

1 Introduction

In the Late Neoproterozoic and Early Cambrian, a long history of subduction and accretion of island arcs occurred along the northern margin of Gondwana (e.g. Murphy et al., 2000). After the protracted period of rifting (e.g. Sanchez-Garcia et al., 2008), the Rheic Ocean opened by the Late Cambrian-Early Ordovician with the separation of several peri-Gondwanan terranes (Avalonia, Carolinia, Ganderia) from the northern margin of Gondwana (e.g. Murphy et al., 2006). This period of rifting and early drifting is recorded in NW Iberia by widespread rift-related igneous activity (e.g. Díez-Montes, 2006; Murphy et al., 2008), and by the coeval accumulation of a thick passive margin sequence (e.g. Aramburu et al., 2002). The Rheic Ocean reached its greatest width (ca. 4000 km) during the Silurian.

Largely on the basis of paleomagnetic data, some authors interpret the location of NW Iberia during the Late Silurian to be part of drifting ribbon continent variously called Armorica or the Hun terrane (e.g. Van Der Voo, 1982; 2002). The drifting of this putative microcontinent away from Gondwana is held to be responsible for the opening of the Paleotethys Ocean and its collision against Laurentia to be responsible for the closure of the Rheic Ocean and the onset of Variscan orogenesis. Other authors, however, place NW Iberia along the northern Gondwana passive margin throughout the Paleozoic (e.g. Robardet, 2003; Fernandez-Suarez et al., 2006; Díez Fernández et al., 2010), implying that subduction of Rheic Ocean lithosphere, which began in the Early Devonian was directed northward i.e. away from the Gondwanan margin.

In the latter scenario, the closure of the Rheic Ocean is recorded by the deformation associated with the final collision between Laurentia and Gondwana and in some ophiolitic suites preserved in the suture between these continents (e.g. Arenas et al., 2007). Continental collision began at ca. 365 Ma (e.g. Dallmeyer, 1997) and continued shortening is thought to have led the extensional collapse of the thickened hinterland at 320 Ma (e.g. Arenas and Catalan, 2003; Martínez Catalán et al., 2009). The latter event is coeval with the development of the non-metamorphic foreland fold-thrust belt of Gondwana (e.g. Perez-Estaun et al., 1994), which is exposed only in the CZ of NW Iberia.

Oroclines were firstly described by Carey (1955) as “an orogenic system that has been flexed in plan to a horse-shoe or elbow shape”. In its first definition, Carey used the word orocline in reference to originally linear orogens that had been bent during a subsequent deformation event. The term orocline was, however, commonly used in the literature as a geometric term for any orogenic curvature, because of the difficulties in constraining the deformation phases and consequently in determining if the orogen had originally been linear (Eldredge et al., 1985). Weil and Sussman (2004) proposed a classification for curved orogens based on kinematics, in contrast to other traditional classifications, which are largely based on the relationship and geometry of displacement and strain trajectories (e.g. Ries and Shackleton, 1976; Marshak, 1988; Ferrill and Groshong, 1993). This classification proposes three kinematic categories of curved orogens: (1) Primary arcs are initially curved belts that have not undergone subsequent rotation around a vertical axis; (2) Progressive arcs are belts that have increased their curvature from an original curved shape during their formation; (3) Oroclines are purely secondary curved orogens, that is to say, they acquired a curved shape from an initially linear, or almost linear, structural grain.

Constraining the kinematics of curved orogens is not a simple task. Paleomagnetic studies have long been used for unraveling vertical-axis rotations (e.g. Irving and Opdyke, 1965; Weil et al., 2000; 2010), but as discussed by Gray and Stakmatakos (1997), the interpretation of paleomagnetic data in complex curved orogenic belts is not straightforward. This difficulty in interpreting the paleomagnetic data arises from the fact that it is not always possible to know the

relative timing of magnetization acquisition (e.g. Weil and Van der Voo, 2002), in which case detailed structural analysis is a helpful alternative tool with which to estimate rotations and to investigate the kinematics of curved orogens. For example, orientation of calcite twins (Kolmeier et al., 2000), strain analysis (e.g. Yonkee and Weil, 2010), anisotropy of magnetic susceptibility (Weil and Yonkee, 2009) or joint analysis (e.g. Engelder and Geiser, 1980) have been used to constrain the kinematic evolution of curved orogens where paleomagnetic data are inconclusive.

Furthermore, the kinematic classification of curved orogens does not consider the mechanisms causing the curvatures. These mechanisms might include: (1) initial configuration of the sedimentary basin (e.g. Mitra, 1997), (2) changes in strength along detachments horizons (Marshak, 2004), (3) buttress effects (Paulsen and Marshak, 1999), (4) indentor tectonics (e.g. Ribeiro et al., 2007), (5) wrench-faulting (e.g. Cunningham, 1993), (6) lateral variations in lithospheric strength across mountain belts (e.g. Willingshofer and Sokoutis, 2009), and (7) buckling mountain belts and ribbon continents (e.g. Johnston, 2001).

The Ibero-Armorican orocline (IAO) of western Europe (Bard et al., 1975) is one of the most striking curved orogenic systems on Earth, tracing a bend of 180° in the Variscan structural grain (Weil et al., 2001) (Fig. 1-11). The IAO is situated within the Western European Variscan Belt, which resulted from Devonian-Carboniferous collision between Gondwana, Laurentia and several microplates (e.g. Martínez-Catalán et al., 1997). The impressive geometry of the IAO was first recognized by Suess (1885) and has been the object of many studies (Brun and Burg, 1982; Dias and Ribeiro, 1995), especially at its core (e.g. Julivert, 1971; Julivert and Arboleya, 1984; Weil et al., 2000). The aforementioned studies have attempted to decipher the curved mountain belt kinematics, and a wealth of different hypotheses, spanning the entire classification of Weil and Sussman (2004), have been proposed: (1) a primary arc inherited from a Neoproterozoic embayment (Lefort, 1979); (2) a progressive arc resulting from indentation of a point-shaped block situated either in Gondwana (e.g. Matte and Ribeiro, 1975; Brun and Burg, 1982; Dias and Ribeiro, 1995) or in Avalonia (Simancas et al., 2009), (3) an oblique collision producing a non-cylindrical orogen (Martínez-Catalán, 1990), (4) a thin-skinned origin produced by a progressive change in the transport direction of the thrust units similar to a photographic iris, (Pérez-Estaún et al., 1988), and more recently (5) a true orocline formed by the rotation around a vertical axis of an originally linear orogen (Weil, 2006; Weil et al., 2000; 2010; Gutiérrez-Alonso, 2004, 2008). The last hypothesis is the one that we have investigated in the present work.

Gutiérrez-Alonso et al. (2004) proposed a thick-skinned kinematic model for the IAO that integrates structural, geochronological, geochemical, magmatic and paleomagnetic data. This model invokes E-W shortening (in present-day coordinates) (i.e., Pérez-Estaún et al., 1991) that produced the initially linear or almost linear Variscan orogen. Subsequent N-S shortening (e.g. Julivert and Marcos, 1973; Weil et al., 2001, 2010; Merino-Tome et al., 2009) led to lithospheric-scale rotation of the orogen limbs. Buckling was accommodated by local and regional thrust rotation and conical folding of the thrust units around E-W axes that produced interference folds in the inner arc when superposed on longitudinal N-S trending structures (Julivert and Marcos, 1973; Julivert and Arboleya, 1984; Álvarez-Marrón and Pérez-Estaún, 1988) and the development of crustal-scale shear zones in the outer arc (Gutiérrez-Alonso et al., 2010). At a lithospheric-scale, the model infers mantle lithosphere thinning below the outer arc and thickening beneath the inner arc based on a tangential longitudinal strain distribution for buckling (Ries and Shackleton, 1976; Gutiérrez-Alonso et al., 2004). Since this thickened lithospheric root is not observed in deep seismic sections (Pérez-Estaún et al., 1994), lithosphere thickening beneath the inner arc might have resulted in a gravitational instability, causing oroclinal development to be followed by mantle lithosphere removal from the lower crust (Gutiérrez-Alonso et al., 2004). Mantle

lithosphere removal would have led to upwelling of the asthenosphere with the associated increase in heat flow. Based on geological (Pastor-Galán et al., 2011) and paleomagnetic constraints (van der Voo et al., 1997; Weil et al., 2000; 2001; 2010) the Ibero-Armorican orocline buckling took place from 310 to 300 Ma in uppermost Carboniferous time.

This model is based on the kinematic observations and consequences of the orocline buckling and does not consider tectono-mechanical aspects of the buckling. Gutiérrez-Alonso et al. (2008) suggest that the self-subduction of the Pangea global plate could have produced the change in the stress field required for orocline buckling. In the latter model, Pangea's oceanic lithosphere subducted northwards beneath the portion of Pangea occupied by Siberia. This scenario would have produced extension in the outer parts of Pangea and shortening near the vertex of the northern Paleotethys subduction zone. According to paleogeographic reconstructions of Stampfli and Borel (2002), the Ibero-Armorican arc was situated in the core of Pangea at the apex of the Paleotethys. Subduction near the apex should have resulted in bending of the lithosphere around a vertical axis. Moreover, the self-subduction model explains (1) dextral faulting in the Ibero-Armorican belt (Gutiérrez-Alonso, 2010), (2) late Carboniferous N-S shortening in the Cantabrian-Asturian arc, (3) large scale extension in the outer parts of Pangea that resulted in radial rift zones, and (4) the rift-to-drift transition of the Cimmerian ribbon continent.

In addition, an intense post-orogenic magmatic event (310 to 290Ma) has been detected during and after orocline buckling. Evidence for elevated heat flow includes: (I) the widespread post tectonic granitoids in the IAO and their particular spatial-temporal distribution (Fernández-Suárez et al., 2000; Fernández-Suárez et al., 2011; Gutiérrez-Alonso et al., 2011b); (II) uncommonly high coal ranks in uppermost Carboniferous continental basins (Colmenero and Prado, 1993; Colmenero et al., 2008); (III) gold mineralizations in the foreland fold-and-thrust belt (Martin-Izard et al., 2000); (IV) remagnetizations due to unexpected heat flow in the late Carboniferous and in the Permian (Weil and Van der Voo, 2002); (V) dolomitization along late breaching and out-of-sequence thrusts (Gasparrini et al., 2003); (VI) post-orogenic elevation of the topography (Muñoz-Quijano and Gutiérrez-Alonso, 2007a; 2007b).

Any tectonic model must also consider the rapid rotation of the structures in the uppermost Carboniferous and the thermal imprints during uppermost Carboniferous and early Permian times. Models assuming an initial curvature (Hirt et al., 1992) or models proposing thin-skinned progressive arc development (Pérez-Estaún et al., 1988) are not supported by paleomagnetic data (Weil et al., 2001; Weil, 2006). Indentor-based models require significant thick-skinned block rotation when the collision took place (i.e. in the Devonian). In addition, the foreland, being the place where the deformation took place during the late stages of orogeny (Dallmeyer et al., 1997), would not have exhibit a linear grain as suggested by the structural and paleomagnetic data.

The broad knowledge of the geology and the lithospheric response after the IAO orocline buckling makes it a natural laboratory in which to study how the mantle lithosphere could have behaved during thick-skinned orocline buckling. From this point of view, the only way to investigate the lithospheric geometry acquired during lithospheric buckling around a vertical axis is to develop models, either numerical or analogue, that can shed light on this process. The analogue modeling performed in this work has illustrates the development of lithospheric roots linked to thick-skinned orocline development.

1.2 Geology of the Cantabrian Zone

The CZ of northern Iberia is situated in the core of the Ibero-Armorican orocline (Weil, 2006; Gutiérrez-Alonso et al., 2004) (Fig. 1-11 of the thesis volume). The CZ is a classical foreland fold-and-thrust belt characterized by thin-skinned tectonics with a transport direction towards the core of the arc (Pérez-Estaún et al., 1988). Deformation in the CZ is characterized by low finite strain values (Gutiérrez-Alonso, 1996; Pastor-Galán et al., 2009), and cleavage is only locally developed. A very low-grade of metamorphism is indicated by illite crystallinity (Gutiérrez-Alonso and Nieto, 1996; Brime et al., 2001) and by conodont colour alteration index studies (Bastida et al., 2004; García-López et al., 2007). The Variscan deformation is diachronous towards the foreland. The first record of instability in the passive margin, due to its loading in the hinterland, is interpreted to have occurred in the upper Devonian (Keller et al., 2008) but the sedimentary record of a fore-bulge and a fore-deep is not evident until the Lower Carboniferous. Deformation began in the Late Mississippian (Dallmeyer et al., 1997) and resulted in the development of several clastic wedges related to the different thrust units.

The CZ consists of thick Neoproterozoic arc-related sequences, unconformably overlain by ca. 4500 m of lower Paleozoic clastic and carbonate platformal strata (Fig. 1-12) that thin towards the east and culminate with a distinctive sequence of Silurian black shale and iron-rich sandstone (Fig. 1-12). Paleocurrent data recorded in these strata indicate that its sediment source was located to the east (Aramburu and García-Ramos, 1993; Shaw et al., in press) but there are no currently exposed potential source rocks. The Devonian and Mississippian succession consists of alternating passive margin carbonate and siliciclastic formations (Fig. 1-12) where several transgressions and regressions have been documented (Keller et al., 2008; Aramburu et al., 2002; Gibbons et al., 2002). This succession is overlain conformably by a 5000 m thick Westphalian (Late Mississippian-Early Pennsylvanian) syn-orogenic sequence dominated by shallow marine and interbedded continental clastic strata followed by unconformably overlying Stephanian (Upper Pennsylvanian) and Permian rocks.

Stephanian strata young westwards (e.g. Colmenero et al., 2008) and show little deformation. They are coal-bearing, continental, sedimentary rocks that show similar stratigraphic and sedimentological characteristics over much of northern Iberia. Given this similarity, it is possible that the Stephanian succession was continuous across much of the western and southern portions of the CZ and the adjacent West Asturian Leonese Zone (Corrales, 1971). According to Pastor-Galán et al. (2011), these rocks do not contain the characteristic Variscan joint pattern that is observed in the older rocks, suggesting deposition after the bulk of Variscan deformation had taken place.

Permian rocks were deposited in small basins (Martínez-García, 1991) that post-date the formation of the Ibero-Armorican orocline (Weil et al., 2010; Pastor-Galán et al., 2011). These strata are only moderately tilted and are not internally deformed. The dominant lithologies are continental red conglomerates, red shales and sandstones, with minor limestones, volcanoclastic rocks and calc-alkaline basaltic lava flows with sparse isolated coal seams (Martínez-García, 1981; Suárez, 1988).

2 Joint Analysis

In order to document systematic joint sets in each of the three studied rock groups (pre-Stephanian, Stephanian and Permian outcrops), 172 measuring stations were analyzed in Stephanian outcrops (between 10 and 38 per outcrop), 64 stations in pre-Stephanian outcrops, and 6 stations in Permian outcrops. All studied rock units in the CZ and WALZ (Fig. 2-4) contain at least two systematic joint sets (Figs. 2-5 to 2-11). At least 30 joints per station were measured following the methodology described by Engelder and Geiser (1980).

The majority of the joints observed in the region depict characteristic plumose decoration of the joint planes when developed in fine grained clastic rocks, with no apparent record of shear. Lack of slip indicators suggests that most of the observed fractures originated as Mode I (tensile) cracks. Nevertheless, some joint surfaces do show evidence of shear re-activation that occasionally produced fibrous minerals on the joint surfaces with sub-horizontal orientation. This is especially true in the Pre-Permian rock sequences.

Intersecting and abutting relationships were carefully observed in the field to establish the relative timing of the different joint sets. Joint sets were separated according to orientation criteria. We assume there is temporal control on the relative timing of the different joint sets from the bracketing unconformities, joint sets present in the youngest rock sequence (i.e., above the bounding unconformities) are subtracted from those joint set populations measured in the oldest rock sequences (i.e., below the bounding unconformities). Thereby distinguishing those joint sets generated prior to the deposition of the overlying unconformable rocks.

All identified joint sets depict sub-vertical dips making them comparable using rose diagrams. Only in outcrops from the southern branch of the CAA do some sets have dips of ca 65°-75° (interpreted to be slightly tilted by the effects of Mesozoic deformation in the area) (Alonso et al., 1996). Backtilting of the aforementioned joint sets was performed and the orientations obtained were statistically identical to their in situ orientations.

Three joint sets are distinguished in the Permian basins (Fig. 2-5), each having a constant orientation. The most prominent set is oriented north-south with a strike of ~170°; secondary and tertiary sets are oriented east-west at ~90°, and northeast-southwest at ~130°. These joint sets have been described over the entire CAA in the pre-Permian rocks with little variation with respect to regional trend.

The joint sets present in Stephanian rocks have a more complicated pattern (Figs. 2-4, 2-8, 2-9, 2-10, 2-11). In addition to the uniform joint sets found in the overlying Permian rocks, all Stephanian outcrops have a joint set that is sub-parallel to the local basin-scale fold axis ("strike set" or "strike-parallel joints" in Engelder and Geiser, 1980) and a second joint set that is sub-normal to the fold-axis. These sets exhibit a variation of less than $\pm 10^\circ$ within individual stations of the same outcrop.

The outcrops studied from south to north are: the La Magdalena outcrop, which trends about W-E (Fig. 2-8A) and The Villablino outcrop, which has a trend of about 110° (Fig. 2-8B). Both outcrops are positioned in the southern limb of the CAA. The Ventana outcrop trends about 140° (Fig. 2-9A) and the Rengos outcrop, which has a regional strike of about 150° (Fig. 2-9B), are both situated in the southern limb of the CAA, but closer to the hinge of the arc. The N-S trending, slightly curved, Cangas del Narcea (Fig 2-10A) and Carballo (Fig. 2-10B) outcrops are located in the hinge of the arc, and the Tineo outcrop has a regional strike of about 30°, and is located slightly to the north of the Cangas del Narcea and Carballo localities (Fig. 2-11).

Three additional smaller outcrops were studied but are not included in the tables (in the

volume in spanish) - the northernmost Arnao (Figs. 2-4 and 2-12A), Buxeiro (Figs. 2-4 and 2-12B) and Combarcio (Figs. 2-4 and 2-12C) outcrops. Only one station in each outcrop was studied due to lack of sufficient exposure. The Arnao and Buxeiro regions have similar strike parallel ($\sim 60^\circ$ and $\sim 40^\circ$ respectively) and strike sub-normal ($\sim 140^\circ$ and $\sim 130^\circ$ respectively) joint sets; whereas, the Combarcio outcrop (Fig. 2-12C) had very limited outcrop and did not yield interpretable results.

In the La Magdalena, Villablino, Cangas and Tineo outcrops, some of the joint sets are indistinguishable from the post-Permian joint sets and are thus not included in further analysis.

In order to compare the Stephanian outcrop joint sets with joint sets preserved in the underlying pre-Stephanian rocks, the pre-Stephanian outcrops are separated into five groups distributed along the trace of the CAA. The groups are arranged based on a consistent structural trend between outcrops (Fig. 2-4). Each group contains data from between 10 and 15 outcrops. Each of the five pre-Stephanian groups corresponds with at least one studied Stephanian outcrop for comparison. The five sectors are: (i) the southern sector, which underlies the La Magdalena, Villablino and part of the Ventana outcrops (Fig 2-13A); (ii) the Rengos sector, which covers the Rengos, Ventana and southern limit of the Carballo outcrops (Fig 2-13B); (iii) the Cangas del Narcea sector, which extends around the Cangas del Narcea and northern portion of the Carballo outcrops (Fig 2-13C); (iv) the Tineo sector, which contains the Tineo, Buxeiro and Combarcio outcrops (Fig. 2-13D); and (v) the north sector, which covers all the pre-Stephanian outcrops north of the Tineo outcrop (Fig. 2-13E).

Both the Permian and Stephanian joint sets have been described in these zones. In addition longitudinal fold-axis parallel and fold-axis normal joint sets are observed and, because of their orientation relative to the folds, are interpreted to be tensile fractures (Hancock, 1985). As observed in Stephanian outcrops, some of the post-Permian and Stephanian joint sets are coincident with the pre-Stephanian sets (see tables in spanish volume).

The overall orientation of the studied joint sets are summarized in figure 2-14, where the pre-Stephanian (A), Stephanian (B) and post-Permian (C) joint pattern traces are depicted. It is noteworthy that the joint sets that were only found in the pre-Stephanian outcrops have a much more curved spatial distribution that mimics the structural trend of the CAA, while the joint sets found in Stephanian outcrops have a more open curvature in their spatial pattern, and those joint sets found in the Permian outcrops have no curvature at all in their spatial pattern.

2.1 Strike test

The strike test (also called an orocline test) (Schwartz and Van der Voo, 1983; Eldredge et al., 1985) evaluates the relationship between changes in regional structural trend (relative to a reference trend), and the orientations of a given geologic fabric element (e.g., cleavage, veins, fractures, lineations, paleomagnetic declination, etc.). This methodology has been mainly used by paleomagnetists (e.g., Schwartz and Van der Voo, 1983; Weil and Van der Voo, 2002) using paleomagnetic declinations, but has recently been adopted by structural geologists to test various kinematic models of orogenic curvature using strain and fracture data (Yonkee and Weil, 2010a), calcite twin data (Kollmeier et al., 2000), and anisotropy of magnetic susceptibility lineations (Weil and Yonkee, 2009). In this paper, the trend of joint sets is compared to the regional structural trend in order to test different kinematic models for CAA development.

Figure 1-5 shows simplified kinematic models for curved orogens using the orientation of systematic joint sets that are products of layer-parallel shortening fabrics. Model 1-5A depicts a primary arc with no correlation between joint orientation and structural trend, which results in a

strike tests with a slope of 0. Model 1-5B is also a primary arc with consistently oriented joint directions, however thrust slip is not uniform but radial, and thus the joint strike test has a slope of 1.0. Model 1-5C depicts a progressive arc with curved thrust slip, where joint orientations progressively rotate with structural trend resulting in a strike tests slope between 0 and 1.0 depending on the amount of curvature present when the joint sets develop. Model 1-5D depicts an orocline (secondary bending of an originally linear belt), which yields a strike test slope of 1. However, since a joint strike test can produce a slope of 1.0 for primary arcs with radial slip and secondary oroclines, the strike test can only be uniquely interpreted if other kinematic constraints are available (Yonkee and Weil, 2010b).

We have performed strike tests for each of the three joint sets: those found in (i) the pre-Stephanian, (ii) Stephanian and (iii) Permian outcrops. All strike tests were done using the refined weighted least-squares method of Yonkee and Weil (2010b).

Strike tests for all three Permian outcrop joint sets (Fig 2-17) give a slope of near 0.0 ($-0.03 \pm .08$, $0.09 \pm .08$, $0.00 \pm .08$), implying that the joint sets in these rocks have experienced no significant secondary rotation.

Figure 2-16 shows strike tests for the Stephanian outcrops that use the mean structural trend of Variscan structures below the Stephanian outcrop as a reference trend. Strike tests were made with the Stephanian outcrop strike sub-parallel (Fig. 2-16 A) and sub-normal joint sets (Fig. 2-16 B). Slopes of $0.72 \pm .18$ and $0.57 \pm .12$ are calculated for the two sets respectively.

The fold-axis parallel joint set (2-17A) and fold-axis normal joint set (2-17B) from the pre-Stephanian outcrops (Fig. 2-17) have strike test slopes close to 1.0 ($1.03 \pm .06$ and $1.16 \pm .10$ respectively). These results indicate a one-to-one correlation between deviations in structural trend and joint set orientation, and suggest that the joint sets pre-date any vertical-axis rotations and that the total deviation in trend of pre-Stephanian outcrop joint sets is about a third greater than that found in the Stephanian outcrop joint sets.

2.2 Discussion

Results from joint set analysis in the three unconformity bounded sedimentary sequences from the CAA reveal the existence of at least three different deformation episodes in which joints developed. These joint sets were generated according to the local or remote stress fields, and in a regional perspective is the result of far-field tectonic stresses (e.g., Gross et al., 1995; Eyal et al., 2001). When used together with other structures, like folds or faults, these joint sets can be extremely valuable in unraveling the geological stress-strain history of a region (e.g., Engelder and Geiser, 1980; Engelder and Gross, 1993). In general, joints develop parallel to the maximum principal stress (σ_1), which in contractional settings that have not been previously deformed, is roughly normal to the axis of the folds that accommodate shortening (Engelder and Geiser, 1980; Whitaker and Engelder, 2006). In the cases where joints develop in rocks that unconformably overlie previously deformed rocks, deformation is non-coaxially and the principle stresses are not necessarily parallel to the axes of the strain ellipsoid (e.g. Alonso, 1989). This relationship can result in joint orientations that are neither normal or parallel to the regional fold axis orientation.

The different orientations of joint sets found in the CAA help to unravel the timing and kinematic history of arc formation. The youngest joint sets generated in the studied region are recorded in the Permian outcrops. The north-south set, and likely the east-west set, are interpreted to be caused by bedding flexure during Alpine northward collision of Iberia with Europe in Cenozoic times (e.g., Álvaro et al., 1979, Alonso et al, 1996). This interpretation is

mainly based on the correlation of joint set orientations with the trend of major structures in post-Carboniferous rocks. The third joint set, which strikes $\sim 130^\circ$, is interpreted to be caused by the opening of the Bay of Biscay during Mesozoic times based on regional tectonic orientations (e.g., Gómez et al., 2002; Rosales et al., 2002). Because deposition of the Permian strata is known to have post-dated formation of the IAA (Weil et al., 2010), these joints have been removed from consideration in the analyses of older rock joint sets.

The Stephanian outcrops record two joint sets that are not found in post-Stephanian Permian outcrops (Fig 2-14). The longitudinal set has an arcuate pattern with lower overall curvature than the trends of the underlying structures (Fig. 2-14A). The orthogonal set shows a radial pattern, sub-perpendicular to the main underlying Variscan structural trend (Fig. 2-14B). Field relations suggest that the two sets usually abutt each other, which is interpreted to represent coeval formation (Fig. 2-6 and 2-7) and the two sets are likely both related to the same stress field as proposed by Caputo (1995; 2010) and Bai et al., (2002). Both sets are oblique to the trend of the local synform fold-axis in the limbs of the CAA and, as stated before, their orientation implies that they did not form in response to the passive bending that deformed the basins (Alonso, 1989), but rather from the regional stress field.

The pre-Stephanian (Neoproterozoic and Paleozoic) outcrops record a complex set of joint sets that include all the Stephanian and younger joint sets as well as at least two older sets that are parallel and perpendicular to the main Variscan structural trend. One of the sets is parallel to the main Variscan structural grain (e.g., fold axis and thrust trends), which mimics the trace of the CAA, and the other set is perpendicular to the arc (Fig. 2-4; 2-14).

To explain the temporal and spatial distribution of joint sets in the region we have assigned each unconformity-bound joint set to stress field responsible for their generation. The youngest sets, which are only found in Permian and younger outcrops, are attributed to the stress field that caused the Cenozoic Alpine deformation and/or the opening of the Biscay Bay. The appropriate tectonic stress field(s) responsible for the joint sets present in the Stephanian and pre-Stephanian outcrops is less straight forward to assign. Given their strong correlation with the arcuate trace of the CAA it is difficult to imagine a process that could have formed in situ joint sets with a primary dispersion of 180° for the pre-Stephanian outcrop sets (strike tests slopes of near unity), and joint sets with between 90° and 125° of primary dispersion for Stephanian outcrops (based on strike test slopes of between 0.5 and 0.7). Given the existing paleomagnetic data that indicate large-scale rotation of Variscan structures during Stephanian and younger times (e.g., Van der Voo et al., 1997; Weil et al., 2000, 2001), it is more conceivable that the joint sets were formed with a regionally linear north-south trend (in present-day coordinates) and were subsequently rotated to their present orientation. Consequently, the pre-Stephanian and Stephanian outcrops record joint sets formed prior to, and penecontemporaneous with, oroclinal bending. Thus, the present orientation of joint sets in pre-Stephanian outcrops formed by ca. 180° of vertical-axis rotation of an approximately linear joint set that was parallel to early longitudinal fold axes; while the Stephanian outcrop joint sets show a rotation of ca. 100° , undergoing between 50-30% of the total oroclinal rotation.

The simplest tectonic scenario that explains these observations indicates that two linear sets of joints formed coevally with the main Variscan shortening in the western part of the CZ during the uppermost Mississippian-early Pennsylvanian. Subsequently, these sets were rotated ca. 90° around vertical-axes prior to deposition of Stephanian sediments. Finally, the arc was tightened another ca. 90° to its present-day curvature between the Stephanian and earliest Permian.

The joint data analyzed herein is interpreted in light of the previously proposed CAA

oroclinal bending model (e.g. Stewart, 1995; Weil et al., 2001, 2010; Weil, 2006; Gutiérrez-Alonso et al., 2004, 2008). This model requires an initial E-W (in present-day coordinates) compression event that produces a near-linear orogen. This compression is followed by a sudden change to N-S shortening (in present-day coordinates) that rotates the limbs of the orogen and is recorded in the latest thrusts of the Cantabrian Zone (Merino-Tomé et al., 2009). This model suggested a brief period of time (around 15 Ma) for oroclinal bending, from the latest Carboniferous to the earliest Permian. Time constraints are based on assigned magnetization ages for rocks sampled in the CZ, importantly the post-arc-parallel folding but pre-orocline paleomagnetic B component that was interpreted as late Carboniferous to early Permian in age (Van der Voo et al., 1997; Weil et al., 2000; 2001). This magnetization has been reinterpreted as Kasimovian in age based on estimated timing of arc-parallel folding inferred from syntectonic sediments. Upper age bounds are given by the eP magnetization found in Permian basins from the northern and southern arms of the larger Ibero-Armorican Arc, thus constraining oroclinal bending to a 10 Ma time interval (see Weil et al., 2010). The relative age progression of joint set formation in the CAA, as constrained by the ages of the bounding unconformities, agrees well with the existing paleomagnetic constraints for closure of the CAA.

According to the joint set strike tests, during Stephanian B-C times the CAA was closed between 50% and 70%, and by lower-most Permian times was completely closed. Assuming a constant bending rate, about 100° of bending took 5 Ma from Stephanian B-C (upper-most Kasimovian ~304 Ma) to the Carboniferous-Permian limit (299 Ma). The remaining curvature of the CAA had to be produced before the generation of the joints in the Stephanian rocks and, if the bending rate was similar to that of Stephanian times, it is likely that the CAA started bending during the Moscovian (around 310 Ma). This chronology reinforces the interpretation of the rapid tectonic lithospheric delamination event proposed by Gutiérrez-Alonso et al., (2004, 2011).

To better illustrate the kinematics model of fracture set evolution, we present an animation (video 1 which can be downloaded from the Data Repository (Anexo D2) with higher resolution; summarized in Figure 2-18). Figure 2-18 A represents the pre-Moscovian to Moscovian pre-oroclinal bending times with the fold-axis parallel and normal joint sets recorded in the pre-Stephanian outcrops. Figure 2-18B represents the lower Kasimovian times with the Leon breaching thrust already formed (Alonso et al., 2009) and approximately 20% bending. Figure 2-18 C shows the initial emplacement of the Picos de Europa and Cuera units (Merino-Tomé et al., 2009), deposition of the Stephanian B-C sediments, development of fold-axis sub-parallel and sub-perpendicular joint sets recorded in Stephanian outcrops, and 50-70% bending. Figure 2-18 D represents the final present-day stage with the addition of post-Permian aged joints imposed on the entire CAA.

The study of systematic joint sets in rock sequences bounded by unconformities, allows for potentially robust timing constraints on joint formation, and can provide constraints on any changes in the regional stress field. Such constraints have the capability of helping unravel the kinematics of regions where other structural criteria are unavailable.

3 Fold analysis

Geometrically, a conical fold is characterized by the trend and plunge of its axis and by the angle between the generatrix of the conical surface and the fold axis, also known as semiapical angle ($\alpha/2$) (Wilson, 1967; Pueyo et al., 2003). Perfect cylindrical folds, which are considered as a special case of a conical fold; have $\alpha/2$ equals 0° . Identification and analysis of conical folds in nature are conducted using stereographic projection of geologic surfaces, typically bedding (π -diagrams) which, when truly representative of a conical surface, scatter along a small circle on the stereonet. Given the nature of geologic orientation data, mathematical methods have been developed to fit measurements to a small circle and to quantify the suitability of the calculated fit (e.g. Ramsay, 1967; Fisher et al. 1987).

Approaches to fitting planar data to a cone typically involve least squares minimisation of a function involving the direction cosines of poles to planes (Ramsay, 1967; Venkitasubramanian, 1971; Cruden and Charlesworth, 1972) and provide estimates for the orientation of the cone axis and the semi-apical angle. Problems associated with these initial methods were resolved by minimising the squares of the actual angular deviations (Kelker and Langenberg, 1982; Fisher et al., 1987) making the minimisation problem non-linear and requiring iterative techniques to determine a solution. The problem may also be solved using the least eigenvector of the orientation matrix (Nidd and Ambrose 1971; Fisher et al., 1987, p. 33) though this approach only works for symmetrical data sets with a semi-apical angle less than 45 degrees. Bingham's distribution on a sphere can be used to find the best-fit great circle to fold data forming a pair of, which is often the case for geological data (Kelker and Langenberg, 1976). Subsequently, using a transformation to spherical coordinates (Stockmal and Spang, 1982), a least-squares best fit was identified for the simulated data of Cruden and Charlesworth (1972). Methods able to cope with elliptical conical folds and statistical tests for distinguishing between circular and elliptical data have also been developed (Kelker and Langenberg, 1987, 1988). Mainstream statisticians have also shown an interest in this problem (Mardia and Gadsden, 1977; Rivest, 1999). Data presented in this paper have been fit using an implementation of the iterative algorithm presented by Fisher et al. (1987) p 140-143 which is based on the method of Mardia and Gadsden (1977) and the improved least-squares algorithm of Gray et al (1980). This method is robust to non-symmetrical data and has been proven to provide accurate solutions to different cases of real and simulated conical folds.

Data consisting of 578 strike and dip measurements were collected from bedding surfaces of Cambrian Limestones and Ordovician quartzites in the CZ and the eastern limit of the WALZ for the geometric analysis of the map-scale folds because these units contain very few smaller folds that would yield data that could obscure large-scale geometries. To compare the data with folds generated in rocks that were not deformed prior to the orocline formation, data were collected in the post-orogenic Stephanian outcrops (Corrales, 1971; Colmenero et al., 2008) around the boundary between the WALZ and the CZ. To obtain the best conical fit folds that have overturned limbs were projected in the lower hemisphere together with the data from the normal limbs .

In the CZ, data were collected in the Somiedo Unit (Fig. 3-1), comprising two main thrust units, the Tameza thrust unit to the east and the Belmonte thrust unit to the west (Fig. 3-2). Within the latter unit, the data were subdivided into three subsets from areas with different initial orientations of the reference surfaces (bedding and thrust surfaces) and thereby resulting in different interference responses. The three areas are the Tameza thrust unit (Fig. 3-2) and the eastern and western parts of the Belmonte unit (Figs. 3-3 and 3-4 respectively; Fig. 3-5). The other two datasets are located in the easternmost part of the WALZ (Fig. 3-6 A and B) and the

unconformable Stephanian outcrops (Fig. 6 B and C).

3.1 Conical folds in the Somiedo unit

The Somiedo unit contains two sets of thrusts : the first generation with a detachment level in the Cambrian limestone, roots to the west into the Neoproterozoic rocks of the NA, and large displacements of tens of km; and the second generation, out-of-sequence thrusts with much smaller displacements (Bastida, et al., 1986; Heredia, 1984; Bastida and Gutiérrez-Alonso, 1989) and formed during second-stage thrusting in the Cantabrian Zone (Pérez-Estaún et al., 1991) or orocline formation (Alonso et al., 2009). The Tameza and Belmonte thrust units (Fig. 3-2B and 3-3B) were passively folded into a large synform named the Los Lagos synform, during development of the antiformal stack to the west, in the NA.

Two different fold sets are recognized in the Somiedo unit (Julivert and Marcos, 1973). The thrust-parallel set, or longitudinal set, contains the classical thin-skinned foreland fold-and-thrust structures (e.g. Dahlstrom, 1969). These folds are cylindrical and have horizontal or sub-horizontal axes, for example, the Viyazón-Reigada syncline (Fig. 3-7). The second radial set deformed all the previous structures, strongly modifying the geometry of the Somiedo unit and, in general, the entire foreland fold-and-thrust belt, producing an asymmetric structural pattern in both flanks of the unit. Six major radial folds with different amount of shortening are recognized in the Somiedo unit (Fig. 3-8; Julivert and Marcos, 1973, Weil et al., 2000). All these radial folds are localized in the eastern flank of the Somiedo unit, but only the two southern folds crosscut the axial trace of the longitudinal Los Lagos synform (Fig. 3-8). On the other hand, in the western flank of the Somiedo Unit, in the western limb of the Los Lagos synform, while the rocks are refolded by the host fold to the six radial folds, they do not deform the limb. (Figs. 3-4, 3-5, 3-6, 3-8).

The E-W-trending fold produces a sharper bend in the cartographic trace of the thrust surfaces even though the folds does not crosscut the axial plane of the Los Lagos synform (Fig. 3-2, 3-3 and 3-8). In the Tameza thrust unit, the radial fold produces a conical shape with a semiapical angle of 30° and axis plunging 70° towards the west (Fig. 3-2A). In the eastern flank of the Los Lagos synform, which base is the Belmonte thrust, is also conical (Fig. 3-3A1) with a similar 30° semiapical angle, but with an axis plunging 80° towards a more northwest direction (308°). In the latter case, the cone shape is more complex and the fit, is better understood when the data are divided into data collected far from the cone apex (Fig. 3-3A-2) and data collected near the cone apex (Fig 3-3A-3). Data near the cone apex exhibits a slightly shallower plunging cone axis, 66° towards the west with a semiapical angle of 30° (Fig. 3-3A-3), as compared to data measured away from the cone apex (Fig. 3-3A-2) with a cone axis plunging 71° towards the north and a very different semiapical angle of 61° .

The presence of lateral ramps, for example the termination of the Tameza sheet under the Somiedo sheet (Fig. 3-8), complicates the ideal cylindrical geometry (Gutiérrez-Alonso, 1992), some of the conical folds have complex forms, including slightly overturned or subvertical flanks where shortening is large, due to fold amplification. Equivalent examples of amplified folds coincident with lateral ramps have been described immediately north of the studied region (Bastida and Castro, 1988) and also to the east in the Ponga region (Alvarez Marrón and Pérez Estaún, 1988; Weil, 2006).

In the western flank of the Somiedo Unit, which corresponds with the western limb of the Los Lagos synform, the radial fold set is represented by a single fold that has much a larger wavelength and can be traced through the whole Cantabrian Arc close to its axial trace. The

shape of this fold is also conical with a semiapical angle close to the values obtained in the eastern flank (31°) and similar axial plunge (75°) but in this case, towards the northeast (Fig. 3-4). We interpret that this difference is principally due to the position of this flank in the outer zone with respect to the rotation axis of the Cantabrian Arc (Figs. 3-3C and 3-4) and also to an initially different bedding attitude, dipping towards the east.

The interference between the longitudinal and radial folds in the Somiedo unit define a type 2 interference pattern of Ramsay (1967), or the third and fourth mode according to Ghosh et al., (Ghosh et al., 1992, 1993; Simón 2004). Given that the axes of the radial set are sub-vertical, the interference patterns draw “worms” instead of “mushrooms” (Julivert and Marcos, 1973).

The overall refold geometry of the Somiedo unit is shown in Figure 3-5 and the 3D interactive model PDF file in the data repository (Anexo D3). This diagram was constructed from serial crossed sections based on the data collected in this study and previous work (Bastida et al., 1984; Bastida and Gutiérrez-Alonso, 1989; Gutiérrez-Alonso, 1992) using the PETREL software.

We interpret these folds to be the result of a vertical axis rotation during the orocline buckling process that formed the Cantabrian Arc. Hence, the sub-horizontal non-outcropping basal detachment of the Somiedo Thrust and all the west-dipping-related imbricated thrust sheets folded conically with different axes and semiapical angles (Figs. 3-2C and 3-3C).

3.2 Geometry of eastern WALZ

West of the NA, in the WALZ, the lower Paleozoic sedimentary rocks can be traced almost continuously depicting the curvature of the Cantabrian Arc for more than 200 km (Fig. 3-6B). We have analyzed the bedding attitude around the bend and the results show great similarity with those obtained for the western flank of the Somiedo unit. As can be observed in the stereonet of Figure 3-6A, the data fit a conical geometry with a semiapical angle of 32° and an axis plunging 75° towards the northwest due to its initial orientation. We interpret that the origin of this conical fold is identical to that of the western flank of the Somiedo unit except that its oppositely dipping initial orientation resulted in the different plunge for the conical fold axis.

3.3 Geometry of the Stephanian basins

To compare the fold geometry in the pre-orocline rocks to the fold geometry to rocks deposited during oroclinal formation, the Stephanian continental deposits were also investigated. The synforms have eastern-shallow-dipping limbs towards the core of the arc, and a western, steep to overturned, limbs away from the core.

Both flanks of the Stephanian outcrops were plotted separately and describe two different conical folds with subvertical axes, ca. 85° (Fig. 3-6C). The eastern flank describes a large semiapical angle (66°) (Fig 3-6C-1), whereas the western flank shows a cone (Fig. 3-6C-2) with an apical angle very similar to the outer section of the Somiedo unit (i.e. 30° , Fig 3-4) and the eastern WALZ (Alonso 1989).

3.4 Discussion and interpretation

The fold interference patterns of the Cantabrian Arc help to understand the mechanisms of folding in the uppermost crust during an orocline buckling process. Conical folds deform rocks such that lines initially parallel to the future cone axis, will no longer be parallel to the cone axis

as the fold develops. Consider a situation where a layer is deforming under the action of a rotation about some axis, then a conical fold develops when the rotation axis is not parallel to the layer. The semiapical angle for the fold increases as the angle between rotation axis and the pole to the layer decreases and reaching a maximum value when the rotation axis and pole are parallel (Fig. 3-9 and 3-10). On the other hand, if the rotation axis is normal to the pole to the layer, then a cylindrical fold develops.

When a horizontal or sub-horizontal layer is folded by a rotation axis at low angles to its pole together with tangential longitudinal strain mechanism (Ries and Shackleton, 1976), two different structural regimes appear. Those portions of the reference surface located in the inner arc of the fold become shortened and develop radial conical folds with sub-horizontal axes plunging towards the rotation axis. In contrast, the portions of the reference surface in the outer arc undergo stretching, which when an initial slope is present yields a large wavelength fold with a conical geometry (Ramsay, 1967 pp.490-517). The difference in behaviour of the outer arc as a function of the absence or presence of a dipping surface is illustrated by comparing Figure 3-9 without the dip and Figure 3-10 with the dip and the large wavelength fold.

In the Cantabrian Arc, the initial geometry is complicated by the existing Variscan fold and thrust belts that created a variety of surface geometries for inclusion in the subsequent conical folding. In the eastern portion of the Somiedo Unit, located in the inner arc, conical folds developed with axes plunging towards the rotation axis. In contrast, the western portion developed a large wavelength, vertical-axis cone (Figs. 3-5, 3-8, 3-10 and 3-11). Some conical folds may have nucleated at previous thrust-related lateral ramps (Bastida and Castro, 1988, Gutiérrez-Alonso, 1992), which provided a favourable location for the radial-fold development as is interpreted for easternmost thrust units within the Cantabrian Zone (Álvarez-Marrón and Pérez-Estaún, 1998; Weil, 2006).

The Cantabrian Arc is interpreted to have been produced by a vertical or near vertical-axis rotation of the Western Europe Variscan belt (Weil et al., 2000; 2001; 2010) and longitudinal tangential strain has been proposed as the main mechanism of deformation (Gutiérrez-Alonso et al., 2004; 2008; Pastor-Galán et al., 2012b), implying that every geologic surface with the exception of the initially vertical surfaces would be conically folded in both inner arc and outer arc.

Figure 3-11 and video (Anexo D2) depict a simplified reconstruction of the geometry of the Cantabrian Arc in the Somiedo Unit. Figure 3-11A shows the geometry of the basal thrust of the Somiedo Unit in the Cantabrian zone, which can be compared with any other surface that was approximately horizontal prior to the orocline development, such as the Stephanian outcrops. The geometry is similar to that obtained folding a horizontal surface around a vertical rotation axis (Fig. 3-9). Figure 3-11B shows the morphology of Los Lagos Synform including the southernmost section reinterpreted from Julivert and Marcos (1973), which is similar to folding a cylindrical fold about a vertical rotation axis (Fig. 3-10). Figure 3-11C shows the overall structure including the easternmost Tameza thrust.

The distribution of radial conical folds, which only occur to the east of the Los Lagos syncline axial trace, as compared to the unique large-wavelength change in bedding strike west of this axial trace (Figs. 3-2 and 3-8) indicate that the local rotation axis for the the Cantabrian Arc is likely located near the Los Lagos axial trace as in the paper models of figures 3-9 and 3-10. The axial trace of Los Lagos syncline acted as the local finite neutral line (Frehner, 2011) in the study area and localizes the change from, outer-arc extension to inner-arc shortening, observed in both limbs of the primary syncline. The deformation distribution is in agreement with a tangential longitudinal strain mechanisms for the development of the studied sector of the Cantabrian Arc (Gutiérrez-Alonso, 2004). In addition, the radial fold set becomes less evident towards the flanks

of the Cantabrian Arc, which are only well preserved at its hinge. Nevertheless, in the hinge to the east of the studied region, the orocline related folds exhibit a less conical geometry (Julivert and Marcos, 1973; Aller and Gallastegui 1994) and display a constant E-W trend instead of a radial divergent disposition due to the larger shortening in the inner arc of the Cantabrian Arc. These data also support the idea of a local upper-crust orocline axis situated in the neighbourhood of the Los Lagos axial plane (Fig. 3-8).

4 Analogue modeling

Analogue modeling is a powerful technique that can be used to study geological processes in 3D (e.g. Ghosh et al., 1995; Zulauf et al., 2003) if the models are properly scaled (Hubbert, 1937, Weijermars and Schmeling, 1986). These models can be used to study tectonic processes from micro- to lithospheric-scale (e.g. Autin et al., 2010; Fernández-Lozano et al., 2011).

Analogue modeling has been previously performed to understand mechanisms leading to curved orogens in several possible scenarios. One of the most common models invoked to produce curved orogens is the collision of a rigid indenter (microplate) with a continental margin, the latter showing a linear structural grain. This scenario has been frequently modeled using simple rock analogues (e.g. Tapponier et al., 1982; Marshak, 1988; Davy and Cobbold, 1988; Zweifler, 1998; Keep, 2000). Other mechanisms propose orogenic curvature to be the result of non-parallel thin-skinned transport at uppermost crustal levels. The approaches used to support such a scenario include: (1) the modeling of the curvature of the Cantabrian Zone in NW Iberia (Julivert and Arboleya, 1984, 1986); (2) the tracking of thickness of the different layers in a foreland fold-and-thrust belt and its relationship with the initiation of thrusts and the possibility of curved resultant geometries (Marshak and Wilkerson, 1992); and (3) the relationships between curved fold-and-thrust belts and previous topography (Marques et al., 2002). However, to the best of our knowledge, there have been no attempts to model lithospheric-scale oroclines developed from an initially linear orogenic architecture.

There are different rock analogues that can be used to model deformation at crustal and lithospheric scales. Granular materials, such as sand or glass beads, are effective in modeling the upper crust, where the strength of quartzofeldspathic rocks is largely controlled by confining pressure (eg. Malavielle, 2010). Viscous rock analogues, on the other hand, are used to model deformation at deeper structural levels, where the strength is controlled by temperature and strain rate. Plasticine can be used to model power-law creep of rocks undergoing dislocation creep (Zulauf and Zulauf, 2004).

4.1 Model set up

The orocline development was modeled as a thick-skinned buckling scenario in order to understand the kinematic and dynamic evolution of the mantle lithosphere and the resultant morphologies during and after orocline formation, and to test if these results are consistent with the available paleomagnetic, structural, petrological and geochronological data and interpretations.

Scaling and construction of the orocline buckling models follow the methodology of Davy and Cobbold (1991) and Cobbold and Jackson (1992) for lithospheric thermomechanical modeling. The experiments were scaled to the natural boundary conditions by selecting proper dimensions and types of analogue materials in order to establish similarity in geometry, kinematics and dynamics. Dynamic and geometric scaling between the model and nature can be achieved when respecting the stress-scale factor:

$$\sigma^* = P^*G^*L^*$$

Where σ refers to stress, P to density, G to gravitational acceleration and L to the length scale. The asterisk refers to the ratio between model and nature.

The analogue materials selected for the experiments are plasticine and sand. In general terms, plasticine is a non linear, strain-rate softening material consisting of a weak organic matrix

and mineral fillers (eg McClay, 1976; Weijermars, 1986; Schöpfer and Zulauf, 2002; Zulauf and Zulauf, 2004). Plasticine at constant temperature shows non-Newtonian creep defined by the flow law:

$$\dot{E}=C \sigma^n ,$$

where \dot{E} is the strain rate, C is a material constant, and n is the stress exponent (McClay, 1976). Plasticine is well recognized as a good analogue to simulate rocks undergoing dislocation creep (Zulauf and Zulauf, 2004), which is as the dominant deformation mechanism in the lower crust and upper mantle (e.g. Carter and Tsenn, 1987; Hirth et al, 2001; Eaton et al, 2009). Dry sand with rounded grains of quartz and feldspar, on the other hand, was used to model the upper crust. Because of low cohesion and internal friction angle, according with the Navier-Coulomb and Byerlee law ($\tau = C + \mu\sigma_n$, where τ is the shear stress at failure, C is the cohesive strength of the material, μ is the coefficient of internal friction and σ_n is the normal stress on the plane at failure) granular materials such as sand are considered as suitable candidates to model the brittle upper crust (Mandl et al., 1977; McClay and Ellis, 1987; Davy and Cobbold, 1991; Rossi and Storti, 2003). Properties are summarized in tables (see tables in spanish volume).

Two different kinds of plasticine were used for the experiments (Fig. 4-1 and 4-4): (1) *Beck's orange*, manufactured by Beck's Plastilin, Gomaringen, Germany, and (2) *Weible red*, made by Weible KG, Schorndorf, Germany. The rheological properties of the selected plasticines (see tables in spanish volume, Fig. 4-2) were studied and well constrained in previous studies (Zulauf and Zulauf, 2004; Tkalcec, 2010) and additional insights are provided by new data from this study. We used a thermal gradient in order to produce the most appropriate power-law conditions for modeling thick-skinned processes. At the temperatures used for the experiments, the stress exponent of the rock analogues is comparable with the stress exponent assumed for lower crust, mantle lithosphere and asthenosphere which ranges from 2 to 5 (Carter and Tsenn, 1987; Karato and Wu, 1993; Freed et al, 2006).

The asthenosphere was modeled with *Beck's orange* plasticine, which flows under nearly steady state under experimental strain rates (Zulauf and Zulauf, 2004). Assuming an average viscosity $\eta_{eff} = 5.15 \cdot 10^{19}$ Pa·s for the asthenosphere (Rydelek and Sacks, 1988; Morency and Doin, 2004), the model is defined with a scale ratio of $M^* = (\eta_{eff})_m / (\eta_{eff})_n = 1.146 \cdot 10^{-17}$ for *Beck's orange* at experimental average temperature, $T = 60^\circ\text{C}$ (where subscripts m and n refer to laboratory model and nature scales, respectively). Density of this plasticine $\rho = 1250$ kg/m³ at $T = 20^\circ\text{C}$ and assuming a density for natural asthenosphere $\rho = 3100$ kg/m³ (Pysklywec and Cruden, 2004) the scale ratio between both densities is $P^* = \rho_m / \rho_n = 0.4$.

Weible red plasticine was used to model the mantle lithosphere. Assuming an average viscosity $\eta_{eff} = 5 \cdot 10^{21}$ Pa·s for the natural mantle lithosphere (e.g. Walcott, 1970; Morency and Doin, 2004; Shi and Cao, 2007; Johnson et al., 2007; Fernández-Lozano et al., 2010), the scale ratio with *Weible red* at the average temperature ($T = 45^\circ\text{C}$) at the experiments run, is $M^* = (\eta_{eff})_m / (\eta_{eff})_n = 1.146 \cdot 10^{-17}$. Considering the density of this plasticine ($\rho = 1400$ kg/m³), its scaling ratio with the natural mantle lithosphere ($\rho = 3360$ kg/m³) is $P^* = \rho_m / \rho_n = 0.41$. The rheological and density data of the analogue materials are listed in Tables (see tables in spanish volume).

Sand and *Beck's orange* plasticine were used for upper and lower crust, respectively. Beck's orange was selected to simulate a more ductile and less dense layer than the material used for the mantle lithosphere, and sand was selected to simulate the brittle upper crust.

During the orocline buckling experiments, the temperature varied between 55°C and 65°C in the analogue asthenosphere, between 45° and 55°C in the analogue mantle lithosphere, between 40°C and 45°C in the lower crust layer, and between 35°C and 40°C in the upper crust

analogue. The thermal diffusivity of plasticine ranges between 0.65 and 0.8 W m⁻¹ K⁻¹ (Touloukian et al, 1970). The activation energy has been determined at 323 ±34 kJ/mol for *Becks orange* plasticine and values between 400 and 500 kJ/mol for other plasticines (Zualuf and Zulauf, 2004). These values are in agreement with the proposed values for modeling the lithosphere by Davy and Cobbold (1991) and Cobbold and Jackson (1992).

4.1.1 Orocline Buckling Experiment

Fourteen elongated layered models (Ob-1 - Ob-14), with varying strain rates, different configurations for the mantle lithosphere, and an approximately constant temperature profile (Fig. 4-2B) were deformed using a thermomechanical apparatus situated in the Geozentrum at Frankfurt University (Fig. 4-3B).

The apparatus used is capable of simulating tectonic processes at different spatial and temporal scales applying stresses and adding a thermal gradient to the rock analogues. The apparatus is based on the design of a three-dimensional coaxial deformation apparatus that works at room temperature (Zulauf et al., 2003). Figure 4-3A depicts a plan-view of the machine, which is comprised of four aluminum plates set up orthogonally (1, 2, 3 and 4, in Fig. 4-3A) built on an aluminum table with a copper plate inlay (5 in Fig. 4-3A). Two independent motors, connected to the plates by worm gears, achieve the movement of plates 1 and 2 in two orthogonal directions, respectively. One of the motors (motor 1) is fixed to the aluminum table plate (7 in Fig. 4-3A). Motor 2 is built on a rail and moves passively along the rail when motor 1 is working (8 in Fig. 4-3A). Plate 3 moves passively when the mobile motor is active, and plate 4 (in Fig. 4-3A) is attached to the fixed working bench. The velocity of the plates is adjustable and ranges from 1 to 800 mm/h. The apparatus includes an oven that is capable to heat the basal copper plate from room temperature to 100° C.

The model's length was limited by the dimension of the apparatus, and was the main constraint for designing the laboratory orocline buckling experiment (Fig. 4-3A). The length scale of the models was set at $l_m = 30$ cm; the length of the Ibero Armorican Arc around its neutral fiber is $l_n \approx 1300$ km, which gives a length scale ratio $L^* = l_m/l_n = 2.3 \cdot 10^{-7}$. Based on this scale and the approximate width of the IAA, that is $W_n \approx 350 - 400$ km, the model's width chosen was $W_m = 8 - 9$ cm. Because there are no constraints on the initial thickness of the mantle lithosphere, three different thicknesses of the model mantle lithosphere ($Ml_m = 1.0, 1.5, 2.0$ cm) were tested in order to analyze their impact on the style of oroclinal buckling and using the same scaling factor represent a mantle lithospheric thicknesses of 40, 60 and 80 km. The crust was kept of the same thickness design in every model consisting of an upper crust, $Uc_m = 0.5$ cm, and a lower crust, $Lc_m = 0.5$ cm, which together represent a slightly thickened continental crust after orogenic shortening (about 40 km in nature). We run our experiments without any extra gravity so the gravity scale ratio is $G^* = G_m/G_n = 1$. Thus, the calculated stress ratio is $\sigma^* = P^*L^*G^* \approx 1 \cdot 10^{-7}$, the time scale ratio for the experiments can now be defined as $T^* = M^*/P^*L^*G^* = t_m/t_n \approx 1,2 \cdot 10^{-10}$ a strain rate ratio given by $\dot{E}^* = \sigma^*/M = 8,7 \cdot 10^9$ and a velocity ratio given by $V^* = \dot{E}^*L^* \approx 2 \cdot 10^3$.

We suggest a strain rate for the IAA of ca. $2 \cdot 10^{-15}$ s⁻¹ based on the N-S shortening (~900km) and the time taken for this process (ca. 10 m.y. following the Weil et al. (2010) and Pastor-Galán et al. (2011) hypothesis). The velocity of the plates was set between 1.5 to 3.5 cm/h, which considering a shortening of ca. 15 cm of the initial length of models, made experiments 5h and 10h induration. The strain rate used to run the buckling experiments varied from $1 \cdot 10^{-5}$ to $5 \cdot 10^{-5}$ s⁻¹. These parameters make a real scale of time, T of between $5 \cdot 10^{-11}$ and $1 \cdot 10^{-10}$; velocity, V of between $1.46 \cdot 10^3$ and $3.4 \cdot 10^3$ and strain rate \dot{E} of between $5 \cdot 10^9$ and $2.5 \cdot 10^{10}$ which is consistent

with the scaling parameters.

The copper plate at the base of the apparatus is the lower boundary and acts a rigid body and limits the downwards developing of the model. This copper plate was spread with a vaseline which is immiscible with *Beck's orange* plasticine to minimize slip of the branches of the experiments during the buckling process. The upper boundary is a free surface (Fig. 4-3C). All models were coated in a 1 centimeter wide wall of *Weible red* plasticine around the *Beck's orange* bottom layer (experimental asthenosphere) in order to avoid the lateral escape of the *Beck's orange* plasticine, which behaves in a very ductile manner at the temperature used in the experiments (Fig.4-3C). This wall is kinematically independent from the behavior of the *Weible red* layer representing the mantle lithosphere and *Beck's orange* and sand layers representing the crust. The wall was removed from the model before computer tomography was applied. The wall acts as a side boundary for the asthenosphere layer, whereas the lithosphere and crust side boundaries are free. During the experiment, the lower part of these walls melted and spread because *Weible red* plasticine can blend with the vaseline, dramatically changing its viscosity. Although the spreading plasticine sometimes touched the side plates (Fig. 4-3C3), their contact had no influence on the development of the deformation of the layers. Most of the models (except Ob6 and Ob7) were run with small wedges at their ends, which triggered the model to bend around a vertical axis and not to develop a train of horizontal folds (Fig. 4-3C).

Strength and viscosity profiles for the model lithosphere compared with a three layered lithosphere Earth strength profile (Davy and Cobbold, 1992) are shown in Figure 4-2D. Ductile strength values are computed for effective viscosities assuming a strain rate of 10^{-5} s⁻¹, which is a reasonable approximation of deformation rates observed in the experiments. The strength profile for the sand crust was calculated following Navier-Coulomb and Byerlee law. Likewise, the effective strength of the ductile crust, mantle lithosphere and asthenosphere will vary in the experiments because of variations in the strain rate and the thermal evolution of the plasticines.

The progress in deformation was monitored by digital plan-view photography (Fig. 4-2D). Digital photographs were taken every minute resulting in 24 frames per second videos at a scale 1 minute/ frame (see data repository, Anexo D4). The temperature was recorded by an electronic logger consisting of four sensors situated in each layer at one side of the model.

4.1.2 Computer tomography

Geometrical data of the deformed models can be obtained either by slicing them mechanically and reconstructing the 3D interior geometry of the different layers, in which case the model is destroyed. Alternatively, the models can be analyzed using computed tomography (CT), which is a non-destructive imaging technique that is sensitive to the different densities of the materials used (Zulauf et al., 2003).

CT is capable of recording high resolution 3D information in a series of 2D slices (e.g. Colleta et al. 1991) and is extremely appropriate for imaging the models produced with plasticine of different densities (Zulauf et al., 2003). The final models were examined using a multislice spiral CT scanner (Phillips CT brilliance 6 slice). The resulting 3D images were used to study the different mantle lithosphere morphologies acquired after orocline buckling. Only one model was sliced after applying a CT scan (Fig. 4-5Ob1). The CT images were represented in virtual reality .wrl files using the software SMOOTH developed at Geozentrum of Frankfurt University. Those files have been transformed into an enhanced pdf, which is more accessible to regular users and are stored in the data repository (Anexo D4).

4.2 Results

To a first order, the results obtained from all 14 experiments show consistent results and are summarized in Tables (see tables in spanish volume). Three different varieties of results, depending on the initial lithospheric thickness in the models, were sorted out and summarized in Fig. 4-5, and are described in the following paragraphs.

After the model had undergone shortening, in each experiment the initially rectilinear models that simulate a linear orogen, buckled around a vertical axis acquiring an arcuate or horseshoe shape, including those experiments that were not equipped with wedges at their edges. In geological terms, regardless of differences in thickness of the layers or in the strain rate, the model mantle lithosphere is shortened in the core of the orocline and is extend in the outer arc.

In all the experiments, shortening of the inner arc was accompanied by conical folding and thickening of the mantle lithosphere, which acted as a lithospheric root of the modeled orocline. Lithospheric root development was accompanied by thrusting and duplication of the lower crust (Fig. 4-5). Every experiment generated a root at least two times, and typically more than 2.5 times thicker than the original mantle lithosphere. A certain degree of vertical growth over the generated root was recognized in all models. This vertical growth varies from 0.5 to 1.0 cm. On the other hand, the extension in the outer arc was produced through almost radial tensile fractures (hereafter radial fractures) and so the thickness of the mantle lithosphere in the outer arc was unchanged.

Another feature, present in every experiment, is the growth of tensile fractures parallel to the shortening direction that crosscut the horseshoe shape of the model. These features were developed mainly in the uppermost part of the mantle lithosphere layer, although they commonly affect the entire layer (hereafter they will be termed parallel fractures). Furthermore, most of the experiments show more or less subvertical shear zones with angles between 30° and 40° with respect to the shortening direction that crosscut the trace of the arc. In three of the experiments (Ob4, Ob9 and Ob14) a shear zone affecting the entire thickness of the model formed. In each case, shear zone displaced the growing root allowing the generation of a new root where the shear zone formerly was located or where its presence decelerated the development of the root.

The resulting morphologies in the experiments are essentially dependent on the initial thicknesses of the model mantle lithosphere. There is no significant impact of the strain rate or the temperature profiles on the geometry of the deformed sample. Nevertheless, variations of these parameters in the experiments were minor in order to preserve the scaling.

Three experiments were run with an initial model mantle lithospheric thickness $MI_m = 1\text{cm}$ (40 km). The results obtained from these experiments are represented by experiment Ob12 (Fig. 4-6). Models with this thickness developed an almost recumbent, isoclinal conical fold as a lithospheric root in the inner arc (Fig. 4-Ob3, Fig. 4-6C), some radial fractures in the outer arc and a great number of parallel fractures over the entire layer (Fig. 4-6B). Although in model Ob12, no oblique shear zones were identified, a small incipient shear zone developed in experiment Ob3 and longer one in model Ob4 (see data repository, Anexo D4).

Five experiments with an initial model mantle lithosphere thickness $MI_m = 1.5\text{ cm}$ (60 km) also developed lithospheric roots in the inner arc as a tight and overturned conical fold. In each experiment the mantle lithosphere in the model was extended in the outer arc by means of radial fractures. Parallel fractures commonly occurred over the entire layer, but they are not as abundant as in the experiments with 1 cm of model mantle lithospheric thickness. Three of the

five experiments produced short (<1cm) shear zones oblique to the compression direction in one of the orocline limbs. Figure 4-6 shows the most representative experiment of this group (for further images see data repository, Anexo D4).

Finally, the 6 experiments with an initial model mantle lithosphere thickness $ML_m = 2$ cm acquired three analogous root morphologies: (1) a very tight upright conical fold (Fig. 4-5-Ob13), (2) a very tight and slightly overturned conical fold (Fig. 4-8) similar to those developed in experiments with thinner mantle lithosphere, or (3) conical fold trains with shapes resembling mullions (Fig. 4-5-Ob1, for further information see data repository, Anexo D4). Radial fractures occurred in the outer arc and few parallel fractures developed. Short (<1cm) oblique to the compression direction shear zones developed in 3 of the models, and model-scale shear zones developed in two of them.

Three experiments responded to the applied stresses by developing model-scale shear zones. Two of them had an initial model mantle lithosphere thickness $ML_m = 2$ cm (Ob9 and Ob14) which developed a root shaped as a tight conical fold. The root of the other model, with $ML_m = 1$ cm (Ob4), turned into a recumbent fold. Model Ob14 (Fig. 4-9) is representative of the morphologies that these models present, such as an incompletely developed root in the inner arc, that is duplicated and off-centered by the effects of the model-size oblique to the compression direction shear zone developed in one of the orocline limbs, and the presence of radial fractures preferentially focused in the outer arc of one limb of the model.

4.3 Discussion

Thermomechanical experiments run over scaled lithosphere models produced results that shed light on the processes taking place during the deformation and evolution of the mantle lithosphere related to orocline buckling. In general, the obtained results allow a better understanding of the lithospheric-scale processes such as thick-skinned orocline buckling and, in a more particular case, the relationship between orocline development and mantle lithosphere removal event that is interpreted to have taken place in the uppermost Paleozoic in the Ibero Armorican Arc. In addition, these experiments show limitations in the development of lithospheric structures due to analogue materials behavior; scaling; and laboratory.

4.3.1 Limitations and interpretation.

During the orocline buckling experiments, independently of the initial setup, the strain pattern in the model is characterized by the extreme shortening in its core and the generation of a root below the inner arc as well as by extension in the outer arc focused in extensional fractures. These features were predicted by Ries and Shackleton (1976, Fig. 4-11A) who suggested tangential longitudinal strain as the main folding mechanism for lithospheric bending. This mechanism has also been proposed by Gutiérrez-Alonso et al. (2004) to explain the different aforementioned geological effects found in nature, especially in the IAA.

The shapes of the buckled model mantle lithosphere in our experiments are strikingly similar to those obtained in other plate convergence analogue and numerical models without pre-existing subduction (e.g. Pysklywec et al., 2002; Luth et al, 2009) and depict a marked conical geometry due to the vertical-axis buckling instead of a cylindrical one (Shemenda and Grocholsky, 1992; Luth et al, 2009) or an unknown 3D geometry, the latter obtained from 2D numerical modeling (Arnold et al., 2001; Pysklywec et al., 2002, 2010; Pysklywec, 2006).

A thickened lithospheric root observed in the experiments could trigger mantle lithosphere removal, conclusions consistent with the models of (Schott and Schmeling, 1998; Pysklywec et al. ; 2002, 2010; Pysklywec, 2006; Morecy and Doin, 2004) among others. However, neither delamination nor dripping of the mantle lithosphere occurred in the experiments due to the size limitations and the analogue materials used in the experiments, which caused the observed vertical crustal growth over the lithospheric root. The thermal conductivity of plasticine is quite low (between 0.65 and $0.8 \text{ W m}^{-1} \text{ K}^{-1}$), which implies that in order to keep realistic scalable thermal gradient the sub-lithospheric mantle in the models could not be thicker than 4 or 5 cm. Hence, the model asthenosphere is not thick enough to avoid the effect of the lower boundary (thermo-copper plate) of the experiments. This lower boundary avoided any further growth, delamination or dripping of the lithospheric root. Moreover, the yield strength of *Beck's orange* plasticine is likely too large to let *Weible red* plasticine detach and sink under natural gravity conditions, at least during the duration of the experiments. The use of a denser plasticine to model the mantle lithosphere would not have modified the results, as the asthenospheric analogue would not have allowed its downward migration. Thus, when the mantle lithosphere could no longer migrate downwards, it started to build a positive bulge along the surface of the model. Without the aforementioned experimental limitations, the process in nature would be the opposite: the dynamic sinking of the lithospheric mantle root would pull down the crust above it resulting in basin formation (Muñoz-Quijano and Gutiérrez-Alonso, 1997a; 1997b).

All of our experiments show extension in the outer arc, mostly produced by radial fractures. This brittle behavior is not expected in natural mantle lithosphere, where deformation should be accommodated by crystal plastic processes (e.g. Karato and Wu, 1993). The development of fractures in the model lithosphere instead of viscous stretching could be due to the plastic failure of the materials. In this way, the curvature in the outer arc becomes more pronounced causing mode 1 tensile fractures in the model mantle lithosphere. Moreover, in nature there is a considerable confining pressure in the mantle, which is incompatible with dilatancy and volume increase due to brittle fracturing. In addition, the lithosphere behaves continuously in nature, adding a lateral confining pressure. In the experiments, the layer representing the lithosphere is discrete and the behavior of the air-plasticine interface is not predictable. This confining pressures are not present in our experiment facilitate the plastic failure of the *Weible red* plasticine.

Apart from the radial fractures, the growth of tensile fractures in the model mantle lithosphere was not expected, as these are not likely to occur in the mantle lithosphere. The parallel fractures developed, maintaining the scaling relationship between the initial lithospheric thickness of the model and the spacing of the fractures as found in other natural examples of brittle rocks (e.g. Narr and Suppe, 1991; Mandal et al., 1994; Gross et al., 1995). These results are indicative of some stage of the experiment with a model mantle lithosphere behaving in a more brittle manner. Enhanced dilatancy due to the lack of confining pressure and subsequently plastic failure of the material is probably the most important reason why these mode 1 tensile axial loading fractures developed in the model mantle lithosphere.

Oblique shear zones (mode 2 fractures) are lacking in every model and just three of the models display model-scale vertical shear zones that control the final morphology of the model mantle lithosphere. All shear zones developed with angles $< 40^\circ$ with respect to the principal shortening direction. Zulauf and Zulauf (2004) documented that in models undergoing high strain rates ($\dot{E} > 10^{-3} \text{ s}^{-1}$) plasticine is not a suitable analogue for viscously deforming rocks because strain tends to be localized along discrete shear corridors inside the plasticine. The orocline buckling experiments of the present study were run under lower strain rates ($\dot{E} = 10^{-5} \text{ s}^{-1}$) and, despite the

local higher strain rates that take place especially in the outer arc of the model, the shear zones of the models mimic natural shear zones observed in most curved orogenic belts. These shear zones are interpreted to develop coevally with the curvature generation (Gutiérrez-Alonso et al., 2010). However, lithospheric-scale shear zones, like those of models Ob4, Ob9 and Ob14 (Fig. 4-9; data repository, Anexo D4), which act as high-strain domains, are not easily recognized in nature, although a more distributed pattern of shear zones likely to occur at lithospheric scale can be recognized (e.g. lithospheric scale shear zones in the North Armorican massifs, Porto-Tomar Shear zone (Gutiérrez-Alonso et al., 2010). In our opinion, the lithospheric scale shear zones probably formed in order to accommodate the lithospheric buckling, but they initiated along the previously formed radial or parallel fractures (see data repository, Anexo D4).

Another limitation is the impossibility of measuring the temperature in the growing root, that was presumably higher than the temperature measured at the side of the model because the root is closer to the thermo plate. It would have been useful to know this temperature in order to compare with numerical and field data, because the mechanism of lithospheric thickening in nature leads to an increase in thermal heat flow, raising the mantle isotherms. The increasing thermal flow drives the chemical and physical process that can trigger mantle lithosphere removal (Leech, 2001).

4.3.2 Implications of the experimental results on the development of the Ibero Armorican Arc

The new experimental results also improve our understanding of the mantle lithosphere behavior during thick-skinned orocline buckling and the dynamics of IAA. Results and structures obtained from the different models are in agreement with the structural, petrological, paleomagnetical and geochronological data described above from the Ibero Armorican Arc.

An interpretation of the whole process of thick skinned orocline buckling and the possible subsequent delamination or dripping of the lithospheric root inferred from structural, petrological and geochemical studies (Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2004; 2011a; 2011b), based on the results of the experiments produced in this work and on the aforementioned geological criteria, is depicted in Figure 4-11. We propose that the effect of mechanical thinning in the outer arc (Fig. 4-11A and C) caused the upwelling of the asthenosphere and the accompanying thermal uplift (Muñoz-Quijano, 2007a, 2007b) due to the increased heat flow and the subsequent lower crustal melting, which is supported by the emplacement of early post-tectonic granitoids at about 310 - 300 Ma (Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2011a; Gutiérrez-Alonso et al., 2011b). In the upper crust, this extension was accommodated by slip along crustal-scale shear zones (Gutiérrez Alonso et al., 2010). In contrast, the shortening in the core of the arc produced a thickened mantle lithospheric root and lower crust duplication by means of underthrusting. Collectively, these processes caused the isostatic depression of the core, which resulted in the sediment discharge into the core of the arc (Gutiérrez-Alonso et al., 2004) as well marine strata as the vestiges of a relict epicontinental sea (Merino Tomé et al., 2009). In the upper crust this shortening was accommodated by radial conical folding (Julivert and Marcos, 1973; Bastida et al., 1984; Gutiérrez-Alonso 1992; Aller and Gallastegui, 1995; Pastor-Galán et al., submitted), reverse and strike-slip faulting (Alonso, 1989; Nijman and Savage, 1989) and out-of-sequence thrusting (Alonso et al., 2009). The shear zones developed in the models at angles < 40° to the compression direction are consistent with the geometry and kinematics of some of the shear zones that cut the IAA's arc trace (Gutiérrez-Alonso et al. 2010).

The mass imbalance between the thickened root underneath the inner arc and the

stretched mantle lithosphere beneath the outer arc presumably caused an important gravitational instability that brought about the detachment and lithospheric removal of the root. Probably, the foundering of the root was initiated in the weakest zone, which permits the asthenosphere to begin to invade the location occupied by the former lithosphere (Pysklywec et al., 2010). The foundering of the sinking root further stretched the mantle lithosphere and probably dragged down part of the remaining, previously thinned outer arc mantle lithosphere, causing the final asthenospheric upwelling. Evidence for this final substitution of the mantle lithosphere by the asthenosphere is recorded by the contrast in the Sm/Nd isotopic signature observed a mantle-derived rocks that pre-date and post-date the postulated mantle lithosphere removal event (Gutiérrez-Alonso et al., 2011a). The enhanced thermal activity caused widespread magmatism from 310 and 300 and the foreland fold-and-thrust belt from 300 to 285 Ma (Valverde-Vaquero, 1999; Fernández-Suárez et al., 2000; Gutiérrez-Alonso et al., 2011b), low pressure - high temperature metamorphism (Fernández-Suárez, 1994; Martínez and Rolet, 1988); widespread Au and Sn-W mineralization (Martín-Izard et al., 2000), unusual high coal ranks (Colmenero et al., 2008) and dolomitizations which become more intense along the out-of sequence thrusts (Gasparini et al., 2003).

The duration of the orocline formation and the subsequent mantle lithosphere removal based on geological data, lasted about 10 m.y. (Weil et al., 2010) starting from ~310 Ma to ~300 Ma (Pastor-Galán et al., 2011) (late Carboniferous). Mantle lithosphere removal, inferred by the igneous activity in the inner arc, which correspond with the foreland fold-and-thrust belt, seem to have started at about 300 Ma (Gutiérrez-Alonso et al., 2011b), when the root was thick enough (Schott and Schmeling, 1998) to destabilize and founder. The mantle lithosphere removal would have lasted between 5 and 10 Ma according to numerical models (Schott and Schmeling, 1998; Pysklywec et al., 2010) and the related magmatic activity lasted for ca. 15 m.y. (Gutiérrez-Alonso et al., 2008).

These data imply that IAA formation was a relatively fast lithospheric-scale tectonic process, with a bulk north-south shortening acting at shortening rates between 5 and 10 cm/year. Such large, continuous and rapid thick-skinned processes could be explained in the context of large-scale plate motions as the scenario proposed by Gutiérrez-Alonso et al. (2008) for the IAA buckling.

5 Geochronology

In recent years, the profusion of published U-Pb detrital zircon age populations from clastic sedimentary rocks has become a powerful tool to unravel the paleogeographic and tectonic evolution of the Earth (Bradley, 2011) and to examine processes such as exhumation rates and related changes in topography during major tectonic events (Lonergan and Johnson, 1998; Stewart et al., 2008; Nie et al., 2010; Weislogel et al., 2010). Several such provenance studies have focused on the Ediacaran to Ordovician sedimentary rocks from the NW Iberian Variscides (e.g. Fernandez-Suarez et al., 1999, 2000; Fernández-Suárez et al., 2002; Gutiérrez-Alonso et al., 2003; Catalan et al., 2004; Díez Fernández et al., 2010) in order to understand the evolution of the northern Gondwana margin during Ediacaran and early Paleozoic times. However, there are only scarce detrital zircon data (Martinez et al., 2008) from mid- to late Paleozoic clastic strata. During this crucial time interval, dramatic changes in tectonic environment occurred in NW Iberia, from a passive margin to a collisional orogen followed by the Late Carboniferous development of a regional oroclinal structure and potential lithospheric delamination, in response to the Carboniferous collision of Laurussia with Gondwana (Weil et al., 2001, 2010; Gutiérrez-Alonso et al., 2004). A detailed analysis of the detrital zircon populations in this time interval provides an opportunity to monitor changes in provenance during continental collision and oroclinal bending of the orogen.

5.1 Sampling strategy

Thirteen samples were collected in the CZ ranging in age from Early Silurian to Early Permian (Figs. 5-1, 5-2 and 5-3). The location of each sample is given in Supplementary file in Anexo D5. Four samples are from the pre-orogenic Silurian to Lower Pennsylvanian sequences: (1) two samples from the platform sequence (PG14; Lower Silurian, Formigoso Formation and PG12; Upper Devonian Fueyo Formation); (2) one sample from the fore-deep succession (G4; Middle-Upper Mississippian Olleros Formation) and (3) one sample from the oldest clastic wedge in the CZ (PG9, Early Pennsylvanian San Emiliano Formation) when deformation started in the west (present day coordinates).

We selected three samples in younger syn-orogenic clastic wedge deposits of Westphalian age (Mississippian-Pennsylvanian) These samples are PG5 (Lena Group, Westphalian A-C), PG4 and PG6 (Sama Group, Westphalian B-D) situated in the Central Coal Basin (CCB; Fig. 5-1). The succession in the CCB is Westphalian A-D (Middle Pennsylvanian) in age and consists of a marine-dominated (mostly siliciclastic with limestone intercalations) lower section (Lena Group) and an upper section with more continental influence (Sama Group) characterized by thick deposits of sandstone, conglomerate and shale with abundant interbedded coal seams.

Additionally we selected four samples to document the effects and after-effects of oroclinal development. One sample is from Stephanian A strata (PG1, ca. 307 Ma; Kasimovian according to the marine stages in Gradstein et al., 2004) that crops out in the Esla unit (Alonso, 1987, 1989; Fig. 5-1), and three samples (PG8, PG11 and PG7) are from Stephanian BC (ca. 305, 304 and 303 Ma, respectively; Fig. 5-1) strata. The latter three samples unconformably overlie different rock units (Fig 2, Fig. 3). Sample PG8 overlies Devonian and Early Carboniferous rocks of the CZ whereas samples PG11 and PG7 (Fig. 5-1), both overlie Ediacaran-Cambrian strata of the Narcea Antiform.

Two samples were collected within the Permian succession (PG2 and PG3, Sotres Formation, ca. 292 and 295 Ma respectively). PG2 was taken in the easternmost sector of the CZ and PG3 in its central part (Figs. 5-1, 5-2 and 5-3).

5.2 Methods

Approximately 120 detrital zircon grains from each sample were separated and extracted using facilities at Complutense (Madrid) and Salamanca universities, then mounted in epoxy resin with zircon standards SL13 (U = 238 ppm) and TEMORA ($^{206}\text{Pb}^*/^{238}\text{U} = 0.06683$). The polished mounts were photographed before the analysis to document each zircon analysis. Individual zircon grains were analysed for U, Th, and Pb isotopes by LA-ICP-MS (Laser Ablation with Inductively Coupled Plasma Mass Spectrometry) at the Museum für Mineralogie und Geologie (Senckenberg Naturhistorische Sammlungen Dresden), using a Thermo-Scientific Element 2 XR sector field ICP-MS coupled to a New Wave UP-193 Excimer Laser System. A teardrop-shaped, low-volume laser cell was used to enable sequential sampling of heterogeneous grains (e.g. growth zones) during time-resolved data acquisition. Each analysis consisted of 15s background acquisition followed by 35s data acquisition, using laser-spot sizes of 15–35 μm . A common-Pb correction based on the interference and background-corrected ^{204}Pb signal and a model Pb composition (Stacey and Kramers, 1975) was carried out where necessary. The criterion for correction was whether the corrected $^{207}\text{Pb}/^{206}\text{Pb}$ lay outside the internal error of measured ratios. Time-resolved signals of the LA-ICP-MS were checked in order to detect disturbances caused by cracks or mineral inclusions. In such cases, analyses were excluded from age calculations. Raw data were corrected for background signals, common Pb, laser-induced elemental fractionation, instrumental mass discrimination, and time-dependent elemental fractionation of Pb/Th and Pb/U using an Excel® spreadsheet program developed by A. Gerdes (Institute of Geosciences, Johann Wolfgang Goethe-University Frankfurt, Frankfurt am Main, Germany). Reported uncertainties were propagated by quadratic addition of the external reproducibility obtained from the standard zircon GJ-1 (ca. 0.6% and 0.5–1% for the $^{207}\text{Pb}/^{206}\text{Pb}$ and $^{206}\text{Pb}/^{238}\text{U}$ respectively) during individual analytical sessions and the within-run precision of each analysis. Analyses with a concordance in the range 90–110% were used for concordia and probability density distribution plots. A total of 1620 analyses were carried out on thirteen samples, of those, 216 were >10% discordant and were discarded. Discordance may originate from Pb loss, addition of common Pb or ablation of different age domains within the zircon. Concordia diagrams and probability density plots (Fig. 5-6) were produced using Isoplot/Ex 3.7 (Ludwig, 2001). For concordant analyses (i.e. analyses whose 2σ error ellipse intercepts the Concordia curve) we used Concordia ages and errors (Ludwig, 1998) as calculated by Isoplot. For discordant analyses (still within the 90–110% concordance range) older than 1000 Ma we use the more precise $^{207}\text{Pb}/^{206}\text{Pb}$ age. For further details on analytical protocol and data processing may be found in Frei and Gerdes (2009).

5.3 Results

The U-Pb data are given in Tables in the supplementary data (Anexo D5) and are represented in the concordia plots (Figures 5-4 and 5-5) and the relative probability plots (Figures 5-6 and 5-7). Additionally, we have plotted all U-Pb age data from the 13 samples in a kernel density plot (Wand and Jones, 1995) using a smoothing wavelength of 15 Ma (Fig. 5-8). The peaks in this plot are used to identify the age/span of the main episodes of zircon-forming events recorded in the detrital population of the studied samples. As the data from the pre-orogenic rocks provide the background necessary to interpret the data from the syn-orogenic and post-orogenic rocks, we discuss the data in order of their depositional age, from oldest to youngest (Fig. 5-2).

5.3.1 Pre-orogenic sequence (PG 14, PG 12, G4, PG9)

In sample PG14 (Fig. 5-2), about 30% of the zircons represent the youngest population and range from 540-850 Ma. Comparable populations, ca. 30% each, yield 900-1150 Ma and 1750-2150 Ma ages. Additionally, 9% of the zircons define an Archean population (2500 – 2800 Ma; Figs. 5-6 and 5-8). Younger pre-orogenic samples (PG12, G4 and PG9) are characterized by a higher proportion (39%-46%) of 540-850 Ma zircons. In sample PG-12, this is also the youngest population (Figs. 5-3 and 5-5). The zircons with ages ranging between 900-1150 Ma are the second most abundant population (26% to 32%) in all three samples. This population is more abundant than in the syn-orogenic and post-orogenic rocks (Fig. 5-6, 5-7 and 5-8). Older Proterozoic zircons represent a comparatively smaller proportion (11 and 21% respectively) whereas the 2500-2800 Ma zircons range between 7% and 14%. In addition, G4 and PG9 contain a small proportion of Late Cambrian-Early Ordovician (475-510 Ma; 1% and 3% respectively) and Mississippian (360-320 Ma; 1% both samples) zircons, the latter representing the youngest population in both samples.

The youngest zircons in samples PG-14 and PG-12 are 523 Ma and 544 Ma, respectively, both much older than their depositional age as established by their fossil content (Bastida, 2004 and references therein). In sample G4, however, the youngest zircon (365 Ma) is similar to its depositional age as established by fauna. The youngest zircon in sample PG9 is 402 Ma, closer to its depositional age than is the case for Silurian and Devonian rocks (see tables in supplementary material, Anexo D5).

5.3.2 Syn-Orogenic sequence (PG5, PG4 and PG6)

The syn-orogenic samples (Fig. 5-4, 5-6, and 5-9) are dominated by a 540-850 Ma population (40%-52%). The second most abundant population in samples PG5 and PG6 is between 900-1150 Ma (21%-24% respectively) whereas in PG4, this population only represents 12%. In the three samples, a Paleoproterozoic (1750-2150 Ma) population represents 10%-20% of the analyses and an Archean population (2500-2800 Ma) represents 5%-8% of analyses. The three samples contain between 3% and 8% of zircons in the 475-510 Ma and 360-320 Ma age intervals respectively.

The youngest zircons in samples PG5, PG4 and PG6 are 326, 325 and 322 Ma respectively, close to the depositional ages established by their fossil content (see tables in supplementary material, Anexo D5).

5.3.3 Syn-Oroclinal sequence PG1, PG8, PG11 and PG7

In three of the samples from post-orogenic but syn-oroclinal strata (PG1, PG8, PG11; Figs. 5-5, 5-7 and 5-9) the main age populations are 540-850 Ma (41%-48%), 900-1150 Ma (27%-30%), 1750 – 2150 Ma (13%-21%) and 2500-2800 Ma (6%-9%). Notably, these samples contain few zircons of the 475-510 Ma age population (<4%) and no 360-320 Ma zircons. The youngest zircons in these samples are 481 Ma, 494 Ma and 474 Ma respectively, i.e. significantly older than their respective stratigraphic ages.

Sample PG7 has a small population (2%) of 360-335 Ma zircons (youngest at 335 Ma) and a slightly higher proportion of 475-510 Ma zircons (8%). Other populations present are similar to those of the other three samples but occur in different proportions (540-850 Ma, 30%; 900-1150 Ma, 9%; 1750 – 2150 Ma, 29%; and 2500-2800 Ma, 10%).

5.3.4 Permian sequence

Permian samples contain a variety of Proterozoic (540-850 Ma, 41%; 900-1150 Ma, 20%-25%; 1750-2150 Ma, 11-20%; and 2500-2800 Ma, 2-7%) and Paleozoic populations (310-290 Ma, 1-10%; 360-320 Ma, 2-5%; 475-510 Ma 2-10%). The youngest zircons (298 Ma in PG3, 290 Ma in PG2) in both samples are approximately the same age as the estimated depositional age of the rocks (ca. 295 Ma).

5.3.5 Main zircon-forming events

Kernel density estimation was used to produce a graph that visually groups statistically similar samples (Sircombe and Hazelton, 2004) in order to measure the similarity of the different samples analysed. In addition, if the analysed samples are similar it is a useful technique to recognize the main zircon-forming events that are recorded by the studied rocks. The kernel density function approximates the shape of the zircon probability density curves at a particular age by taking into account the age uncertainties and the influence of estimated ages within close proximity of a given age. The average distance between the smoothed probability curves is then employed as a measure of dissimilarity, and displayed on a graph. This approach can provide a statistical means for assessing the extent of those similarities. With the Kernel density function and using a wavelength of 15 m.y., seven populations can be distinguished in the whole detrital zircon population of the 13 samples analysed (Fig. 5-8): (1) an Archean population (2500 – 2800 Ma) with a maximum at 2615 Ma; (2) a 1750 – 2150 Ma, with two maxima at 2000 and 1875 Ma; (3) a 900 to 1150 Ma population with a maximum in 1025 Ma; (4) a 540-850 Ma population with a maximum at 630 Ma; (5) a Late Cambrian to Early Ordovician population (475 – 510 Ma) with a maximum at 495 Ma; (6) a Mississippian population (360 to 320 Ma) with a maximum at 338 Ma and a Late Carboniferous to Early Permian (310-290 Ma) population with a maximum at 300Ma.

5.5 Discussion

Many of the main populations (peaks) identified in the Kernel density diagram (Fig. 5-8) can be related to established different-zircon forming events. The Late Archean and Proterozoic populations are common features in many clastic sequences worldwide. For example, the 2500-2800 Ma population has been linked to global-scale orogenic events associated with the formation of a Late Archean supercontinent (known by different names, Vaalbara, Superia, Sclavia or Kenorland; Bradley et al., 2011 and references therein). Similarly, the 1750 – 2150 Ma population is coincident with the suggested time for the amalgamation of the supercontinent Nuna (Bradley et al., 2011). The samples contain more dominant populations that are coeval with Grenville orogenesis (900-1100 Ma; Evans, 2009), responsible of the formation of the supercontinent Rodinia and with the Cadomian-Pan-African orogeny (540-850 Ma) which occurred along the northern margin of Gondwana (Fig 5-6, 5-7, 5-8 and 5-9; Murphy et al., 2006).

The Late Cambrian-Early Ordovician population corresponds with widespread magmatic events along the northern margin (Amazonia, West Africa) of Gondwana associated with the opening of the Rheic Ocean. Examples of such magmatism in NW Iberia include the “Ollo de Sapo” volcano-sedimentary formation (Fig 5-8; Díez-Montes et al., 2006).

The Mississippian zircon population corresponds with the exhumation of collisional and extensional igneous rocks developed during the Variscan Orogeny (e.g. Fernández-Suárez et al., 2000). The Late Carboniferous-Early Permian population may reflect zircons formed during intra-crustal magmatism triggered by a thermal event produced by thinning (outer arc) and and

delamination (inner arc) of the lithospheric mantle as a consequence of oroclinal buckling in NW Iberia (Gutiérrez-Alonso et al., 2004; 2011a; 2011b; Pastor-Galán et al., in press)

Previous U-Pb studies in detrital zircons of sedimentary rocks in NW Iberia have constrained the paleogeographic position and tectonic evolution of Iberia from Ediacaran to Ordovician times. Based on the match between detrital zircon populations and potential sources, these studies conclude that NW Iberia was located near northern Africa in the Ordovician (e.g. Fernández-Suárez et al., 2002; Bea et al., 2010; Díez-Fernández, 2010). Our data yield similar populations in Silurian to Late Devonian strata, suggesting derivation from the same source. Collectively, these data are consistent with a long-lived (Ordovician-Lower Devonian) passive margin setting that was not affected by any significant zircon forming event and/or changes in the source areas. There is no evidence for the development of a magmatic arc during this time interval, indicating that closure of the Rheic Ocean could not have been accommodated by subduction beneath the Gondwanan margin (i.e. Stampfli and Borel, 2002). Instead, these data are consistent with the hypothesis of a northerly-directed subduction of the Rheic oceanic lithosphere beneath Laurussia (Martínez-Catalán and Arenas, 2003; Nance et al., 2010 and references therein). These data also indicate that NW Iberia was part of the passive northern margin of Gondwana from the Late Cambrian-Early Ordovician opening of Rheic Ocean (e.g. Díez-Montes, 2006; Avigad et al., 2012; Fig. 5-10) until the onset of collision between Gondwana and Laurussia which began in the Late Devonian (e.g. Dallmeyer et al., 1997; Fig. 5-10).

The zircon populations of the CZ strata in the Late Cambrian-Late Devonian time interval have approximately the same percentages of Proterozoic populations (ca. 30% of 540-850 Ma zircons, ca. 30% zircons of 900-1150 Ma and ca. 40% of 1750 – 2150 Ma) and minor Archean zircons (Figs. 5-6, 5-7, 5-8 and 5-9). These populations are similar to those in detrital rocks of the same age in central north Africa, except for the slightly higher proportion of 900-1150 Ma zircons found in this study (20%-25% for central north Africa; 25%-30% for NW Iberia (Meinhold et al., 2011) but different to those in western north Africa, which lack 900-1150 Ma zircons (e.g. Abati et al., 2010; Avigad et al., 2012). The populations found in the CZ also match the known bedrocks in central north Africa during this interval of time: the Saharan Craton and Arabian-Nubian shield (e.g. Loizenbauer et al., 2001; Abdelsalam et al., 2002; Be'eri-Shlevin et al., 2009; Stern et al., 2010; Morag et al., 2011; Avigad et al., 2012). These data indicate that central north Africa is the nearest likely paleoposition for the CZ from Ordovician to Late Devonian.

In contrast to the stability of source regions from the Lower Ordovician to the Late Devonian, in the Carboniferous rocks two new zircon populations occur, one spanning from the Late Cambrian to Early Ordovician (475-510 Ma), the other is a Mississippian population (315-359 Ma). In addition some scarce Silurian zircons are also present. The Mississippian population is not present in three of the Stephanian samples (Fig. 5-7 and 5-9).

Compared to the passive margin pre-orogenic rocks, the syn-orogenic and post-orogenic Carboniferous clastic rocks contain a higher proportion of Cadomian zircons (540-850 Ma, 52% in sample PG4; Fig. 5-3 and 5-5) as well as a significant Late Cambrian-Early Ordovician (510-475 Ma) zircon population. These changes may reflect exhumation of Cadomian basement rocks of NW Iberia during Variscan deformation, a conclusion consistent with (i) regional syntheses indicating that the Variscan mountains started to form in the late Devonian (Fig. 5-10; Dallmeyer et al., 1997), and (ii) stratigraphic studies indicating that this interval was accompanied by a change from platformal to foreland basin deposition (Keller et al., 2008). The U-Pb data suggest that the sediment supply involved the recycling of the previously deposited sediments and erosion of igneous rocks formed during the Late Cambrian-Early Ordovician (opening of the Rheic Ocean) (Figs. 5-8 and 5-10), (Murphy et al., 2008; (Montes et al., 2010). In addition, the

presence of Late Ordovician, Silurian and Devonian zircons (450-395 Ma) is interpreted to be the result of the erosion of rocks within the Rheic Ocean suture zone, where zircons of these ages occur (Sánchez-Martínez, 2007) and are located at least ca. 300 km to the west (present day coordinates). The Mississippian population (359-316 Ma) is attributed to the exhumation of the collisional and extensional igneous rocks developed during the Variscan orogeny (e.g. Fernández-Suárez et al., 2000).

Syn-oroclinal rocks have zircon populations very similar to those found in the Neoproterozoic clastic rocks in NW Iberia (Fernández-Suárez et al., 2000). Compared with the Carboniferous pre-orogenic and the syn-orogenic strata, the syn-oroclinal rocks contain a lower proportion of Late Cambrian-Early Ordovician Ma zircons, and virtually no Silurian-Devonian or Mississippian zircons (with the exception of a 2% population in sample PG 7, Figs. 5-7 and 5-9). These minor differences with pre-orogenic and syn-orogenic rocks are interpreted to reflect recycling of local strata, possibly facilitated by the uplift associated with coeval out-of-sequence thrusts (Alonso et al., 2009; Fig 5-1) and by reactivation of Variscan structures during oroclinal buckling (Pastor-Galán et al., 2012). Samples PG1, PG11, PG8 are located in strata that overlie Ordovician, Silurian, Devonian and Carboniferous rocks (Fig. 5-1 and 5-9) and contain similar populations to those of the pre-orogenic rocks as well as a small input of Late Cambrian-Early Ordovician zircons (Fig 5-6, 5-7 and 5-9). These detrital zircons may have been derived from the recycling of pre-Pennsylvanian strata, in which Mississippian zircon population is scarce or absent. However, PG7 contains Late Cambrian-Early Ordovician and Mississippian populations in similar proportions to those of syn-orogenic foreland basin deposits of the CCB as well as a small 900-1150 Ma zircon population (Fig 5-7 and 5-9). These differences may reflect its location in the CZ. PG7 is located over Cambrian and Ediacaran rocks (Fig. 5-1) thus receiving sediments derived from the recycling of uplifted Cambrian and Ediacaran rocks and from Ediacaran, Late-Cambrian-Ordovician and Mississippian igneous rocks.

Permian strata contain very similar zircon populations to the Carboniferous pre-orogenic and syn-orogenic strata, but also contain Early Permian zircons, attributed to erosion of the Permian volcanic rocks interbedded with the siliciclastic strata. However, sample PG2, located furthest to the east, has only minor Late Cambrian-Early Ordovician and Mississippian populations (Figs. 5-7 and 5-9). The populations of PG2 and PG3 can be explained by recycling of the CZ strata to the west and the erosion of the presumably exposed Gondwana basement to the east (in present day coordinates and not currently exposed). Therefore, outcrops situated to the east (in present-day coordinates) received a higher sediment input from eastern Gondwana (PG2, Fig. 5-1) whereas outcrops situated to the west (in present days coordinates) show more input from the recycling of the pre-orogenic passive-margin rocks, as well as from the syn-orogenic foreland rocks (PG3, Fig 5-1). The formation of the Permian basins and associated volcanism is attributed to isostatic uplift (Muñoz-Qujano and Gutiérrez-Alonso, 2007) caused by lithospheric delamination (Gutiérrez-Alonso et al., 2004; 2008; 2011a,b; Pastor-Galán et al., in press).

6 Conclusions

The findings and contributions of my research have been documented and discussed throughout the various chapters of this PhD dissertation. The main conclusions of this thesis are summarized in the following paragraphs.

In the Iberian-Armorican Arc, and especially its core, the Cantabrian Orocline, exposes the main structural and stratigraphic features necessary to investigate the kinematics and timing of oroclinal buckling. The analysis of joint patterns in the Cantabrian Orocline shows the presence of at least three different stress-strain fields related to the formation of tension fractures:

A) During the east–west (in present-day coordinates) compression, related to the collision between Gondwana and Laurussia and the development of the Variscan foreland fold–thrust belt, two sets of joints, both parallel and normal to the main Variscan structures, were formed. The geometric enveloping surface of these joint sets was passively folded around a vertical axis during the oroclinal buckling together with the whole orogen.

B) During the north–south compression that resulted in oroclinal bending of the Cantabrian Orocline, two new different joint sets developed with initial orientations N-S and E-W when the orocline was already bent between a 30% and 50%. Therefore, the enveloping surface of these sets presents a curvature between a 50% and a 70% of the overall curvature of the Cantabrian Orocline. These sets have been recognized in the Upper Pennsylvanian continental sedimentary rocks.

C) During post-Permian times and probably related to the different Mesozoic and Tertiary Alpine processes three different sets developed. These joint sets present no relative vertical axis rotation.

Joint patterns in the Cantabrian Orocline indicate that the Cantabrian Orocline was closed between 30% and 50% prior to Stephanian times (Kasimovian to Upper Gzelian, *ca.* 306-303 Ma.) and was completely bent by the lowermost Permian. These kinematic constraints indicate that oroclinal bending of the Cantabrian Orocline occurred between the middle Moscovian and the Carboniferous–Permian boundary (*ca.* 310 to 299 Ma). The results are consistent with previous paleomagnetic data (Weil *et al.*, 2010) and with the ductile structures found in the outer arc of the IAO (Gutiérrez-Alonso *et al.*, 2010). The closure rate suggested for the OIA in angular terms is *ca.* $15^{\circ}/10^{-6}$ year to $20^{\circ}/10^{-6}$ year, and the north-south shortening rate would be greater than 10 cm/year, which makes the IAO buckling a faster process when compared with Bolivian orocline ($4^{\circ}/10^{-6}$ year and *ca.* 9 cm/year, Allmendinger *et al.*, 2005), Calabrian orocline ($11^{\circ}/10^{-6}$ year and 5,5 cm/year, Cifelli, 2008; Johnston y Mazzoli, 2009) or the Alaskan oroclinal (10°*10⁻⁶ year and 4 cm/year, Johnston, 2001).

The buckling of a linear orogen at a continental scale necessarily causes structures at all lithospheric levels. Thus, strike-slip shear zones developed in the middle crust (Gutiérrez-Alonso *et al.*, 2010) whereas in the upper crust of the IAO the following structures were formed: 1) joint sets (described above), 2) out-of-sequence thrusts (Alonso *et al.*, 2009), 3) radial folds to the Variscan structural main trend with conical geometry. The detailed study of conical folds in curved mountain belts resulting from the interference of superposed orthogonal shortening events is a powerful tool for characterizing the sequence of tectonic events that produce oroclinal. Conical folds develop with different semiapical angles and axis attitudes depending on the initial orientations of the geological surfaces and their position with respect to the vertical rotation axis responsible for the orocline formation.

Considering the Cantabrian Orocline, the geometry of the conical folds in the Somiedo

unit and the innermost section of the WALZ indicate that the local vertical rotation axis should be placed somewhere near the axial trace of the syncline described by the Belmonte thrust (Los Lagos synform). All the rocks within the Somiedo unit located to the east of this axial trace are situated in the shortening domain of the orocline. The axial trace of this syncline acted as the neutral line, where rocks are neither shortened nor stretched. Finally, all rocks to the west of this neutral fibre are located in the stretching zone. The conical folds studied in the Cantabrian Arc indicate the general usefulness of conical folds in recognizing and interpreting orocline development in other curved mountain belts.

Although a detailed study of the crustal structures generated during the oroclinal buckling has been possible thanks to the existing outcrops (Gutiérrez-Alonso *et al.*, 2010; this thesis), the study of the lithospheric response to a thick skinned oroclinal buckling can only be done through indirect methods such as geochronology (Gutiérrez-Alonso *et al.*, 2011a), geophysics (Martínez-Catalán, 2011), geochemistry (Gutiérrez-Alonso *et al.*, 2011b) or numerical and analogue modeling (*e.g.* Muñoz-Quijano y Gutiérrez-Alonso, 2007a; 2007b; this thesis). Analogue modeling of orocline buckling is a powerful tool to reproduce lithospheric-scale processes similar to those observed in nature at laboratory scale. The experiments provided useful information indicating that during thick skinned orocline buckling of the lithosphere the longitudinal tangential strain is the main mechanism of deformation (Gutiérrez-Alonso, 2004), causing extension in the outer arc and significant shortening in the inner arc. The extension of the mantle lithosphere in the outer arc is probably accommodated by viscous stretching, while extension in the crust is accommodated by crustal-scale shear zones. On the other hand, the inner arc shortening is accommodated in the mantle lithosphere by folding and underthrusting, developing a well defined lithospheric root as has been observed in other mechanisms of lithospheric thickening (*e.g.* Cloetingh *et al.*, 1999; 2002; Pysklywec *et al.*, 2010; Fernández-Lozano *et al.* 2011).

Delamination and detachment of the mantle lithosphere root developed during orocline buckling was possible under some of the conditions selected for the performed analogue modeling experiments. At several temperature conditions the detachment did not take place, likely because the selected plasticines have a yield strength that does not permit sinking. However, the experiments in which detachment occurred showed a geometry in which dripping and delamination seem to act as the same time. Additionally, the geometry obtained is similar to that observed in the natural lithosphere under the Vrancea Arc (Carpathinas; Fillerup *et al.*, 2010). Following the scaling parameters for lithospheric processes (Davy and Cobbold, 1991) the detachment process would take 15 m.y. to be completed and 8 m.y. to reach an intermediate stage. These data are consistent with results of previous numerical and analogue modeling experiments (Schott and Schmelling, 1998; Pysklywec and Cruden, 2004) and the geochronological and geochemical results obtained in the Iberian Armorican Orocline (Gutiérrez-Alonso *et al.*, 2011a; 2011b).

Great changes in the morphology of the lithosphere, both lithospheric thickening and thinning, should bring about a significant topographic response. If the oroclinal buckling in the European Variscides was a thick-skinned feature, profound topographic changes must have taken place in a manner analogous to what happens in rift zones, or in places where the lithosphere has been thickened. In rift zones the lithosphere is very thin and therefore, the asthenosphere is situated close to the surface, which produces a substantial thermal flow that in turn produces a topographic uplift. On the other hand, in the places where the lithosphere has been thickened, the gravitational forces induce subsidence and creation of sedimentary basins (*e.g.* the Focsani Basin, Romania).

Detrital zircon studies can complement regional syntheses in helping to deduce

paleogeographic location and nature of source areas, the occurrence of major tectonic events such as terrane dispersal and continental collisions, as well as the crustal response to lithospheric-scale processes such as oroclinal buckling and lithospheric delamination.

U-Pb geochronological analysis of detrital zircons in thirteen samples of the CZ of the NW Iberian Variscan belt reveal that this portion of Iberia was part of the northern passive-margin of Gondwana from the Ordovician to Late Devonian, until the onset of collision between Gondwana and Laurussia. Zircon populations in these samples show important similarities with zircons found in coeval detrital rocks from central-north Africa (e.g. Meinhold *et al.*, 2011). Additionally, the populations found in NW Iberia are consistent with a possible Saharan/NE African provenance. We suggest that from the Ordovician to the Late Devonian, NW Iberia was situated along the northern Gondwanan passive margin, close to the paleoposition of central north Africa and the Saharan craton. Additionally, the Carboniferous-Permian samples studied record the topographic changes produced during the Variscan orogeny, the Cantabrian Orocline formation and the subsequent detachment of the lithospheric mantle.

Thus, considering the previous studies and those carried out as part of this PhD thesis, we can conclude that a plausible overall interpretation is that the Variscan orogen was folded around a vertical axis during the Pennsylvanian during a period that lasted about 10 m.y. The structures developed during the formation of IAO buckling suggest that this process occurred due to a large change in the stress field from E-W to N-S (in present day coordinates), which suggests that the folding of the orogen was produced by the mechanism of buckling. The buckling process affected the whole lithosphere, which would have been deformed by a dominant mechanism of longitudinal-tangential strain. According to the experimental analogue models, the root formed in the lithospheric-mantle beneath the core of oroclinal was probably caused by lithospheric folding. This root became gravitationally unstable at around the Carboniferous-Permian boundary. At that time it could begin to develop a Rayleigh-Taylor instability ending with the detachment and sinking of the lithospheric-mantle in the asthenospheric-mantle in a process that could be lithospheric delamination, dripping or a mixture of both.