

# Zircon Inheritance Reveals Exceptionally Fast Crustal Magma Generation Processes in Central Iberia during the Cambro-Ordovician

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*The Variscan basement of the Central Iberian Zone contains abundant Cambro-Ordovician calc-alkaline to peraluminous metagranites and metavolcanic rocks with two notable features: first, they were apparently produced with no connection to any major tectonic or metamorphic event; second, they have an unusually high zircon inheritance. U–Pb dating combined with cathodoluminescence imaging reveals that about 70–80%, in some samples nearer 100%, of the zircon grains contain inherited pre-magmatic cores, despite the temperature reached by the magmas (about 900°C, calculated using the Ti-in-zircon thermometer) being high enough to dissolve all the available zircon (from the rock's zircon saturation temperature, 770–860°C). The fact that the dissolution of zircon was so incomplete can only be attributed to the kinetics of heat transfer to and from the magmas. Three-dimensional modeling of zircon dissolution behavior in melts with a composition similar to the Iberian Cambro-Ordovician magmas indicates that the survival of zircons from the suggested late Pan-African protolith would be possible only if melt production was rapid, specifically less than 10<sup>4</sup> years, and probably about 2 × 10<sup>3</sup> years, from the beginning of melting (700°C) to the thermal peak (900°C). Melt production was followed by fast magma transfer to upper crustal levels resulting either in surface eruption or in the emplacement of small (<400 m thick) sills or laccoliths. We suggest that these elevated rates of crustal melting could only have been caused by intrusion of mantle-derived mafic magmas, most probably at the base of the crust. This scenario is consistent with a rifting regime in which crust and mantle were mechanically decoupled; this would explain the scarcity of contemporaneous crustal deformation. Furthermore, fast melting rates in the lower crust followed by fast melt transportation to the*

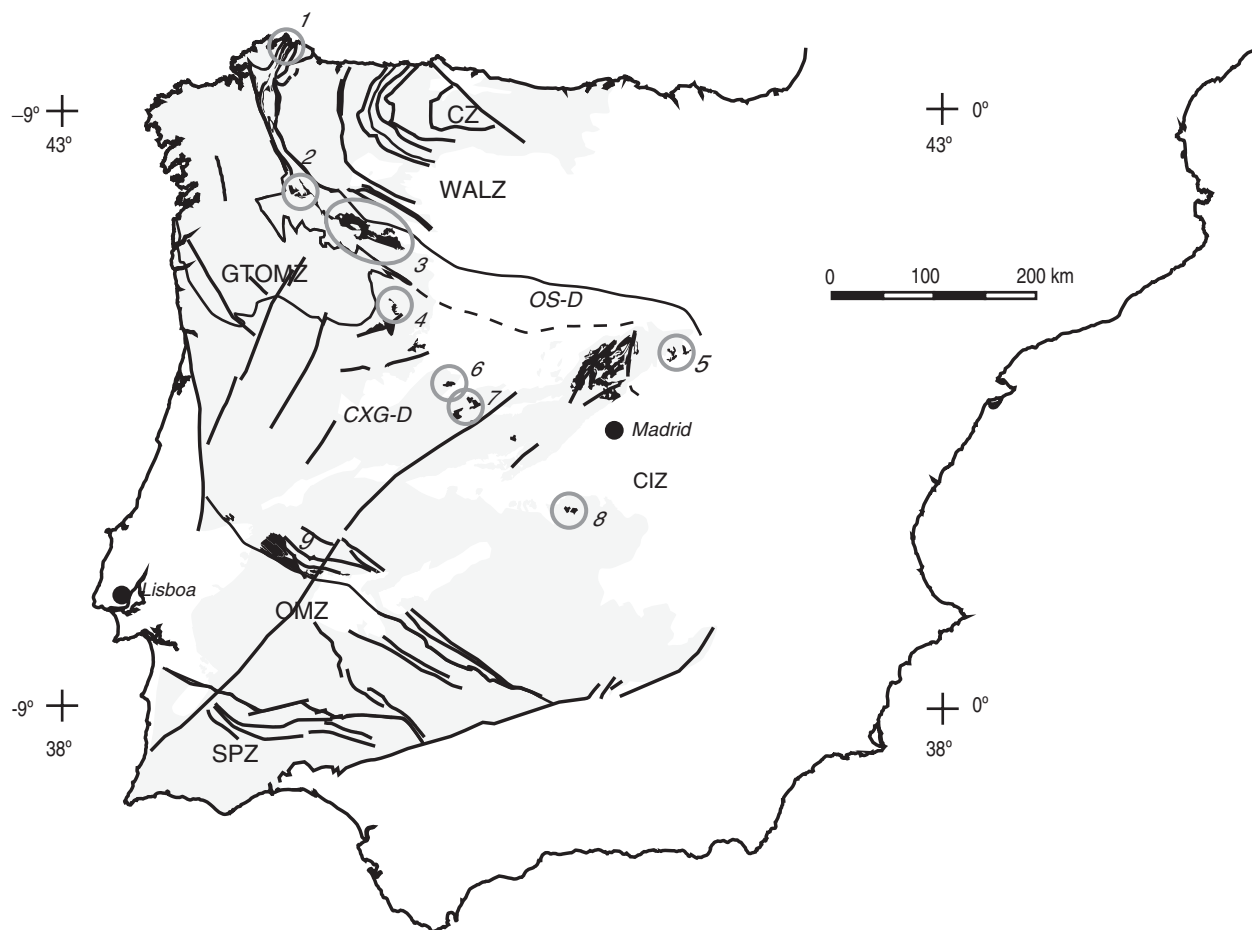
*upper crust could also explain the lack of contemporaneous metamorphism. The speed of the partial melting process resulted in the production of felsic magmas that inherited the geochemical characteristics of their granitoid crustal protolith. This explains the apparent contradiction between the calc-alkaline to peraluminous geochemical characteristics of the magmas and the inferred extensional (i.e. rift-related) tectonic setting. Our model is compatible with the hypothesis of fragmentation and dispersal of terranes from the northern margin of Gondwana that led to the opening of the Rheic and Galicia–South Brittany oceans and, ultimately, caused the detachment of the Iberian microplate from Armorica and Gondwana during the early Paleozoic.*

KEY WORDS: igneous petrology; migmatite; granite; geochemistry; crustal contamination; ICP-MS; laser ablation

## INTRODUCTION

The pre-Variscan basement of the Central Iberian Zone (Fig. 1) contains numerous small bodies of Cambro-Ordovician granitoids and felsic volcanic rocks, which were strongly deformed and variably metamorphosed during the Variscan. These rocks have two notable characteristics. First, they are not obviously connected with any major tectonic or metamorphic event (Gutiérrez-Marco *et al.*, 2002). Second, they have an unusually high proportion of inherited zircon: in the 18 massifs we have studied so far, no less than 70–80%, and in some samples nearer

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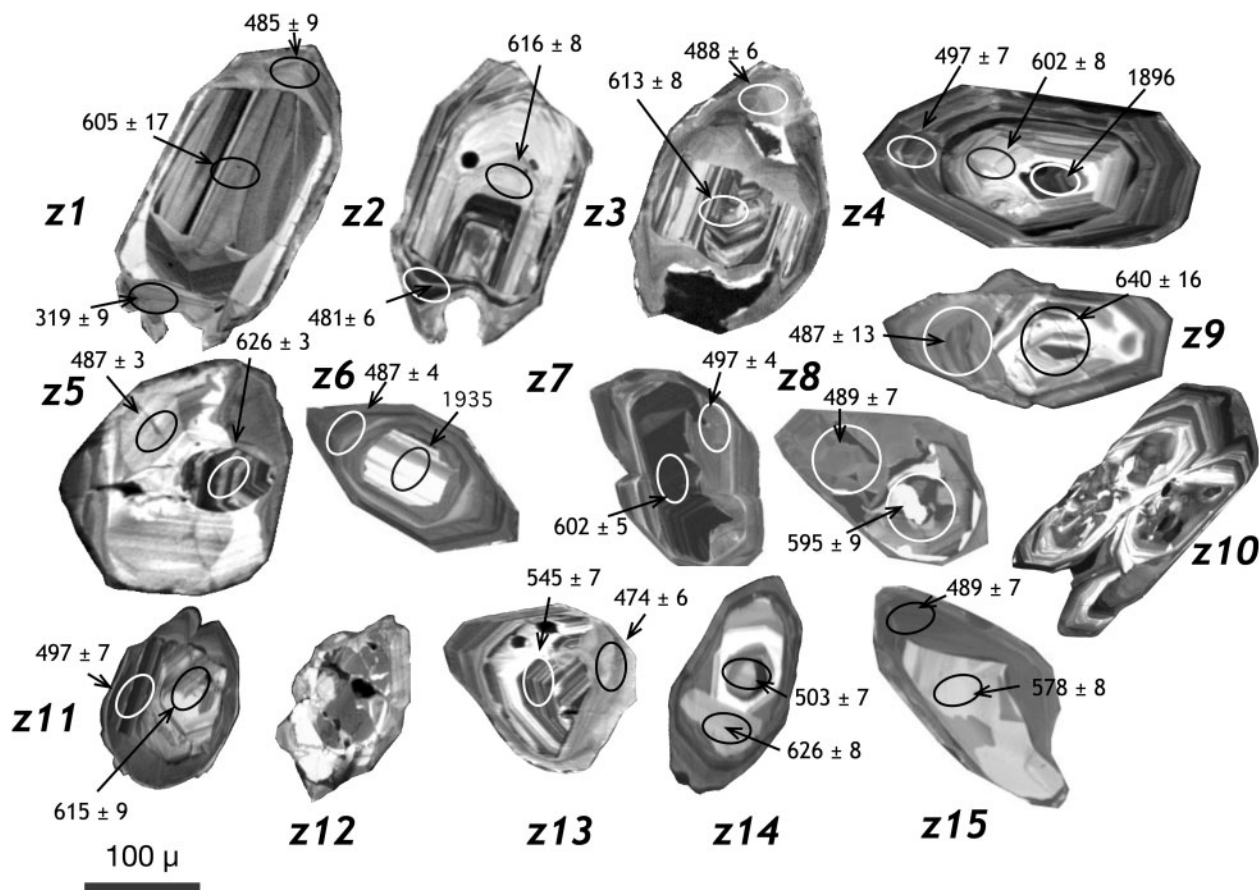
**Fig. 1.** Geological map of the Iberian peninsula showing the location of the Iberian massif, in grey, and the Cambro-Ordovician rocks of the Central Iberian Zone (CIZ) in black. Other peralkaline to peraluminous Cambro-Ordovician rocks occur in the Ossa Morena Zone (OMZ) and the allochthonous terranes of the Galicia-Trans-Os-Montes-Zone (GTOMZ). CZ, Cantabrian Zone; WALZ, Western Asturias-Leon Zone; SPZ, South Portuguese Zone; OS-D, Olla de Sapo Domain; CXG-D, Schist-Graywacke Complex Domain. The massifs enclosed by a grey oval are those studied here (see Fig. 4). 1, El Barquero and Mañón; 2, Puebla de Trives; 3, Sanabria and Viana-Covelo; 4, Villadepera and Miranda do Douro; 5, Hiendelaencina and Antuña; 6, San Pelayo; 7, Bercimuelle and Castellanos; 8, Mochares and Pollán. 9 indicates the Carrascal-Portalegre granitoids and the Urra Formation studied by Solá *et al.* (2005, 2006).

100%, of zircon grains contain pre-magmatic cores (Figs 2–4; see also Solá *et al.*, 2006). This high degree of inheritance, greatly exceeding what is to be expected for common felsic volcanic rocks and granites (e.g. Miller *et al.*, 2003), seems to be a common feature of the Cambro-Ordovician rocks of western Europe (e.g. Laumonier *et al.*, 2004; Teipel *et al.*, 2004; Helbing & Tiepolo, 2005), which must, undoubtedly, reflect some petrogenetic peculiarity of their evolution.

Pre-magmatic zircons survive when the magma temperature is not high enough to dissolve all the available zircon, or when kinetic effects hinder its dissolution. In Iberia, the former can easily be excluded because the peak magmatic temperature recorded by the Cambro-Ordovician rocks exceeds their zircon saturation temperatures (Figs 5 and 6). Among the kinetic factors capable of preventing or delaying zircon dissolution, those associated

with shielding by major phases (Bea, 1996) or with limited volume melt-reservoirs (Watson, 1996) can also be discarded, because these magmas were highly mobile, as indicated by their upper crustal emplacement and lack of restitic material. Most probably, therefore, the elevated zircon inheritance was caused by fast heat transfer to the protolith and fast cooling of the resulting magmas. Understanding how this occurred will lead to a better understanding of the petrogenesis and geodynamic significance of the Cambro-Ordovician magmatism of Central Iberia and, by inference, of western Europe.

The principal objective of this study is to determine the minimum heating and cooling rates that might have caused such a high degree of zircon inheritance. To do this, we used the 3D instant dissolution rate model for spherical zircons in felsic melts developed by Watson (1996, equation 17). This permits calculation of the change



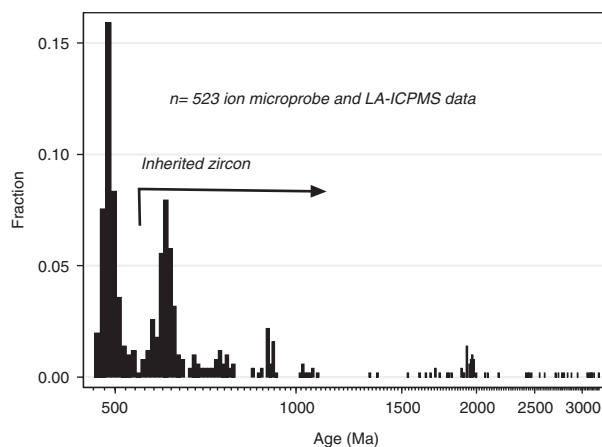
**Fig. 2.** Representative examples of zircons from the Cambrian–Ordovician metagranites and metavolcanic rocks. The numbers indicate the age (in Ma) obtained using either ion-microprobe (ellipses) or LA-ICPMS (circles). The ages of the cores are mainly Ediacaran and the rims Cambro-Ordovician. In some cases there are also narrow, discontinuous overgrowths with Variscan ages (e.g. Z1).

in radius of zircon crystals suspended in a melt as a function of heating and/or cooling rate. To apply Watson's equation using realistic assumptions, we first identified the probable protolith from the ages recorded in the inherited zircon cores. This allowed us to estimate, at least approximately, the bulk composition of the source and the modal abundance, shape and grain-size distribution of pre-magmatic zircons. Then, to estimate a lower limit for the maximum temperature reached by the Cambro-Ordovician magmas, we determined the concentration of Ti and the U–Pb age of zircons by laser ablation inductively coupled plasma mass spectrometry (LA-ICPMS). This permitted us to apply the Ti-in-zircon thermometer (Watson & Harrison, 2005; Watson *et al.*, 2006) and to calculate the crystallization temperature of both neo-formed and inherited zircons. The results of applying Watson's equation indicate that the generation of crustal magma in Central Iberia during Cambro-Ordovician times occurred through repeated fast pulses, each of which lasted no more than a few thousand years. Such felsic magmatism can be explained by a process of mafic

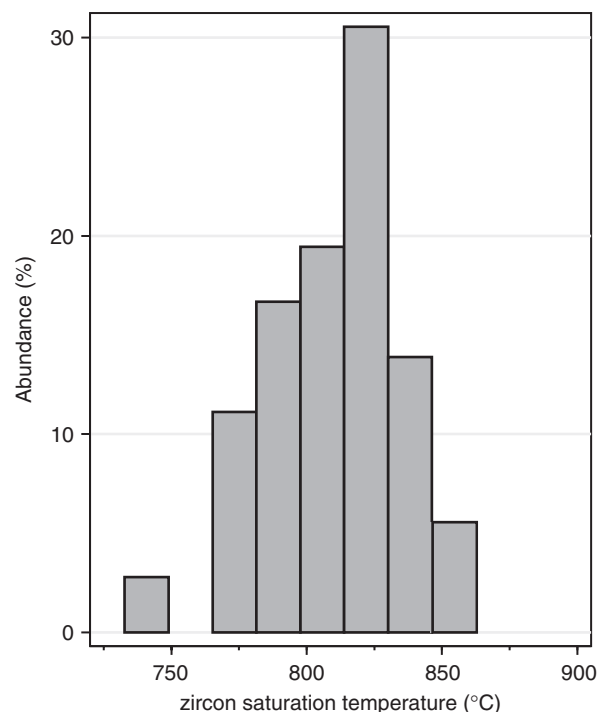
magmas underplating along linear arrays of lithospheric fractures that originated during the separation of the Iberian microplate from Armorica and Gondwana.

## GEOLOGICAL SETTING AND PETROGRAPHY

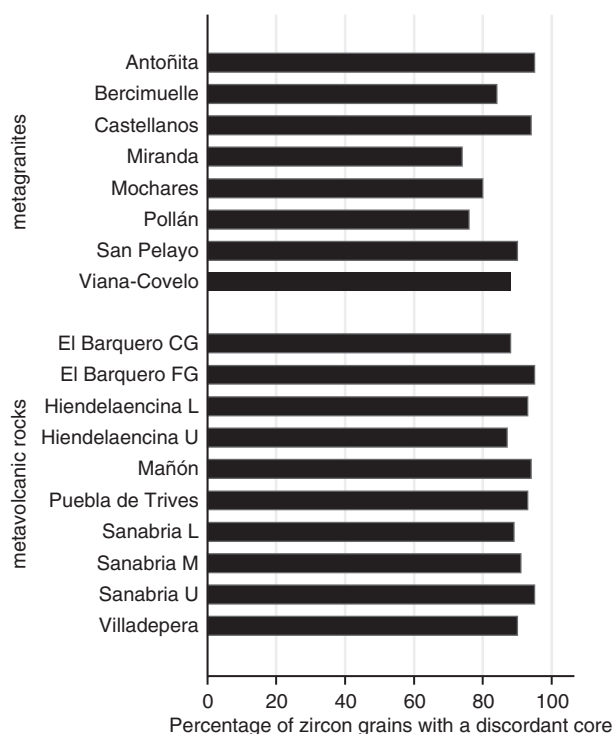
The pre-Variscan basement of the Central Iberian Zone contains three main belts of Cambro-Ordovician igneous rocks (Fig. 1): the metavolcanic rocks and metagranites of the Ollo de Sapo Formation (Parga-Pondal *et al.*, 1964; see a recent overview by Díaz Montes *et al.*, 2004), the metagranites of the northernmost zone of the Schist–Graywacke Domain (e.g. Viallette *et al.*, 1987; Valverde-Vaquero & Dunning, 2000; Bea *et al.*, 2003) and, further south near the boundary with the Ossa Morena Zone, the Carrascal–Portalegre granitoids and the metavolcanic rocks of the Urrea Formation (e.g. Solá *et al.*, 2006). What follows is a short petrographic description of these rocks, largely based on the lithologies of the Ollo de



**Fig. 3.** Histogram of zircon U–Pb concordant or subconcordant ion-microprobe and LA-ICPMS age data for zircon from the Central Iberian Cambro-Ordovician rocks. Most grains were analyzed at the core and the rim. Variscan ages found occasionally in narrow overgrowths of zircons from migmatized samples (e.g. Fig 2, z1) are excluded. The abundance of Precambrian ages and, among these, the vast dominance of Ediacaran (605–615 Ma) ages, should be noted. Data sources: Electronic Appendix I; C. Talavera (unpublished).



**Fig. 5.** Zircon saturation temperatures calculated using the expression of Watson & Harrison (1983) for representative samples of the metavolcanic rocks (22) and the metagranites (19). The two groups have an identical average and therefore are presented together. Data sources: Montero *et al.* (2007) and our unpublished data. All Zr data are XRF determinations on fused discs. The average is 826°C, similar to the 'hot' granites of Miller *et al.* (2003).

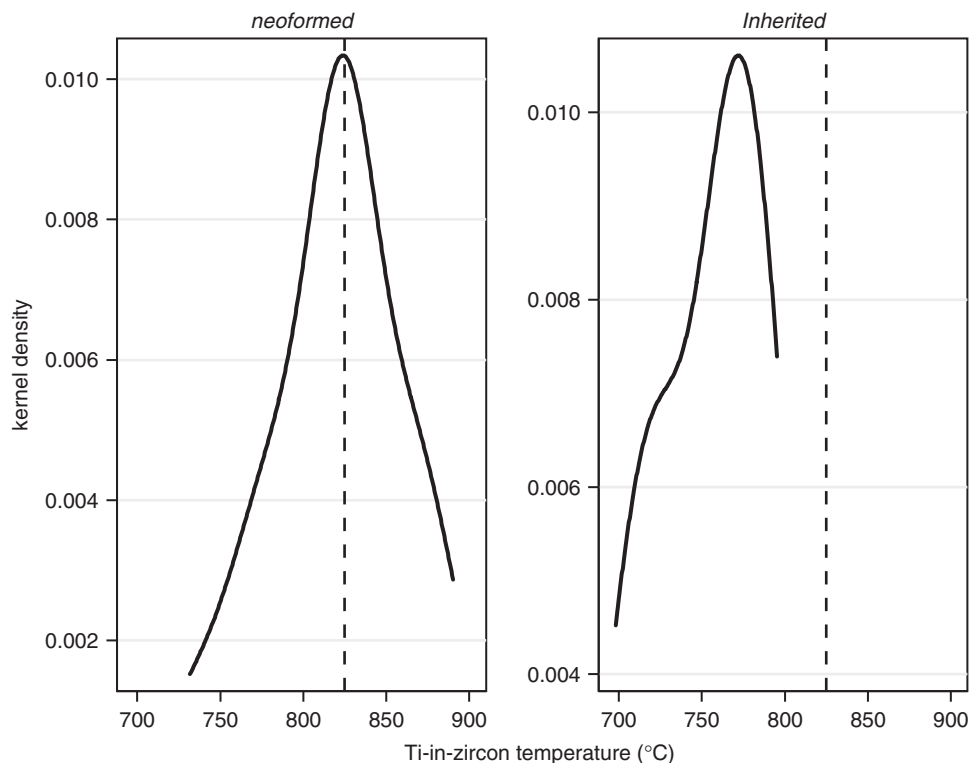


**Fig. 4.** Percentage of zircon grains with pre-magmatic cores in 18 Cambro-Ordovician massifs of Central Iberia, calculated from a minimum of 50 zircon grains from each massif and a total of about 1800 zircon grains. The location of the massifs is shown in Fig. 1. CG, coarse grained; FG, fine grained. L, lower; M, middle; U, upper. This figure indicates the stratigraphic position of the samples in the thicker sequences of the Olla de Sapo Formation.

Sapo Formation. With minor modifications, this is also applicable to the other two belts.

The metavolcanic rocks originally consisted of dacitic to rhyolitic ignimbrites and tuffs (Navidad *et al.*, 1992). They are currently represented by augen-gneisses interbedded with micaceous schists, sandstones and quartzites. Overlying them is a siliciclastic series of Ordovician to Early Devonian age, and underlying them is a mainly metapelitic sequence of probable Early Cambrian age. Petrographically they are distinctive because of the presence of huge (>10 cm) K-feldspar megacrysts, locally with rapakivi structure, euhedral oligoclase phenocrysts (up to 3 cm) and rounded, and frequently embayed, phenocrysts of quartz (up to 1.5 cm), which, when the metamorphic grade is low, sometimes have a noticeable blue color resulting from inclusions of sagenitic rutile. The phenocrysts are surrounded by a fine-grained groundmass of quartz, K-feldspar, muscovite, biotite and rare albite. Accessory minerals include apatite, zircon, ilmenite, magnetite, monazite, rare xenotime and irregularly distributed Fe–Cu sulfides.

The metagranites crop out as small laccoliths or sill-like bodies with a thickness of 300–400 m emplaced within the underlying metasedimentary sequence. They consist of



**Fig. 6.** Kernel density distribution plot of Ti-in-zircon temperatures (Watson & Harrison, 2005; Watson *et al.*, 2006) of neoformed and inherited zircons of the Miranda do Douro orthogneiss. This sample was used because it contains large and inclusion-free zircon grains capable of being analyzed with a 60  $\mu\text{m}$  diameter laser beam. The neoformed grains (22 determinations) peak at 820°C and reach a maximum of 890°C. The inherited grains (nine determinations) peak at 753°C and do not exceed 795°C. The quasi-Gaussian distribution and the absence of outliers indicate that no Ti-rich inclusions were ablated during analysis.

coarse-grained augen-gneisses, locally with abundant aplo-pegmatitic dikes and rare metasedimentary xenoliths. The major mineralogy consists of large crystals of K-feldspar (up to 5–7 cm) frequently with abundant inclusions of oligoclase and biotite, within a coarse-grained groundmass formed of quartz, oligoclase, K-feldspar, biotite, muscovite and occasional tourmaline, cordierite and garnet. The accessory minerals consist of apatite, ilmenite, minor magnetite, zircon, monazite, and rare xenotime and huttonite. The metagranites frequently crop out inside Variscan thermal domes and are, therefore, often noticeably migmatized (Bea *et al.*, 2003). When the migmatization was metatextitic, this caused the development of narrow, discontinuous rims of Variscan age over the Cambro-Ordovician zircons (Fig. 2; Bea *et al.*, 2006b). When the migmatization was diatextitic, the Variscan rims became thicker and newly formed Variscan zircons may appear (Bea *et al.*, 2003).

The crystallization age of these rocks, obtained by U–Pb ion microprobe and LA-ICPMS, and  $^{207}\text{Pb}/^{206}\text{Pb}$  stepwise sequential evaporation, ranges from 496 to 483 Ma for the metavolcanic rocks and from 488 to 474 Ma for the metagranites (Solá *et al.*, 2005, 2006; Bea *et al.*, 2006b;

Montero *et al.*, 2007; F. Bea, unpublished data). Both rock types have chemical composition similar to felsic peraluminous to calc-alkaline igneous rocks with  $\text{K}_2\text{O} > \text{Na}_2\text{O}$  and  $\text{Fe}/(\text{Fe} + \text{Mg}) \approx 0.49\text{--}0.6$ . The metagranites are granodioritic to granitic with aluminium saturation index (ASI)  $\approx 1.07\text{--}1.29$ ,  $^{87}\text{Sr}/^{86}\text{Sr} \approx 0.7048\text{--}0.7112$  and  $\epsilon\text{Nd}(t) -2$  to  $-4$ , and the metavolcanic rocks are rhyodacitic to dacitic with ASI  $\approx 1.21\text{--}1.49$ ,  $^{87}\text{Sr}/^{86}\text{Sr} \approx 0.7069\text{--}0.7118$  and  $\epsilon\text{Nd}(t) -3.5$  to  $-5$  (Montero *et al.*, 2007).

## ZIRCON INHERITANCE, SATURATION TEMPERATURES AND TITANIUM THERMOMETRY

We separated zircon from 18 massifs in the two northernmost belts of Cambro-Ordovician rocks in Central Iberia; three metagranites and 10 metavolcanic rocks from the Ollo de Sapo Domain, and five metagranites from the Schist–Graywacke domain (Fig. 1). These samples have been previously studied by cathodoluminescence (CL) imaging and dated with the U–Pb (ion microprobe and LA-ICPMS) and the  $^{207}\text{Pb}/^{206}\text{Pb}$  stepwise evaporation methods. A complete description of the procedures

employed and the precision attained has been given by Bea *et al.* (2006b) and Montero *et al.* (2007). Solá *et al.* (2006) carried out similar studies on the metavolcanic rocks of the Urro–Portalegre Formation.

The above investigations have revealed an unusually large proportion of inherited zircon crystals. If we accept that discordant cores such as those shown in Fig. 2 are relict (which was confirmed wherever U–Pb spot analyses are available; see also Fig. 3 and Electronic Appendix 1, available for downloading from <http://www.petrology.oxfordjournals.org>), the examination of some 1800 zircons by CL imaging has shown that the proportion of grains with inherited cores varies from 75% to 95% in the metagranites, and from 87% to 95% in the metavolcanic rocks (Fig. 4). To the authors' knowledge, such a high degree of inheritance is common in low melt-fraction migmatites (e.g. Montero *et al.*, 2004) or low-*T* pegmatoid mobilizates (e.g. Gilotti & McClelland, 2005) but is extremely rare in high-level granites or rhyodacitic volcanic rocks.

Miller *et al.* (2003) have demonstrated that zircon inheritance and zircon saturation temperatures ( $T_{Zr}$ ) are antithetic and divided North American granites into 'hot' (average  $T_{Zr}$  = 837°C, little or no inheritance) and 'cold' (average  $T_{Zr}$  = 766°C; high zircon inheritance, most often clustering around 50%) types. With minor modifications, this two-fold categorization seems applicable worldwide, irrespective of granite age and typology, implying that the inverse relation between inheritance and  $T_{Zr}$  is probably a reflection of the conditions that very often occur during granite petrogenesis. Remarkably, the Central Iberian Cambro-Ordovician igneous rocks do not follow this rule, because they have simultaneously an average  $T_{Zr}$  of 826°C (Fig. 5), characteristic of 'hot' granites, and a zircon inheritance that not only matches but nearly duplicates that of 'cold' granites. This unusual combination suggests disequilibrium processes that do not normally occur in granite magmas.

To understand the atypical processes involved in Iberian Cambro-Ordovician magma generation, we must consider the following. Zircon solubility in common crustal melts depends on melt composition and temperature, but it is almost independent of pressure and water content (Watson & Harrison, 1983). Because the variations to be expected in the melt bulk-composition have less influence than the variations to be expected in temperature (e.g. melts with 68 wt % and 72 wt % SiO<sub>2</sub> dissolve 148 and 131 ppm Zr at 800°C, but 253 and 223 ppm Zr at 850°C), the first hypothesis that might explain the abnormally elevated zircon inheritance of the Cambro-Ordovician magmas is that they were never sufficiently hot to dissolve all the zircon grains entrained from the source, these being exceptionally abundant for some unspecified reason.

This hypothesis can be evaluated as follows. Under equilibrium conditions, the temperature for total zircon dissolution in a magma roughly corresponds to  $T_{Zr}$ , which, in the present case, averages 826°C and does not surpass 870°C (Fig. 5). The minimum temperature attained by the magma, on the other hand, can be estimated by applying the Ti-in-zircon thermometer (Watson & Harrison, 2005; Watson *et al.*, 2006) to the Cambro-Ordovician rims of the zircon grains. To this end, we analyzed the low-abundance (5.5%) but interference-free <sup>49</sup>Ti isotope, plus <sup>238</sup>U, <sup>207</sup>Pb and <sup>206</sup>Pb (to ascertain the age) and <sup>92</sup>Zr and <sup>29</sup>Si (as internal standards) with a LA-ICPMS system, ablating 60 µm diameter spots. Details of the analytical procedure have been given by Bea *et al.* (2006a). The study was carried out on a metagranite, the Miranda do Douro orthogneiss (Bea *et al.*, 2006b), which has the largest and most inclusion-free zircon crystals of all the studied bodies. In all other samples, zircon grains were either too small or too inclusion-rich for reliable Ti analysis. The results of the Miranda do Douro study (Fig. 6) reveal that whereas the crystallization temperatures of the inherited zircons peak at about 770°C and never surpass 800°C, the neoformed Cambro-Ordovician grains peak at 820°C and some values approach 900°C, which would then represent the minimum temperature reached by the magmas. If we accept these results (see discussion), we must conclude that the temperature attained by the Cambro-Ordovician magmas was certainly high enough to dissolve all entrained zircons. The reasons why so many of them survived should be, therefore, related to the zircon dissolution kinetics.

In a melting protolith the main factors delaying the dissolution of zircon are: shielding by major phases, small and isolated melt reservoirs, and fast heating and cooling rates (Watson, 1996). As the Iberian Cambro-Ordovician magmas were highly mobile, capable of eruption at the Earth's surface, the melt fraction should have been high enough to ensure total interconnectivity of the melt pores (i.e. the system behaved as an infinite melt reservoir). Similarly, the temperatures recorded by zircon indicate that biotite, which is the major mineral with the greatest tendency to include accessories (Bea, 1996), was involved in melting reactions so that the fraction of zircon shielded from the melt must have been low. Consequently, the only acceptable explanation for the observed high zircon inheritance is incomplete dissolution owing to the short life-span of the magmatic pulses. Before trying to determine the duration of these with Watson's equation, it is necessary to have an idea about the nature and composition of the protolith.

## THE NATURE OF THE PROTOLITH

The distribution of 523 concordant or nearly concordant ion-microprobe and LA-ICPMS ages for the

Cambro-Ordovician rocks of Central Iberia (Fig. 3) shows a polymodal distribution with a mode at *c.* 490 Ma that we interpret as the age of crystallization, a second mode at *c.* 610 Ma that we interpret as the age of the predominant protolith, and some minor modes at older ages. The marked dominance among the inherited components of the *c.* 610 Ma age, as well as the chemical and Sr and Nd isotope bulk-rock composition, led Montero *et al.* (2007) to suggest that the protolith of the Cambro-Ordovician magmas of Iberia mostly consisted of intermediate to felsic calc-alkaline igneous rocks of late Pan-African age (or younger, immature sediments derived from them). As a result of vigorous Variscan crustal reworking, these rocks are poorly exposed in Iberia, except for some small dioritic to granitic massifs in the Merida region (Bandrés *et al.*, 2004). Calc-alkaline plutonism at 615 Ma, however, was one of the most important magmatic events of the neighboring Anti-Atlas region of Morocco (Gasquet *et al.*, 2005), a region attached to Iberia during the Ediacaran (e.g. Ennih & Liegeois, 2001, 2003) that was subsequently little affected by the Variscan orogeny. Merida and Morocco granitoids can, therefore, give an idea, at least approximately, of the grain-size distribution of zircon, the Zr concentration and the bulk-rock composition of the protolith, which are needed for application of Watson's equation.

The data of Bandrés *et al.* (2004) indicate that the Pan-African granitoids of Merida contain zircon grains with maximum dimensions of  $270 \times 100 \times 100 \mu\text{m}$  (a volume equivalent to a  $86 \mu\text{m}$  radius sphere) and have an average concentration of Zr of 125 ppm. This value, however, is probably underestimated because the samples were analyzed after acid digestion; our X-ray fluorescence (XRF) data for the same rocks reveal concentrations between 150 and 270 ppm Zr. Additionally, the *c.* 600 Ma granites of Morocco, with average XRF Zr concentrations of 227 ppm, contain zircon grains with dimensions most frequently around  $150 \times 70 \times 50 \mu\text{m}$  (volume equivalent to a  $56 \mu\text{m}$  radius sphere) and only the largest ones reach  $300 \times 120 \times 90 \mu\text{m}$  (volume equivalent to a  $92 \mu\text{m}$  radius sphere) (unpublished data of the authors). Neither the zircon grain size nor the Zr concentration of these rocks is exceptional, but instead both are close to what one would expect for common granodiorites and granites.

## ESTIMATION OF LIMITS FOR HEATING AND COOLING RATES FROM WATSON'S EQUATION

Watson's equation (1996, equation 17) for calculating the instant dissolution rate of spherical zircon crystals is

$$(dr/dt) \times 10^{17} = -U \left[ (1 \cdot 25 \times 10^{10}/r) \exp(-28380/T) + 7 \cdot 24 \times 10^8 \exp(-23280/T) \right]$$

where  $dr/dt$  is the instant dissolution rate (cm/s),  $r$  is the radius of a spherical zircon crystal (cm),  $T$  is the absolute temperature (K), and  $U$  is the difference between the current Zr concentration of the melt and the concentration required for zircon saturation according to the experimental model of Watson & Harrison (1983). In partially molten systems, calculating  $U$  requires knowledge of the Zr concentration and bulk-rock composition of the protolith, and the volume of the melt reservoirs.

The equation can be used stepwise to calculate the variations of the zircon radius as a function of time for a given heating (or cooling) gradient. This requires inputting the new  $T$ , the new zircon radius that resulted from the previous step, and the new  $U$  value calculated considering the amount of zircon dissolved in the previous step and the change in zircon solubility caused by the variation in  $T$  and melt composition.

## Heating rates

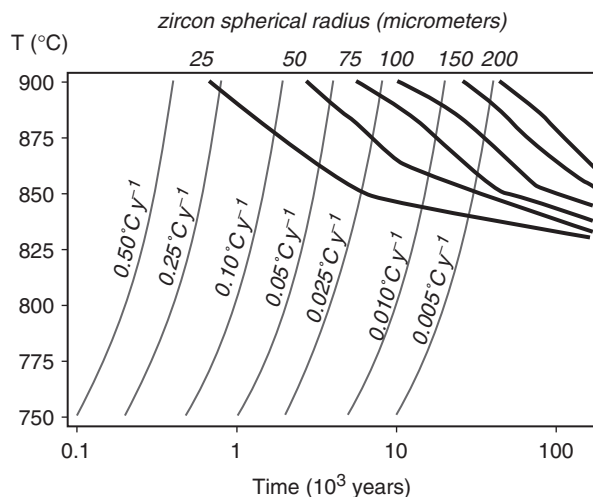
We applied the equation to a melting protolith similar in composition to the Pan-African rocks described in the previous section using the following initial conditions:

- (1) beginning of melting occurs at  $700^\circ\text{C}$ ;
- (2) maximum temperature reached by the magma is  $900^\circ\text{C}$ ;
- (3) volume of melt reservoirs is infinite;
- (4) major element composition of the melt (to calculate zircon solubility) is calculated using the equations of Winther (1995) for a granodioritic protolith at 8 kbar and 2%  $\text{H}_2\text{O}$ ;
- (5) Zr concentration in the protolith (residing only in zircon) is 225 ppm.

Figure 7 shows the calculated time–temperature coordinates at which spherical zircons with radius of 25, 50, 75, 100, 150 and  $200 \mu\text{m}$  will dissolve totally in the melt as a function of the heating rate. The following features stand out. Zircons with a spherical radius of  $50 \mu\text{m}$  (i.e. with a volume similar to that of the population most commonly found in the protolith) would survive only if heating occurred at a rate of  $0.1^\circ\text{C}/\text{year}$  or higher. Zircons with a radius of  $100 \mu\text{m}$  (i.e. with a volume larger than the largest zircons of the protolith) would survive only if the heating rate was  $0.025^\circ\text{C}/\text{year}$  or higher. In summary, the survival of protolith zircons requires less than 10 000 years, probably around 2000 years, from the beginning of melting ( $700^\circ\text{C}$ ) to the thermal peak ( $900^\circ\text{C}$ ).

## Cooling rates

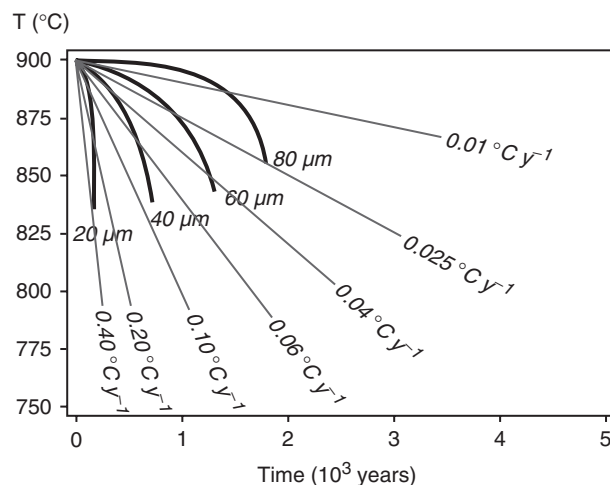
The above calculations represent only half of the history: zircon grains that were partially dissolved during heating will continue to dissolve during cooling as long as the melt does not become zircon saturated. Therefore, we calculated again the time–temperature coordinates at which spherical zircon survivors with radius of



**Fig. 7.** Disappearance curves (bold continuous lines) of spherical zircon grains as a function of the heating rate (fine continuous lines) calculated with Watson's equation. (See text for the calculation parameters.) It should be noted that heating rates are represented as curves because the horizontal coordinate (time) is logarithmic. The intercept of a heating rate curve with the disappearance curve of zircon with a given radius marks the point at which that zircon will be totally dissolved into the melt. For example, zircons with a spherical radius of 50  $\mu\text{m}$  would not dissolve totally if the heating rate is 0.10 °C/year, but would dissolve if the heating rate is 0.05 °C/year or less. Remarkably, zircons with a radius of 100  $\mu\text{m}$ , which is larger than the largest found in the probable protolith of the Cambro-Ordovician magmas of Central Iberia, will dissolve if the heating rate is slower than 0.025 °C/year; that is, if heating from the beginning of melting (700 °C) to the thermal peak (900 °C) occurred in about 10<sup>4</sup> years. (See text for discussion.)

20, 40, 60 and 80  $\mu\text{m}$  will totally dissolve in the melt as a function of the cooling rate. To estimate the Zr concentration in the melt at 900 °C we considered that all protolith zircon grains had a spherical radius of 100  $\mu\text{m}$  and that, during heating, the radius of the dissolving crystal was coupled to the Zr concentration of the melt so that the total concentration of Zr in the system (melt plus crystals) was always equal to that of the protolith: 225 ppm. The results are shown in Fig. 8, from which it follows that the survival of 20, 40, 60 and 80  $\mu\text{m}$  zircons requires linear cooling rates from 900 °C to 830 °C faster than 0.4, 0.1, 0.05 and 0.025 °C/year, respectively. As the cooling rates of erupted felsic magmas are often much faster than these (e.g. Harris *et al.*, 2002), we can conclude that most survivors would have a chance of cooling with little size reduction if the magma crystallized quickly by being rapidly transported from the melting region to Earth's surface.

The situation, however, is different if the magmas crystallized at depth, such as in the case of the metagranites. According to González Lodeiro (1981), Iglesias Ponce de León & Ribeiro (1981) and Lancelot *et al.* (1985), these rocks occur as high-level sills or laccoliths about 300–400 m thick. To assess the behavior of zircon under these conditions we calculated the 1D cooling paths of



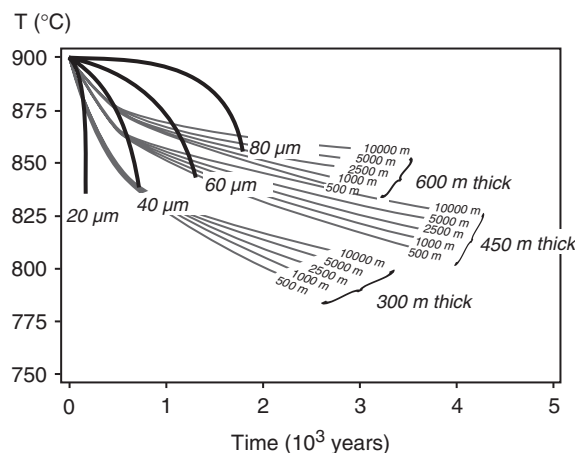
**Fig. 8.** Disappearance curves (bold continuous lines) of spherical zircon survivors as a function of the cooling rate (fine continuous lines) calculated with Watson's equation. (See text for the calculation parameters.) As before, the intercept of a given cooling rate curve with the disappearance curve of zircon with a given radius marks the point at which that zircon will be totally dissolved into the melt. For example, zircons with a spherical radius of 80  $\mu\text{m}$  would not dissolve totally if the cooling rate is 0.025 °C/year or faster. In the rapid cooling regime of volcanic conditions even the smallest zircon grains would not dissolve during cooling.

granitic sills with the same initial  $T$  of 900 °C and different thickness (300, 450 and 600 m), which were emplaced at different depths (500, 1000, 2500, 5000 and 10 000 m), and compared them with the curves of zircon disappearance during cooling as calculated with Watson's equation.

The results of this calculation are shown in Fig. 9, from which the following features stand out. During cooling of a 300 m thick body, independent of the depth of intrusion (to 10 km), all zircons with a radius larger than 40  $\mu\text{m}$  will survive. If the sill thickness increases to 450 m, only zircons with a radius larger than 70  $\mu\text{m}$  would survive. If the sill thickness increases to 600 m, even the zircons with a radius of 80  $\mu\text{m}$  will disappear unless the sill is emplaced at a depth less than 2 km. It seems, therefore, that the critical parameter governing zircon survival in granite magmas emplaced in the upper crust is the thickness of the magmatic body, with the depth of intrusion seemingly playing a secondary role.

## DISCUSSION

The applicability of the above calculations to geological systems depends on the validity of the numerical model of zircon dissolution, the proper choice of the initial conditions and model parameters, and the deviations caused by (1) the residence of Zr in minerals other than zircon and (2) the variable grain size and non-spherical shape of zircon crystals. These circumstances can be evaluated as follows.



**Fig. 9.** Disappearance curves (bold continuous lines) of spherical zircon survivors as a function of the cooling rate calculated as in Fig. 8, compared with the 1D cooling curves of granitic sills of different thickness emplaced at depths from 500 to 10 000 m. It should be noted that even 40  $\mu\text{m}$  zircons can survive if the sill thickness is 300 m. In 600 m thick sills, 80  $\mu\text{m}$  zircons can survive if the sill was emplaced at a depth of 2000 m or less.

Watson's (1996) equation relies on two factors: (1) zircon solubility in silicic melts, which for non-peralkaline liquids depends primarily on the temperature and the melt major-element composition (Watson & Harrison, 1983); (2) Zr diffusion in the melt, which depends additionally on the  $\text{H}_2\text{O}$  content of the melt. The equation treats temperature and melt composition as independent variables (the latter for calculating zircon solubility), but it assumes a constant 3%  $\text{H}_2\text{O}$  for the melt. In principle, this assumption might be a serious limitation. In practice, however, it does not critically affect the model; first, because crustal magmas rarely have less than 2–3%  $\text{H}_2\text{O}$  (e.g. Clemens, 1984; Carrington & Harley, 1996), and, second, because the effects on Zr diffusion mostly occur in the first 2–3% dissolved  $\text{H}_2\text{O}$  (Harrison & Watson, 1983).

It should also be considered that Watson's equation is a simplification. According to Watson, however, it deviates <10% from the results of more rigorous, moving boundary, finite-difference methods (Watson, 1996). Therefore, as this deviation is tolerable for our purposes and nothing indicates that the Cambro-Ordovician magmas of Central Iberia were exceptionally  $\text{H}_2\text{O}$ -poor, we can accept the numerical foundation for calculating the curves of zircon disappearance plotted in Figs 8 and 9.

The choice of a Pan-African protolith is justified by the dominance of inherited 600–620 Ma ages (Fig. 3). Moreover, this selection is not critical because the zircon grain-size distribution and the Zr concentration estimated for this protolith are typical for common granitoids. Only if the protolith had zircons with a spherical radius larger than 150  $\mu\text{m}$ , or a bulk-rock concentration of Zr greater than 400–500 ppm, would we expect significant

departures from the model. Because such features are mostly limited to peralkaline rocks, and these are unknown among the 600–620 Ma magmatism of NW Gondwana (e.g. Gasquet *et al.*, 2005), we can safely exclude this possibility. Neither is the major-element composition of the melt critical, because it was necessarily silicic and variations of less than 5%  $\text{SiO}_2$  have little effect.

In contrast, the determination of the peak temperature reached by the melts is crucial, especially if it is overestimated. The Ti-in-zircon thermometer requires  $\text{TiO}_2$  activity equal to one (Watson *et al.*, 2006). In the present case, the presence of primary ilmenite and rutile inclusions indicates that such a condition is satisfied. If not, it would have caused underestimation, which would not invalidate our conclusions but, instead, indicate even faster heat-transfer rates. More important, perhaps, is that Ti-in-zircon temperatures can be easily overestimated if there are minute inclusions of Ti-bearing minerals or glass within the analyzed volume. The problem is especially serious when Ti is determined using a LA-ICPMS system such as the one used here, which to obtain reasonable  $^{49}\text{Ti}$  sensitivity requires ablation of craters with a diameter of 60  $\mu\text{m}$  and a depth of about 40  $\mu\text{m}$ . Nevertheless, the careful selection, under the microscope, of the areas to be analyzed and, especially, the nearly Gaussian distribution of the results, notably exempt of outliers (see Fig. 5), indicates that inclusions have caused little trouble in the present case. Consequently, we can assume that the maximum temperatures recorded by the Ti-in-zircon thermometer (900°C) represent a minimum estimate of the magma's thermal peak, an assumption totally consistent with a large body of experimental data indicating that temperatures of this order are required for generating large volumes of silicic crustal magma in vapor-absent conditions (e.g. Clemens, 2003, and references therein).

A final consideration is that Watson's equation assumes that all zircon grains are spheres of the same size, and that all Zr resides in zircon. Real rocks, however, have non-spherical zircon crystals, and these are of different shapes and sizes. Real rocks also have a variable fraction of Zr residing in minerals other than zircon, such as titanite, amphibole or garnet (e.g. Bea *et al.*, 2007). Certainly, all these differences may affect the zircon dissolution rate. Shapes other than a sphere would increase it, as a sphere represents the smallest surface/volume ratio. Nevertheless, the existence of a large variety of zircon sizes could delay the dissolution of the largest grains because the Zr concentration in the melt would increase rapidly owing to the fast dissolution of the smallest grains. Zircon dissolution would also be delayed if phases other than zircon release Zr to the melt but, on the other hand, it would be accelerated if a Zr-bearing mineral such as garnet appears as a product of melting reactions and extracts Zr from the melt. However, on balance, none of

these effects are likely to be of great importance and, what is more, they tend to mutually cancel, so we can accept that the results of Watson's equation acceptably match the real situation; that is, that the magmatic pulses that generated the Central Iberian Cambro-Ordovician magmas were very fast, probably taking around 2000 years from the beginning of melting until their eruption or emplacement as thin sills or laccoliths at upper crustal levels.

## GEODYNAMIC IMPLICATIONS

Rapid melt generation and crystallization of the Cambro-Ordovician magmas constrains the possible geodynamic setting in which they were formed, a highly controversial matter not only in Iberia but throughout the European Variscides (e.g. Crowley *et al.*, 2000). In Iberia, apart from the peraluminous to calc-alkaline rocks described here, which mostly occur in the Central Iberian Zone, there are several small massifs of peralkaline granitoids and a few gabbros that are restricted to the Ossa Morena Zone and the allochthonous complexes of the Galicia Tras-os-Montes Zone. Whereas there is a general agreement that the peralkaline rocks and associated gabbros originated in a rifting environment (e.g. Ribeiro, 1987; Ribeiro & Floor, 1987; Santos Zalduegui *et al.*, 1995; Montero *et al.*, 1998; Montero & Floor, 2004), the peraluminous to calc-alkaline rocks of Central Iberia, solely by virtue of their chemical signature, have been interpreted by several workers as evidence of an active margin setting (e.g. Gebauer *et al.*, 1993; Valverde-Vaquero & Dunning, 2000; von Raumer *et al.*, 2003).

However, the link between the geochemical signature and geodynamic setting is not definitive and may be equally explained as a legacy from their protoliths, as proposed for the Cambro-Ordovician rocks of the northern Bohemian Massif by Klimas-August (1990) and Floyd *et al.* (2000). In the present case, the fast melting and magma-transport rates inferred from the elevated zircon inheritance are enough to cause that effect: first, because the short duration of the whole process would surely have negatively affected the efficiency of melt–restite segregation, especially if there were no syn-magmatic deformation (e.g. Bea *et al.*, 2005); second, because the fast melting rates cause the effective partition coefficients to converge to one, despite their equilibrium values (Bea, 1996). In these conditions, is not surprising that both the chemical and isotopic signature of the resulting magmas would be close to that of their late Pan-African protoliths and, therefore, useless for geodynamic discrimination purposes.

The geodynamic environment proposed for the Cambro-Ordovician magmas must be compatible with the generation of crustal melts at the elevated rates inferred here and, at the same time, account for the absence of any perceptible orogenic event. With respect to the first point, it should be considered that the only heating

mechanism capable of melting crustal materials at the required rate is the advection of heat by mafic magmas, as revealed by the numerical analysis of Huppert & Sparks (1989). Other crustal-heating mechanisms have much larger time constants, from  $10^5$ – $10^6$  years for the displacement of isotherms caused by tectonics, burial or erosion (e.g. Chapman & Furlong, 1992; Zen, 1995; Huerta *et al.*, 1998) to  $(1\text{--}3) \times 10^7$  years for radiogenic heating (e.g. Vanderhaeghe & Teyssier, 2001; Bea *et al.*, 2003). The analysis of Huppert & Sparks (1989) also predicts that felsic magmas generated following the intrusion of mafic magmas would have peak temperatures of 900°C, abundant pre-magmatic crystals and a highly porphyritic character, a set of features found in the Central Iberia Cambro-Ordovician rocks that are difficult to explain by any other mechanism. This gives additional support to the idea that heat for crustal melting was supplied by mantle-derived mafic magmas. Additionally, the imperceptible hybridization between these and the felsic magmas (e.g. Montero *et al.*, 2007) suggests that the meltable crustal material just overlay the mafic intrusions, where the density difference and the quick solidification of the mafic magma at the contact would make mixing unlikely (Huppert & Sparks, 1989). This locates the mafic intrusions at the crust–mantle interface. Lastly, the fast melt transport to upper crustal levels points to extensional rather than compressional forces. All these reasons, therefore, strongly suggest that the Central Iberian Cambro-Ordovician magmas were generated during the rifting of continental crust caused by an upwelling mantle plume, which probably occurred during the early Paleozoic fragmentation and dispersal of terranes from the northern margin of west Gondwana (Crowley *et al.*, 2000; Matte, 2001) and ultimately led to the formation of the Iberian microplate.

In this scenario crustal deformation depends to a significant extent on the mechanical coupling between mantle and crust (Burov & Guillou-Frotier, 2005); if the coupling is weak, the concentration of plume-related extension in the mantle lithosphere has little effect on the crust. This might explain the scarcity of contemporaneous deformation. The lack of any Cambro-Ordovician metamorphic imprint in all exposed midcrustal sections of Central Iberia may also be explained by the swiftness of the generation and emplacement of the crustal magmas. Metamorphism involves conductive heat transfer, a process inherently slow that requires much more than a few thousand years to be perceptible at a crustal scale. Here, the heat advected to the lower crust by mantle magmas was first consumed by melting reactions and then quickly transported to the uppermost crustal sections by the so-produced magmas, thus causing a negligible thermal impact on most of the crustal section above the melting zone except, perhaps, the lowermost 1000–2000 m.

## CONCLUSIONS

The most important conclusions of this paper can be summarized as follows.

The Cambro-Ordovician igneous rocks of Central Iberia, dacites to rhyolites and high-level granites, contain about 70–80%, and in some samples nearer 100%, of zircon grains with inherited pre-magmatic cores. The elevated zircon survival occurred despite the fact that peak temperature of the magmas, estimated with the Ti-in-zircon thermometer at 900°C or higher, surpassed the rock's zircon saturation temperature. This was the result of the swiftness of the magmatic pulses. Modeling the dissolution of zircon suspended in a melt as a function of heating and cooling rates indicates that the pulses lasted only a few thousand years, probably about 2000 years, from the beginning of melting to final emplacement.

Considering the time constants involved, these rates of crustal recycling can only be achieved by anatexis induced by the intrusion of hot mantle-derived magmas into the crust. This mechanism also explains why the crustal magmas are highly porphyritic and have reached peak temperatures of 900°C or higher. The imperceptible hybridization between mafic and felsic magmas indicates that the locus of the mafic intrusions was at the crust–mantle interface. The fast melt transport to upper crustal levels points to extensional rather than compressional forces, and explains the negligible metamorphic imprint of this event on mid-crustal sections.

The calc-alkaline to peraluminous signature of the Cambro-Ordovician magmas, which has been considered as proof of a subduction environment, was inherited from their late Pan-African protolith owing to the swiftness of the melt-generation process. First, the short time involved led to a low efficiency of felsic melt–restite segregation, especially in the absence of syn-magmatic deformation. Second, the fast melting rates led the effective partition coefficients to depart from their equilibrium values and converge to one. Accordingly, the geochemical signature of the resulting magmas cannot be invoked as a proof of a subduction setting.

The most probable setting for the generation of the Central Iberian Cambro-Ordovician magmas is a continental rifting environment in which crust and mantle were mechanically decoupled, as indicated by the scarcity of contemporaneous deformation. Crustal melting was triggered by the intermittent arrival of batches of mafic magmas at the mantle–crust interface along linear arrays of lithospheric fractures. The heat advected to the lower crust was first consumed by melting reactions and then quickly transported to the uppermost crustal sections by the so-produced magmas, causing negligible metamorphism of the crustal section above the melting zone except in the first 1000–2000 m.

Our interpretation is in good agreement with the idea of fragmentation and dispersal of terranes from the northern margin of west Gondwana during the early Paleozoic caused by, among other factors, an upwelling mantle plume (Crowley *et al.*, 2000), which led to the opening of the Rheic Ocean and Galicia–South Brittany oceans (Matte, 2001) and, ultimately, detached Iberia from Gondwana and Armorica.

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## SUPPLEMENTARY DATA

Supplementary data for this paper are available at *Journal of Petrology* online.

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