data from both arcs yielded expected Early Permian paleomagnetic pole positions for stable Iberia, with no indication of vertical-axis rotation since the Early Permian. Consequently, the Early Permian was argued to mark termination of oroclinal buckling. This result placed a well-constrained time window of about 10 Ma for oroclinal buckling (Fig. 7).

4.2. Structural data

Along with bulk rotation, the kinematic evolution of an orogenic belt is recorded in the spatial–temporal development of its three-dimensional displacement field, which also includes bulk translation (slip on major faults) and internal strain (accommodated by mesoscopic to grain-scale structures) (Weil and Sussman, 2004). Understanding the kinematic evolution of orogenic belts is challenging due to complexities in determining all components of the displacement field over various spatial and time scales (e.g., Gray and Stamatakos, 1997; Hindle and Burkhard, 1999; Hindle et al., 2002; Marshak, 1988; Mitra, 1994; Thibert et al., 2005). Many orogenic

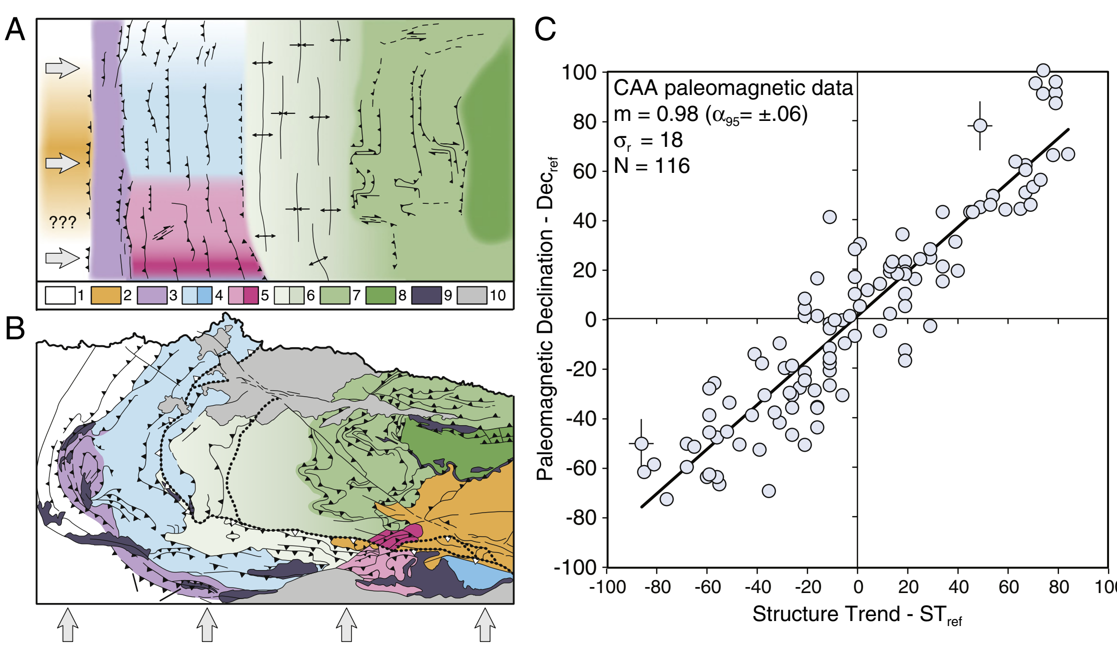


Fig. 6. Cartoon map of palinspastically restored Cantabrian–Asturian Arc showing (A) the region's general geometry after initial east–west (in present-day coordinates) compression (D1 deformation phase). Block arrows represent the schematic orientation of the ancient stress-field during deformation. Fill patterns represent different tectonic units of the Cantabrian–Asturian Arc as depicted in Fig. 3B: (1) West Asturian–Leonese; (2) Pisuerga–Carrión; (3) Narcea Antiform; (4) Somiedo/Correcillas–Valsurbio; (5) Esla–Lois–Cigüera; (6) Sobia/Bodón–Central Coal Basin; (7) Ponga; (8) Picos de Europa; (9) Unconformable Upper Carboniferous rocks; and (10) Mesozoic and Tertiary cover. (B) Present-day configuration of Cantabrian–Asturian Arc after oroclinal buckling, which resulted in counterclockwise rotation of the southern limb Pisuerga–Carrión, Narcea Antiform, and Somiedo/Correcillas–Valsurbio units, clockwise rotation of the northern limb and buckling, superposed and radial folding, and thrust reactivation in the central core and hinge of the arc (mainly the Sobia/Bodón–Central Coal Basin, Ponga and Picos de Europa units) as well as the southeastern Esla–Lois–Cigüera units. (C) Paleomagnetic strike test for data from the Cantabrian–Asturian Arc fold-nappe province (Van der Voo et al., 1997; Weil et al., 2001), which gives a slope of $0.98 \pm .06$. Best-fit slopes, m, number of sites, N, α 95% confidence intervals, $\alpha 95 = 2\sigma CI$, and standard deviation of residuals, σ_r are all given.

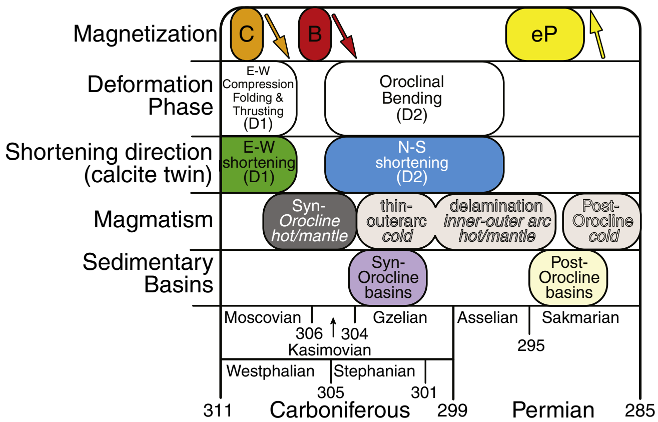


Fig. 7. Proposed timeline for the temporal relationship between the successive magnetizations recorded in the Cantabrian Orocline and their relationship to the two main phases of oroclinal formation, the acquisition of twin strains (and associated stresses) during deformation, the age of various magmatic pulses, and the age of sedimentary basins.
Data from Alonso (1987), Kollmeier et al. (2000),Weil et al. (2001), Colmenero et al. (2008), Weil et al. (2010), and Gutiérrez-Alonso et al. (2011a, 2011b).

belts display widespread layer-parallel-shortening fabrics that formed early in their deformation histories, which provide a record of early stress patterns (e.g. Geiser, 1988; Geiser and Engelder, 1983; Gray and Stamatakos, 1997; Hogan and Dunne, 2001; Mitra, 1994; Ong et al., 2007; Yonkee and Weil, 2010a). However, early layer-parallel-shortening fabrics are commonly modified during subsequent deformation. A commonly observed modification is vertical-axis rotation that reorients fabrics, and complicates the reconstruction of paleo-stress and paleo-strain fields and thus interpretation of the kinematic evolution. By integrating multiple datasets, such as mesoscopic structural patterns, anisotropy of magnetic susceptibility, strain determinations, calcite twin analysis, and paleomagnetism, limitations of individual sets can be minimized and robust kinematic models developed (Fig. 5) (Yonkee and Weil, 2010b).

Calcite twinning analysis in the Cantabrian–Asturian Arc of northern Spain by Kollmeier et al. (2000) provided a complementary dataset that established paleo-stress directions recorded during the two main phases of shortening that produced early longitudinal folding and thrusting, and eventual secondary oroclinal buckling (Fig. 8). Because calcite twinning typically occurs during the earliest phases of shortening due to the low value of critical resolved shear stress required to initiate twinning (about 10 MPa), the results provide a robust proxy for early paleo-stress directions within orogenic systems (e.g., Craddock et al., 1988; Ferrill, 1991, 1993; Ferrill and Groshong, 1993; Harris and van der Pluijm, 1998). This is especially true since calcite twinning is a strain hardening process that restricts later twin overprinting (e.g., Donath and Fruth, 1971; Kilsdonk and Wiltschko, 1988; Teufel, 1980). Calcite twin analysis from the Cantabrian Zone of the Cantabrian–Asturian Arc (Fig. 2) revealed two discrete shortening trends: 1) an early D1 layer-parallel-shortening stress oriented perpendicular to regional structural trends that has an in situ radial pattern around the Cantabrian–Asturian Arc (Fig. 8a), and 2) a secondary D2 overprinted shortening direction that is roughly parallel to the regional structural trend (Fig. 8b) (Kollmeier et al., 2000). Both datasets have a linear (strike test slope of 1.0) relationship between changes in shortening direction trend and structural trend. D1 shortening trends remain perpendicular to structural trend around the arc, while D2 shortening trends remain parallel (Fig. 8). This led Kollmeier et al. (2000) to argue for a two stage tectonic model of D1 E–W shortening (in present-day coordinates) that produced a roughly N–S linear thrust belt, which was followed by a D2 phase the resulted in N–S directed shortening and thrust sheet rotation (Figs. 7 and 8c). Due to the high angle of the two shortening events, the strain hardening effects of calcite twinning did not interfere with the preservation and recording of the two discrete events in the Paleozoic carbonates of NW Iberia.

To further test and constrain the stress field changes established from calcite twin analysis, Pastor-Galán et al. (2011) documented the spatial and temporal distribution of systematic tensile joints from multiple rock units exposed throughout the Cantabrian–Asturian Arc (Fig. 9). Though caution is needed when using joint systems as kinematic markers in polydeformed rock units, systematic joint sets can provide a stress field record of past deformation (e.g., the Ouachita salient (Whitaker and Engelder, 2006), the Appalachian plateau (Engelder and Geiser, 1980), the Idaho–Wyoming salient (Yonkee and Weil, 2010a), the Variscan belt in Wales (Dunne and North, 1990), and the Pyrenees (Turner and Hancock, 1990)) especially when they affect synorogeneic deposits limited by angular unconformities that provide temporal constraints on the development of the different joint sets (Pastor-Galán et al., 2011). There are multiple joint sets present throughout the Cantabrian–Asturian Arc, thus, caution is needed when linking the spatial pattern of joints across a region to a specific tectonic history (e.g., Dunne and North, 1990; Engelder and Geiser, 1980). Fortunately, there are several well-dated angular unconformities within the Paleozoic passive margin and syn-orogenic strata preserved in NW Iberia, which allowed Pastor-Galán et al. (2011) to isolate discrete joint sets to temporally bound rock units. In this way, the regional development of successive joint sets in the Cantabrian–Asturian Arc was unraveled, as the occurrence of angular unconformities constrained the timing of joint formation to pre- and post-unconformity sets. Overall, three groups of sedimentary rocks separated by angular unconformities were studied; each constrained to predate, be coeval with, and to postdate orocline formation (Fig. 9).

All joint sets were interpreted to be products of the local and remote stress fields, and are therefore representative of far-field tectonic stresses (e.g., Eyal et al., 2001; Gross et al., 1995). The youngest joint sets generated in the Cantabrian–Asturian Arc are recorded in Permian outcrops (Fig. 8a). These sets, which are found across the arc and show no regional trend variability (a strike test slope of 0), were interpreted as being caused by bedding flexure during Alpine collision of Iberia with the rest of Europe in Cenozoic times (e.g., Alonso et al., 1996; Álvaro et al., 1979), and by the opening of the Bay of Biscay during Mesozoic times (e.g., Gómez et al., 2002; Gong et al., 2008). These post-orocline joint sets corroborate the study by Weil et al. (2010) who used paleomagnetic data to establish that deposition of Permian strata post-dated formation of the larger Cantabrian Orocline.

The joint sets documented from syn-tectonic Stephanian outcrops recorded two additional joint sets that were not found in the post-orocline Permian outcrops (Fig. 9b) (throughout this paper the absolute and relative timescale of Davydov et al., 2004 is used). A longitudinal set has an arcuate pattern with lower overall curvature than the trends of the underlying structures while an orthogonal set has a radial pattern, sub-perpendicular to the main underlying structural trend (Fig. 9b). These joint sets were interpreted to have formed during N–S (in present-day coordinates) D2 shortening, which resulted in oroclinal buckling. Strike tests of the syn-tectonic Stephanian joint sets indicate that they were formed penecontemporaneous with arc–limb rotation (strike test slopes of between 0.57 and 0.72, indicating 30–40% of oroclinal buckling occurred prior to joint formation) and thus preserve a snapshot view into the oroclinal buckling deformation history.

Observations from pre-Stephanian (Neoproterozoic and Paleozoic) outcrops underlying the Stephanian outcrops record a complex set of joint sets that include all the Stephanian and younger joint sets as well as older sets that are parallel and perpendicular to the main Variscan structural trend (Fig. 9c). These joint sets were interpreted to have formed during E–W (in present-day coordinates) D1 shortening,

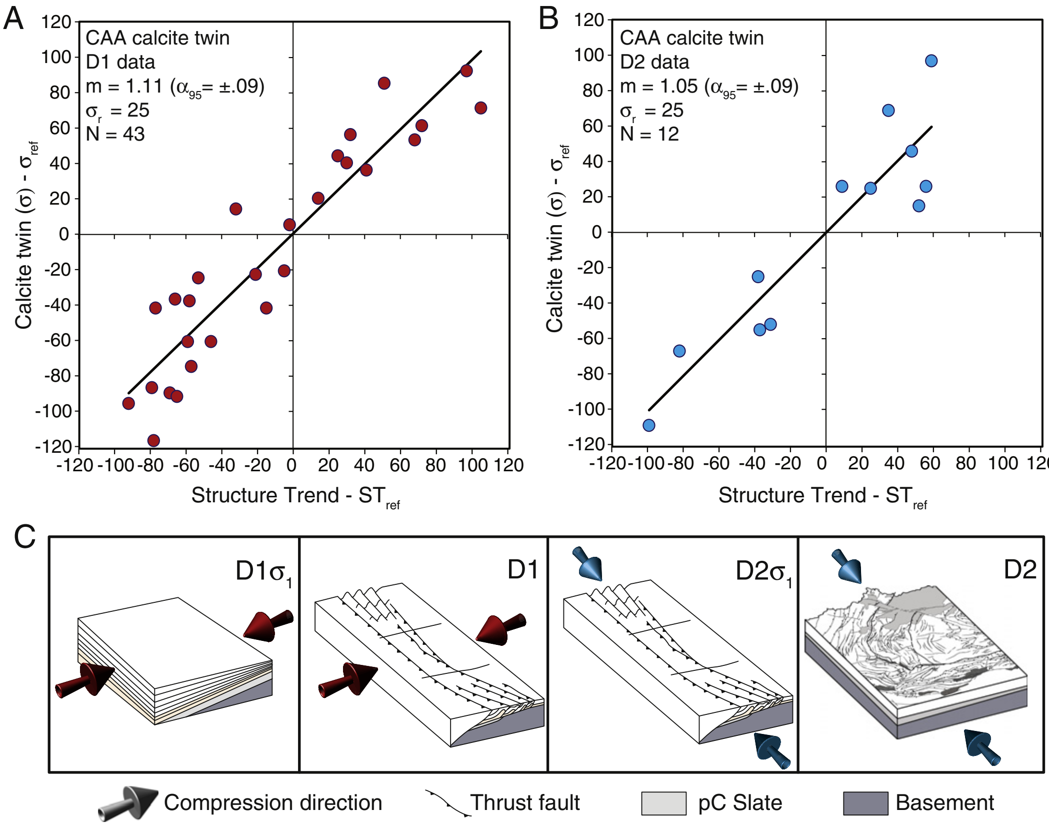


Fig. 8. (A and B) Strike tests and (C) schematic block diagrams summarizing the evolution of the Cantabrian–Asturian Arc based on acquisition of calcite twin strains, and associated stresses, during two discrete deformation phases associated with oroclinal buckling. Data and block diagram modified from Kollmeier et al. (2000). (A) Structural strike test for data from calcite twin strains interpreted to represent D1 east–west shortening (in present-day coordinates using a $\sigma_{ref}=244°$) (slope of 1.11±.09). (B) Structural strike test for data from calcite twin strains interpreted to represent D2 north–south shortening (in present-day coordinates using a $\sigma_{ref}=176°$) (slope of 1.05±.09). Statistical parameters are as in Fig. 6c.

which resulted in initial longitudinal folding and faulting. Strike tests of the pre-Stephanian joint sets indicate that there is a direct correlation between changes in structural trend around the Cantabrian–Asturian Arc and changes in the orientations of preserved D1 joint sets (strike test slopes of 1.03 and 1.16) (Fig. 9c). Consequently, the current orientation of pre-Stephanian joint sets are a consequence of about 180° of vertical-axis rotation of approximately linear joint sets that started parallel, and perpendicular with, early longitudinal fold axes; while the Stephanian outcrop joint sets record about 100° of rotation (Fig. 9b).

Combining the age constraints for the timing of joint set formation from existing angular unconformities, with strike tests that quantified relative amounts of rotation experienced by each subsequent set (Fig. 9), Pastor-Galán et al. (2011) argued that during the Kasimovian the Cantabrian–Asturian Arc closed between 30% and 50%, and by the lower-most Permian was completely closed. Assuming a simplified constant rotation rate, about 100° of arc buckling took about 5 Ma from the upper-most Kasimovian to the Carboniferous–Permian boundary. The initiation of rotation in the Cantabrian–Asturian Arc had to be before the generation of the joints in the Stephanian rocks and, if the buckling rate was similar to that of Stephanian times, suggests that buckling began in the Moscovian (around 310 Ma) (Fig. 7).

In the outer arc portion of the Cantabrian Orocline, Aerden (2004) used three-dimensional microstructural analysis of porphyroblasts to argue for secondary oroclinal buckling. By documenting the geometric relationships between inclusion trails and regional structural trends Aerden (2004) was able to isolate the multiple deformation phases that affected NW Iberia during the late stages of the Variscan orogeny. Specifically it was determined that the strikes of the dominant matrix foliation in the analyzed samples paralleled the overall structural trend of the Cantabrian Orocline; whereas, the strikes of their inclusion trails were independent of their location around the arc, and instead correlated with successive crustal shortening events during the Variscan orogeny. These observation were used to support a two-stage shortening model, which resulted in oroclinal development by late-stage modification of an originally N–S to NE–SW trending orogen (Aerden, 2004), although this method does not provide discrete temporal constraints on the deformation events.

Though not a classic macro- or microscopic structural fabric, paleocurrent data was recently used as a geometric marker for tracking strain in the coupled Cantabrian and Central Iberian orocline system (Shaw et al., 2012a). Paleocurrent data (e.g., cross bed foresets, ripple crests, ball and pillow structures, slump folds, and incised channels) was collected from outcrops across the Iberia in the lower Ordovician Armorican Quartzite (Fig. 10). This quartzite is a prominent unit of the lower Paleozoic Gondwana margin and is predominantly comprised of thick-bedded clean quartzites, but also contains beds of mature sandstones with silt and shale intercalations (Aramburu, 1989; Aramburu and García-Ramos, 1993; Gutiérrez-Marco et al., 2002 and references therein; Gutiérrez-Alonso et al., 2007). Stratigraphic characteristics suggest a nearshore shallow water depositional environment under the range of tidal, shore current, and storm influences (e.g., Gutiérrez-Marco et al., 2002).

When restored to paleohorizontal, measured paleocurrent directions fanned around the coupled oroclines, which when restored for oroclinal buckling, indicates a formerly linear margin with a consistent westward paleocurrent direction (Fig. 10). Strike test analysis of site mean current directions reveals a best-fit linear model for all of the data, indicating a one-to-one correlation of changes in current direction with present-day changes in structural trend (slope of 1.2) (Fig. 10a). This palinspastic restoration implies that the Rheic Ocean

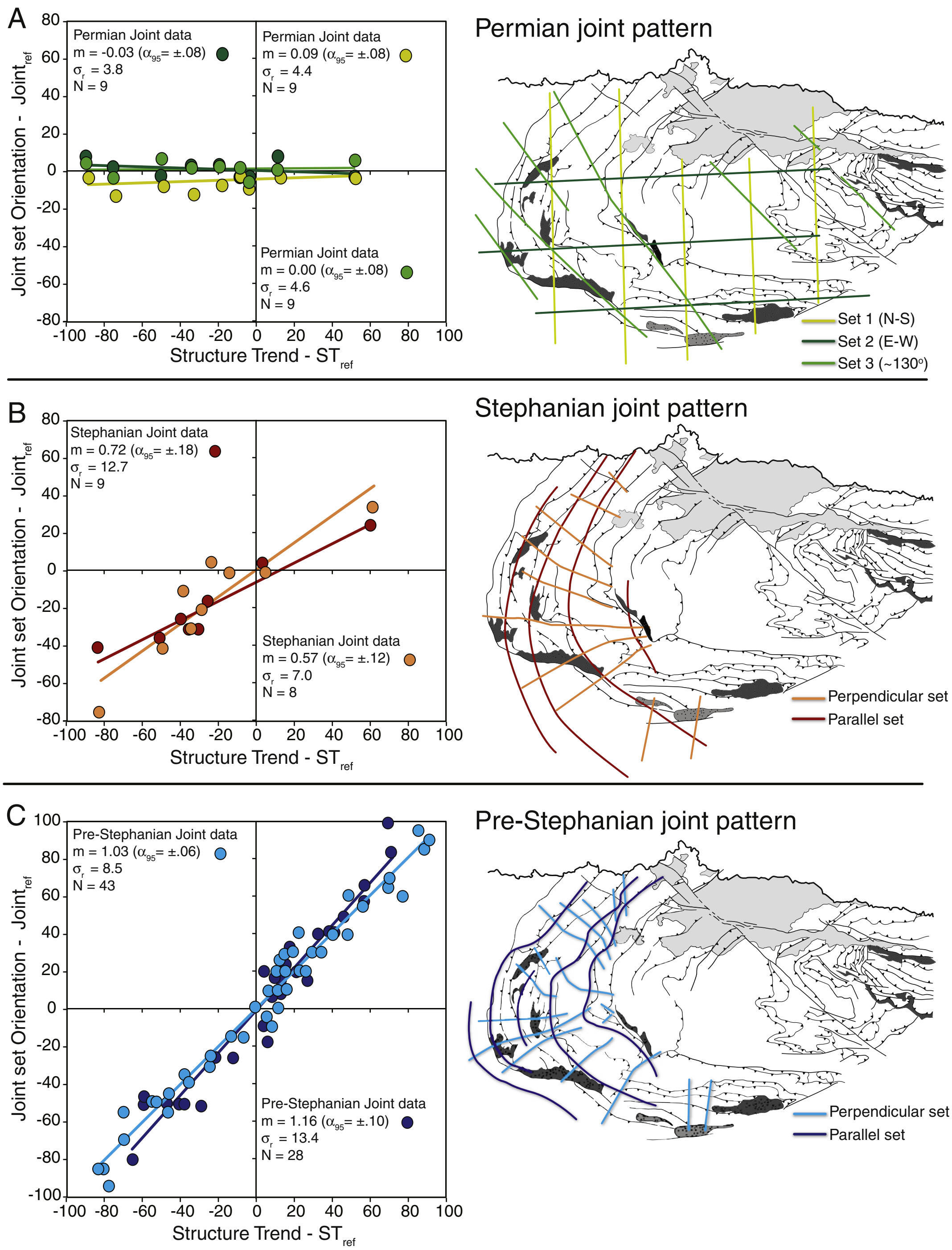


Fig. 9. Strike tests and schematic cartoons of the orientation of different joint sets from data presented in Pastor-Galán et al. (2011). (A) Representative data of the three joint set orientations constrained to be Permian and younger. Note the lack of correlation between structural trend and joint orientation (slopes of $-0.03 \pm .08$, $0.09 \pm .08$, $0.00 \pm .12$). (B) Representative data of the two joint set orientations constrained to be Stephanian in age. Note the correlation between structural trend and joint orientation, which indicates joint formation during oroclinal buckling (slopes of $0.72 \pm .18$ and $0.57 \pm .12$). (C) Representative data of the two joint set orientations constrained to be Pre-Stephanian in age. Note the one-to-one correlation between structural trend and joint orientation, which indicates joint formation prior to oroclinal buckling (slopes of $1.03 \pm .06$ and $1.16 \pm .10$). Statistical parameters are as in Fig. 6c.

opened to the west (in present-day coordinates), and was bound by a landmass to the east (Fig. 10b) (the Gondwana continent). Unfortunately, these data do not provide further time constraints on orocline formation of the Central Iberian Orocline.

To better understand the localized structural response to oroclinal buckling, Van der Voo et al. (1997), Weil et al. (2000), Weil (2006), and Weil et al. (2012) performed detailed paleomagnetic studies on individual structural domains from throughout the core of the Cantabrian Orocline. These studies used the preserved record of syn-tectonic magnetizations to decipher the rotational and kinematic history of individual thrust sheets, and described in detail how rotation, and hence oroclinal buckling might have been accommodated at the local scale (Fig. 11a–c). Due to the complexity of superimposed-folding in the Cantabrian–Asturian Arc, these studies determined the optimal tectonic correction for individual site mean directions within discrete structural domains. Each sampling site was evaluated in the context of the local structures to determine the best possible correction to undo post-magnetization tilts and rotations. Deformation axes were then

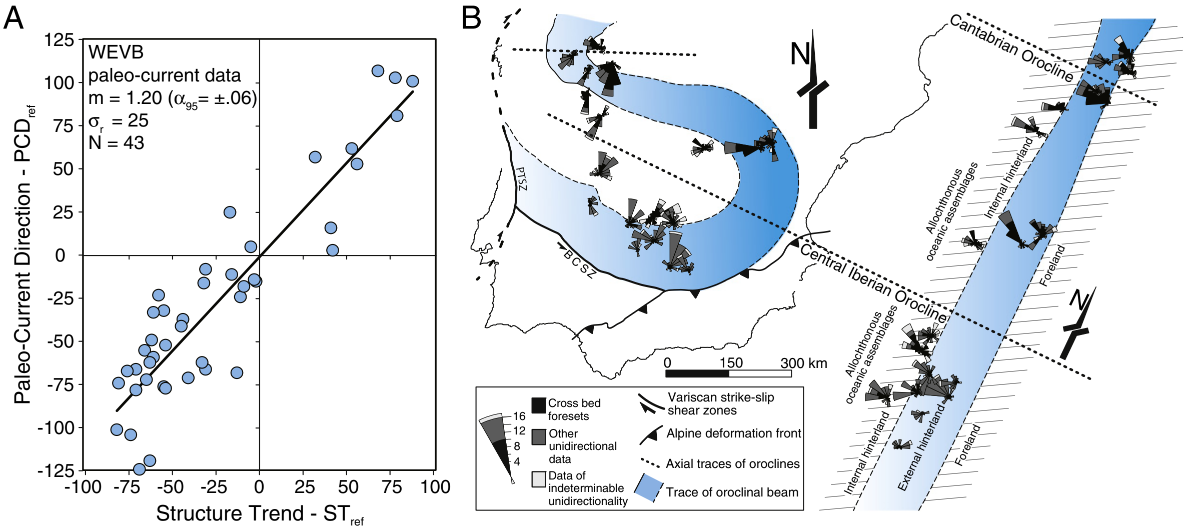


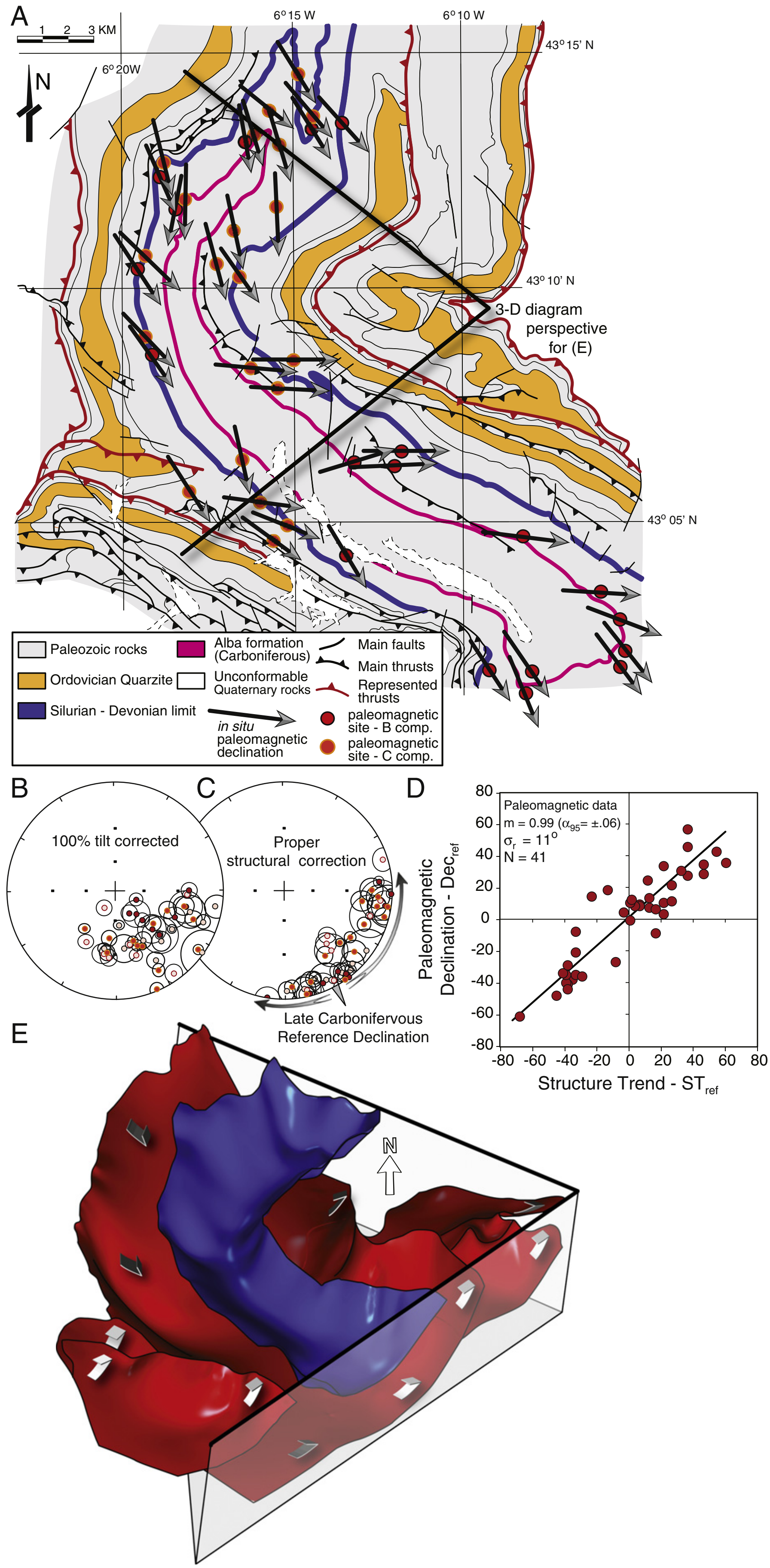
Fig. 10. Strike test and schematic cartoon of the orientation of paleo-current indicators around the oroclines of Iberia from Shaw et al. (2012a). (A) Paleo-current strike test for averaged data culled from Shaw et al. (2012a), showing a one-to-one correlation between changes in flow direction and structural trend around two tectonic bends (slope of 1.20 ± .06). Statistical parameters are as in Fig. 6c. (B) Reconstruction of the Iberian double orocline to its originally linear shape, which produces a uniform direction of offshore current for an originally linear orogen.

determined by calculating the best-fit rotation axes to cluster in-situ magnetic vectors to a known reference direction. The observed girdle distributions represented the differential rotation of D1 fold limbs during D2 oroclinal buckling (Fig. 11c). These studies argued for an early D1 folding and faulting event that produced longitudinal folds that today are arc parallel (based on earlier work by Julivert and Marcos, 1973). This was followed by D2 deformation that produced a 'radial' fold set (Julivert and Marcos, 1973) characterized by variably plunging fold axes that are imposed on existing D1 structural fabrics, and in the hinge zone produced conical folds (Fig. 11e) (Pastor-Galán et al., 2012b). In this way, the early longitudinal structures control the orientation, and position, of subsequent fold axes, which was accommodated by fault reactivation during D2 shortening, bed steepening, increased thrust stacking and arc tightening. A typical example comes from the study of the Lagos del Valle syncline by Van der Voo et al. (1997) (Fig. 11). This study found systematic clockwise and counterclockwise rotations totaling at least 120° within a single km-scale open syncline positioned in the hinge zone of the Cantabrian–Asturian Arc (Fig. 11a). Two remagnetization components were reported that were acquired during Variscan deformation, but prior to oroclinal buckling (the B and C components); thus, the magnetizations record 100% of the observed variability in structural trend seen today (Fig. 11c), and when restored produce a linear N–S trending synclinorium (Van der Voo et al., 1997).

Due to the apparent control of existing geologic surfaces on the accommodation of secondary vertical-axis rotation in the Cantabrian–Asturian Arc, Pastor-Galán et al. (2012b) investigated the effects of conical folding during superimposed episodes of bed rotations about horizontal and steep axes, with a focus on the hinge zone of the Cantabrian–Asturian Arc (Fig. 11d). They concluded that reactivating cylindrical longitudinal D1 folds resulted in two different structural regimes. Within the inner portion of secondarily bent folds, shortening occurs by the formation of radial conical folds with shallow west-plunging axes towards the vertical-axis of rotation (Fig. 11e) (Pastor-Galán et al., 2012b). Whereas in the outer portion of secondarily bent folds, there is D1 axis-parallel stretching and formation of large wavelength superposed folds with a distinct conical geometry (after Ramsay, 1967).

The Narcea Antiform, a major Variscan anticlinorium, delineates the trace of the Cantabrian Orocline and provides further support for a secondary origin for the Cantabrian Orocline (Fig. 2b). The anticlinorium is cored by Neoproterozoic rocks that are carried on major thrusts that lie along and define the foreland–hinterland boundary (Gutiérrez-Alonso, 1996). Thrust faults of the Cantabrian Zone foreland root into these major thrusts. The Narcea Antiform, including its major thrusts, outcrops along a continuous curved trace parallel to the Cantabrian Orocline structural grain. The major thrusts consist of shear zones that are up to 2 km wide, and which developed under low grade metamorphic conditions (Gutiérrez-Alonso and Nieto, 1996) during the upper Serpukhovian (321 Ma., Dallmeyer et al., 1997). Kinematic indicators in the shear zones show a centripetal pattern of hanging wall translation towards the core of the arc, a pattern that can only be acquired if the thrust related shear zones were rotated around a vertical axis subsequent to their emplacement.

Palinspastic restoration of individual structural domains (e.g., Lagos del Valle syncline in Van der Voo et al., 1997; Proaza anticline in Weil et al., 2000) and tectonic domains (e.g., the Ponga Units in Weil, 2006; the Esla Unit Weil et al., 2012) all indicate that the present-day structural sinuosity in the Cantabrian–Asturian Arc is a consequence of secondary rotation of originally linear features, and the modification and tightening of originally curvilinear features (e.g., hanging wall lateral/oblique structures) (Fig. 6a and b). These paleomagnetic and structural observations indicate that the Variscan tectonic history in the Cantabrian–Asturian Arc involved at least two temporally discrete deformation phases. D1 resulted in thrusting and folding related to west-to-east tectonic transport (in present-day coordinates) that lasted into the Kasimovian. Thrusting during this initial phase resulted in locally complex footwall geometries characterized by frontal and oblique/lateral ramps — particularly in the core zone of the Cantabrian–Asturian Arc (e.g., Ponga and Esla units (Alonso, 1987; Alvarez-Marron and Pérez-Estaún, 1988)) (Fig. 6a). D1 folding was followed by late Variscan D2 deformation that buckled originally linear, north–south (in present-day coordinates) trending thrusts and hanging wall folds (mainly by conical folding in the hinge zones of the arc), and modified drape folds associated with D1 frontal/lateral/oblique ramp intersections (Fig. 6b). Thrust sheet modification was accommodated by reactivation of lateral/oblique ramps as frontal ramps, reactivation of frontal ramps as oblique ramps, and overall tightening of D1 folds, commonly by conical folding of existing D1 arc-parallel fold limbs (Pastor-Galán et al., 2012b; Van der Voo et al., 1997; Weil et al., 2000, 2001; Weil et al., 2012). These structural modifications were a kinematic requirement for accommodating north–south shortening associated with oroclinal buckling.


A
1 2 3 KM
N
6° 20W
6° 15 W
6° 10 W
43° 15' N
43° 10' N
43° 05' N
3-D diagram perspective for (E)
Paleozoic rocks
Ordovician Quartzite
Silurian - Devonian limit
Alba formation (Carboniferous)
Unconformable Quaternary rocks
Main faults
Main thrusts
Represented thrusts
in situ paleomagnetic declination
paleomagnetic site - B comp.
paleomagnetic site - C comp.
B
100% tilt corrected
C
Proper structural correction
Late Carbonifervous Reference Declination
D
Paleomagnetic data
m = 0.99 (α95= ±.06)
σr = 11°
N = 41
Paleomagnetic Declination - Decref
Structure Trend - STref
E
N

5. Oroclinal lithospheric delamination model

The above review of paleomagnetic, structural, and geologic data indicate that the Cantabrian Orocline is a true secondary orocline as first envisioned by Carey in 1955. At this scale it seems unlikely, if not dynamically impossible, for the Cantabrian Orocline to be a thin-skinned crustal structure, as the space problems alone from rotating the inner Cantabrian–Asturian Arc though a 180° bend cannot be easily reconciled. Thus, one of the more challenging questions concerning the formation of the Cantabrian Orocline is the evolution of its three-dimensional geometry, especially with regards to the extent of the lithosphere affected by observed near-surface vertical-axis rotations. One of the main consequences of collisional orogeny is the initial thickening of the mantle lithosphere (e.g. Pysklywec et al., 2002, 2010). Modeling has shown that if lithospheric thickening is extensive enough, the lower lithosphere will become gravitationally unstable, and ultimately detach and cause rapid mechanical thinning of the mantle lithosphere beneath the orogenic belt (e.g. Bird, 1978; Collins, 1994; Davy and Cobbold, 1991; Houseman and Molnar, 1997; Houseman et al., 1981; Molnar and Houseman, 2004; Molnar et al., 1998; Morency and Doin, 2004; Nelson, 1992; Pastor-Galán et al., 2012c; Pysklywec, 2006; Pysklywec et al., 2002, 2010; Schott and Schmeling, 1998; Turner et al., 1992). Some of the predicted by-products of delamination are syn-orogenic or post-orogenic lithospheric thinning, high thermal gradients, domal uplift and near-surface extension (e.g., Ducea and Saleeby, 1998; Fernández-Suárez et al., 2000; Lee et al., 2006; Levin et al., 2000; Muñoz-Quijano and Gutiérrez-Alonso, 2007; Nelson, 1992; Saleeby and Foster, 2004; Whalen et al., 1996). The two most important geodynamic processes responsible for mechanical thinning of lithosphere in orogenic environments are: (i) slab break-off of subducted lithosphere (e.g. Gerya et al., 2004; Regard et al., 2008; van Hunen and Allen, 2011; von Blanckenburg and Davies, 1995), and (ii) lithospheric delamination (e.g., Bird, 1978, 1979; England and Houseman, 1989; Morency and Doin, 2004; Nelson, 1992). More recently Gutiérrez-Alonso et al. (2004, 2012) argued that oroclinal buckling at the crustal-scale can also result in differential thinning and thickening of the mantle lithosphere and may lead to lithospheric delamination. This model, based on the kinematic requirements of oroclinal buckling of the Cantabrian Orocline as dictated by the data presented above, infers that during oroclinal buckling, the mantle lithosphere thins below the outer arc and thickens beneath the inner arc (Fig. 12). These predictions were built upon those of Ries and Shackleton (1976) who first proposed a tangential longitudinal strain distribution for oroclinal buckling. A predicted thickened lithospheric root is not observed in deep seismic sections from the Cantabrian Orocline (Pérez-Estaún et al., 1994), and thus, if the model of lithosphere thickening beneath the inner arc is correct, its absence today suggests that a gravitational instability occurred that resulted in removal of the mantle lithosphere from the lower crust (Gutiérrez-Alonso et al., 2004). Today, a potentially similar process is occurring under the Vrancea Arc, in the Romanian Carpathians, where an intense swarm of deep earthquakes (60 to 200 km deep; e.g. Chalot-Prat and Girbacea, 2000 Gvirtzman, 2002;) is interpreted to be caused by ongoing mantle lithosphere delamination located under the maximum curvature region of the Vrancea orocline. More recent seismic profiling and tomography confirm the existence of an actively delaminating lithospheric root in the region (e.g. Fillerup et al., 2010; Ismail-Zadeh et al., 2012; Knapp et al., 2005).

The oroclinal-lithospheric-delamination hypothesis further predicts that removal of mantle lithosphere leads to upwelling of the asthenosphere, with an associated increase in crustal heat flow. It has been well established that extensive magmatism accompanied formation of the Cantabrian Orocline, and provides evidence for the thick-skinned, lithospheric-scale response to buckling predicted by Gutiérrez-Alonso et al. (2004) (Fig. 13) (Gutiérrez-Alonso et al., 2011a, 2011b). Syn-orogenic Variscan granitoid magmatism was active from 345 Ma to 315 Ma and records the building and collapse of the Variscan belt (Fernández-Suárez et al., 2000) associated with major D1 shortening and subsequent orogenic extension and construction of a roughly linear N–S tending orogen (Fig. 13a). Ensuing magmatism comprised of intrusive and volcanic rocks were emplaced from 310 to 285 Ma, and are associated with, and slightly post-date oroclinal buckling (Fig. 13a–c) (Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2011a). The late-stage magmatic record consists of mantle and crustal derived melts that show systematic changes in their age, spatial distribution, petrology and geochemistry, and include unique foreland magmatism in the core of the Cantabrian Orocline (Fig. 13d) (Gutiérrez-Alonso et al., 2011b).

Magmatism associated with oroclinal buckling began in the orogenic hinterland with intrusion of mantle and lower crustal derived mafic melts from 310 to 305 Ma (Fig. 13b). These mafic rocks and their accompanying granitoids are interpreted as a by-product of decompression mantle and lower crustal melting, caused by lithospheric extension around the outer arc of the orocline during buckling. Thinning of the lithosphere in the outer arc resulted in a concomitant rise of the asthenosphere, and eventual intrusion of gabbroic melts. Together, these responses would have elevated the regional geothermal gradient. This increase in thermal energy would then result in melting of middle-upper crustal rocks still hot from Variscan orogeny, leading to intrusion of felsic, crustal derived, magmas into the outer arc of the orocline between 305 and 295 Ma (Fig. 13c and d) (Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2004, 2011b).

In the inner arc of the orocline, magmatism did not begin until 300 Ma, and did not end until 285 Ma (Fig. 13c and d). Magmatism in the foreland core of the orocline began at about 295 Ma with the intrusion of mantle and lower crust-derived mafic rocks and granitoids and with widespread volcanism. This phase culminated with production of felsic, crustal-derived leucogranites in the foreland (Gutiérrez-Alonso et al., 2011b). The delayed onset of magmatism within the foreland is interpreted to reflect initial thickening of the lithospheric mantle in the core of the orocline, forming an orogenic root that subsequently became gravitationally unstable. Delamination and sinking of the unstable root facilitated upwelling of hot asthenospheric mantle beneath the foreland core of the orocline, giving rise to mantle derived mafic magmatism and melting of the lower crust. The subsequent felsic melts are attributed to melting of the fertile (pelite- and greywacke-rich) middle crust upon upwards migration of the thermal anomaly above the high-standing asthenosphere (Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2011b).

The Sm/Nd isotopic ratios from mantle-derived rocks from NW Iberia provide additional evidence of mantle lithosphere involvement during orocline development (Fig. 13e) (Ducea, 2011; Gutiérrez-Alonso et al., 2011a). Pre-Variscan mantle-derived volcanic rocks

Fig. 11. (A) Geologic map of the Lagos de Valle syncline in the core of the Cantabrian–Asturian Arc. In situ paleomagnetic vectors for B component (red circles) and C component (orange circles) plotted from Van der Voo et al. (1997). (B) Equal area stereonets of 100% structurally corrected and syn-tectonic corrected paleomagnetic site means and associated α95 cones of confidence. Red filled (open) circles represent lower (upper) hemisphere projection of B component magnetizations acquired after early D1 folding, but prior to oroclinal buckling. Orange filled (open) circles represent lower (upper) hemisphere projection of C component magnetizations acquired during early D1 folding, and prior to oroclinal buckling. Note high degree of scatter in 100% structurally corrected data, indicating a syn-tectonic acquisition. (C) Equal area stereonets of B (C component) component corrected to 5% (40%) untilting as in Van der Voo et al. (1997). Note that proper structural correction brings paleomagnetic site means to a consistent inclination, but with a high degree of declination scatter away from the Late Carboniferous reference direction for Iberia (gray triangle) as a consequence of significant vertical-axis rotation. (D) Paleomagnetic strike test for data from the Lagos de Valle syncline, which gives a slope of $0.99 \pm .06$. (E) Three-dimensional conical fold model for central core of Lagos de Valle syncline (after Pastor-Galán et al., 2012b).

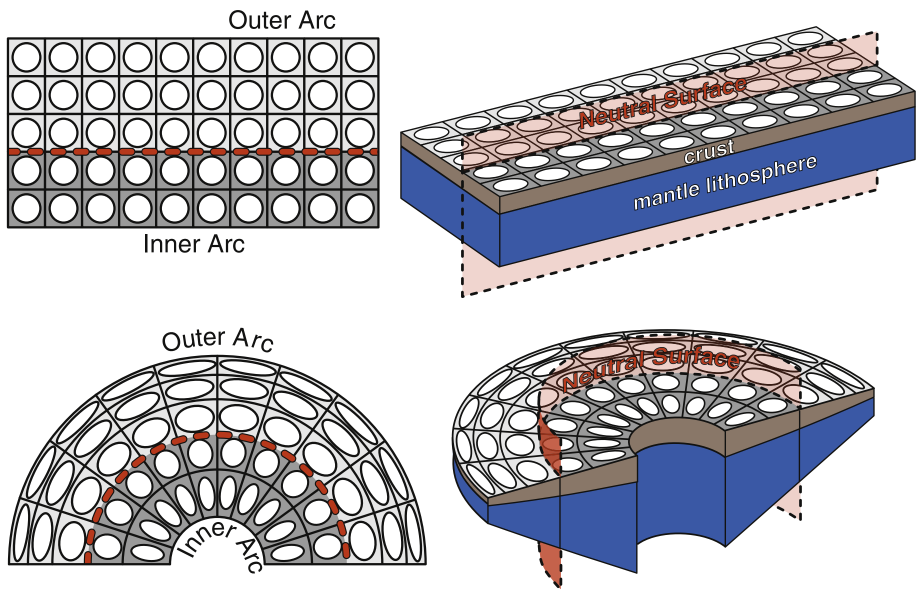


Fig. 12. Schematic diagrams depicting the effect of lithospheric buckling around a vertical axis and the resultant strain field (modified tangential longitudinal strain). Strain ellipses depict arc-parallel shortening in the inner arc and arc-parallel stretching in the outer arc. The different behavior of the mantle lithosphere in the inner and outer arcs and the increase in thickness of mantle lithosphere below the inner arc and thinning below the outer arc are highlighted.
Modified from Ries and Shackleton (1976).

indicate that the mantle lithosphere in NW Iberia was emplaced, or metasomatized, at about 1.0 Ga while post-Variscan mantle-derived magmatic rocks yield neodymium model ages (TDM) of about 300 Ma. This dramatic change in mantle lithosphere age indicates that orocline formation was coeval with removal of an older mantle lithosphere that was subsequently replaced by a new, juvenile mantle lithosphere. The mantle-derived melts that were formed during orocline buckling were contaminated by crustal sources, and yield model ages that span the inferred age of the underlying pre-Variscan lithosphere and the new juvenile lithospheric mantle (Gutiérrez-Alonso et al., 2011a). The resultant contamination indicates that melting of the continental mantle lithosphere and lower crust, and the subsequent mixing with upwelling asthenosphere, is likely responsible for generating the new lithospheric mantle (Fig. 13e) (Gutiérrez-Alonso et al., 2011a).

Overall, this protracted and voluminous phase of magmatism in NW Iberia had a profound effect on the thermal structure of the upper crust. Throughout the inner arc there is pervasive replacement and void-filling dolomitization of Carboniferous limestones that are post-dated by calcite cements (Gasparrini et al., 2003, 2006; Schneider et al., 2008). Geochemical analysis indicates the dolomitizing fluids were hypersaline and hydrothermal in nature, and field observations indicate that the dolomitization occurred after D1 folding and faulting (Gasparrini et al., 2006). Though most dolomitization fluid temperatures are estimated in the 100 to 160 °C range, there are some reactions related to talc deposits that formed at temperatures estimated to have been as high as 400 °C (Tornos and Saphiro, 2000).

A rock magnetic and petrographic study of Paleozoic carbonates from the Cantabrian–Asturian Arc was undertaken to further understand the relationship between hydrothermal orogenic fluids and the pervasive occurrence of remagnetizations during the late Paleozoic Variscan orogeny in northern Spain (Weil and Van der Voo, 2002). As indicated earlier, the Paleozoic carbonates from the Cantabrian–Asturian Arc contain at least three ancient late Paleozoic magnetizations. Scanning Electron Microscopy (SEM) of magnetic extracts from Cantabrian–Asturian Arc carbonates reveal abundant authigenic Fe-oxides and ubiquitous evidence of fluid flow driven chemical reactions that resulted in the formation of new Fe-oxides. Fluid reactions occurred along cracks and grain boundaries and within void space, and were genetically associated with Fe-rich clay and calcite–dolomite transformation reactions, or as oxidation of Fe-sulfide framboids. Together, the SEM observations and accompanying rock magnetic experiments revealed that the three late Paleozoic remagnetizations experienced by Cantabrian–Asturian Arc carbonates are chemical remanent magnetizations facilitated by the presence of thermally activated fluids associated with late Paleozoic Variscan deformation and oroclinal buckling (Tohver et al., 2008; Weil and Van der Voo, 2002).

In addition, abnormally high coal ranks are found throughout the core of the Cantabrian Orocline, especially near faults that bound syn-tectonic Stephanian basins in the Cantabrian–Asturian Arc (Colmenero and Prado, 1993; Colmenero et al., 1996, 2008). Sedimentological observations indicate that the high rank of these coals could not have been reached at the inferred depth of burial unless an extra heat source was involved (Colmenero et al., 2008). Heat transfer likely occurred via fluids circulating through networks of deep fractures and faults that partially accommodated orocline related tangential shortening. Further evidence for hot fluids associated with coal deposits is provided by the re-equilibration of fluid inclusions in quartz veins in Stephanian continental coal bearing basins (Ayllon et al., 2003; Frings, 2002) and the unusual occurrence of igneous intrusions in some coal seams that converted coal into anthracite (Colmenero and Prado, 1993; Knight, 1983).

Abundant hydrothermal mineralization characterizes the Cantabrian–Asturian Arc, including voluminous Zn–Pb deposits in the Picos de Europa region (Gómez-Fernández et al., 2000), and ca. 270–290 Ma gold mineralization (Martín-Izard et al., 2000) that is spatially associated with zones of low P–high T metamorphism (Arenas and Martínez-Catalán, 2003), epizonal plutonism (Valverde Vaquero et al., 1999) and Stephanian volcanism (e.g., the 303 ± 7 Ma Niao Andesite, Knight et al., 2000; Valverde Vaquero et al., 1999). Field relationships have been used to argue that gold mineralization is genetically linked to accommodation structures related to shortening in the inner arc and extension in the outer arc during oroclinal buckling (Jahoda et al., 1990).

Other geological evidence for increased heat flow in the Cantabrian Orocline includes: (i) late-Variscan hinterland uplift and coeval normal faulting, similar to that described for the Gibraltar Arc of southern Iberia during Late Miocene times (Duggen et al., 2003); (ii) foreland-directed gravity-driven movement of large thrusts (Martínez-Catalán et al., 2003); (iii) an increased flux of continental molassic sediments shed from the thermally elevated outer

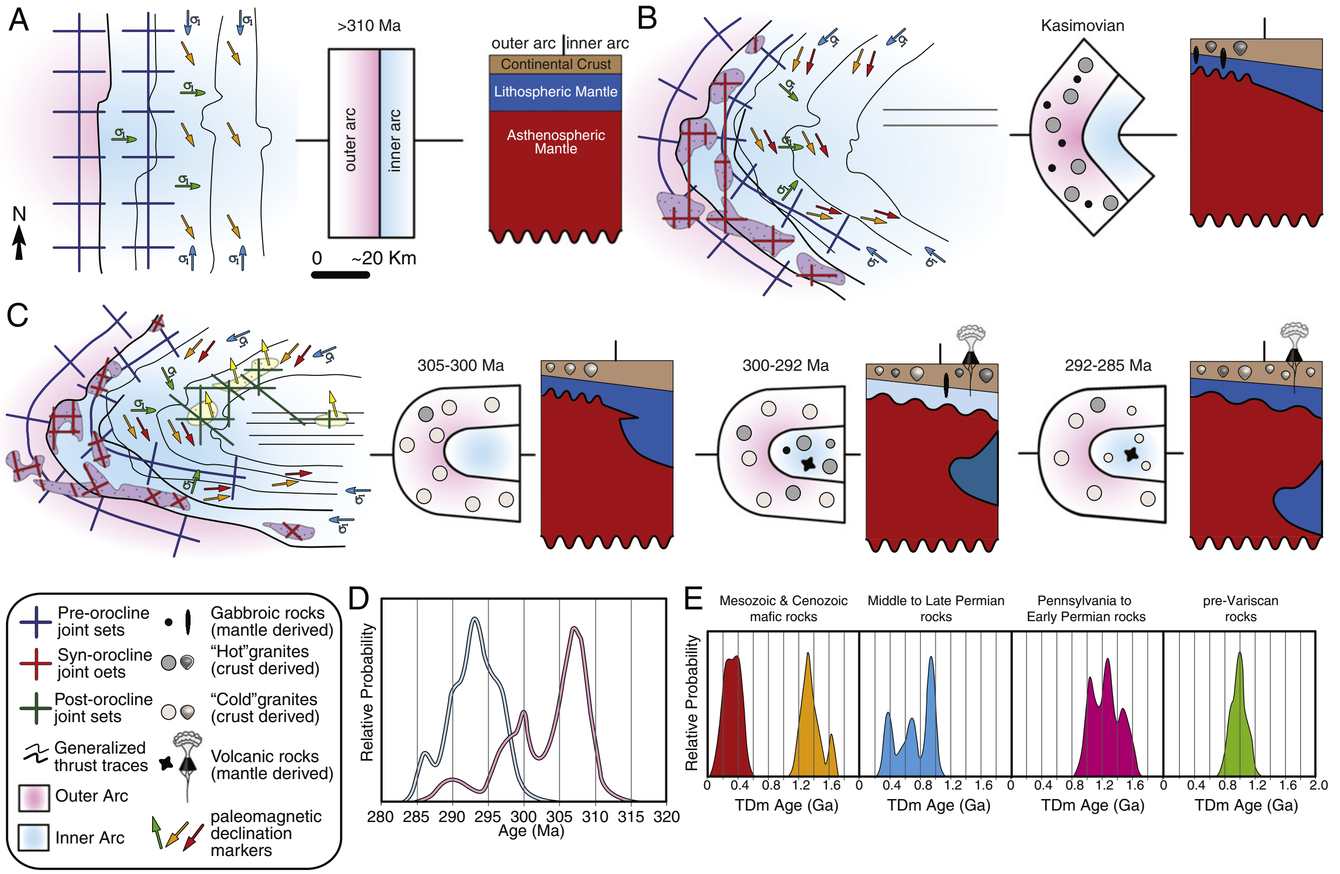
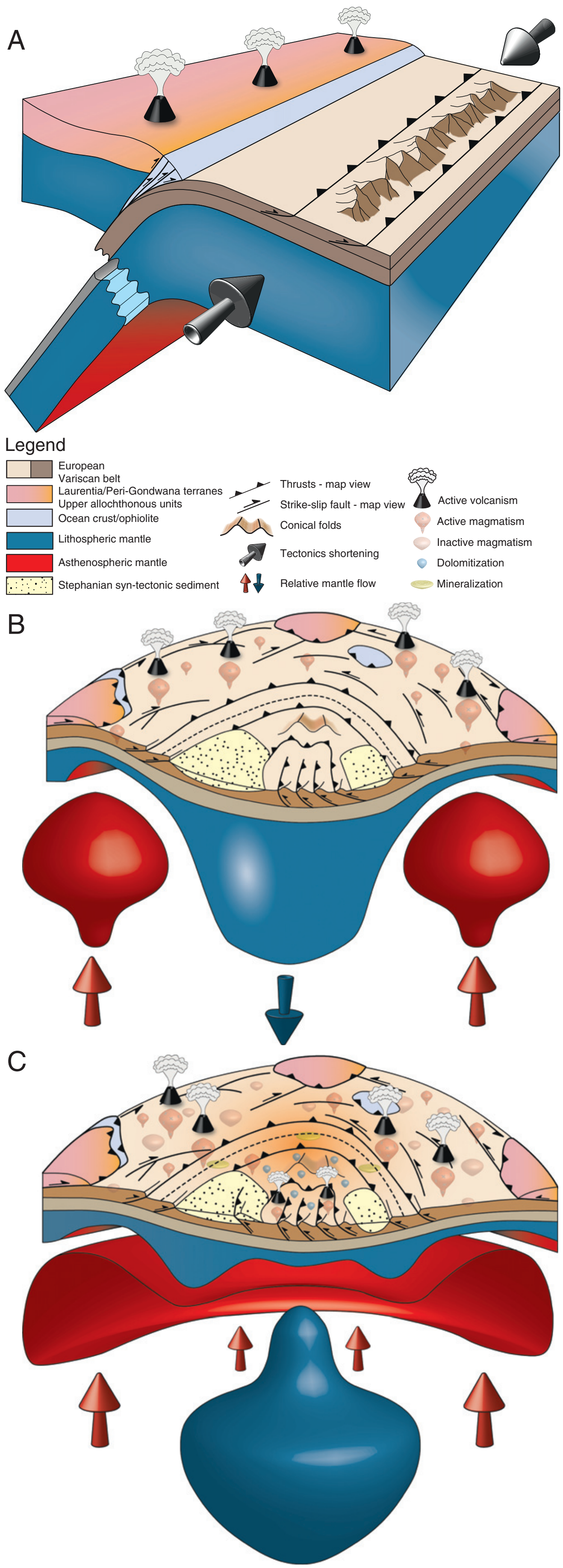


Fig. 13. Cartoon sketches summarizing the development of joint sets (Pastor-Galán et al., 2011), acquisition of multiple magnetizations (Weil et al., 2001, 2010), the orientation of calcite twin strains (Kollmeier et al., 2000) and the episodes of magmatism during formation of the Cantabrian Orocline. Outer (inner) arc region highlighted in pink (blue). (A) Shows the joints, calcite twin strains and paleomagnetic vectors interpreted to develop contemporaneous with formation of a nearly linear Variscan orogen in pre-Moscovian and Moscovian times. Plan view and profile view models for lithospheric structure of inner and outer arc during initial stage of east–west shortening (in present-day coordinates). (B) Depicts the early phase of orocline development during Kasimovian times, with around 20% of present-day curvature already attained. Plan view and profile view models depict thinning of the lithosphere in the outer arc, and thickening in the inner arc, with concomitant magmatism in the thinned outer arc. (C) Shows the final present-day geometry of the orocline and the orientation of the Early Permian paleomagnetic vectors showing no rotation, the orientation of pre-, syn and post-orocline joint sets, and the current orientations of measured calcite twin strains and syn-tectonic remagnetization vectors. Plan view and profile view models for lithospheric structure for three time slices during final phases of oroclinal buckling, which depict lithospheric thickening, eventual foundering and delamination, and the resultant magmatic pulses intruded into the crust. (D) Probability age distributions for magmatic rocks from the inner (blue) and outer (pink) arc regions. Note the general younging trend of magmatic ages from outer to inner arc during oroclinal buckling and related lithospheric delamination. (E) Model ε_{Nd} age probability plots grouped as: Mesozoic and Cenozoic mafic rocks, Middle to Late Permian rocks, Pennsylvanian to Early Permian samples, and pre-Variscan samples.
Data and schematic cartoons taken and modified from Kollmeier et al. (2000), Weil et al. (2001), Weil et al. (2010), Pastor-Galán et al. (2011), Gutiérrez-Alonso et al. (2011a, 2011b).

arc into the inner arc (Gutiérrez-Alonso et al., 2004; Pastor-Galán et al., 2012a); (iv) the presence of intramontane basins that were rapidly filled by Stephanian continental clastic sedimentary strata; (v) widespread volcanic activity (Bruguier et al., 2003a,b; Capuzzo et al., 2003); and (vi) thermal resetting of pre-existing metamorphic fabrics (Dallmeyer et al., 1997). Additional evidence of increased heat flow in the inner arc is indicated by the development of a subhorizontal cleavage (Aller, 1981) and localized mineral metamorphic parageneses that contain pumpellyte and muscovite + chlorite + chloritoid (Marín, 1997; Raven and van der Pluijm, 1986).

Dramatic changes in the thermal structure and thickness of the Cantabrian Orocline lithosphere during oroclinal buckling likely resulted in major topographic changes of the Earth's surface (Jiménez-Munt and Platt, 2006). Synchronicity of topographic changes and oroclinal buckling is provided by timing relationships of syn-tectonic basin deposits (Colmenero et al., 1996, 2002, 2008; Corrales, 1971and references therein). Uplift of the structurally thinned outer arc lithosphere by upwelling hot asthenosphere occurred coeval with subsidence of the thickened inner arc lithospheric, and produced a regional topographic slope from a high in the outer arc to a low in the inner arc (Fig. 14). This orocline-induced topographic gradient is recorded in thick conglomerate-rich continental deposits of Stephanian age preserved throughout the inner arc (Colmenero et al., 2008). Subsequent foundering of the lithospheric root under the inner arc (Fig. 14b and c) and its replacement by hotter, more buoyant, asthenospheric mantle, resulted in a topographic inversion that is recorded by the unconformity between Lower Permian and Stephanian sediments in the region (Martínez-García, 1991). These topographic changes agree with simple isostatic balance models of Muñoz-Quijano and Gutiérrez-Alonso (2007) and are in agreement with provenance shifts revealed by detrital zircon analysis (Pastor-Galán et al., 2012a).

The above geologic, geochronologic and geochemical data constrains the genetic links between mantle replacement and orocline formation in the Cantabrian Orocline; and consequently, the buckling of the Cantabrian Orocline must have involved the whole lithosphere (Gutiérrez-Alonso et al., 2004). Buckling of the entire lithosphere about a vertical axis requires a folding mechanism similar to the longitudinal tangential strain model of Ries and Shackleton (1976) (Fig. 12), with shortening in the inner arc and arc parallel extension in the outer arc accommodates oroclinal buckling at the lithospheric-scale. The transition from outer arc to inner arc in the Ibero-Armorican Arc is approximately the boundary between the hinterland and the foreland in NW



Iberia (Gutiérrez-Alonso et al., 2004). Geological and structural observations indicate that tangential arc-parallel shortening increases from the transition zone toward the inner core of the arc (Julivert and Marcos, 1973). This shortening was accommodated by formation of conical folds with axial traces radial to arc curvature (Aller and Gallastegui, 1995; Bastida et al., 1984; Gutiérrez-Alonso, 1992; Julivert and Marcos, 1973; Pastor-Galán et al., 2012b). These conical folds were caused by a decrease in shortening towards the outer arc and are observed in the field as re-folded pre-existing thrusts and related folds formed during Carboniferous E–W shortening (Gutiérrez-Alonso, 1992; Pastor-Galán et al., 2012b; Weil, 2006; Weil et al., 2000, 2012). In addition, to superposed folding and thrust reactivation in the inner arc, numerous strike-slip and reverse faults with complex movement histories in the southern branch of the arc played an important role in accommodating tangential shortening (Alonso et al., 2009; Marcos, 1968; Rodríguez Fernández and Heredia, 1988). These faults also played an important role in the location and distribution of Stephanian molassic basins (e.g., Alonso, 1987; Marcos, 1968; Nijman and Savage, 1989).

Arc-parallel stretching in the outer arc was accommodated by strike-parallel ductile elongation, which led to a pervasive subhorizontal stretching lineation in the deeper crustal levels (Brun and Burg, 1982; Matte and Ribeiro, 1975) and strike-slip faulting in the shallower levels. These strike-slip faults have displacements ranging from kilometers to tens of kilometers and a conjugate pattern in which the dextral component dominates over the sinistral component. Many of these faults have associated releasing stepovers that acted as emplacement loci for orocline related late-Variscan granitoids (e.g. Aranguren et al., 2003). In general, the number of strike-slip faults decreases towards the neutral surface where they are progressively replaced by low-grade dextral shear zones.

In order to better understand the lithospheric-scale behavior and response to oroclinal buckling, Pastor-Galán et al. (2012c) performed analog modeling of orocline formation. The analog experiments were performed with multiple rheologic layers of plasticine, and imaged with 3D tomography. Model response indicated extension and thinning of lithospheric mantle in the outer arc and accumulation of lithospheric mantle in the core of the arc, which, depending on the initial lithospheric mantle thickness, resembled the shape of a fold that was duplicated by a fault-like structure. The observed model geometries obtained by Pastor-Galán et al. (2012c) are in accordance with theoretical models proposed by Ries and Shackleton (1976) and Gutiérrez-Alonso et al. (2004). The observed generation of an overthickened lithospheric mantle root under the orocline core is fundamental to explaining the spatial and temporal distribution of magmatic activity (Fernández-Suárez et al., 2002a,b; Gutiérrez-Alonso et al., 2011b).

The above discussion emphasizes the kinematic observations, and possible consequences of buckling a lithospheric orocline during the final stages of Variscan orogenesis. What has not been directly addressed, and which is not the focus of this review, is the geodynamic mechanism(s) and tectonic scenario responsible for the buckling. Given that the kinematic observations indicate a secondary origin for the Cantabrian Orocline, the question remains — what caused the necessary far-field stress change from compression normal to the orogenic belt to sub-parallel to it? In light of the orocline model, several ideas have been proposed to date: (i) the self subduction of Pangea (Gutiérrez-Alonso et al., 2008), (ii) buckling of a ribbon continent between Laurussia and Gondwana during the final amalgamation of Pangea (Johnston and

Fig. 14. Schematic block diagram illustrating orocline development starting with (A) a linear belt resulting from a Gondwana–Laurentia collision. (B) Oroclinal buckling caused lithospheric stretching in the outer arc, with associated magmatism, and thickening beneath the inner arc (modified from Gutiérrez-Alonso et al., 2004). (C) The final stage of oroclinal buckling depicting delamination and collapse of thickened lithospheric root beneath the inner arc, replacement of sinking lithosphere by upwelling asthenospheric mantle, and associated magmatism in the inner and outer arc regions.

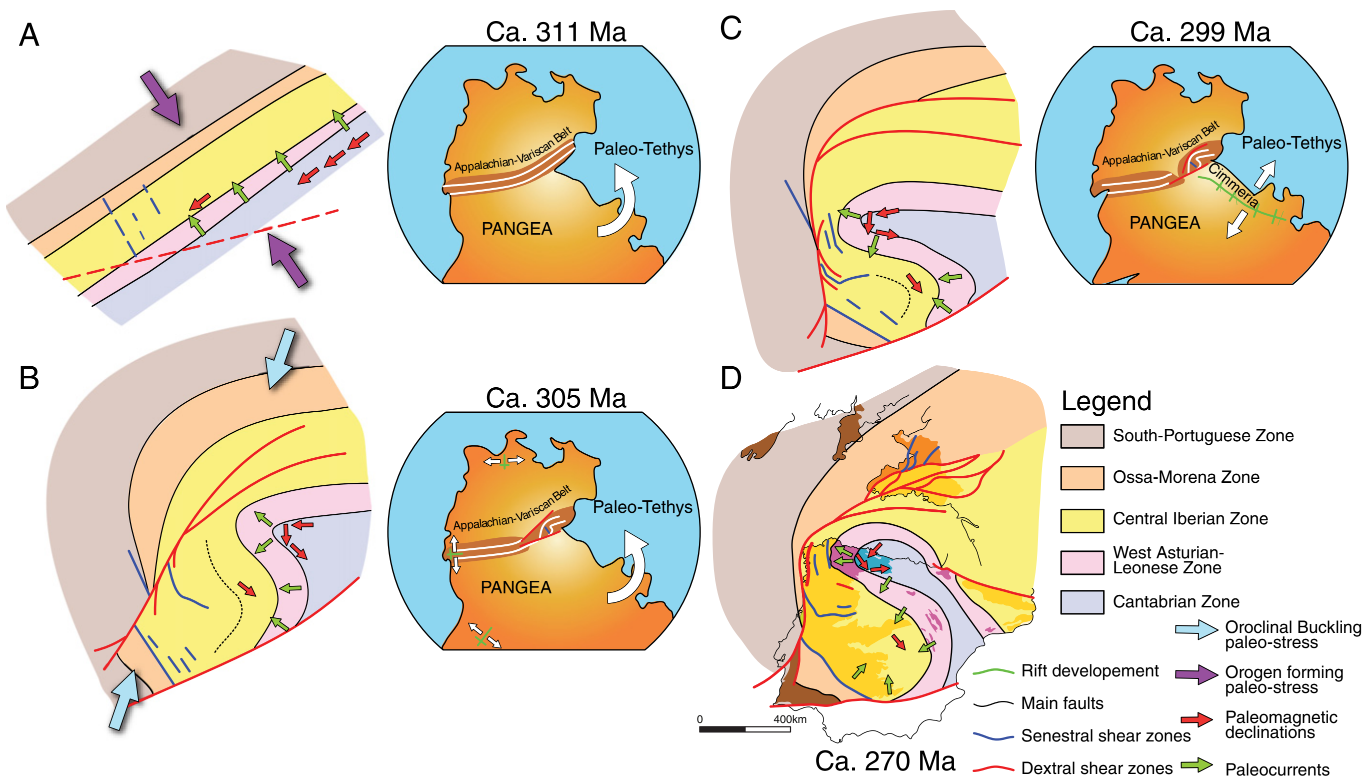


Fig. 15. Tectonic evolution of the Variscan oroclines in Iberia and simplified paleogeographic evolution during Upper Pennsylvanian times. A) Latest stage of Western Europe Variscan Belt development, highlighted by a nearly linear orogen in which B and C magnetizations and paleocurrents are shown. Dashed lines represent future shear zones. B) Buckling of the Variscan orogen related to a change in the regional stress-field probably related with a relative movement between the southern and northern sides of Pangea (see text for further details). At this time the orogen was buckled about a 50%, shear zones developed in the outer arc, drawing a "C shape" while in the inner arc an "S shape" is developing due to disharmonic folding. C) The orogen has been completely buckled. D) Final architecture of the Variscan orogen before the opening of the Biscay Bay.

Gutierrez-Alonso, 2010; Weil and Gutiérrez-Alonso, 2007; Weil et al., 2010), and (iii) a continental-scale post-Variscan dextral shear zone (Martínez-Catalán, 2011, 2012; Martínez-García, 2012). The first idea takes into account the penecontemporaneous development of Cantabrian Orocline with final amalgamation of the Pangea supercontinent. Within this larger paleogeographic framework it is possible to produce dramatic stress-field changes at the regional scale. If one takes into account the northward subduction of the Paleotethys spreading ridge beneath Siberian Pangea before rifting of the Cimmerian ribbon continent, then Pangea would have to have begun consumption of its own oceanic lithosphere. This scenario would result in shortening near the apex of the northern Paleotethys subduction zone, and extension in the outer portions of the Pangea plate, which paleogeographically was the position of the Cantabrian Orocline. The inferred kinematic consequences implied by this scenario explains the late dextral faulting seen throughout the Cantabrian Orocline, penecontemporaneous Late Carboniferous and Early Permian compression, and the large-scale extension in the outer zone of Pangea that ultimately resulted in creation of radial rift zones, peripheral Pangea seas, and rift-to-drift of the Cimmerian ribbon continent. The second, ribbon continent model, is also related to the larger paleogeographic configuration of major crustal blocks during final Pangea amalgamation. In this model the Cantabrian Orocline is considered a buckled ribbon continent, similar to SAYBIA of the Western North American Cordillera (Johnston, 2001, 2008), which consisted of microcontinental blocks brought together prior to or during closure of the Rheic Ocean. The inspiration for this model comes from the requirement of bounding free-surfaces to allow for the space considerations, amounts of translation and the geometric restrictions of buckling the lithosphere. The ribbon continent would have been detached on its western (Rheic Ocean) and eastern (Paleotethys) flanks, with major dextral strike-slip faulting between it and Gondwana that accommodated oroclinal buckling. The ribbon continent was probably pinned to a larger continental block on at least one edge to allow buckling during tectonic transport, analogous to SAYBIA (Johnston, 2001, 2008). This hypothesis assumes that the dramatic rotations observed in the Western European Variscan Belt were accommodated by major crustal detachments, many of which have yet to be identified. The final dextral shear model is based on the presence of a continental-scale dextral shear zone that reshaped the Variscan belt through wrenching along a diffuse plate-scale shear zone that produced a heterogeneous distribution of deformation. In this model the large shear zone, which would have run across present-day western Europe, acted as an inter-continental transform fault that included a series of en echelon shear zones, and that was kinematically connected to the convergent boundaries of the Appalachians and the Urals (Martínez-Catalán, 2011, 2012). The diverse geodynamic and tectonic implications of these models highlight the remaining uncertainty in the larger tectonics framework for the final stages of supercontinent amalgamation, and highlight the need for more work to be done on the evolution and development of the late Variscan oroclines into a global plate tectonic perspective.

Fig. 15 is a synthesis of existing knowledge and work in progress about the mechanisms and tectonic scenario that drove oroclinal buckling in Iberia including the formation of the Cantabrian and Central-Iberian oroclines. A change in the stress-field from a N–S to an E–W regime (in present day coordinates, Fig 15A and B) during Upper Pennsylvanian produced buckling of the Variscan orogen near the apex of the recently amalgamated Pangea supercontinent. This change in stress-field (simplified in the Pangea maps in Fig. 15) would drive a disharmonic buckling of the orogen (Pastor-Galán, 2012) producing both Iberian oroclines with an "S shape" while forming a greater and simple "C shape" in the outermost orocline, described by the trace of the South-Portuguese and Ossa-Morena zones beneath the Atlantic margin (Fig. 15).

6. Conclusions

The Paleozoic Variscan orogeny was a protracted collisional event that involved the dispersal and eventual amalgamation of multiple continents and micro-continents and culminated with the final amalgamation of the Pangea supercontinent. The data review presented above strongly supports a secondary oroclinal buckling model of an originally near-linear convergent margin during the last stages of Variscan deformation in the late Paleozoic. Closure of the Rheic Ocean between Gondwana and Laurussia resulted in E–W shortening (in present-day coordinates) in the Carboniferous, which produced a near linear N–S trending, east verging orogenic belt. Subsequent N–S shortening near the Carboniferous–Permian boundary resulted in oroclinal buckling.

Petrologic, geologic, geochemical and geochronologic data all point to penecontemporaneous magmatic and tectonothermal activity occurring synchronous with oroclinal buckling over a short 10 Ma time window at the end of the Carboniferous. Temporal and spatial relationships link these processes with thinning in the outer arc, thickening in the inner arc, and ultimately foundering and delamination of the mantle lithosphere under western Europe. Such cause and effect linkages help explain many previously enigmatic geologic events related to a post-orogenic Variscan Europe.

The Cantabrian Orocline, provides a unique location and opportunity for understanding the processes that take place when a lithospheric-scale orocline forms. The core of this arc in northern Spain has been studied for over a century and from multiple points of view, and the immense amount of existing data offers an exceptional opportunity to continue to develop ideas about the processes involved in orocline development, which can ultimately be used to better understand other active and ancient oroclines from around the world.

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