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Provenance analysis of the Paleozoic sequences of the northern Gondwana margin in NW Iberia: Passive margin to Variscan collision and orocline development

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ABSTRACT

The Cantabrian Zone of NW Iberia preserves a voluminous, almost continuous, sedimentary sequence that ranges from Neoproterozoic to Early Permian in age. Its tectonic setting is controversial and recent hypotheses include (i) passive margin deposition along the northern margin of Gondwana or (ii) an active continental margin or (iii) a drifting ribbon continent. In this paper we present detrital zircon U–Pb laser ablation age data from 13 samples taken in detrital rocks from the Cantabrian Zone sequence ranging from Early Silurian to Early Permian in depositional age. The obtained results, together with previously published detrital zircon ages from Ediacaran–Ordovician strata, allow a comprehensive analysis of changing provenance through time. Collectively, these data indicate that this portion of Iberia was part of the passive margin of Gondwana at least from Ordovician to Late Devonian times. Zircon populations in all samples show strong similarities with the Sahara Craton and with zircons found in Libya, suggesting that NW Iberia occupied a paleoposition close to those regions of present-day northern Africa during this time interval. Changes in provenance in the Late Devonian are attributed to the onset of the collision between Gondwana and Laurussia.
Additionally, the Middle Carboniferous to Permian samples record populations consistent with the recycling of older sedimentary sequences and exhumation of the igneous rocks formed before and during the Variscan orogeny. Late-Devonian to Permian samples yield zircon populations that reflect topographic changes produced during the Variscan orogeny and development of the lithospheric scale oroclinal buckling.

1. Introduction

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; Gutierrez-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.

In this paper we present detrital zircon ages from 13 samples from the Cantabrian Zone (CZ) of NW Iberia, in the foreland of the Variscan belt of NW Iberia whose depositional age spans the passive margin stage of northern Gondwana from Silurian to upper Devonian times, the continental collision (Variscan orogeny) from upper Devonian to upper Carboniferous times and the development of the Cantabrian Orocline (CO) (Gutiérrez-Alonso et al., 2012) during the latest Carboniferous and the early Permian (Johnston and Gutiérrez-Alonso, 2010). The objective of our study is threefold: (i) to characterize the sources of the sediments during its Silurian–Devonian passive margin stage in order to constrain its paleogeography; (ii) to provide constraints on the

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exhumation and erosion of the different units involved in the Variscan orogenic event; and (iii) to account for the provenance changes related with the formation of the CO as well as the potentially major topographic changes triggered by the hypothesized lithospheric detachment event. In addition, by comparing the Silurian–Devonian data with previously published data for the Ediacaran–Ordovician, changes in provenance during the evolution of the passive margin can be assessed.

2. General geology

2.1. Regional setting

In the Late Neoproterozoic and Early Cambrian, a long history of subduction and accretion of island arcs occurred along the northern margin of Gondwana (Murphy et al., 2000; Linnemann et al., 2008; Nance and Linnemann, 2008; Pereira et al., 2012). After the protracted period of rifting (Sanchez-Garcia et al., 2008; Pereira et al., 2012), the Rheic Ocean opened by the Late Cambrian–Early Ordovician with the separation of several peri-Gondwanan terranes (Avalonia, Carolina, Ganderia) from the northern margin of Gondwana (Murphy et al., 2006; Nance et al., 2010). This period of rifting and early drifting is recorded in NW Iberia by widespread rift-related igneous activity (Díez-Montes, 2006; Valverde-Vaquero et al., 2006; Gutiérrez-Alonso et al., 2007; 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 (Nance et al., 2010 and references therein).

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 (Van Der Voo, 1982, 1988; van der Voo, 1993; Tait et al., 1994; Tait, 1999; Stampfli and Borel, 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 (Robardet, 2003; Linnemann et al., 2004, 2008; Barreiro et al., 2006; Fernandez-Suarez et al., 2006; Gutiérrez-Alonso et al., 2008; Martínez Catalán et al., 2009; 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 (Dallmeyer et al., 1997) and continued shortening is thought to have led the extensional collapse of the thickened hinterland at 320 Ma (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.

The CO was developed after closure of the Rheic Ocean and the development of the Variscan orogen. Gutiérrez-Alonso et al. (2004) propose a thick-skinned model for oroclinal development which involves lithospheric-scale rotation of the orogen limbs, with extension in the outer arc resulting in thinning of the mantle lithosphere, and coeval shortening in the inner arc (Julivert and Marcos, 1973; Julivert and Arboleya, 1986; Alvarez-Marron and Perez-Estaun, 1988; Gutiérrez-Alonso et al., 2010; Pastor-Galán et al., 2012a). Lithosphere thickening beneath the inner arc would have resulted in gravitational instability causing detachment and removal of the mantle lithosphere from the lower crust, in turn resulting in upwelling of the asthenosphere thereby triggering voluminous Late Carboniferous–Permian magmatism in the Variscan fold-and-thrust belt as a result of an associated increase in heat flow (Fernandez-Suarez et al., 2000; Gutiérrez-Alonso et al., 2004, 2011a,b; Fernández-Suárez et al., 2011). The hypothesized high heat flow may also explain (i) uncommon high coal ranks in the uppermost Carboniferous continental basins (Colmenero and Prado, 1993; Colmenero et al., 2008); (ii) gold mineralization in the foreland fold-and-thrust belt (Martin-Izard et al., 2000); (iii) remagnetization recorded in Late Carboniferous–Permian strata (Weil and Van der Voo, 2002); (iv) dolomitization along late breaching and out-of-sequence thrusts (Gasparrini et al., 2006) and (v) post-orogenic topographic elevation (Muñoz-Quijano and Gutiérrez-Alonso, 2007).

2.2. Geology of the Cantabrian Zone

The CZ of northern Iberia is situated in the core of the CO (Gutiérrez-Alonso et al., 2004; Weil, 2006) (Fig. 1A and B). 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 (Gutierrez-Alonso and Nieto, 1996; Brime et al., 2001) and by conodont color 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. 2) that thin towards the east and culminate with a distinctive sequence of Silurian black shale and iron-rich sandstone (Fig. 2). Paleocurrent data recorded in the lower paleozoic strata indicate that its sediment source was located to the east (Aramburu and García-Ramos, 1993; Shaw et al., 2012) but there are no currently exposed potential source rocks. In the CZ, the Devonian and Mississippian succession consists of alternating passive margin carbonate and siliciclastic formations (Fig. 2) where several transgressions and regressions have been documented (Aramburu et al., 2002; Gibbons and Moreno, 2002; Keller et al., 2008). 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 are younger westwards (e.g. Colmenero et al., 2008) and show little deformation. They are coal-bearing, continental, clastic rocks including conglomerates, sandstones and pelities (Fig. 2) 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 Stephanian 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 CO (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, volcaniclastic rocks and calc-alkaline basaltic lava flows with sparse isolated coal seams (Martínez-García, 1981; Suárez, 1988).