Composite origin of an early Variscan transported suture: Ophiolitic units of the Morais Nappe Complex (north Portugal)

C. Pin,[1] J. L. Paquette,[1] B. Ábalos,[2] F. J. Santos,[3] and J. I. Gil Ibarguchi[3]

Received 17 March 2006; revised 15 June 2006; accepted 26 June 2006; published 2 September 2006.

[1] The two structural units that constitute the ophiolitic complex of the Morais allochthonous complex display different geochemical features. Amphibolites from the lower unit are nearly homogeneous, chemically similar to normal mid-ocean ridge basalts, with low Th/Nb and highly radiogenic Nd isotope signature (εNd = +8.7), and might derive from igneous protoliths formed in an oceanic ridge setting 447 ± 24 Myr ago (Sm-Nd whole rock isochron). In contrast, the upper unit consists largely of metaperidotites together with gabbroic rocks and minor amphibolites and felsic veins. U-Pb zircon ages of the upper unit range from 405 ± 1 Ma to 396 ± 1 Ma. These rocks are geochemically more variable but share in common high Th/Nb ratios and high initial εNd (from +7.5 to +8.9) interpreted to reflect generation above an intraoceanic subduction zone. This implies that the Morais ophiolitic complex is composite and consists of two distinct, relatively thin tectonic slivers that were detached from contrasting oceanic domains before being finally imbricated. This occurred broadly coeval, at the regional scale, with eclogite-facies metamorphism documented in several separate tectonic units of NW Iberia. On the basis of this space and time association, we interpret the early stage of the Variscan (in the broad sense) evolution (circa $\geq$ 405–400 Ma) in terms of intraoceanic subduction, followed by an arc-continent collision, circa 395 Ma. This model may account for the long-recognized polycyclic evolution of the north Portuguese catazonal complexes without resorting to the existence of a high-grade Precambrian basement. **Citation:** Pin, C., J. L. Paquette, B. Ábalos, F. J. Santos, and J. I. Gil Ibarguchi (2006), Composite origin of an early Variscan transported suture: Ophiolitic units of the Morais Nappe Complex (north Portugal), *Tectonics*, *25*, TC5001, doi:10.1029/2006TC001971.

[1]Département de Géologie, Centre National de la Recherche Scientifique, Université Blaise Pascal, Clermont-Ferrand, France.
[2]Departamento de Geodinámica, Universidad del País Vasco, Bilbao, Spain.
[3]Departamento de Mineralogía-Petrología, Universidad del País Vasco, Bilbao, Spain.

0278-7407/06/2006TC001971

1. Introduction

[2] The proper identification and dating of oceanic lithosphere remnants provides first-order constraints on the plate tectonic interpretation of collisional orogens. However, complete ophiolite sequences, commonly interpreted to represent oceanic lithosphere, formed at mid-ocean ridges or suprasubduction environments, are relatively rare in ancient orogens. For example, large and thick ophiolitic nappes are notoriously absent from the European Variscan Belt, where only sparse and usually highly dismembered units, reminiscent of true ophiolites, have been found [e.g., *Pin*, 1990]. In part, this might be due to erosion, which removed most of the higher structural levels, including oceanic units obducted during early collisional episodes. Also, a strong tectonic and thermal overprinting occurred during the major orogenic stage in the Carboniferous, which obscured or even erased original features in most of the internal zones of the orogen.

[3] The Trás-os-Montes (NE Portugal) region of the NW Iberian Massif offers an opportunity to study early stage processes associated with ophiolite generation and emplacement. In this area, high-grade polymetamorphic complexes are preserved in the core of two late stage Variscan synforms at Bragança and Morais. These are interpreted as erosional remnants of a large thrust sheet emplaced onto low-grade Silurian sediments [*Ribeiro et al.*, 1964; *Ries and Shackleton*, 1971; *Anthonioz*, 1972]. Imbricated between the high-grade upper units and their Silurian substratum, an intermediate unit composed of prograde amphibolite-facies mafic rocks occurs in the Bragança and the Morais synforms and has been referred to as ophiolitic unit [e.g., *Ribeiro et al.*, 1990] or, in this study, ophiolitic complex [after *Pereira*, 2001; *Pereira et al.*, 2004]. Available Ar/Ar data suggest that this pile of tectonic units was brought up to shallow levels, and remained in the upper crust from the Mid-Devonian (circa 385 Ma) onward [*Dallmeyer et al.*, 1991]. This implies that the upper part of the nappe stack behaved as a rigid body, and that transport onto the Variscan foreland took place in a piggyback style through the activation of thrusts in the para-autochthon [*Marques et al.*, 1996]. For that reason, Variscan overprinting (in the 360–330 Ma time span) is believed to be very weak or even absent in the upper units, thereby providing a good opportunity to study the early geodynamic processes, including petrogenesis of the inferred oceanic slivers. As yet, the ophiolitic units have not been investigated in detail from a geochemical perspective. This contribution presents U-Pb zircon along with trace element and Nd isotope data obtained on the metaigneous ophiolitic rocks of the Morais

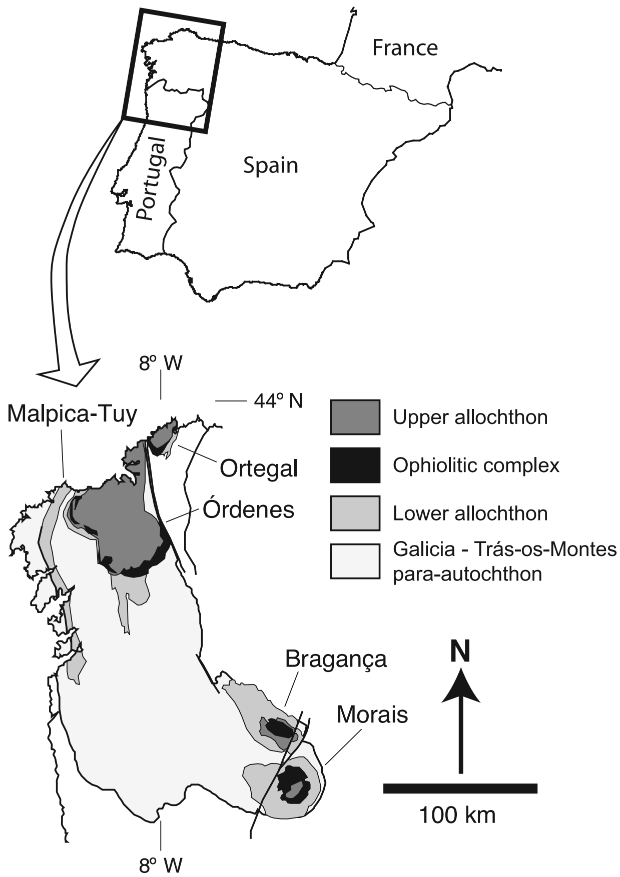

Figure 1. Geological map of the NW Iberian massif showing the location of the five allochthonous complexes: Cabo Ortegal, Órdenes, Morais, Bragança. and Malpica-Tuy and their principal lithological units.

Complex, in order to constrain their age and geodynamic setting. The possible tectonic implications of these results are also explored and discussed.

2. Geological Setting and Previous Work

[4] Continental and oceanic lithosphere slices were subducted and subsequently obducted onto western Gondwana during the Variscan orogeny constituting the so-called Allochthonous complexes of the northwestern Iberian Massif. These are the complexes of Cabo Ortegal, Órdenes, and Malpica-Tuy in Spain and of Bragança and Morais in Portugal (Figure 1), all of them forming regional-scale klippen within late orogenic Variscan synforms. The complexes comprise three principal structural elements that can be correlated at a regional scale [e.g., *Martínez Catalán et al.*, 1997]: the so-called Upper Allochthon on top, the intermediary ophiolitic complex, and the Lower Allocthon at the bottom (Figure 1).

[5] The Iberian autochthon (Galicia-Trás-os-Montes para-autochthon; Figure 1) and the Lower Allochthon are separated by thrust contacts. Both structural components probably represent colliding elements of the continental margin of Gondwana, now composing the footwall to the true suture, delineated by the ophiolitic complex. The Lower Allochthon consists of abundant metasediments and felsic igneous rocks, the latter in part related to an Ordovician continental rifting episode [*van Calsteren et al.*, 1979]. These rocks are thought to have undergone crustal subduction under the accretionary wedge formed by the ophiolitic complex and the Upper Allochthon [*Martínez Catalán et al.*, 1996; *Rodríguez Aller et al.*, 2003].

[6] The hanging wall to the suture (Upper Allochthon) consists of several thrust sheets that can be grouped into (1) a lower, high-pressure ensemble of granulites, gneisses, eclogites, and ultramafic rocks and (2) an upper, intermediate-pressure unit including a thick sequence of metasediments, large orthogneiss massifs and variably recrystallized mafic intrusives [*Martínez Catalán et al.*, 1996, 2002; *Ábalos et al.*, 2003]. The igneous protoliths are Late Cambrian to Early Ordovician in age in all cases. A nearly coeval first tectonothermal event around circa 490 Ma is suggested by the age of monazite in metasediments and orthogneisses of the upper, intermediate-pressure units, while a distinct prograde metamorphic episode reaching high-pressure granulite to high-temperature eclogite facies conditions occurred at circa 390–400 Ma in the lower, high-pressure units [*Ordóñez Casado et al.*, 2001; *Fernández Suárez et al.*, 2002; *Santos Zalduegui et al.*, 2002, and references therein].

[7] The ophiolitic complex, the inferred Variscan suture, is situated in a structurally intermediate position between the Lower and the Upper Allochthons. It is formed by various thrust sheets made of ophiolitic rocks displaying variable outcropping extension in each of the complexes (Figure 1). This complex is interpreted to reflect stacking during the closure of a Paleozoic ocean (Rheic).

[8] Good descriptions of the ophiolitic complex metabasites and associated rocks in the Morais allochthon are given by *Anthonioz* [1972], *Ribeiro* [1974], *Ribeiro et al.* [1987], and *Ribeiro and Pereira* [1997] and are synthesized in the Portuguese Geological Survey maps at the 1:50,000 and 1:200,000 scales [*Pereira et al.*, 1999; *Meireles*, 2000a, 2000b; *Pereira*, 2001; *Pereira et al.*, 2003]. Here, the ophiolitic complex is sandwiched between two allochthons of continental origin; an upper allochthon (upper allochthonous complex in Figure 2a) comprising the Lagoa Micaschists on top and the Lagoa Orthogneisses underneath, and the lower allochthonous complex. The thrust contact between the upper and the ophiolitic allochthonous complexes is delineated in places by mappable slices of granulites and garnet peridotites (for example, to the SW of Morais and the SE of Lagoa; compare Figure 2a). Blastomylonitic mafic granulites associated with quartz-feldspar gneisses of uncertain origin also occur in a different structural position isolated by hectometer-scale duplexes and out-of-sequence tectonic horses under the lower thrust contact of the ophiolitic complex to the NW of Morais (Figure 2a).

3. Structure of the Morais Ophiolitic Complex

[9] The ophiolitic rocks comprise areally widespread amphibolites and, in decreasing order of abundance, ser-

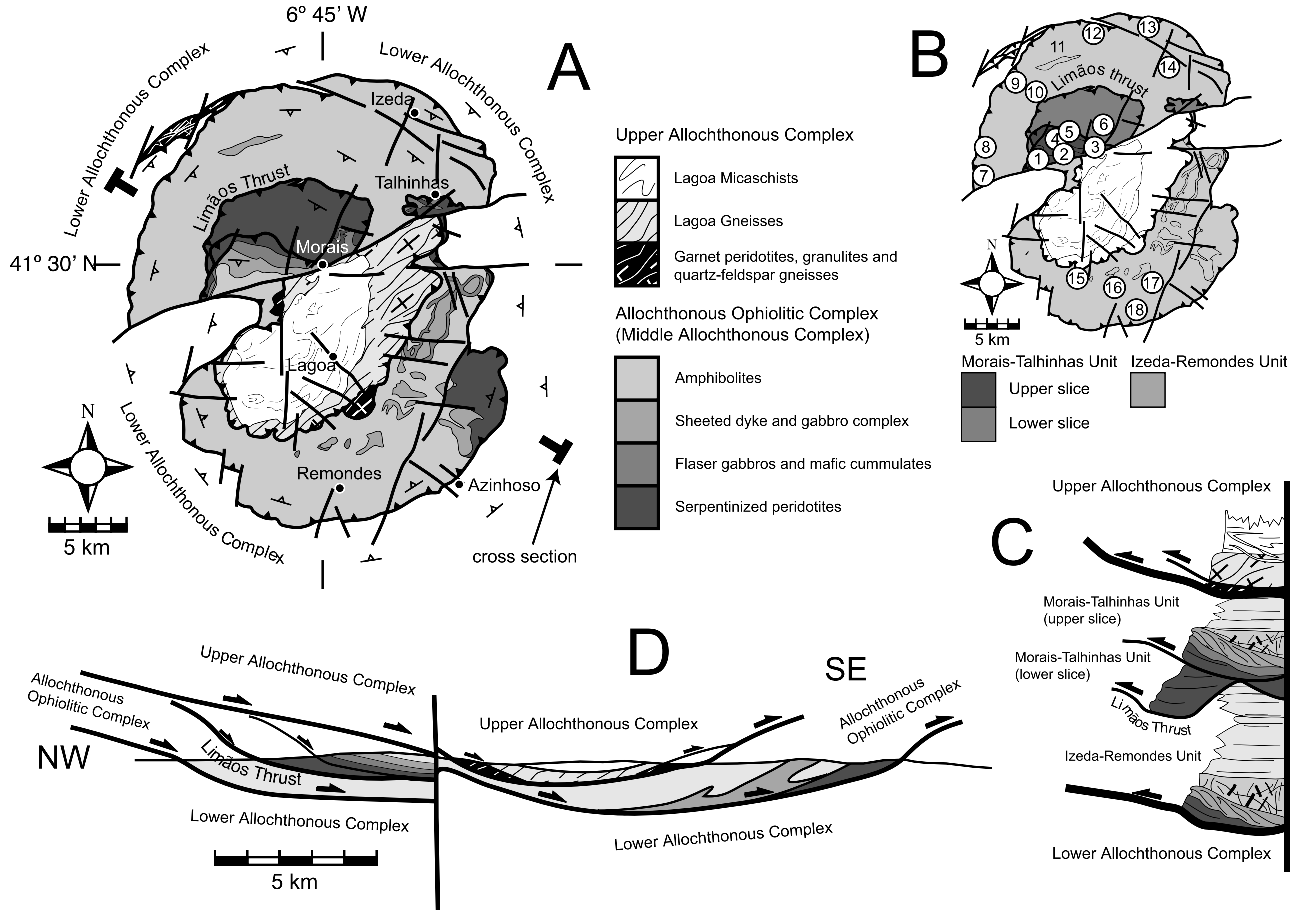

Figure 2. (a) Geological sketch map of the middle and upper allochthonous units of the Morais Complex simplified after the 1:200,000 map compiled by *Pereira* [2001]. (b) Sampling sites in the allochthonous ophiolitic complex of Morais. Morais-Talhinhas Unit: 1 (samples Mo1, Mo2), 2 (samples Mo3a, Mo3b, Mo3c, Mo4), 3 (sample Mo5), 4 (samples Mo6, Mo7, Mo8, Mo9, Mo10), 5 (samples Mo11, Mo12), 6 (sample Mo13). Izeda-Remondes Unit: 7 (sample Iz1), 8 (sample Iz2), 9 (sample Iz3), 10 (sample Iz4), 11 (sample Iz5), 12 (sample Iz6), 13 (sample Iz7), 14 (sample Iz8), 15 (samples Iz9, Iz10, Iz11), 16 (sample Iz12), 17 (samples Iz13, Iz14), 18 (sample Iz15). Categorization of the intervening thrust slices after *Marques et al.* [1992] and *Pereira et al.* [2004]. (c) Lithostratigraphic log showing the internal constitution and superposition order of ophiolitic rock units and thrust slices, notably in the allochthonous ophiolitic complex. Gray shadings for the allochthonous ophiolitic complex and patterns for the upper allochthonous complex relate to rock types and units as indicated in the legend of Figure 2a. (d) Interpretative cross section of the Morais Complex, modified after *Marques et al.* [1992], redrawn along a direction subparallel to the regional Variscan mineral and stretching lineations related to metamorphic nappe emplacement (D1 [cf. *Ribeiro*, 1974; *Ribeiro et al.*, 1990]). Shades of gray for the allochthonous ophiolitic complex and patterns for the upper allochthonous complex relate to rock types and units as indicated in the legend of Figure 2a.

pentinized peridotites, sheeted dike/gabbro complexes, flaser gabbros and mafic cumulates. *Ribeiro et al.* [1990] grouped the outcrops of these rocks into three thrust slices. The lowermost slice corresponds to the Izeda-Remondes Unit, whereas the two others constitute the Morais-Talhinhas Unit [*Marques et al.*, 1992; *Pereira et al.*, 2004] (compare Figures 2b and 2c). The lower sliver of the Morais-Talhinhas Unit is a tectonic horse made of serpentinized peridotites, whereas the upper one is interpreted to preserve a complete, though attenuated, ophiolitic sequence (Figures 2c and 2d). Except for nearly monolithological lower slice of the Morais-Talhinhas Unit, mafic ophiolitic rocks exhibit an internal organization in which serpentinized peridotites at the base of the thrust slices are overlain in turn by flaser gabbros/mafic cumulates, sheeted dike/gabbro complexes, and layered amphibolites (Figures 2a and 2c). This pattern is mostly undisturbed throughout the Morais Complex, except in its southeastern sector, where large-scale synmetamorphic deformation gave rise to the development of kilometer-scale folds with overturned limbs in the Izeda-Remondes Unit (Figure 2d). However, the general organization of ophiolitic rocks ''pseudostratigraphy'' is maintained in this unit from the outer (structurally lower) to the inner (structurally higher) outcrops. The sparse geochemical data available on the widespread amphibolitic rocks were interpreted to reflect enriched mid-ocean ridge basalts (MORBs) compositions (Indian MORB and enriched MORB (EMORB) types), suggesting an oceanic basin of rapid expansion and little extension [*Pereira et al.*, 2004].

[10] The internal structural organization and tectonic evolution of the metabasites and associated rocks of the Morais ophiolitic complex forms two principal sheets separated by the Limãos thrust (Figure 2), which corresponds to a ductile thrust contact [*Ribeiro et al.*, 1990; *Marques et al.*, 1992; *Pereira et al.*, 2004]. Each of these sheets exhibits a partial ophiolite lithostratigraphy with serpentinized peridotites at the base, overlain in turn by mafic cumulates, flaser gabbros, sheeted dike/gabbro complexes and amphibolites (Figure 2d). The units contain a subhorizontal penetrative main foliation developed under upper amphibolite facies conditions, with WNW-ESE mineral and stretching lineations [*Anthonioz*, 1972; *Ribeiro*, 1974]. The lineation is interpreted to document a top-to-the ESE sense of shear [*Pereira et al.*, 2004].

[11] The lower sheet (Izeda-Remondes Unit) is dominated by amphibolites. The structural base of the ophiolite stratigraphy crops out discontinuously at the nappe sole or in the cores of overturned synmetamorphic folds only in the SE part of the complex (Figures 2a and 2d). These rocks, together with mylonitized amphibolites along the outer ring, constitute the structural sole (a few hundred meters thick) of the Izeda-Remondes sheet or ophiolitic nappe. The ophiolite sole was subjected to pervasive dynamic greenschist facies mylonitization. Further retrogression was associated with localized deformation, represented by late, sparse and discrete ductile shear zones oblique to the principal (penetrative) foliation of amphibolites. These events led to structural and metamorphic inversions in the SE part of the Morais Complex and reflect the intense stretching of the Izeda-Remondes sheet parallel to the direction of tectonic transport (Figure 2d).

[12] In addition to its deeper metamorphic overprinting, the upper sheet (Morais-Talhinhas Unit) differs from the underlying one because it is dominated (both areally and volumetrically) by ophiolite lithotypes other than amphibolite. Metamorphic foliations are gently dipping throughout and the general structure is only disturbed by the imbrication of the lowermost, ultramafic elements (giving rise to two mappable slices; Figure 2d). By contrast with the underlying Izeda-Remondes Unit, the development of a thick structural sole is not obvious, though pervasive retrogression (serpentinization) of ultramafic rocks is ubiquitous. The outcrop trace of the basal ophiolite thrusts in this unit delineates a structural pattern of laterally accreted units whose thickness normal to the regional foliation is larger than the actual structural separation between the sole and roof bounding thrusts (Figure 2d). The Morais-Talhinhas ophiolite lithologies beneath the layered amphibolites sequence consist of relatively thick units and are better preserved than in the Izeda-Remondes Unit. The synmetamorphic sheet elongation and the concomitant thinning normal to the regional foliation orientations are subjects of debate, and might be related either to contractional deformations during plate convergence and nappe emplacement or to later extensional collapse of the nappe stack [cf. *Dallmeyer and Gil Ibarguchi*, 1990].

4. Petrography and Mineralogy

[13] We present here a brief description of the petrography and mineral chemistry of representative samples of metabasic, felsic and ultramafic lithotypes from both the structurally lower (Izeda-Remondes, samples Iz) and the upper (Morais-Talhinhas, samples Mo) units of the Morais ophiolitic complex. Full mineral compositions are available in the auxiliary material.[1]

4.1. Amphibolites

[14] Amphibolite is the dominant rock type of the Morais allocthonous ophiolitic complex. These are layered, mineralogically homogeneous rocks with a nematoblastic microstructure. Relic igneous textures suggest derivation from dolerite or gabbroic protoliths. Fine-grained leucocratic differentiates are also found, while epidote-rich, plagioclase-rich, and nearly monomineralic amphibole-rich mafic rocks are rare. Low-grade amphibolite- to greenschist-facies metabasites showing intensely foliated fabrics and dynamically recrystallized microstructures with relic ophitic textures predominate in the lowermost structural levels of the Izeda-Remondes Unit [*Pereira et al.*, 2003]. Garnet-bearing amphibolites are found in the Morais-Talhinhas Unit.

[15] Amphiboles are unzoned magnesiohornblende to tschermakitic hornblende in common amphibolite and ferrotschermakite in the garnet-bearing type; actinolitic hornblende is secondary. Unzoned plagioclases range from An_{14} to An_{35} in common amphibolite and show reverse zoning (An_{23-33}) in the garnet-bearing rocks. Garnets have appreciable amounts of (calculated) Fe^{3+} and show nearly flat cores and a decrease in Mn and increase in Mg content near rims, e.g., $Alm_{56}Grs_{28}Sps_5Andr_4Prp_7$ (cores), $Alm_{55}Grs_{24}Sps_3Andr_8Prp_{10}$ (rims). Sparse epidote shows decreasing pistacite contents from core to rim in common amphibolites and increase in the garnet-bearing type. Rutile occurs as minute crystals within amphibole and as larger grains with rare magnetite cores in the matrix; rutile may be replaced by titanite or by ilmenite, which may be in turn replaced by titanite. Fe-sulphide and apatite occur as accessories.

4.2. Metagabbros and Metadolerites

[16] Gabbroic amphibolites, Fe-Ti-rich and garnet-rich (meta)gabbros, and flaser gabbros, ranging to microgabbroic or doleritic types form extensive outcrops in the Morais-Talhinhas Unit. Amphibolites with a plano-linear fabric are gradational with deformed gabbros. Clinopyroxene is Al-rich and Na-poor diopside, and orthopyroxene has approximately equal amounts of Fe and Mg. Amphibole replacing clinopyroxene is magnesiohornblende to parga-

[1]Auxiliary materials are available at ftp://ftp.agu.org/apend/tc/2006tc001971.

sitic hornblende; cummingtonite replaces orthopyroxene; secondary amphibole is actinolitic hornblende. Plagioclase forms unzoned to normally zoned porphyroclasts (An_{37-26}) and unzoned neoblasts (An_{28-30}). Garnet coronas between ilmenite and plagioclase have typical compositions ranging from $Alm_{56}Grs_{22}Sps_{12}Andr_{5}Prp_{50}$ close to plagioclase, to $Alm_{65}Grs_{15}Sps_{6}Andr_{6}Prp_{8}$ next to ilmenite. The latter composition is similar to that of poorly zoned minute idiomorphic and skeletal garnets in the matrix or included in plagioclase. Garnet in Fe-Ti rich (meta)gabbros is strongly zoned and rich in Ca: $Alm_{32}Grs_{51}Sps_{6}Andr_{9}Prp_{1}$ (core), $Alm_{42}Grs_{37}Sps_{5}Andr_{5}Prp_{11}$ (rim). Epidote, clinozoisite, rutile, apatite, Fe-sulphide, magnetite, ilmenite and rare allanite are accessory minerals. Titanite replaces rutile and ilmenite.

4.3. Felsic Rocks

[17] Fine-grained felsic rocks with poorly oriented granoblastic microstructure form sporadic meter to decameter size outcrops in metabasites of both ophiolitic units. Those of Morais-Talhinhas may contain garnet and/or amphibole. They also occur within ultramafic rocks, for example a garnet and amphibole-bearing hectometer-scale felsic dike at Pedras Brancas in metaperidotite structurally overlying the amphibolites of Morais-Talhinhas [*Anthonioz*, 1972]. These rocks might result from the recrystallization of either former felsic differentiates or partial melts related to intraoceanic shear zones (e.g., those in ultramafic units). Amphibole is Na-rich magnesiohornblende to hastingsitic hornblende. Unzoned plagioclase is albite with rare cores of oligoclase-andesine. Subidiomorphic, inclusion-poor garnets preserve growth zoning, e.g., $Alm_{60}Grs_{14}Sps_{15}Andr_{4}Prp_{7}$ (cores), $Alm_{72}Grs_{13}Sps_{5}Andr_{4}Prp_{6}$ (rims), whereas skeletal crystals have compositions similar to those of idiomorphic garnet rims. Epidote forms discrete grains or rims on allanite. Rare biotite is rich in Fe and Ti. Ilmenite, titanite, magnetite, allanite, apatite and zircon are accessory minerals.

4.4. Ultramafic Rocks

[18] Ultramafic rocks form extensive outcrops at the base of the Izeda-Remondes and Morais-Talhinhas units. These rocks are not dealt with in detail in the present study. Ultramafic rocks in Morais are mainly dunite and harzburgite with minor podiform chromite and carbonate veins [*Pereira et al.*, 2004]. They exhibit primary compositional layering probably related to their tectonic evolution in a lithospheric mantle realm. Hornblende is often an initial metamorphic recrystallization phase, while tremolite, serpentine, talc and chlorite are late retrogression minerals. Dismembered flaser gabbros occur as small lenses in the peridotites. The ultramafic samples studied are strongly serpentinized harzburgite with rare relics of olivine and spinel.

5. Metamorphic Evolution

[19] P-T conditions calculated for two-pyroxene, garnet-bearing assemblages (garnet-pyroxene-plagioclase-quartz thermobarometry [*Powell*, 1985; *Eckert et al.*, 1991]) in coronitic metagabbros of the Morais-Talhinhas Unit are unrealistically low: ~300–500°C and 0.3 GPa. This suggests that metamorphic garnet did not reach chemical equilibrium with variably modified igneous pyroxene in these rocks. Instead, garnet-amphibole-plagioclase assemblages appear to record progressive equilibrium in various types of metabasites and felsic rocks. Using garnet-amphibole-plagioclase thermobarometry [*Powell*, 1985; *Kohn and Spear*, 1990] and assuming all iron as Fe^{2+} and neoblast or homogeneous plagioclase compositions, P-T estimates for metabasites range from ~630 ± 35°C, 0.95 ± 0.1 GPa, for garnet cores, to ~685 ± 30°C, 1.0 ± 0.1 GPa, for garnet rims, while felsic rocks associated with metabasites have yielded equilibrium conditions of ~720 ± 50°C, 1.3 ± 0.1 GPa, for garnet cores, to ~615 ± 30°C, 1.15 ± 0.1 GPa, for garnet rims. Garnet-amphibole-bearing felsic rocks found in ultramafic lithologies attest to slightly higher pressure equilibrium conditions of up to 1.4 GPa at ~600°C, thereby pointing to a deeper provenance for this subunit.

[20] P-T conditions for the Izeda-Remondes Unit are not as well constrained as for the Morais-Talhinhas Unit. Calculated temperature conditions for amphibole-plagioclase assemblages in amphibolites of Izeda-Remondes are similar to those of Morais-Talhinhas: ~650 ± 55°C (plagioclase-amphibole geothermometer of *Blundy and Holland* [1990] and *Holland and Blundy* [1994]), while a medium-pressure metamorphic regime, with pressures >0.5 GPa, may be assumed on the basis of amphibole compositions [cf. *Raase*, 1974; *Laird and Albee*, 1981].

[21] Following the generalized medium- to high-grade amphibolitic recrystallization, the ophiolitic rocks of both units experienced a decrease in P and T down to low-grade amphibolite- to greenschist-facies conditions. This produced abundant epidote, replacement of rutile by titanite and sparse growth of actinolitic amphibole and chlorite after primary amphibole, garnet and pyroxene in appropriate lithologies. P-T conditions for this stage are poorly constrained, although a cooling process coeval with decompression, probably related to active tectonic uplift, may be inferred from the low Al^{VI}/Si ratios and NaM4 contents of secondary amphiboles [cf. *Raase*, 1974].

6. Geochronology

6.1. Sm-Nd Data

[22] All but two of the eleven samples from the Izeda-Remondes unit (Figure 2b) define a linear array (mean square weighted deviation of 1.4) when plotted on a $^{143}Nd/^{144}Nd$ versus $^{147}Sm/^{144}Nd$ diagram (Figure 3 and Table 1). This alignment, if interpreted as an isochron, provides an age of 447 ± 24 with an initial $^{143}Nd/^{144}Nd$ ratio corresponding to $\varepsilon Nd = +8.7 \pm 0.1$ (calculated following the method of *Fletcher and Rossman* [1982]). The two samples discarded from the isochron calculation (Iz5 and Iz13) have εNd_{450} values of +10.7 and +11.2, respectively, implying that they were extracted from a mantle domain which had an stronger time-integrated depletion of the light rare earth elements (LREE), compared to the

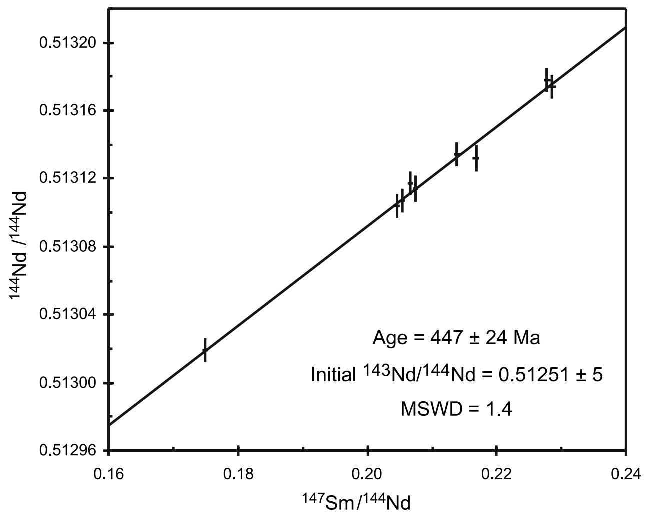

Figure 3. Sm-Nd isotopic correlation diagram for whole rock samples of group I metabasites from the Morais ophiolitic complex. Age calculation excluding samples Iz5 and Iz13 (see text).

isotopically more homogeneous source of the rest of amphibolite samples (see below, elemental Nd isotope chemistry).

6.2. U-Pb Data

[23] Abundant zircon crystals were extracted from three rocks belonging to the upper ophiolitic unit. One coarse grained flaser gabbro sample from near Sobreda forms part of a large flaser gabbro to amphibolite unit. The other two are felsic rocks, one is a garnet-bearing rock within peridotite at Pedras Brancas, and the other is a felsic rock without garnet near Paradinha. A slightly weathered sample of felsic rock was also taken as a loose block near Chacim (Iz1). Analytical procedures are outlined in the auxiliary material, and the results are listed in Table 2.

[24] Zircons from the two felsic rocks (Paradinha and Pedras Brancas dike) are morphologically similar, with a light yellow color and euhedral, short prismatic crystals. Four multigrain fractions from the Paradinha sample were analyzed. Their radiogenic Pb content is low, from 11 to 16 ppm (Table 2). Within analytical uncertainty, all fractions plot on the Concordia curve and indicate an age of 404.5 ± 0.4 Ma (Figure 4a). This concordant age is interpreted to date the crystallization of the zircons from the felsic melt.

Table 1. Sm-Nd Analytical Data

Sample	Sm	Nd	$^{147}Sm/^{144}Nd$	$^{143}Nd/^{144}Nd$	εN_0	εN_{400}	$\varepsilon_{500}N_{005}$
			Group I Metabasites				
Iz2	3.84	11.2	0.2066	0.513117 ± 7	9.3	8.8	11.4
Iz3	5.97	15.6	0.2055	0.513107 ± 7	9.2	8.7	11.2
Iz4	3.2	8.93	0.2168	0.513132 ± 8	9.6	8.6	11.1
Iz5	4.32	12.8	0.2085	0.513132 ± 6	9.6	9.3	11.8
Iz6	5.7	16.1	0.2139	0.513134 ± 7	9.7	8.8	11.3
Iz7	3.56	10.4	0.2075	0.513114 ± 8	9.3	8.7	11.3
Iz8	4.61	13.6	0.2046	0.513104 ± 7	9.1	8.7	11.2
Iz9	1.68	4.47	0.2278	0.513178 ± 7	10.5	9	11.5
Iz10	2.89	9.98	0.175	0.513019 ± 7	7.4	8.5	11.1
Iz13	3.74	9.82	0.2304	0.513229 ± 7	11.5	9.8	12.3
Iz14	3.23	8.54	0.2286	0.513174 ± 7	10.5	8.8	11.4
			Group II Cumulate Gabbros				
Mo5	1.03	2.64	0.2355	0.513128 ± 7	9.6	7.6	10.1
Mo6	0.88	2.31	0.2301	0.513154 ± 9	10.1	8.4	10.9
Mo8	3.17	7.52	0.2546	0.513244 ± 8	11.8	8.9	11.4
Iz12	1.17	3.46	0.2046	0.513035 ± 8	7.7	7.3	9.9
			Group II Noncumulate Mafic Rocks				
Mo1	3.28	9.14	0.2172	0.513141 ± 7	9.8	8.8	11.3
Mo2	13.8	50.8	0.1645	0.512956 ± 7	6.2	7.9	10.4
Mo10	2.36	6.97	0.2041	0.51308 ± 8	8.7	8.3	10.9
			Group II Amphibolites				
Iz11	5.29	24.2	0.1319	0.512817 ± 7	3.5	6.8	9.3
Iz15	4.65	19.2	0.1466	0.512825 ± 9	3.7	6.2	8.7
			Group II Felsic Rocks				
Mo3a	5.8	21.9	0.1604	0.512931 ± 7	5.7	7.6	10.1
Mo3b	9.32	32	0.176	0.512962 ± 7	6.3	7.4	9.9
Mo3c	6.42	23.5	0.1649	0.512982 ± 9	6.7	8.3	10.9
Mo4	4.51	15.9	0.1719	0.512851 ± 7	4.2	5.4	7.9
Mo13	8.48	31	0.1655	0.512986 ± 8	6.8	8.4	10.9
Iz1	16	60.4	0.1604	0.512980 ± 6	6.7	8.5	11.1

Table 2. U-Pb Analytical Data[a]

Fraction	Weight, μg	Concentration, ppm		Atomic Ratios					Apparent Ages, Ma		
		U	Pb rad	$^{206}Pb_b/^{204}Pb$	$^{208}Pb^b/^{206}Pb$	$^{206}Pb_c/^{238}U$	$^{207}Pb_c/^{235}U$	$^{207}Pb_c/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$
Mo3a Felsic Rock, Paradinha											
1 Ab	65	237	16.2	984	0.1467	0.06482 (0.13)	0.4899 (0.16)	0.05481 (0.08)	405	405	405
2 Ab	77	171	11.1	3275	0.1119	0.06472 (0.21)	0.4891 (0.22)	0.05481 (0.06)	404	404	404
3 Ab	57	201	12.9	4101	0.1013	0.06457 (0.50)	0.4877 (0.52)	0.05478 (0.16)	403	403	403
4 Ab	135	243	15.9	2188	0.1208	0.06400 (1.09)	0.4831 (1.10)	0.05474 (0.09)	400	400	402
Mo7 Fe-Ti Gabbro, Sobreda											
5 Ab	101	277	19	5411	0.1919	0.06381 (0.10)	0.4813 (0.11)	0.05471 (0.04)	399	399	400
6 Ab	182	245	16.7	10066	0.1843	0.06381 (0.10)	0.4812 (0.10)	0.05469 (0.02)	399	399	400
7 Ab	72	192	13.6	3984	0.2349	0.06371 (0.20)	0.4803 (0.21)	0.05468 (0.06)	398	398	399
8	48	444	30.1	1950	0.1836	0.06304 (0.12)	0.4752 (0.13)	0.05468 (0.06)	394	395	399
Mo13 Felsic Rock, Pedras Brancas											
9 Ab	96	188	12.1	2863	0.1242	0.06348 (0.77)	0.4784 (0.78)	0.05466 (0.09)	397	397	398
10 Ab	85	338	22.2	3438	0.1461	0.06339 (0.68)	0.4776 (0.68)	0.05464 (0.06)	396	396	398
11 Ab	160	315	20.6	5750	0.1384	0.06339 (0.47)	0.4776 (0.48)	0.05464 (0.05)	396	396	398
12 Ab	121	271	18.4	1148	0.1601	0.06327 (0.21)	0.4766 (0.22)	0.05463 (0.05)	396	396	397
13 Ab	174	338	22.1	7125	0.1419	0.06319 (0.65)	0.4758 (0.65)	0.05461 (0.03)	395	395	396
Iz1 Felsic Rock, Chacim											
14 Sd	123	276	18.6	1731	0.1761	0.06272 (0.13)	0.4728 (0.14)	0.05467 (0.05)	392	393	399
15	56	503	37.4	419	0.2438	0.06241 (0.31)	0.4711 (0.49)	0.05475 (0.35)	390	392	402

[a]Analytical uncertainties in parentheses (%) are given at the 2σ level; Ab, zircon fraction mechanically abraded before dissolution [*Krogh*, 1982]; Sd, first step of selective dissolution [*Mattinson*, 1994]. Detailed analytical techniques are reported in the auxiliary material.

[b]Measured ratio.

[c]Calculated ratio (compare analytical techniques).

[25] Zircons from the Sobreda gabbro are light yellow, large and anhedral with a broken aspect interpreted as reflecting late stage crystallization. Four multigrain fractions were analyzed. On the Concordia diagram (Figure 4b), the three abraded fractions are concordant, while the fourth unabraded is slightly discordant. Pb and U concentrations are broadly similar in all the samples. All data points plot on a line giving an upper intercept age of 399.8 ± 0.6 Ma, interpreted as dating the crystallization of the zircons during the igneous emplacement of the gabbro.

[26] The five multigrain fractions analyzed for the Pedras Brancas dike are concordant and define an age of 396.5 ± 0.4 Ma (Figure 4c) Again, this age is interpreted to reflect the igneous crystallization of the zircons, at the time of emplacement of the felsic dike into ultramafic country rocks.

[27] The Chacim felsic rock provided rare and small zircon grains with cracks and inclusions not well suited for high-quality U-Pb dating. Nevertheless, two fractions were analyzed (Table 2 and Figure 4b). In both cases, the data points are discordant, but provide broadly similar results with $^{207}Pb/^{206}Pb$ apparent ages around 400 Ma. Albeit imprecise, this age probably reflects the igneous crystallization of the zircons.

7. Elemental and Nd Isotope Chemistry

7.1. Methods

[28] Major and trace element data for the main petrographic types within the Morais ophiolite thrust complex are given in the auxiliary material. The metabasites as a whole, including amphibolites, metagabbros and metadolerites, show relatively low values of alkalis and high Fe_2O_3 contents with respect to MgO, suggestive of the Fe enrichment trend of the tholeiitic rock series. This is also shown by triangular diagrams designed for the discrimination of metabasites based on "immobile" trace elements. In the Ti/100-Zr-Yx3 diagram [*Pearce and Cann*, 1973], the majority of samples plot close to the B field of ocean floor basalts (some outliers of Ti-rich metagabbros plot out of the fields of reference). In the Nbx2-Zr/4-Y diagram [*Meschede*, 1986] most samples, albeit showing considerable dispersion, plot in the field for normal MORB (NMORB) and volcanic arc basalts, with again some minor lithotypes plotting outside this area.

[29] A more specific assessment of magmatic affinities and, by inference, source characteristics and tectonic setting of igneous emplacement, can be achieved by using a larger set of incompatible trace elements selected on the basis of their resistance to alteration and metamorphism, namely, high field strength elements, REE, and Th [see *Jenner et al.*, 1991, and references therein]. These data are used to calculate several critical ratios of pairs of elements of similar, or slightly different, degrees of incompatibility during mantle melting and fractional crystallization (i.e., Th/Nb, La/Nb, Zr/Nb, La/Sm, La/Lu), which are useful for tracing petrogenetic processes and magma sources. In particular, Th and Nb are key elements for evaluating the geodynamic environment of mafic metaigneous rocks [e.g., *Elthon*, 1991; *Jenner et al.*, 1991]. These elements are

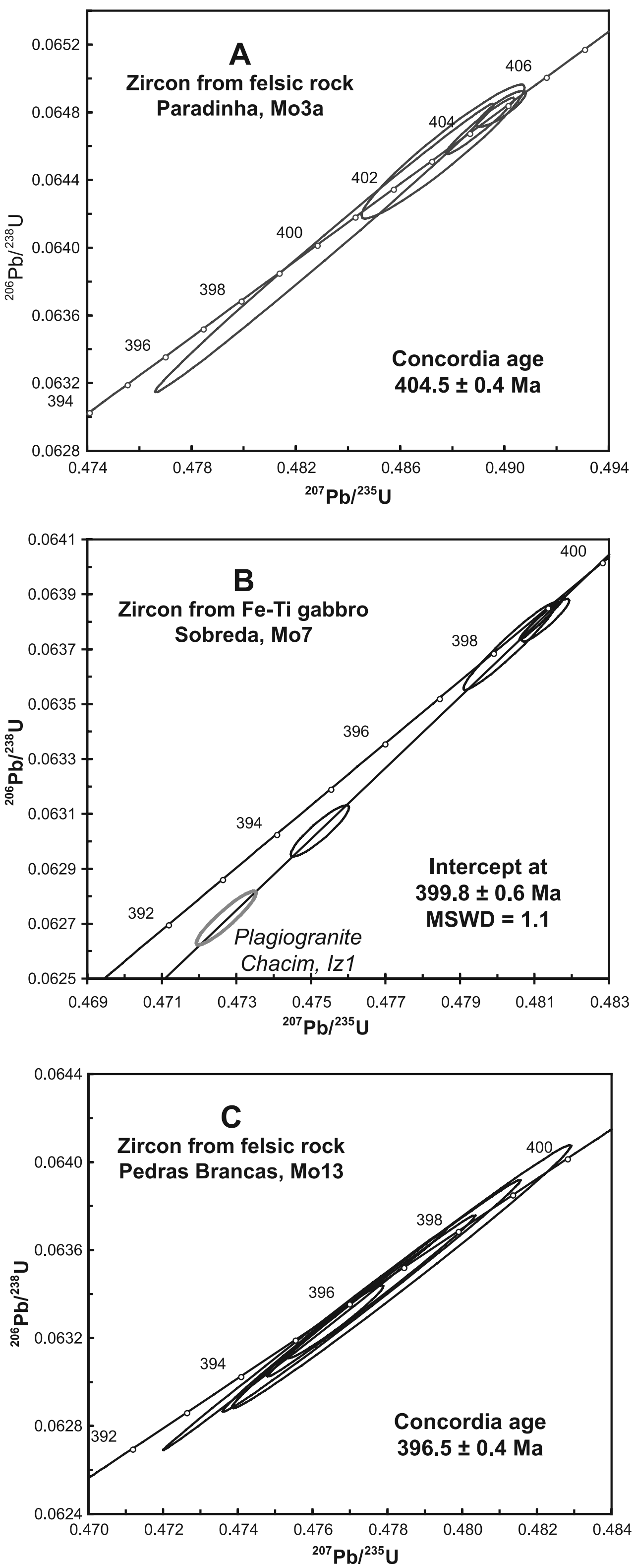

Figure 4. U-Pb Concordia diagrams for (a and b) the two felsic rocks (Pedras Brancas and Paradinha) and (c) one metagabbro (Sobreda) of the Morais ophiolitic complex (ellipse errors of 2σ, see analytical techniques).

partitioned to a very similar degree during ''dry'' melting of the upper mantle responsible for basaltic magmatism in both mid-ocean ridge and within-plate settings. This keeps their ratio to a low value, close to the chondritic one (Th/Nb $\approx$ 0.07). In contrast, Th and Nb are strongly fractionated in subduction zone environments, because Th is more easily transferred from the subducted slab of oceanic lithosphere to the overlying mantle wedge through H_2O-rich fluids and/or silicic melts [e.g., *Pearce et al.*, 1984; *Pearce*, 1991]. For this reason, anomalously high Th/Nb ratios are observed in magmas generated above subduction zones. Caution is needed, however, because high Th/Nb ratios are a typical feature of materials from the continental crust [e.g., *Taylor and McLennan*, 1985]. This implies that within-plate basalts, or mafic magmas emplaced during continental breakup episodes, having suffered crustal contamination may also develop elevated Th/Nb ratios mimicking those of subduction-related magmas [e.g., *Dupuy and Dostal*, 1984; *Pin and Marini*, 1993]. Because they provide a sensitive tool for monitoring contamination by mature continental materials, Nd isotope data used in combination with Th-Nb systematics may help in unraveling these alternative scenarios [e.g., *Pin and Paquette*, 1997]. Incompatible trace element data are also displayed graphically as multielement patterns (''spiderdiagrams'') normalized with respect to chondritic abundances in order to highlight their relative fractionation and any anomalous behavior of a given element (e.g., Eu, Nb). Although potentially fairly mobile during seawater alteration and metamorphism, Sr was also plotted and used with due care, in order to get some insight into plagioclase fractionation effects.

7.2. Results

[30] Two main groups of metabasite samples may be distinguished according to their multielement normalized patterns, and Th/Nb ratios:

[31] 1. A group of metabasites with typical NMORB characteristics (e.g., increasing relative depletion of the most incompatible elements, reflected by low Th/Nb and La/Sm ratios), and broadly homogeneous, highly radiogenic Nd isotope compositions. These are referred to as group I samples hereafter (see the auxiliary material).

[32] 2. Group II of metabasites has significantly higher Th/Nb ratios, highlighted in most samples by negative Nb anomalies, and more variable, albeit still strongly radiogenic, Nd isotope signatures. A few gabbroic samples with similarly elevated Th/Nb ratios display a positive Nb anomaly, associated with conspicuous large positive anomalies of Ti. Some results are also presented for ultramafic rocks (see auxiliary material).

7.2.1. Group I

[33] This group comprises NMORB-like metabasites, and consists of both pristine amphibolites from the middle part of the ophiolitic thrust complex, and variably retrogressed and deformed types taken from lower structural levels. All these samples belong to the so-called Izeda-Remondes lower subunit.

[34] Major element contents are similar to those of basaltic liquids, with low K_2O and low to moderate contents

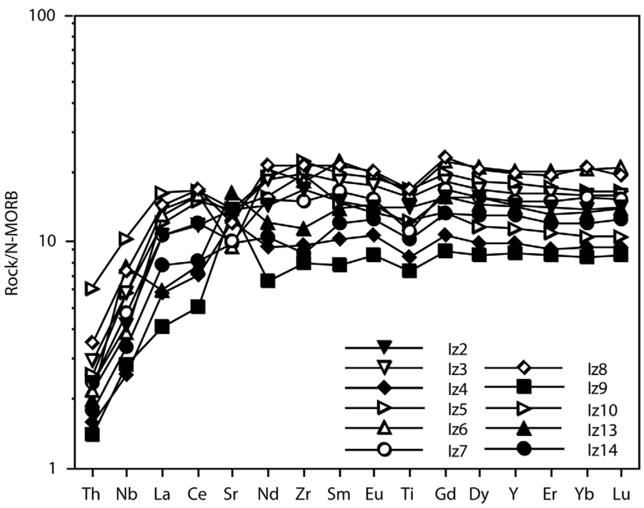

Figure 5. Multielement diagrams normalized to chondrite values [*Nakamura*, 1974; *Sun*, 1980] for group I metabasites of the Morais ophiolitic complex.

of compatible elements like Ni or Cr. Rare earth element (REE) contents correspond to ~10–20 times chondritic abundances for the middle and heavy rare earths, giving flat to gently sloping normalized patterns. A strong depletion in light relative to heavy rare earths reflected by $(La/Lu)_N$ ratios of 0.48–0.86, and a pronounced fractionation of LREE with increasing relative depletion of the lightest lanthanides $(La/Sm)_N$ 0.53–0.71, is characteristic. Only the amphibole-rich sample Iz10 displays a mild relative enrichment of LREE, associated with a more distinct fractionation of the middle and heavy REE. Eu anomalies are minor or absent. Such patterns and ratios are similar to those of typical mid-ocean ridge basalts of the NMORB type ($(La/Lu)_N = 0.57$, $(La/Sm)_N = 0.59$ [*Sun and McDonough*, 1989]). Likewise, Ti/V ratios display relatively high values well within the field (20–50) of mid-ocean ridge and back-arc basin basalts, and higher than the range (10–20) typical for island arc tholeiites [*Shervais*, 1982]. Besides the style of REE fractionation, another index of the degree of mantle source depletion is provided by Zr/Nb ratios [e.g., *Erlank and Kable*, 1976]. All but one sample (Iz9, with Zr/Nb = 31), have very high values (>45) further demonstrating that the basaltic protoliths of Morais amphibolites were extracted from highly depleted mantle sources.

[35] Chondrite-normalized multielement plots of group I are characterized by relatively flat patterns, except for La, Nb and Th which show increasing degrees of depletion (Figure 5), as typically observed in NMORBs. Negative Ti anomalies are also observed, while Zr does not show significant anomaly, except in sample Iz10. Most samples display negative Sr anomalies which, in the absence of Eu anomaly, might reflect chemical leaching during alteration and metamorphism. Samples Iz4, Iz9 and Iz13 (with 18.7 wt% Al_2O_3) that preserve relic igneous textures with plagioclase phenocrysts (now porphyroclasts) have a positive anomaly suggestive of accumulation processes. By comparison with all other rocks analyzed in this study, and despite some internal variations, the samples from group I (including Iz10), are readily identified by their low Th/Nb ratios (0.06–0.09), close to the chondritic value (~0.07 [*Sun*, 1980]). These are interpreted to characterize basalts generated in extensional environments unrelated to subduction processes, and not affected by contamination by crustal materials, an inference corroborated below by Nd isotopes.

[36] Sm-Nd isotope data are given in Table 1, together with initial Nd isotope composition, expressed using the ε notation, following correction for in situ radiogenic decay of ^{147}Sm assuming igneous formation ages of 400 Ma and 500 Ma, respectively. The 400 Ma age is our inferred lower limit for igneous emplacement. However, as argued hereafter, the amphibolites did not share simple cogenetic relationships with gabbros and felsic rocks, and might be significantly older than 400 Ma. For this reason, εNd values were also calculated for an age of 500 Ma, which corresponds to a widespread episode of rifting and continental breakup throughout Variscan Europe [e.g., *Pin and Marini*, 1993; *Crowley et al.*, 2000]. Because ^{147}Sm/^{144}Nd ratios in most samples do not deviate strongly from the chondritic value (0.1966), our conclusions are not sensitive to the uncertainty about the true geological age of the igneous protoliths.

[37] All samples display high εNd_i values in the +8 to +9 range, implying that their mantle source was depleted in Nd relative to Sm on a time-integrated basis, as typical for the present-day NMORB source reservoir. It is notable that the amphibole-rich sample Iz10, with slightly less depleted trace element characteristics (i.e., higher La/Sm and La/Lu, combined with lower Zr/Nb ratios) is reminiscent of EMORBs. However, this sample has the same Nd isotope composition (εNd_{450} = +8.7) as the other, NMORB-like samples, suggesting that its mantle source might have been enriched by local, low-degree partial melts shortly before, or at the time of magma generation.

[38] In summary, the amphibolites of the Morais ophiolitic complex, and more specifically, those exposed in the structurally lower Izeda-Remondes Unit, show elemental and Nd isotope similarities with oceanic basalts formed in normal mid-ocean ridge settings, without any recognizable interaction with components derived from the continental crust. An imprecise 447 ± 24 Ma Sm-Nd whole rock isochron age is tentatively interpreted to date their generation in a mature oceanic basin, apparently free of subduction zone influence.

7.2.2. Group II

[39] The second group of samples of the ophiolitic complex includes the metabasites with relatively high Th/Nb ratios and closely associated low-volume felsic rocks. These rocks were largely sampled in the structurally higher Morais-Talhinhas Unit of the ophiolite thrust complex. The rocks dated by the U-Pb zircon method belong to this second group.

[40] The mafic rocks comprise gabbros or microgabbros including an associated doleritic dike and one garnet-bearing amphibolite from the Morais-Talhinhas Unit, two amphibolites and a strongly deformed flaser gabbro from the Izeda-Remondes Unit. They have greater major element

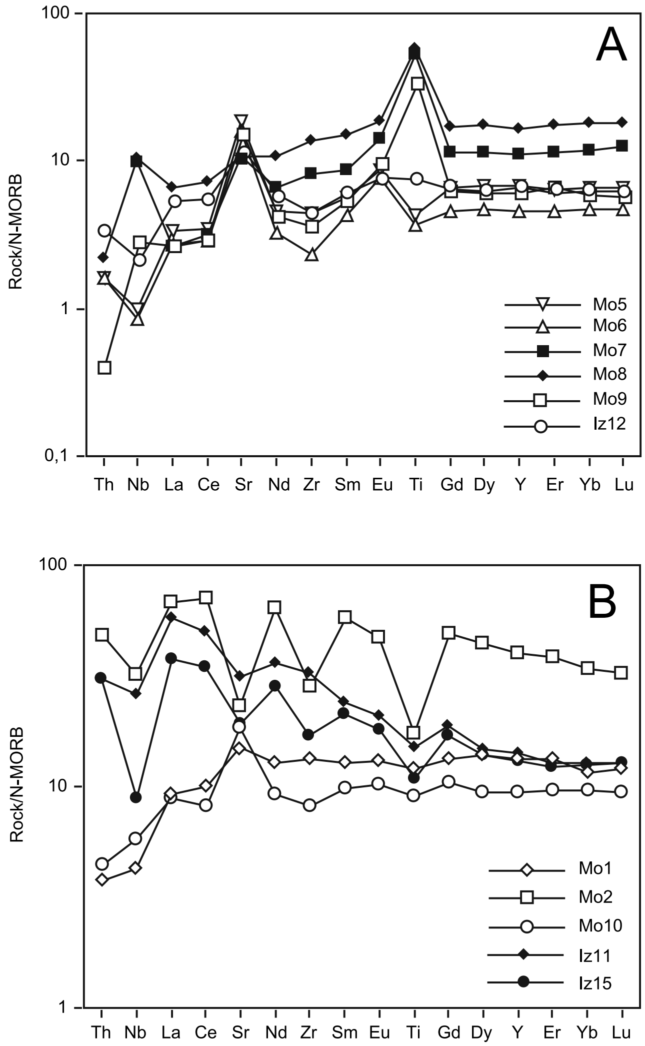

Figure 6. Multielement diagrams normalized to chondrite values [*Nakamura*, 1974; *Sun*, 1980] for (a) cumulative gabbros and (b) noncumulative mafic rocks and amphibolites, of group II rocks from the Morais ophiolitic complex.

chemical heterogeneity than group I amphibolites (see auxiliary material), partly reflecting the obvious cumulate nature of most of these gabbros.

7.2.2.1. Cumulate Gabbros

[41] Most gabbroic samples have low, unfractionated heavy REE (HREE) concentrations (5–15 times chondrite) and a strong depletion in LREE (generally more important than that shown by NMORB metabasites). Conspicuous positive Eu anomalies document plagioclase accumulation in these rocks, as do positive Sr anomalies on chondrite-normalized multielement plots (Figure 6a). A few gabbros also display large positive Ti anomalies corresponding to the presence of cumulus ilmenite. The positive Nb anomalies shown in these peculiar samples are interpreted to reflect the accumulation of Fe-Ti oxides and to be devoid of further petrogenetic significance regarding magma sources. All the other cumulates display variably pronounced, negative Nb anomalies, with Th/Nb ratios ranging from 0.13 to 0.26. Highly radiogenic initial Nd isotope compositions (εNd_{400} from +7.5 to +8.9) point to the time-integrated depletion of their magma source (Table 1).

7.2.2.2. Noncumulate Mafic Rocks

[42] Other mafic samples from the Morais-Talhinhas Unit lack evidence for significant crystal accumulation, and are interpreted to broadly approach former liquid compositions. Among them, a garnet-bearing flaser gabbro most likely corresponds to an evolved magma, or to an early stage, relatively low-degree partial melt. This sample shows a LREE-enriched pattern, with a distinct fractionation of the HREE and a negative Eu anomaly (Figure 6b). Deep negative anomalies of Nb, Sr, Zr, and Ti are also observed, with a high Th/Nb ratio (0.22). Along with a low Zr/Nb ratio (14), the LREE enrichment points to a relatively enriched source for that sample. However, the high value of εNd_{400} (+7.9) precludes a significant role for crustal assimilation, and suggests derivation from a mantle domain that was strongly depleted in LREE on a time-integrated basis. This implies that any enrichment of the source should have occurred just prior to, or at the same time as magma genesis. A crosscutting doleritic dike from the same outcrop shows a flat REE pattern from Nd to Lu, and an increasing depletion of the more incompatible elements Ce, La, Nb, and Th, with high Zr/Nb (50), and a faint negative Nb anomaly (Th/Nb = 0.13). Albeit markedly different from its host gabbro for incompatible trace elements, this dike has, within analytical uncertainty (±0.2 ε unit), the same Nd isotope composition (εNd_{400} = +7.5). Another microgabbroic sample shows broadly similar trace element characteristics (Figure 6b) combined with even more radiogenic isotope signatures (εNd_{400} = +8.6).

7.2.2.3. Amphibolites

[43] A few amphibolites collected in the Izeda-Remondes Unit (Iz11 and Iz15) share the same trace element features as the previous gabbroic rocks, that is, high Th/Nb ratios (corresponding to negative Nb anomalies in chondrite-normalized multielemental diagrams; Figure 6b), combined with variably enriched REE patterns, and largely positive εNd_{400} values (from +6.5 to +7.3). Sample Iz15, rich in plagioclase, also displays a negative Zr anomaly and may correspond to an increased deformation state of strongly transformed gabbros like Iz12 nearby (Figure 6b). Sample Iz11, a banded amphibolite with epidote-rich layers collected near the top of the Izeda-Remondes Unit is rich in Al_2O_3, CaO and LREE. Given its close spatial association with the amphibole-rich metabasite Iz10 (>80% amphibole), it might correspond to hydrothermally altered sections of the mafic sequence.

7.2.2.4. Felsic Rocks

[44] Except for sample Iz1 from the Izeda-Remondes Unit (Chacim albitite [*Anthonioz*, 1972]) of rather uncertain setting due to poor exposures, all the felsic samples analyzed in this work belong to the Morais-Talhinhas Unit and occur generally as small (a few cm to a few dm wide) leucocratic veins within gabbroic host rocks. Sample Mo13 (Pedras Brancas dike [*Anthonioz*, 1972]) differs from the other felsic rocks as it represents a ∼5-m-thick, 100-m-long

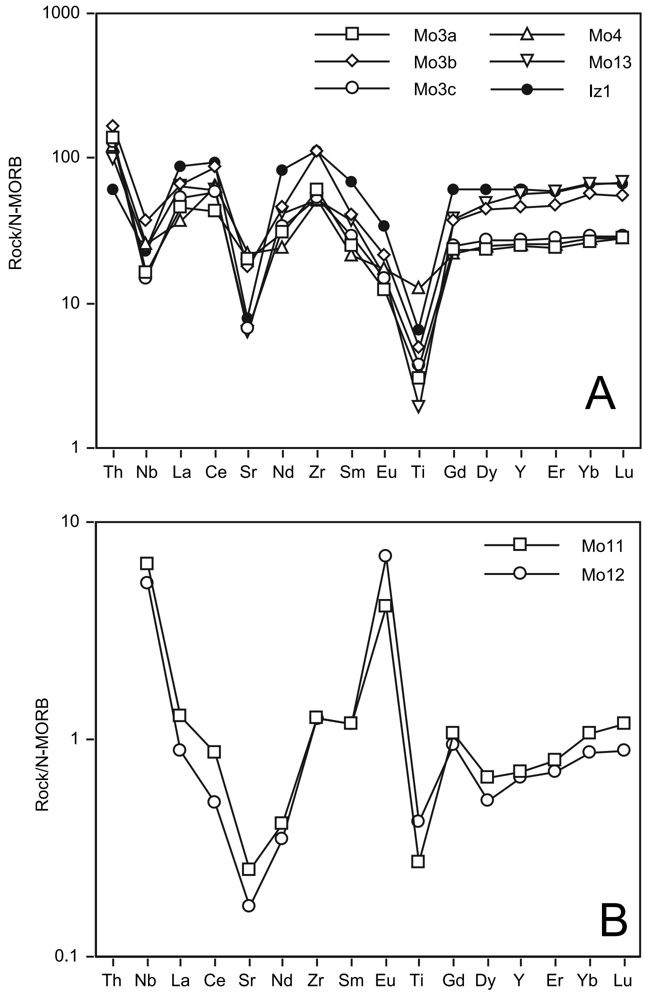

Figure 7. Multielement diagrams normalized to chondrite values [*Nakamura*, 1974; *Sun*, 1980] for (a) felsic rocks from group II and (b) ultramafic rocks of the Morais ophiolitic complex.

dike of foliated leucocratic rock crosscutting a serpentinized ultramafic unit, without any associated mafic rock type.

[45] With two exceptions, these rocks are nearly homogeneous in composition with SiO_2 contents ranging from 70 to 77 wt %, Na_2O between 6 and 7 wt %, and extremely low K_2O (< 0.2 wt %), as typical for oceanic plagiogranites [e.g., *Coleman and Donato*, 1979]. The REE patterns show a distinct enrichment of LREE with, in most cases, significant positive anomalies of Ce (Figure 7a). This style of LREE fractionation with $(La/Sm)_N > 1$ differs from the LREE depletion generally observed in plagiogranites from large ophiolites such as Troodos and Oman [e.g., *Coleman and Donato*, 1979] and suggests derivation from a LREE-enriched precursor. The heavy REE give either flat or variably fractionated patterns with increasing relative enrichment of HREE. Together with deep Eu anomalies in all samples (Eu/Eu* 0.4–0.55), the contrasting fractionation patterns of the light and heavy REE produce characteristic, V-shaped chondrite-normalized spectra (Figure 7a).

[46] In multielement chondrite-normalized diagrams, a positive Zr anomaly of variable size is observed, corresponding to the precipitation of zircon in these silicic magmas. All the felsic rocks exhibit important negative anomalies for Nb, Sr, and Ti, which may be interpreted to reflect low-pressure fractionation of an assemblage of plagioclase (compare large negative Eu anomalies), amphibole and/or clinopyroxene, plus Fe-Ti oxide, if the felsic rocks evolved from more mafic parental magmas by fractional crystallization (Figure 7a).

[47] Such a crystal fractionation model, interpreting plagiogranites as residual liquids left after about 85–90% crystallization of a MORB-like parent, has been favored in many cases [e.g., *Borsi et al.*, 1998] and might be supported by spatial association of felsic veinlets with largely cumulate gabbros within the same Morais-Talhinhas Unit. However, we cannot rule out an alternative interpretation of shear-zone-related partial melting of hydrothermally altered (amphibolitized) basaltic protoliths, as proposed by *Spray and Dunning* [1991] and *Flager and Spray* [1991]. This model appears to be compatible with experimental constraints on dehydration melting of low-K amphibolite [e.g., *Beard*, 1995]. Following separation from its solid residue, the partial melt could further evolve through fractional crystallization. Possibly, this model could better account for the generation, in a dynamic environment, of the Pedras Brancas felsic rock which occurs as a relatively voluminous, isolated dike embedded within ultramafic country rocks.

[48] Initial Nd isotope data for all but the intermediate sample (Mo4, with $\varepsilon Nd_{400} = +5.5$), range from +7.6 to +8.8. These strongly radiogenic signatures overlap those of the associated metagabbros with high Th/Nb ratios, characterized by εNd_{400} between +7.9 and +8.9 (Table 1). They preclude any significant contribution from continental materials, and clearly document the ensimatic origin of subordinate felsic rock types of the Morais ophiolite thrust complex. Unraveling the petrogenetic processes responsible for the generation of these highly differentiated melts is beyond the scope of this study, but the frequent occurrence of positive Ce anomalies is noteworthy, as unusual, highly oxidizing conditions are required to stabilize Ce^{4+}. It may be suggested that materials with a positive Ce anomaly, such as those occurring in oceanic environments (e.g., subducted metalliferous sediments such as manganese nodules [e.g., *Fleet*, 1984]) played a role in the genesis of the felsic derivatives. The fact that sample Mo4 displays both the largest Ce anomaly and the lowest εNd_{400} value (+5.5) measured throughout the Morais ophiolitic units, is tentatively interpreted to reflect igneous generation via partial melting of oceanic crust including a component with a positive Ce anomaly and relatively less radiogenic Nd such as oceanic sediment.

7.2.2.5. Ultramafic Rocks

[49] Two samples of serpentinized harzburgites from the Morais-Talhinhas Unit have rare earth contents lower than or close to chondritic abundances, with positive Eu anomalies (Eu/Eu* 3.6–6.5), and their U-shaped patterns point to substantial enrichment in LREE ($(La/Sm)_N$ = 0.75–1.08)

(Figure 7b) possibly related to the percolation of LREE enriched fluids.

8. Geological Implications of Geochemical Data and Inferred Tectonic Settings

[50] *Ribeiro et al.* [1990] defined for the first time the existence of two subunits within the Morais ophiolitic complex: the Izeda-Remondes unit at the bottom and the Morais-Talhinhas unit on top. These authors considered these two units to show a similar sequence repeated by a reportedly minor tectonic contact, the Limãos thrust. *Dallmeyer and Gil Ibarguchi* [1990, p. 874] put forward, complementarily, that the Morais ophiolitic complex is composed of two contrasting structural units: a lower one dominated by variably foliated amphibolite, and an upper one largely constituted of metaperidotite and metagabbro. Our results document the different incompatible trace element signatures of the mafic rocks from these two units, implying that the Limãos thrust is not merely a minor ductile thrust duplicating the same sequence. The amphibolites from the lower unit are nearly homogeneous and similar to NMORBs (with low Th/Nb and a highly radiogenic Nd isotope signature). Their igneous protoliths might have been formed in an oceanic ridge setting 447 ± 24 Ma ago, based on an imprecise Sm-Nd whole rock isochron. In contrast, the largely gabbroic rocks and minor felsic veins from the upper unit are geochemically and isotopically more variable, but share high Th/Nb ratios and high initial εNd values. These characteristics are interpreted to reflect generation above an intraoceanic subduction zone. U-Pb zircon ages measured in this unit range from 405 ± 1 Ma to 396 ± 1 Ma.

[51] On the basis of these differences, it appears that the amphibolites and the metagabbros of the Morais ophiolitic complex do not share simple cogenetic relationships, and do not correspond to superposed levels of a single oceanic section. This implies that the Morais ophiolitic complex is composite and consists of two distinct, relatively thin tectonic slices that were detached from contrasting oceanic domains before being telescoped and finally imbricated along the Limãos thrust (Figure 2d). In the scope of the various types recognized among arc-related ophiolites [e.g., *Leitch*, 1984; *Moores*, 1998], the upper unit with slab-related geochemical affinities might be interpreted as a fragment of ophiolite produced at the inception of island arc formation and/or during arc evolution, whereas the lower, presumably older, NMORB-like unit should be interpreted differently. It has geochemical characteristics consistent with a fragment of normal oceanic crust entrapped either in the fore arc or in the back-arc region. Alternatively, this unit might represent a basement sliver decoupled from the oceanic crust of a downgoing plate, e.g., by reactivation of ancient normal faults in response to flexural bending, and incorporated into the overlying accretionary wedge. This mechanism of transfer of 1- to 3-km-thick slivers from the downgoing slab to the overriding plate was proposed on the basis of seismic imaging of the modern New Britain accretionary wedge [*Bernstein-Taylor et al.*, 1992a, 1992b].

[52] In so far as the imprecise 447 ± 24 Ma Sm-Nd age provides a reliable estimate for the formation of the metabasalts of the lower tectonic unit, and in the absence of other evidence, we suggest that a piece of oceanic crust was accreted along a normal ridge during Ordovician and/or Silurian times (based on the timescale of *Gradstein et al.* [2004]). This is in agreement with the large-scale geological record of the Iberian Massif [e.g., *Martínez Catalán et al.*, 1997, and references therein] and the Variscan realm in general, which point to a broadly extensional regime throughout this period [e.g., *Pin and Marini*, 1993]. Several authors, however, favor a circa 500 Ma arc setting for some Galician complexes [e.g., *Andonaegui et al.*, 2002, and references therein]. The precise 405 ± 1 Ma and 400 ± 1 Ma zircon ages measured in the mafic section of the upper tectonic unit, on a felsic vein and a cumulate gabbro, respectively, are believed to date a second, independent igneous event, close to the boundary between Early and Middle Devonian [*Gradstein et al.*, 2004]. On the basis of geochemical features, we propose that this igneous event occurred in a suprasubduction zone (SSZ) setting, possibly during the incipient stage of building of an island arc.

[53] Taking into account its peculiar field occurrence, the leucocratic dike (Pedras Brancas) emplaced at 396 ± 1 Ma within the ultramafic section of the Morais-Talhinhas Unit is tentatively interpreted to reflect partial melting, triggered by intraoceanic shearing, of underthrust mafic protoliths at the time of initial tectonic displacement of the SSZ ophiolitic unit. Alternatively, the 396 Ma age could provide a lower limit to magmatism in the roots of the inferred island arc. Albeit chemically much more evolved, the Pedras Brancas felsic rock is reminiscent of a leucogabbro from the Sierra de Careón ophiolitic unit, in the eastern sector of the equivalent Órdenes Complex (NW Spain). Indeed, this rock has a similar geological setting as a leucogabbro dike within serpentinites, and gave the same age of 395 ± 2 Ma [*Pin et al.*, 2002] interpreted to reflect ophiolite generation in a suprasubduction zone environment. The Pedras Brancas dike records high-pressure (HP) equilibration ($\sim$1.4 GPa), but relatively low T, suggesting that it might have been emplaced in the "refrigerated" roots of an island arc, close to a subduction channel. Within analytical uncertainty, its 396 ± 1 Ma U-Pb zircon age is not significantly different from the 392 ± 7 Ma Ar/Ar age measured on green hornblende separated from a coarse-grained metagabbro from the same Morais-Talhinhas Unit, and only 12 ± 6 Ma younger than the 384 ± 5 Ma Ar/Ar age measured for a green hornblende extracted from an amphibolite of the Izeda-Remondes lower structural unit [*Gil Ibarguchi and Dallmeyer*, 1991]. This suggests that uplift of the SSZ Morais-Talhinhas Unit to relatively shallow crustal levels and cooling below $\sim$500°C occurred within a fairly short time span and further indicates that tectonic telescoping with the NMORB-like Izeda-Remondes Unit was achieved within circa 10–20 Myr after igneous generation.

9. A Tentative Geodynamic Scenario for the Morais and Bragança Allochthonous Units

[54] Together with structural features and P-T estimates for the tectonometamorphic episodes, the radiometric ages and geodynamic settings inferred from geochemical data provide basic constraints for attempting to elaborate a geodynamic model, including data available from neighboring units and the closely related and complementary Bragança Massif.

9.1. Subduction and Amalgamation of Younger and Older Oceanic Lithosphere

[55] The observation that generation of SSZ ophiolite, and/or inception of arc magmatism, occurred shortly before deformation and metamorphism is in keeping with the evolution of many ophiolites worldwide, which are tectonically emplaced onto continental crust by collision with a continental margin about 10–20 Myr after their igneous formation [e.g., *Gealey*, 1980; *Moores*, 1982, 1998]. Moreover, we emphasize that the tectonometamorphic overprinting of the Morais ophiolite was broadly coeval at the regional scale with the development of high-temperature/high-pressure (HT-HP) metamorphism in other, initially separate domains now thrust into the same nappe stack. This high-grade metamorphic episode is documented by the occurrence of eclogite-bearing paragneisses and high-pressure granulites in tectonic units lying at a higher structural level in the nappe stack of several allochthonous complexes of NW Iberia, especially the Cabo Ortegal Complex. In the Morais Complex, high-grade units are rare, represented by three small bodies of retrogressed mafic granulites imbricated between the ophiolitic thrust complex and overlying, medium-grade uppermost tectonic unit (Lagoa orthogneisses and micaschists), and various hectometer-scale duplexes in the contact with the underlying lower allochthonous complex (Figure 2a). The high-grade rocks are better exposed in the neighboring Bragança Complex, where a circa 390 Ma U-Pb zircon age has been published recently on an eclogite pod within paragneisses [*Roger and Matte*, 2005]. This age is in agreement with estimates obtained in the more thoroughly studied Cabo Ortegal Complex [e.g., *Santos Zalduegui et al.*, 1996; *Ordóñez Casado et al.*, 2001, and references therein]. Ar/Ar hornblende ages measured in the high-grade unit and the underlying ophiolitic unit of the Bragança Massif [*Dallmeyer et al.*, 1991] suggest that both units were assembled prior to 385–390 Ma and subsequently cooled below $\sim$500°C. On the basis of the high-pressure estimates and "clockwise metamorphic path typical of rocks formed in a subduction zone and later exhumed" [*Marques et al.*, 1996, p. 748], these high-grade units are interpreted to represent tectonic slivers of deeply subducted oceanic and attenuated continental crust. Initially, these units probably belonged to a passive margin of Early Paleozoic age, thinned by faulting and ductile stretching and intruded by mafic bodies during the rifting process [e.g., *Pin and Vielzeuf*, 1983; *Pin and Marini*, 1993].

[56] Therefore two critical pieces of the Early Devonian (circa 400 Ma) "geodynamic puzzle" might be identified in north Portugal and NW Spain:

[57] 1. The remnants of the upper plate of an intraoceanic subduction zone; these would consist of the suprasubduction zone ophiolitic units documented in the Morais-Talhinhas Unit (this study) and the Sierra de Careón Unit [*Díaz García et al.*, 1999; *Pin et al.*, 2002] in the Órdenes Complex. We also include in this group, tentatively, the podiform chromitite-bearing ultrabasic rocks from the Bragança Complex, that were interpreted as fragments of island arc mantle by *Bridges et al.* [1995].

[58] 2. The remnants of the lower plate of a subduction system, represented by HP granulites and eclogite-bearing metasediments exposed in the Bragança Complex and Galician occurrences. Bearing in mind the presence of abundant sediments in the subducted units, and the Ar/Ar ages in the 390–380 Myr range, interpreted as reflecting rapid uplift to shallow levels, it is likely that the subduction was not steady state in Early Devonian times, but instead was halted by collision with a passive continental margin [*Pin et al.*, 2002].

9.2. An Arc-Continent Collision Model for Eo-Variscan NW Iberia

[59] We propose, as a working hypothesis, an arc-collision model for the early Variscan (circa 410–380 Ma) geodynamic evolution of the Morais and Bragança complexes, as already suggested by *Bridges et al.* [1995] and *Von Raumer et al.* [2003]. In this scenario, depicted in Figure 8a, a plate convergence regime prevailed in Ordovician through Devonian times, with an intraoceanic subduction zone dipping away from a northern continental block (in palinspastically restored coordinates; see Figure 8) that rifted from Gondwana during a prior (Early Paleozoic?) period of plate separation. This subduction zone was fringed, at least locally, by an island arc facing the continental block. As a result of subduction zone retreat, this island arc migrated [cf. *Hamilton*, 1988, 1995], thereby approaching the passive continental margin conjugate to the Gondwana margin (Central Iberian Zone). When the arc/fore-arc system first impinged promontories of the continental margin, local extension related to continuing subduction hinge retreat within oceanic embayments of the initial passive margin promoted the generation of small SSZ ophiolites in the upper plate [see *Pin et al.*, 2002], following a model first proposed by *Edelman* [1988; cf. *Hamilton*, 2003]. At the same time, the salients of the former passive margin were overridden by the ophiolitic leading edge of the upper plate, which ramped onto the continent as the final result of the prior intraoceanic subduction. As a result of the flexural rigidity of the subducting lithosphere and the buoyancy force generated by the subducting continental crust [e.g., *Cloos*, 1993], the overriding oceanic plate failed under compression, probably along the weakened arc/fore-arc lithosphere boundary, causing the arc to override the fore-arc (NMORB like) region [cf. *Tang and Chemenda*, 2000]. Concomitantly, HP-HT metamorphism (generating eclogites and "type 1" high-pressure granulites

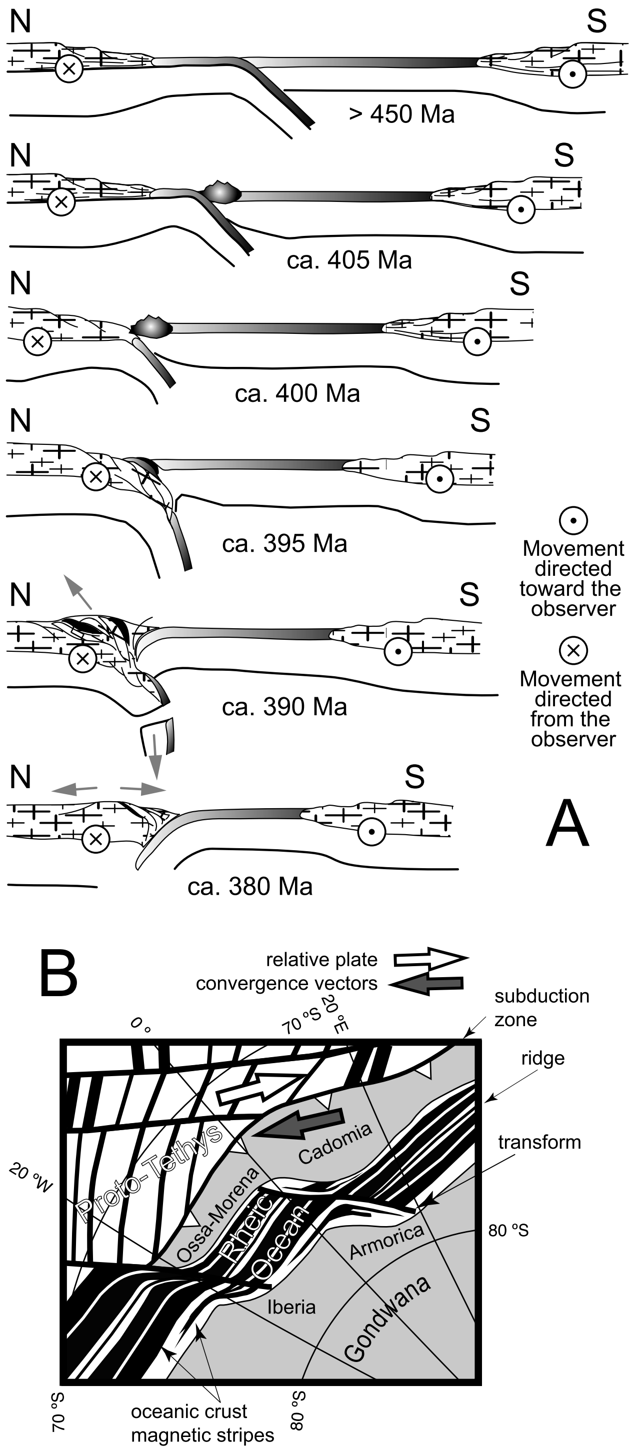

Figure 8. (a) Geodynamic models proposed for the genesis and further evolution of the ophiolitic units of the Morais Complex (see text for explanations). (b) Paleotectonic plate reconstruction of the Cadomia/Ossa-Morena peri-Gondwanan realm for the Early Ordovician (circa 480 Ma) inspired by *Von Raumer et al.* [2003] model. Dominant transcurrent dextral plate movements highlighted with arrows.

of *Marques et al.* [1996]) occurred in the deeply buried units of the attenuated crust and overlying sediments of the former passive continental margin. As progressively thicker, nonsubductible continental crust entered the subduction zone, the subduction eventually stopped due to increasing gravitational resistance, and the stage of arc-continental margin collision proper was reached. This continental locking of subduction generated severe compressional strain and shortening in the lower plate, with nappe formation and folding, that resulted in the formation of a collision-related accretionary complex at the lithospheric scale. At least part of the deeply subducted (~50 km) crustal units eventually got decoupled from their mantle underpinnings, and rose buoyantly toward shallow levels ("buoyant escape" or "eduction") as extrusions in the former interplate zone, possibly following mechanisms such as those proposed by, e.g., *Andersen et al.* [1991] and *Chemenda et al.* [1996] to model exhumation of subducted continental material [see also *Brueckner and van Roermund*, 2004]. In this way, HP units from the lower continental plate might have been ductilely imbricated within lower pressure units from the overlying accretionary wedge, including the previously "obducted" SSZ and NMORB oceanic units.

[60] At greater depth in the mantle, the arc-continent collision was probably accompanied, or shortly followed, by rupture and further sinking into the upper mantle of the dense, eclogite-facies oceanic lithosphere segment of the subducted plate (slab break-off) which presumably provided the major force (slab pull) driving subduction. An isostatic rebound of the buoyant, subducted continental lithosphere ensued, further contributing to the rapid tectonic exhumation of the HP units and the concomitant extensional collapse of the overlying accretionary wedge. This might also have caused further ductile imbrication of tectonic flakes from both the upper and lower plate of the former subduction system, and could explain, at least in part, the occurrence of ophiolitic fragments as very thin tectonic units.

[61] As a consequence of continuing plate convergence at a larger scale, the stacking of the collisional nappe pile was followed, or even accompanied, by a jump in the locus of active convergence and subduction polarity reversal, as observed in many arc-continent collision systems [e.g., *Hamilton*, 1988; *Teng et al.*, 2000; *Konstantinovskaia*, 2001]. This reversal might be viewed as a plate-scale back thrusting of the collided arc, resulting from indentation of the weak subducted continental margin by the strong lithospheric mantle of the former upper plate. In this hypothesis, the oceanic lithosphere plate that lay behind the initial island arc (remnant oceanic basin) began to subduct beneath the recently accreted island arc-passive margin system [cf. *Sommer and Katzung*, 2006]. From this stage onward, the nappes previously stacked during the arc-continent collision and now uplifted to mid to upper crustal level, acted as the upper plate of a new subduction system dipping away from the Gondwana (Central Iberian) margin. The retreat of this newly formed subduction might have generated an extensional regime in the overlying region (nappe stack), and promoted its further uprise to shallow level, as proposed by *Thomson et al.* [1998] for exhumation

of HP rocks in the Aegean region. This second-stage subduction probably consumed older and older oceanic crust during the Late Devonian (from circa 385 Ma to circa 360 Ma [*Gradstein et al.*, 2004]). It might even have operated until the end of the Early Carboniferous, as suggested by the 336 ± 2 Ma Ar/Ar plateau age [*Gil Ibarguchi and Dallmeyer*, 1991] measured on white mica from blueschist facies rocks of the distal parts of the Gondwana margin, insofar as this age does not reflect late stage resetting of the Ar/Ar chronometer. In any case, the peak conditions recorded in these rocks (>11 kbar, 420°C) and the very low temperature/pressure gradient they document, are suggestive of a cold subduction zone environment, such as those developed where relatively fast convergence of old oceanic plate has been ongoing for tens of millions of years [*Cloos*, 1993]. In this scheme, the Morais and Bragança thrust complexes might have behaved as the backstop of the newly formed, east verging accretionary wedge, produced by progressive westward underthrusting of the Gondwana margin along deeper and deeper, and more external active thrust planes.

9.3. Variscan Polycyclic Geodynamic Evolution Versus Polyorogenic Tectonism

[62] The tentative model proposed above provides an attempt to interpret in plate tectonic terms the still rather poorly understood early stages of collisional tectonics recorded within the allochthonous complexes and may account for some characteristics of the geological evolution of this segment of the Variscan Belt. In particular, a rapid tectonic exhumation is implied by the 385–375 Ma ages [*Dallmeyer et al.*, 1991] measured throughout the tectonic pile. The short duration inferred for the major tectonometamorphic evolution (from circa 395 to circa 390 Ma) is in good agreement with the timescale documented for typical arc-continent collision, from initial impingement to final collapse, in the modern analogue of Taiwan [e.g., *Chemenda et al.*, 1997; *Huang et al.*, 2000]. Also, it is in accordance with the dated amalgamation of units from disparate tectonic settings that occurred at an eo-Hercynian orogenic channel in the Cabo Ortegal Complex [*Ábalos et al.*, 2003] and with their subsequent, shared synmetamorphic reworking.

[63] Nappe juxtaposition, internal deformation and high-pressure metamorphism took place under polyphase tectonism. A number of "geometrical" deformation phases have been unraveled by different authors in the allochthonous complexes of NW Iberia, including those of Bragança and Morais [*Anthonioz*, 1972; *Ribeiro*, 1974; *Marques et al.*, 1996, and references therein] that have been related to either a polycyclic evolution, involving ancient Precambrian events, or a polyphase, monocyclic evolution during the Variscan orogeny. In the latter case, geodynamic models invoke deformation in either discrete or contrasted tectonic regimes [e.g., *Martínez Catalán et al.*, 2002], or in a single, kinematically simple but polyphase, tectonic regime [e.g., *Azcárraga et al.*, 2003; *Ábalos et al.*, 2003].

[64] From our point of view, the long-recognized polycyclism of the Trás-os-Montes high-grade complexes can be explained readily without invoking a Precambrian orogeny, for which radiometric or stratigraphic evidence is otherwise missing (see *Dallmeyer and Tucker* [1993] and discussion by *Ribeiro et al.* [1993]). In our model, the early evolutionary stages of the upper allochthonous units (ascribed to a pre-Variscan cycle 1 by *Marques et al.* [1996]) are decoupled both in space and time from Variscan events (ascribed to their cycle 2), since they occurred along the opposite margin of an oceanic domain, and several tens of millions of years earlier, although this time span could vary largely from place to place, depending on the width of the remnant oceanic basin, and the rate and obliquity of convergence during Late Devonian and Early Carboniferous times.

[65] The first cycle, here dated as Silurian to Early Mid-Devonian, involved arc-continent collision probably constrained by local continental margin complexities. This probably did not produce areally extensive major elevation above sea level, as suggested by the lack of large amounts of coeval synorogenic sediments. In marked contrast, the second cycle (mostly Early Carboniferous) corresponded to the final suturing of oceanic domains, and was concluded by continent-continent collision. This gave rise to the erosion and deposition of large volumes of synorogenic and post-orogenic sediments, as recorded in the Variscan foreland basin.

[66] The oldest synorogenic sediments are alternating shales and graywackes of the Gimonde Formation, in the transition zone of the autochthonous-lower parautochthonous sequences of the Bragança Massif [*Ribeiro and Pereira*, 1997]. These sediments were deposited during the Early Frasnian (circa 385 Ma) and contain reworked spores of Lochkovian and Praguian age [*Pereira et al.*, 1999]. In the absence of a provenance study, it is not clear whether these sediments were derived from an east facing (in present coordinates) accretionary wedge belonging to the upper plate, or alternatively, from a flexural bulge of the lower lithospheric plate (bearing the Gondwana margin). In our model it began to subduct toward the west after the arc-continent collision, as a result of polarity reversal. No detrital components of high-grade metamorphic and/or ophiolitic provenance have been reported from the Gimonde Formation, or from the coeval San Vitero Formation in Spain [*González Clavijo and Martínez Catalán*, 2002].

[67] The structural record of the high-grade unit of the Bragança Complex [*Marques et al.*, 1996] and of the ophiolitic unit at Bragança and Morais [*Marques et al.*, 1992] appears to reflect opposite kinematics for the early (pre-Variscan) and late stage (Variscan in the strictest sense) compression stages. Additionally, an intervening extensional regime was proposed by *Marques* [1994] and *Marques et al.* [1996].

[68] The complexity of the superposed deformations described by these authors can be reinterpreted in the context of our model by ascribing their earliest, syneclogite facies deformation phase (their D1) to the subduction of the passive margin of the "western" continental block beneath the overriding arc. Their subsequent metamorphism and syngranulite facies pervasive deformation, (D2, with a top-

to-the-west sense of movement of hanging wall blocks), could reflect the major compressional phase of the inferred arc-continent collision and the early stage of uplift of a subducted crustal sliver through buoyant escape. The forthcoming noncoaxial shear deformations (referred to as D3 to D5 by *Marques et al.* [1996]) occurred under progressively decreasing P and T conditions, and might have accompanied the uplift and rapid exhumation of the subducted passive margin, after the oceanic segment of the subducting slab broke off. This process was completed at circa 385–390 Ma in the Bragança Complex, as suggested by hornblende Ar/Ar ages [*Dallmeyer et al.*, 1991]. On the basis of a similar 384 ± 5 Ma Ar/Ar hornblende age in the Morais-Talhinhas upper ophiolitic unit [*Dallmeyer and Gil Ibarguchi*, 1990], the ductile deformations of ophiolitic amphibolites for which a top-to-the-east sense of shear has been associated can be related to a comparable tectonic setting after the onset of back thrusting, induced by indentation of the subduction hanging wall block (attenuated continental lithosphere) by the still converging and downgoing oceanic plate (Figure 8). This hypothesis contrasts with that proposed by *Pereira et al.* [2004] and *Ribeiro et al.* [2004, and references therein], in which kinematic reversals are ascribed to separate orogenic cycles. Taking into account modern plate tectonic analogues in oblique subduction settings, we suggest that kinematic changes such as those recorded by the ductile deformation of Morais ophiolites can be explained in terms of back thrusting and arc subduction transfer in a context of uniform oblique plate convergence.

[69] *Marques et al.* [1992] reported the occurrence of late, conjugate shear zones in the Bragança Complex with an overall geometry and kinematics they ascribed to ductile normal faults developed in extensional regimes. *Azcárraga et al.* [2003] described similar structures in the Cabo Ortegal Complex, but taking into account their low-angle geometrical relationship with respect to the regional foliation, they related such conjugate shear zones to intense along-lineation stretching associated with nappe emplacement under retrogressive metamorphic conditions. These structures denote subhorizontal intense stretching and subvertical thinning.

[70] During the subsequent tectonic evolution, the ophiolitic and overlying units were kept to shallow structural levels, below ~500°C, and did not suffer significant internal deformation. This evolution was characterized by nappe emplacement with east directed components. Geometrically, such a tectonic movement is moderately oblique to the orogenic trend in the area, which is in good accordance with an oblique subduction-collision setting. Such a setting has been proposed based upon comparable arguments in other allochthonous complexes in Spain (e.g., Malpica-Tuy and Cabo Ortegal). The latest orogenic deformations are recorded by structures reflecting shortening normal to the trend of the orogen together with wrench tectonism along strike [e.g., *González Clavijo and Martínez Catalán*, 2002]. During these events, active deformation was localized toward the footwall of the structural units underlying the ophiolite and prograded toward more proximal domains of the Central Iberian passive margin [*Martínez Catalán et al.*, 1997]. In this regard, low-T and relatively high-P metamorphism of the lowermost allochthonous unit and of the para-autochthonous sequences was dated circa 330–340 Ma [*Gil Ibarguchi and Dallmeyer*, 1991; *Dallmeyer et al.*, 1997].

10. Summary and Conclusion

[71] The trace element and isotopic (U-Pb, Sm-Nd) study documents significant differences between the two structural subunits defined in the Morais ophiolite complex. Amphibolites from the lower, strongly deformed Izeda-Remondes Unit are homogeneous and similar to NMORBs, with low Th/Nb and a highly radiogenic Nd isotope signature. They might derive from igneous protoliths formed in an oceanic ridge setting possibly at 447 ± 24 Ma (Sm-Nd whole rock isochron). In contrast, the upper unit (Morais-Talhinhas) consists largely of metaperidotites together with gabbroic rocks and minor felsic veins, whose U-Pb zircon ages range from 405 ± 1 Ma to 396 ± 1 Ma. Although geochemically more variable, rocks from this unit have high Th/Nb ratios and elevated εNd values, interpreted to reflect generation above an intraoceanic subduction zone. These data demonstrate that the Morais ophiolite complex is composite. It consists of two distinct, relatively thin, tectonic units that were derived from distinct oceanic domains before their tectonic superposition. The tectonometamorphic overprinting of the ophiolitic units, including imbrication with high- and medium-grade nappe units, was completed prior to circa 385 Ma.

[72] Special emphasis is put on the space and time association, at the regional scale (i.e., Bragança and Cabo Ortegal complexes), of suprasubduction zone ophiolitic units with eclogite-bearing metasedimentary units, interpreted to document metamorphism of a subducted passive continental margin. On the basis of this association, we propose to interpret the early stage of the eo-Variscan evolution (circa $\geq$ 405–400 Ma) in terms of intraoceanic subduction, followed by arc-continent collision ~395 Myr ago. During this episode, tectonic slivers from both the upper and lower plates were imbricated, probably in deep domains of a lithospheric-scale shear zone (a subduction channel). The arc-continent collision probably involved a relatively narrow strip of terrain, and apparently did not produce extensive topographic relief in view of the scarcity of coeval synorogenic sediments. The collision zone might have had significant elongation along strike, as suggested by a broadly analogous evolution inferred in the northern French Massif Central [*Pin and Paquette*, 2002].

[73] The final emplacement of the ophiolite-bearing nappes onto the Gondwana margin (i.e., Central Iberian Zone) occurred during a separate event in early Carboniferous times (circa 360–330 Ma), when the complete suturing of former oceanic domains led to continent-continent collision. During this second tectonic cycle, the nappe stack behaved passively and was transported in a piggyback style, on top of eastward propagating thrust nappes forming the lower allochthonous units and the para-autochthonous units of pre-Mesozoic northwestern Iberia.

[74] In this model, the genetically composite ophiolite thrust complex exposed in the Morais and Bragança allochthonous massifs represents remnants of a transported suture created during early stage arc-continent collisional events. It is devoid of any simple, direct tectonic significance with regard to the Variscan (Carboniferous) events. The igneous generation of the suprasubduction zone unit is interpreted to record early stages of intraoceanic subduction (405–400 Ma). This was followed shortly by major tectonometamorphic overprinting as a result of impingement with a passive continental shelf (arc-continent collision), ~395 Myr ago, well before the total closure of oceanic domains. This early compressional tectonic episode caused the development of HP-HT metamorphism in the partially subducted passive margin, the complex imbrication of tectonic flakes detached from various structural levels of both lower and upper plates, and was followed by rapid synconvergence exhumation. The subsequent evolution (i.e., telescoping with the Central Iberian domain) appears to reflect a separate event which began at least 30 Myr later, close to the Devonian-Carboniferous boundary. This geodynamic model involves two broadly polyphase, but independent tectonometamorphic events, and may account for the long-recognized polycyclic features of the Morais and Bragança complexes.

[75] **Acknowledgments.** We are grateful to F. Cantagrel, C. Bosq, and K. David for help provided during the chemical and isotopic analyses. This study was supported by travel funding supplied in the scope of the "Picasso" Spanish-French cooperation project and represents a contribution to IGCP projects 453 and 497. Financial support by the Spanish MEC (CGL2005-03157) is acknowledged.

References

Ábalos, B., P. Puelles, and J. I. Gil Ibarguchi (2003), Structural assemblage of high-pressure mantle and crustal rocks in a subduction channel (Cabo Ortegal, NW Spain), *Tectonics*, *22*(2), 1006, doi:10.1029/2002TC001405.

Andersen, T. B., J. F. Jamtveit, J. F. Dewey, and E. Swensson (1991), Subduction and eduction of continental crust: Major mechanisms during continent-continent collision and orogenic extensional collapse: A model based on the southern Norvegian Caledonides, *Terra Nova*, *3*, 303–310.

Andonaegui, P., J. González del Tánago, R. Arenas, J. Abati, J. R. Martínez Catalán, M. Peinado, and F. Díaz García (2002), Tectonic setting of the Monte Castelo gabbro (Órdenes Complex, north-western Iberian Massif): Evidence for an arc-related terrane in the hanging wall to the Variscan suture, in *Variscan-Appalachian Dynamics: The Building of the Late Paleozoic Basement*, edited by J. R. Martínez Catalán et al., *Spec. Pap. Geol. Soc. Am.*, *364*, 37–56.

Anthonioz, P. M. (1972), Les complexes polymétamorphiques Précambriens de Morais et Bragança (NE Portugal): Étude pétrographique et structurel, *Memoria 20*, 112 pp., Serv. Geol. de Portugal, Lisbon.

Azcárraga, J., B. Ábalos, and J. I. Gil Ibarguchi (2003), On the relationship between km-scale sheath folds, ductile thrusts and minor structures in the basal high-pressure units of the Cabo Ortegal complex (NW Spain), *J. Struct. Geol.*, *24*, 1971–1989, doi:S0191-8141(02)00005-6.

Beard, J. S. (1995), Experimental, geological, and geochemical constraints on the origins of low-K silicic magmas in oceanic arcs, *J. Geophys. Res.*, *100*, 15,593–15,600.

Bernstein-Taylor, B. L., K. S. Kirchoff-Stein, E. A. Silver, and D. Reed (1992a), Large-scale duplexes within the New Britain accretionary wedge: A possible example of accreted ophiolitic slivers, *Tectonics*, *11*, 732–752.

Bernstein-Taylor, B. L., K. M. Brown, E. A. Silver, and K. S. Kirchoff-Stein (1992b), Basement slivers within the New Britain accretionary wedge: Implications for the emplacement of some ophiolitic slivers, *Tectonics*, *11*, 753–765.

Blundy, J. D., and T. J. B. Holland (1990), Calcic amphibole equilibria and a new amphibole-plagioclase geothermometer, *Contrib. Mineral. Petrol.*, *104*, 208–224.

Borsi, L., R. Petrini, L. Cortesogno, L. Gaggero, and L. Crispini (1998), Modelling of plagiogranite genesis for the evolution of the Ligurian-Piedmont oceanic basin (Ligurian Alps, Italy), *Neues Jahrb. Mineral. Monatsh.*, *5*, 208–220.

Bridges, J. C., H. M. Prichard, and C. A. Meireles (1995), Podiform chromitite-bearing ultrabasic rocks from the Bragança Massif, northern Portugal: Fragments of island arc mantle?, *Geol. Mag.*, *132*, 39–49.

Brueckner, H. K., and H. L. M. van Roermund (2004), Dunk tectonics: A multiple subduction/eduction model for the evolution of the Scandinavian Caledonides, *Tectonics*, *23*, TC2004, doi:10.1029/2003TC001502.

Chemenda, A. I., M. Mattauer, and A. N. Bokun (1996), Continental subduction and a mechanism for exhumation of high-pressure metamorphic rocks: New modelling and field data from Oman, *Earth Planet. Sci. Lett.*, *143*, 173–182.

Chemenda, A. I., R. K. Yang, C. H. Hsieh, and A. L. Groholsky (1997), Evolutionary model for the Taiwan collision based on physical modelling, *Tectonophysics*, *274*, 253–274.

Cloos, M. (1993), Lithospheric buoyancy and collisional orogenesis: Subduction of oceanic plateaus, continental margins, island arcs, spreading ridges, and seamounts, *Geol. Soc. Am. Bull.*, *105*, 715–737.

Coleman, R. G., and M. M. Donato (1979), Oceanic plagiogranites revisited, in *Trondhjemites, Dacites and Related Rocks*, edited by F. Barker, pp. 149–168, Elsevier, New York.

Crowley, Q. G., P. A. Floyd, J. A. Winchester, W. Franke, and J. G. Holland (2000), Early Palaeozoic rift-related magmatism in Variscan Europe: Fragmentation of the Armorican Terrane Assemblage, *Terra Nova*, *12*, 171–180.

Dallmeyer, R. D., and J. I. Gil Ibarguchi (1990), Age of amphibolitic metamorphism in the ophiolitic unit of the Morais allochthon (Portugal): Implications for early Hercynian orogenesis in the Iberian Massif, *J. Geol. Soc. London*, *147*, 873–878.

Dallmeyer, R. D., and R. D. Tucker (1993), U-Pb zircon age for the Lagoa augen gneiss, Morais Complex, Portugal: Tectonic implications, *J. Geol. Soc. London*, *150*, 405–410.

Dallmeyer, R. D., A. Ribeiro, and F. Marques (1991), Polyphase Variscan emplacement of exotic terranes (Morais and Bragança Massifs) onto Iberian successions: Evidence from $^{40}Ar/^{39}Ar$ mineral ages, *Lithos*, *27*, 133–144.

Dallmeyer, R. D., J. R. Martínez Catalán, R. Arenas, J. I. Gil Ibarguchi, G. G. Alonso, P. Farias, F. Bastida, and J. Aller (1997), Diachronous Variscan tectonothermal activity in the NW Iberian Massif: Evidence from $^{40}Ar/^{39}Ar$ dating of regional fabrics, *Tectonophysics*, *277*, 307–337.

Díaz García, F., R. Arenas, J. R. Martínez Catalán, J. González del Tánago, and G. Dunning (1999), Tectonic evolution of the Careón ophiolite (northwest Spain): A remnant of oceanic lithosphere in the Variscan belt, *J. Geol.*, *107*, 587–605.

Dupuy, C., and J. Dostal (1984), Trace element geochemistry of some continental tholeiites, *Earth Planet. Sci. Lett.*, *67*, 61–69.

Eckert, J. O., Jr., R. C. Newton, and O. J. Kleppa (1991), The ΔH of reaction and recalibration of garnet-pyroxene-plagioclase-quartz geobarmeters in the CMAS system by solution calorimetry, *Am. Mineral.*, *76*, 148–160.

Edelman, S. H. (1988), Ophiolite generation and emplacement by rapid subduction hinge retreat on a continent-bearing plate, *Geology*, *16*, 311–313.

Elthon, D. (1991), Geochemical evidence for formation of the Bay of Islands ophiolite above a subduction zone, *Nature*, *234*, 140–143.

Erlank, A. J., and E. J. D. Kable (1976), The significance of incompatible elements in Mid-Atlantic Ridge basalts from 45°N, with particular reference to Zr/Nb, *Contrib. Mineral. Petrol.*, *54*, 281–291.

Fernández Suárez, J., F. Corfu, R. Arenas, A. Marcos, J. R. Martínez Catalán, F. Díaz García, J. Abati, and F. J. Fernández (2002), U-Pb evidence for a polyorogenic evolution of the HP-HT units of the NW Iberian Massif, *Contrib. Mineral. Petrol.*, *143*, 236–253.

Flager, P. A., and J. G. Spray (1991), Generation of plagiogranite by amphibolite anatexis in oceanic shear zones, *Geology*, *19*, 70–73.

Fleet, A. J. (1984), Aqueous and sedimentary geochemistry of the rare earth elements, in *Rare Earth Element Geochemistry*, edited by P. Henderson, pp. 343–373, Elsevier, New York.

Fletcher, I. R., and K. J. R. Rossman (1982), Precise determination of εNd from Sm-Nd isochron, *Geochim. Cosmochim. Acta*, *46*, 1983–1987.

Gealey, W. K. (1980), Ophiolite obduction mechanism, in *Ophiolites*, edited by A. Panayiotou, pp. 228–243, Geol. Surv. Dep., Nicosia, Cyprus.

Gil Ibarguchi, J. I., and R. D. Dallmeyer (1991), Hercynian blueschist metamorphism in North Portugal: Tectonic implication, *J. Metamorph. Geol.*, *9*, 539–549.

González Clavijo, E., and J. R. Martínez Catalán (2002), Stratigraphic record of preorogenic to synorogenic sedimentation, and tectonic evolution of imbricate units in the Alcañices synform (northwestern Iberian Massif), in *Variscan-Appalachian Dynamics: The Building of the Late Paleozoic Basement*, edited by J. R. Martínez Catalán et al., *Spec. Pap. Geol. Soc. Am.*, *364*, 17–35.

Gradstein, F. M., J. G. Ogg, A. G. Smith, W. Bleeker, and L. J. Lourens (2004), A new geologic time

scale, with special reference to Precambrian and Neogene, *Episodes*, *27*, 83–100.
Hamilton, W. B. (1988), Plate tectonics and island arcs, *Geol. Soc. Am. Bull.*, *100*, 1503–1527.
Hamilton, W. B. (1995), Subduction systems and magmatism, in *Volcanism Associated With Extension at Consuming Plate Margins*, edited by J. L. Smellie, *Geol. Soc. Spec. Publ.*, *81*, 3–28.
Hamilton, W. B. (2003), An alternative Earth, *GSA Today*, *13*(11), 4–12.
Holland, T. J. B., and J. D. Blundy (1994), Non-ideal interactions in calcic amphiboles and their bearing on amphibole-plagioclase thermometry, *Contrib. Mineral. Petrol.*, *116*, 433–447.
Huang, C. Y., P. B. Yuan, C. W. Lin, T. K. Wang, and C. P. Chang (2000), Geodynamic processes of Taiwan arc-continent collision and comparison with analogs in Timor, Papua New Guinea, Urals and Corsica, *Tectonophysics*, *325*, 1–21.
Jenner, G., G. R. Dunning, J. Malpas, M. Brown, and T. Brace (1991), Bay of Islands and Little Port complexes, revisited: Age, geochemical and isotopic evidence confirm suprasubduction zone origin, *Can. J. Earth Sci.*, *28*, 1635–1652.
Kohn, M. J., and F. S. Spear (1990), Two new geobarometers for garnet amphibolites, with applications to southeastern Vermont, *Am. Mineral.*, *75*, 89–96.
Konstantinovskaia, E. A. (2001), Arc-continent collision and subduction reversal in the Cenozoic evolution of the northwest Pacific: An example from Kamchatka (NE Russia), *Tectonophysics*, *333*, 75–94.
Krogh, T. E. (1982), Improved accuracy of U-Pb zircon ages by the creation of more concordant systems using an air abrasion technique, *Geochim. Cosmochim. Acta*, *46*, 636–649.
Laird, J., and A. Albee (1981), Pressure, temperature and time indicators in mafic schists: Their application to reconstructing the polymetamorphic history of Vermont, *Am. J. Sci.*, *281*, 127–175.
Leitch, E. C. (1984), Island arc elements and arc-related ophiolites, *Tectonophysics*, *106*, 177–203.
Marques, F. G., A. Ribeiro, and E. Pereira (1992), Tectonic evolution of the deep crust: Variscan reactivation by extension and thrusting of Precambrian basement in the Bragança and Morais Massifs (Trás-os-Montes, NE Portugal), *Geodin. Acta*, *5*, 135–151.
Marques, F. O. (1994), Estudo tectónico das rochas infracrustais do manto de soco do SW do Maçico de Bragança (Trás-Os-Montes), thesis, 196 pp., Univ. Lisboa, Lisbon, Portugal.
Marques, F. O., A. Ribeiro, and J. Munhá (1996), Geodynamic evolution of the Continental Allochthonous Terrane (CAT) of the Bragança Nappe Complex, NE Portugal, *Tectonics*, *15*, 747–762.
Martínez Catalán, J. R., R. Arenas, F. Díaz García, F. J. Rubio Pascual, J. Abati, and J. Marquínez (1996), Variscan exhumation of a subducted Paleozoic continental margin: The basal units of the Ordenes Complex, Galicia, NW Spain, *Tectonics*, *15*, 106–121.
Martínez Catalán, J. R., R. Arenas, F. Díaz García, and J. Abati (1997), Variscan accretionary complex of northwest Iberia: Terrane correlation and succession of tectonothermal events, *Geology*, *25*, 1103–1106.
Martínez Catalán, J. R., F. Díaz García, R. Arenas, J. Abati, P. Castiñeiras, P. González Cuadra, J. Gómez Barreiro, and F. Rubio Pascual (2002), Thrust and detachment systems in the Ordenes Complex (northwestern Spain): Implications for the Variscan-Appalachian geodynamics, in *Variscan-Appalachian Dynamics: The Building of the Late Paleozoic Basement*, edited by J. R. Martínez Catalán et al., *Spec. Pap. Geol. Soc. Am.*, *364*, 163–182.
Mattinson, J. M. (1994), A study of complex discordance in zircons using step-wise dissolution techniques, *Contrib. Mineral. Petrol.*, *116*, 117–129.
Meireles, C. (2000a), Carta geológica de Portugal à escala 1:50,000, *Not. Explicativa Folha 3-D (Espinhosela)*, Inst. Geol. Min. de Portugal, Lisbon.
Meireles, C. (2000b), Carta Geológica de Portugal à escala 1:50,000, *Not. Explicativa Folha 4-C (Deilãoa)*, Inst. Geol. Min. de Portugal, Lisbon.
Meschede, M. A. (1986), A method for discrimination between different types of mid-ocean ridge basalts and continental tholeiites with the Nb-Zr-Y diagram, *Chem. Geol.*, *56*, 207–218.
Moores, E. M. (1982), Origin and emplacement of ophiolites, *Rev. Geophys.*, *20*, 735–760.
Moores, E. M. (1998), Ophiolites, the Sierra Nevada, "Cordillera", and orogeny along the Pacific and Caribbean margins of North and South America, *Int. Geol. Rev.*, *40*, 40–54.
Nakamura, N. (1974), Determination of REE, Ba, Fe, Mg, Na, and K in carbonaceous and ordinary chondrites, *Geochim. Cosmochim. Acta*, *38*, 757–775.
Ordóñez Casado, B., D. Gebauer, H.-J. Schäfer, J. I. Gil Ibarguchi, and J. J. Peucat (2001), A single Devonian subduction event for the HP/HT metamorphism of the Cabo Ortegal Complex within the Iberian Massif, *Tectonophysics*, *232*, 359–385.
Pearce, J. A. (1991), Ocean floor comes ashore, *Nature*, *354*, 110–111.
Pearce, J. A., and J. R. Cann (1973), Tectonic setting of basic volcanic rocks determined using trace element analyses, *Earth Planet. Sci. Lett.*, *19*, 290–300.
Pearce, J. A., S. J. Lippard, and S. Roberts (1984), Characteristics and tectonic significance of suprasubduction zone ophiolites, in *Marginal Basins*, edited by B. P. Kokelaar and M. F. Howells, *Geol. Soc. Am. Spec. Publ.*, *16*, 77–94.
Pereira, E. (Coordinator) (2001), Carta geológica de Portugal à escala 1:200,000, Inst. Geol. Min. de Portugal, Lisbon.
Pereira, E., A. Ribeiro, and J. Rebelo (2003), Carta Geológica de Portugal à escala 1:50,000, *Not. Explicativa Folha 11-B (Mogadouro)*, Inst. Geol. Min. de Portugal, Lisbon.
Pereira, E., A. Ribeiro, P. Castro, and D. Oliveira (2004), Complexo ofiolítico Varisco do Maciço de Morais (NE de Trás-os-Montes, Portugal), in *Complejos Ofiolíticos en Iberoamérica*, edited by E. S. Pereira, R. Castroviejo, and F. Ortiz, pp. 267–286, Proyecto XIII.1–CYTED, Madrid.
Pereira, Z., C. Meireles, and E. Pereira (1999), Upper Devonian palynomorphs of NE sector of Trás-os-Montes (central Iberian Zone), paper presented at XV Reunión Geológica del Oeste Peninsular, IUGS-UNESCO, Badajoz, Spain.
Pin, C. (1990), Variscan oceans: Ages, origins and geodynamic implications inferred from geochemical and radiometric data, *Tectonophysics*, *177*, 215–227.
Pin, C., and F. Marini (1993), Early Ordovician continental break-up in Variscan Europe: Nd-Sr isotope and trace element evidence from bimodal igneous associations of the southern Massif Central, France, *Lithos*, *29*, 177–196.
Pin, C., and J. L. Paquette (1997), A mantle-derived bimodal suite in the Hercynian Belt: Nd isotope and trace element evidence for a subduction-related rift origin of the Late Devonian Brévenne volcanics, Massif Central (France), *Contrib. Mineral. Petrol.*, *129*, 222–238.
Pin, C., and J. L. Paquette (2002), Le magmatisme basique calco-alcalin d'âge dévono-dinantien du Nord du Massif Central, témoin d'une marge active hercynienne: Arguments géochimiques et isotopiques Sr-Nd, *Geodin. Acta*, *15*, 63–77.
Pin, C., and D. Vielzeuf (1983), Granulites and related rocks in Variscan median Europe. A dualistic interpretation, *Tectonophysics*, *93*, 47–74.
Pin, C., J. L. Paquette, J. F. Santos Zalduegui, and J. I. Gil Ibarguchi (2002), Devonian supra-subduction zone ophiolite related to incipient collisional processes in the Western Variscan Belt: The Sierra de Careón Unit, Órdenes Complex, Galicia, in *Variscan-Appalachian Dynamics: The Building of the Late Paleozoic Basement*, edited by J. R. Martínez Catalán et al., *Spec. Pap. Geol. Soc. Am.*, *364*, 57–71.
Powell, R. (1985), Regression diagnostics and robust regression in geothermometer/geobarometer calibration: The garnet-clinopyroxene geothermometer revisited, *J. Metamorph. Geol.*, *3*, 231–243.
Raase, P. (1974), Al and Ti contents of hornblende, indicators of pressure and temperature of regional metamorphism, *Contrib. Mineral. Petrol.*, *45*, 231–236.
Ribeiro, A. (1974), *Contribution à l'étude tectonique de Trás-os-Montes Oriental*, vol. 24, 168 pp, Serv. Geol. Portugal, Lisbon.
Ribeiro, A., and E. Pereira (1997), Introducão à geologia do NE de Trás-os-Montes, in *Evolucão Geológica do Maçico Ibérico e seu Enquadramento Continental, Premeeting Field Trip Guide*, coordinated by M. E. Preto Gomes, 77 pp., Univ. de Trás-os-Montes e Alto Douro, Vila Real, Portugal.
Ribeiro, A., C. Cramez, and J. Rebelo (1964), Sur la structure de Trás-os-Montes oriental, *C. R. Acad. Sci.*, *258*, 263–265.
Ribeiro, A., R. Dias, E. Pereira, H. Merino, F. Sodré Borges, F. Noronha, and M. Marques (1987), Guide-book for the Miranda do Douro-Porto excursion, paper presented at Conference on Plate Tectonics and Deformation, Univ. de Oviedo, Gijón-Oviedo, Spain.
Ribeiro, A., E. Pereira, and R. Dias (1990), Structure of Centro-Iberian Allochthon in Northern Portugal, in *Pre-Mesozoic Geology of Iberia*, edited by R. D. Dallmeyer and E. Martínez, pp. 220–236, Springer, New York.
Ribeiro, A., E. Pereira, and F. A. Marques (1993), Discussion of Dallmeyer, R.D. and Tucker, R.D., 1993. U-Pb zircon age for the Lagoa augengneiss, Morais complex, Portugal: Tectonic implications, *J. Geol. Soc. London*, *150*, 405–410.
Ribeiro, A., E. Pereira, and P. Castro (2004), Morais Complex, in *Cr-PGE Mineralization, Petrology and Tectonics of the Allochthonous Complexes of NW Spain and Portugal*, pp. 10–16, 30–37, Ital. Agency for the Environ. Protect. and Tech. Serv., Rome.
Ries, A., and R. M. Shackleton (1971), Catazonal complexes of north-western Spain and north Portugal: Remnants of a Hercynian thrust plate, *Nature Phys. Sci.*, *234*, 65–69.
Rodríguez Aller, J., M. A. Cosca, J. I. Gil Ibarguchi, and R. D. Dallmeyer (2003), Strain partitioning and preservation of eclogitic $^{40}Ar/^{39}Ar$ ages during exhumation of a subducted crust (Malpica-Tui allochthon, NW Spain), *Lithos*, *70*, 111–139, doi:10.1016/S0024-4937(03)00095-7.
Roger, F., and P. Matte (2005), Early Variscan HP metamorphism in the western Iberian allochton. A 390 Ma U/Pb age from the Bragança eclogite (NW Portugal), *Int. J. Earth Sci.*, *94*, 173–179, doi:10.1007/s00531-005-0466-3.
Santos Zalduegui, J. F., U. Schärer, and J. I. Gil Ibarguchi (1996), Isotope constraints on the age and origin of magmatism and metamorphism in the Malpica-Tuy Allochthon, *Chem. Geol.*, *121*, 91–103.
Santos Zalduegui, J. F., U. Schärer, J. I. Gil Ibarguchi, and J. Girardeau (2002), Genesis of pyroxenite-rich peridotite at Cabo Ortegal (NW Spain): Geochemical and Pb-Sr-Nd isotope data, *J. Petrol.*, *43*, 17–43.
Shervais, J. W. (1982), Ti-V plots and the petrogenesis of modern and ophiolitic lavas, *Earth Planet. Sci. Lett.*, *59*, 101–118.
Sommer, M., and G. Katzung (2006), Saxo-Thuringia in the Variscan belt from a geodynamic point of view, *Terra Nova*, *18*(1), 63–71, doi:10.1111/j.1365-3121.2005.00664.x.
Spray, J. G., and G. W. Dunning (1991), A U/Pb age for the Shetland Islands oceanic fragment, Scottish Caledonides: Evidence from anatectic plagiogranites in "layer 3" shear zones, *Geol. Mag.*, *128*, 667–671.
Sun, S. S. (1980), Chemical composition and origin of the Earth's primitive mantle, *Geochim. Cosmochim. Acta*, *46*, 179–192.
Sun, S. S., and W. F. McDonough (1989), Chemical and isotopic systematics of oceanic basalts: Implications for mantle composition and processes, in

Magmatism in the Ocean Basins, edited by A. D. Saunders and M. J. Norry, *Geol. Soc. Spec. Publ.*, 42, 313–345.
Tang, J. C., and A. I. Chemenda (2000), Numerical modelling of arc-continent collision: Application to Taiwan, *Tectonophysics*, *325*, 23–42.
Taylor, S. R., and S. M. McLennan (1985), *The Continental Crust: Its Composition and Evolution*, 312 pp., Blackwell, Malden, Mass.
Teng, L. S., C. T. Lee, Y. B. Tsai, and L. Y. Hsiao (2000), Slab break-off as a mechanism for flipping subduction polarity, *Geology*, *28*, 155–158.
Thomson, S. N., B. Stöckhert, and M. R. Brix (1998), Thermochronology of the high-pressure metamorphic rocks of Crete: Implications for the speed of tectonic processes, *Geology*, *26*, 259–262.
van Calsteren, P. W. C., N. A. I. M. Boelrijk, E. H. Hebeda, H. N. A. Priem, E. Den Tex, E. A. T. Verdurmen, and R. H. Verschure (1979), Isotopic dating of older elements (including the Cabo Ortegal mafic-ultramafic complex) in the Hercynian orogen of NW Spain: Manifestations of a presumed early Palaeozoic mantle-plume, *Chem. Geol.*, *24*, 35–56.
Von Raumer, J. F., G. M. Stampfli, and F. Bussy (2003), Gondwana-derived microcontinents: The constituents of the Variscan and Alpine collisional orogens, *Tectonophysics*, *365*, 7–22, doi:10.1016/S0040-1951(03)00015-5.

B. Ábalos, Departamento de Geodinámica, Universidad del País Vasco, Aptdo. 644, E-48080 Bilbao, Spain. (benito.abalos@ehu.es)

J. I. Gil Ibarguchi and F. J. Santos, Departamento de Mineralogía-Petrología, Universidad del País Vasco, Aptdo. 644, E-48080 Bilbao, Spain.

J. L. Paquette and C. Pin, Département de Géologie, CNRS, Université Blaise Pascal, 5 rue Kessler, F-63038 Clermont-Ferrand, France.