

Silurian/Ordovician asymmetrical sill-like bodies from La Codosera syncline, W Spain: A case of tholeiitic partial melts emplaced in a single magma pulse and derived from a metasomatized mantle source

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Abstract

A Silurian/Ordovician extensional event in the southernmost sectors of the Central Iberian Zone is inferred from the Sm/Nd isochron obtained (436 ± 17 Ma) after the diabase sills from the La Codosera syncline. From the geochemical and mineralogical points of view, the diabase sills are subalkaline and range between high-Mg tholeiite diabases to tholeiite andesites. LREE enrichment, an Nb negative anomaly, the absence of a Ta trough and a high Nd isotope signature ($\epsilon_{Nd} = +6$) are the most relevant geochemical features. The diabase bodies are up to 330 m in thickness and were sampled from bottom to top along several different sections, permitting the definition of an accumulation of clinopyroxene, olivine and plagioclase close to chilled margins at the bottom, and abundant pegmatoid layers at the top. Chemical profiles and mass-balance modelling suggest that the bulk rock and chilled margin compositions are not dissimilar, defining an unusual S-type vertical compositional profile for large (>50 m thick) sills, which in turn strongly suggests a single magma pulse and a probable gravitational settling. Assuming chilled margin samples as the parental magma, as well as Cr-enriched samples as cumulate layers, a two-stage liquid line of descent has been established, the first one consisting of a clinopyroxene-plagioclase-olivine cumulate assemblage. A second stage in relation to the depletion in Ti, Fe and V is accounted for by ilmenite fractionation, along with that of clinopyroxene, plagioclase and olivine fractionation. Thermobarometric estimations reveal that the clinopyroxene (around 1100 °C and 197 MPa) was a late mineral phase, whereas the plagioclase (around 1200 °C) was pre- to syn-emplacement, in agreement with the presence only of plagioclase phenocrysts in the chilled margins and the very abundant positive Eu anomaly. The energy constraint modelling is consistent with the lack of a significant assimilation process owing to the high temperature contrast between the country rock and the magma itself. The compositional characteristics of chilled margins enable them to be ruled out as primary melts in equilibrium with mantle olivine (Fo_{91}), a certain amount of olivine fractionation being required, which might have occurred in magma conduits *en route* to shallow emplacement levels in the crust. Spinel lherzolitic xenoliths from the European Cenozoic alkaline magmatism appear to be unsuitable protoliths to account for the chilled margin compositions. Instead, a hybrid mantle source consisting of a small amount of OIB-mantle component (5 wt.%) and a depleted end-member mantle component seems to be a plausible protolith, resulting in a good fit with the fractionation-corrected chilled margin trends for 10% of partial melting.

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1. Introduction

The Iberian sector studied (Fig. 1a) includes a wide variety of mafic sheet-like bodies (Fig. 1b) that have aroused little geological interest. Only a few works were carried out in the seventies and eighties, including field relationships, petrographic studies, and basic geochemistry (García de Figuerola et al., 1974; Carnicero and Castro, 1983). Most of these bodies (dykes 1 to 5 in Fig. 1b) became intruded into Variscan granites and Late Neoproterozoic–Lower Cambrian metasediments. García de Figuerola et al. (1974) interpreted these relationships as being connected to a post-Variscan rifting process. A similar conclusion can be drawn from the Taliga dykes (number 8 in Fig. 1b), where the K–Ar geochronology reveals a Permian age (Galindo et al., 1990). However, in poorly studied sheet-like bodies outcropping in Variscan synclines, recent studies have revealed the presence of a strong deformation that developed a post-crystalline planar fabric (López-Moro et al., 2005). Considering the metasedimentary country rock, this magmatic episode could be assigned to a broad time-span lasting from Ordovician to Carboniferous times. In any case, it is certain that the diabase bodies of synclines represent a rifting event prior to the Variscan collisional orogeny. The mafic sill-like bodies from La Codosera Syncline (LCS), and to a lesser extent those from the Carmonita syncline (numbers 6 and 7 in Fig. 1b), show a more or less simple geometry and good exposures in some quarries (e.g. the so-called Granito Negro Villar, López-Moro et al., 2003). This is why they were selected in order to characterize the magmatism and to understand the significance of these basic rocks.

The first aim of this study is to date the mafic dyke swarms selected by means of an appropriate isotope system, such as Sm–Nd, which is assumed not to have been reset during the Variscan overprinting.

Several sections of some dyke-like intrusions were extensively sampled to gain insight into the intrusion history and the evolution of melts, since accumulation is a common process in sill-like intrusions. The detailed geochemical profiles developed in this study will permit us to approach the recently debated mechanism of emplacement in sills (Latypov, 2003a); that is, multi-intrusion, a single pulse related to Soret fractionation, or a single pulse with gravitational settling. The magmatic evolution in open or closed systems and the liquid line of descent of some sills is discussed. Likewise, systematic sampling of most bodies from the synclines proved to be an essential technique for pinpointing a cogenetic link between them. Additionally, the mineral chemistry should provide an opportunity to assess

certain intensive variables of these magmas during their emplacement and subsolidus events.

Finally, the characteristic chemical and isotope signature of diabase sills from La Codosera syncline is investigated to identify the most suitable protolith by modelling with different mantle components.

2. Geological setting

The area studied is located in the so-called Central Iberian Zone (CIZ) (Fig. 1a), which is the innermost part of the Iberian Variscan Belt, equivalent to the South Armorican Zone, the French Central Massif and the Moldanubian Zone. However, owing to the proximity to the Ossa Morena Zone (OMZ) some authors consider the zone studied to be transitional because it shows a certain affinity with the CIZ in the north and the OMZ in the south (Campos and Gumiel, 1990). Based on stratigraphic criteria, this transitional zone has been defined and designated as the Lusitan–Marianic Zone by other authors (Herranz et al., 1999).

Taking into account tectonic criteria, an old Cadomian suture must have been reactivated during the Variscan events, leading to the Badajoz–Córdoba Shear Zone (BCSZ), considered to be the limit between the CIZ and the OMZ (Farias et al., 1987), the zone studied being located to the north of the BCSZ (Fig. 1a).

The sedimentary succession in the CIZ is represented by materials from the Late Neoproterozoic or Early Cambrian to the Carboniferous, whereas in the sector studied the sedimentation only reaches Devonian sediments.

A first-order unconformity between the materials of the Cambrian and Lower Ordovician is very characteristic in the area and has been attributed to Sardinian extensional events (Valverde-Vaquero and Dunning, 2000). An important feature is an Ordovician–Silurian boundary marked by a stratigraphic discontinuity. This latter is linked to a second important extensional event (rifting) that affected the Iberian autochthon (Gutiérrez-Marco et al., 1990; Quesada, 1991; Castro et al., 2002).

In the CIZ, three main Variscan deformation phases (D1, D2 and D3) are usually considered to have been responsible for the main structures. In the area studied, the main phase was D1, which produced subvertical folds, with a steep slaty cleavage (dominant cleavage). The D2 deformation phase produced steeply inclined, gently-to-moderately plunging, asymmetrical folds with a crenulation cleavage. This was finally followed by an extensive late fracturing.

During late Ordovician times important volcanic activity took place close to the SE of the CIZ; it was

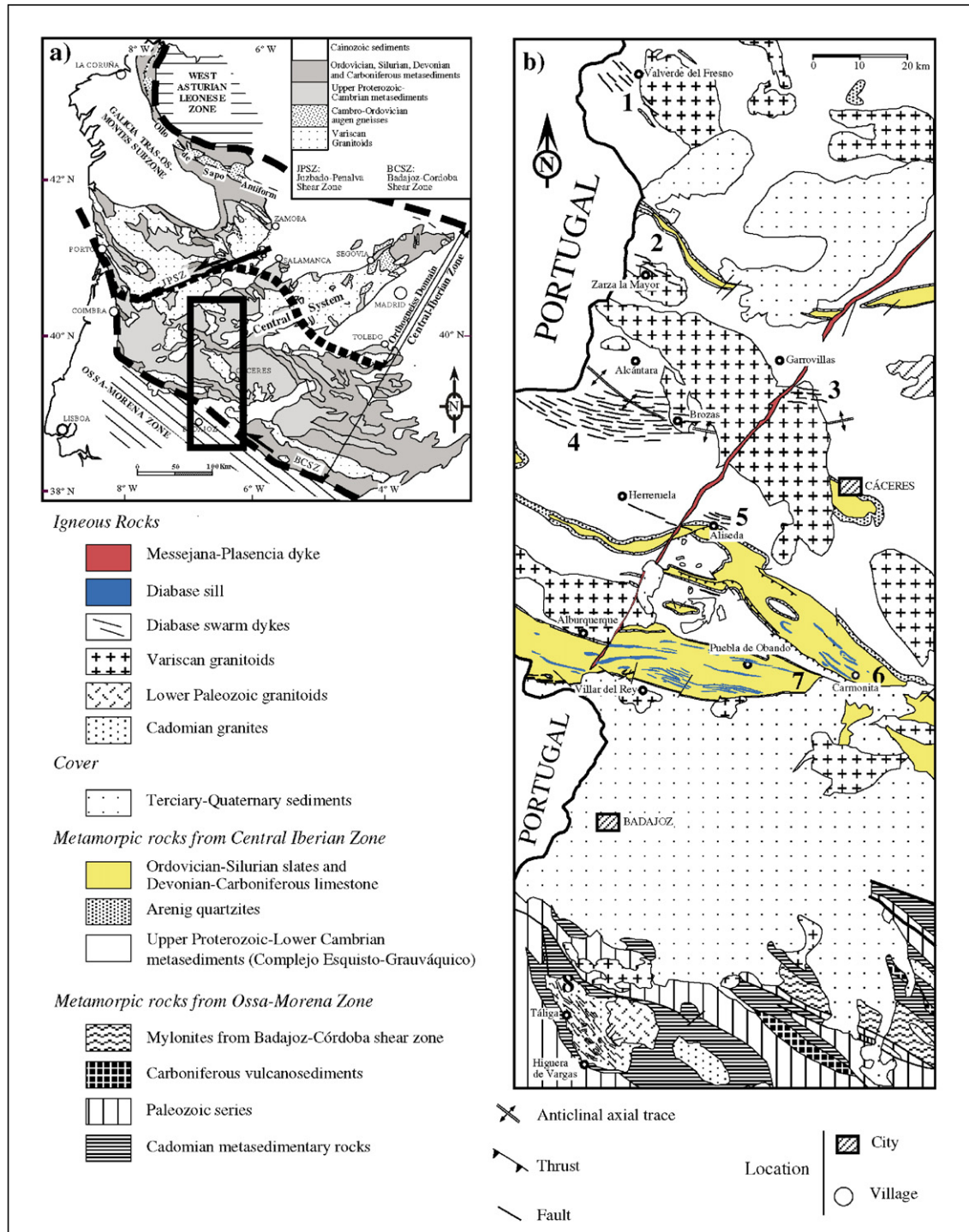


Fig. 1. a) Geological sketch map of the Iberian Massif. b) Geological map showing the location of the following diabase dyke swarms or sills in the Iberian Massif (modified after García de Figuerola et al., 1974; Apraiz et al., 1999; Gumiel et al., 2002): 1) Valverde del Fresno-Villamiel dyke swarms; 2) Zarza la Mayor dyke swarms; 3) Río Tago dyke swarms; 4) Brozas dyke swarms; 5) Aliseda dyke swarms; 6) Carmonita syncline sills; 7) La Codosera syncline sills; 8) Táliga dyke swarms. c) Enlarged area (on next page) showing the location and number of sills studied and samples analysed as well as the location of the drill holes.

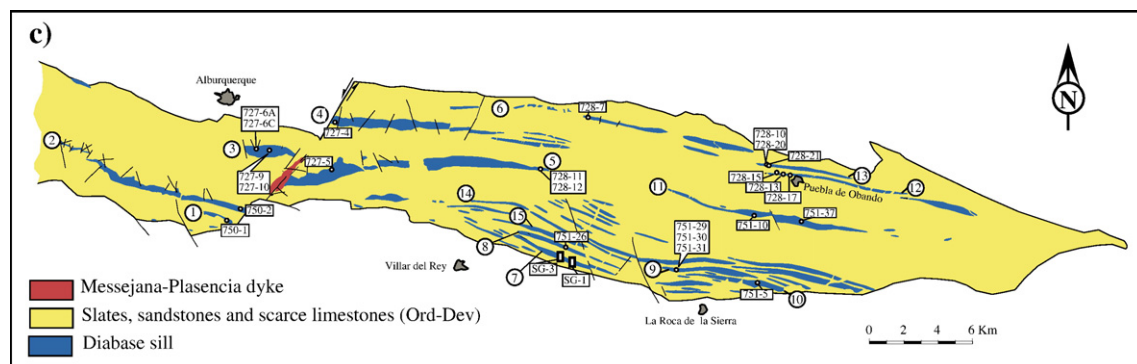


Fig. 1c (continued).

broadly of rhyolitic composition, with minor basalts related to proto-rift stages preceding later Silurian extensional events (Gutiérrez-Marco et al., 1990; Castro et al., 2002). In contrast, Silurian igneous rocks have been reported to be most abundant in the northern part of the CIZ, with abundant acidic volcanics and rare subvolcanic dolerites. The latter vary from tholeiitic to alkaline-transitional basalts (Ribeiro, 1987; Navidad and Bea, 2004).

Regarding post-Variscan magmatism in the CIZ, five occurrences of diabase dykes in low-grade metamorphic areas have been distinguished in the north of the zone studied (Fig. 1b, numbers 1 to 5), the thicknesses and lengths of the individual dykes ranging from 50 cm to 8 m and up to 5 km, respectively. Finally, the so-called Messejana-Plasencia megadyke, of Jurassic age (200 Ma), should be noted; this crosscuts the Variscan structures and all diabase dyke swarms (Fig. 1b and c) (García de Figuerola et al., 1974; Cebriá et al., 2003).

Regarding the study area, the dominant nature of magmatism is basic (diabases, Fig. 1c). Diabase rocks occur as sill-like intrusions, showing steep dipping, and striking N80°E to N120°E. They are up to 3 km in length, with a maximum thickness of 300 m (López-Moro et al., 2001). They outcrop in a D1 Variscan synclinorium (La

Codosera syncline, Sanderson et al., 1991), consisting of Arenig quartzites, Ordovician-to-Devonian slates, and very scarce limestone beds, the latter considered to be of Devonian age (Gumiel et al., 2002). A strong ductile deformation can be recognized, particularly on the southern flank of the structure, due to the influence of the BCSZ, yielding a strong postcrystalline planar fabric as well as mylonitic rocks. Until now, the age of these diabase sills was unclear and only rough approximations have been proposed: from pre-Jurassic times (IGME, 1982) to pre-to syn-orogenic (Gumiel et al., 2002). Other basic magmatism linked to Variscan synclines has been reported: Tamames (Martín-Herrero et al., 1990), Herrera del Duque (Olivé-Davó et al., 1989), Puebla de Don Rodrigo (Mateo et al., 1989), Alcañices (Arenas, 1988; Ancochea et al., 1988) and Carmonita (Gumiel et al., 2002). The latter (number 6 in Fig. 1b) shows folded diabase sills due to Variscan events; they have a marked postcrystalline planar fabric, and a thickness and a wallrock nature similar to those studied in this work. Their age is unknown, but a pre-Variscan age for their emplacement seems reasonable.

In the south of the area studied, in the OMZ, a mafic dyke swarm (number 8 in Fig. 1b) was emplaced in Cadomian granitoids. These rocks are continental

Table 1
Sm–Nd isotope analysis of the diabase sills from the La Codosera Synform

Sill	4	10	8	9	
Sample	727/4	751/5	751/26	751/29	751/31
Grain size	Medium	Medium	Fine	Fine	Coarse
Sm (ppm)	4.21	2.78	9.94	6.10	13.81
Nd (ppm)	15.71	9.76	42.77	22.94	55.62
$^{147}\text{Sm}/^{144}\text{Nd}$	0.1623	0.1723	0.1406	0.1609	0.1501
$^{143}\text{Nd}/^{144}\text{Nd} \pm 2\sigma$	0.5128300 ± 5	0.5128744 ± 8	0.5127835 ± 9	0.5128403 ± 6	0.5128108 ± 7
ϵ_{Nd}^0	+3.74	+4.61	+2.84	+3.95	+3.37
T_{DM} (Ma)	734	743	638	698	661

T_{DM} : Depleted mantle model age.

tholeiites (Carnicero and Castro, 1983) and have been dated with the K–Ar technique at 250 ± 5 Ma (Galindo et al., 1990). Recently, rift-related Late Neoproterozoic–Early Palaeozoic amphibolites have been reported in the OMZ (Sánchez-García et al., 2003; Gómez-Pugnaire et al., 2003), with a compositional variation from tholeiitic to mildly alkaline.

3. Geochronology

Five samples of diabases were selected for Sm–Nd isotope studies and geochronological purposes (Table 1). The samples correspond to four different sills, three of which are located in the southern limb of the La Codosera syncline, while the other one is in the northern limb. Likewise, two of the samples correspond to the same sill (Table 1). An isochron was obtained disregarding the sample from the northern limb, with an age of 436 ± 17 Ma (MSWD=0.4, $I_{\text{Nd}}^{436} = 0.51238$, $\varepsilon_{\text{Nd}}^{436} = +6.0$), (Fig. 2). This age is Silurian, but very close to the Ordovician–Silurian boundary (435 Ma). The values of $\varepsilon_{\text{Nd}}^{436}$ (ca. +6.0) lie between those of the Depleted Mantle and the CHUR and they are very similar to those of OIB or HIMU-oceanic island basalts.

4. Petrography

The grain-size variation in the sills studied (>50 m in thickness) is asymmetrical, with abundant pegmatoids at the top but absent at the base. Fine-grained facies are prevalent at the bottom and poorly developed at the top (Fig. 3). The facies located at the top half of the dyke are coarse-to medium-grained diabases, more leucocratic

than at the bottom half, and they contain the most Fe-enriched mafic minerals (Fig. 3), indicating a major degree of differentiation. This zoning pattern is similar to that reported by Mangan et al. (1993) in the York Haven diabase sheet, and it can be used as an indicator of sill polarity. All the facies have a similar mineralogy, with clinopyroxene, plagioclase, ilmenite and brown amphibole as the major constituents. Accessory minerals are quartz, apatite, and minor quantities of zircon. Secondary minerals are green amphibole, chlorite, saussurite, albite, calcite, titanite, zoisite/clinozoisite, and minor quantities of sulfides (mainly pyrrhotite, but also chalcopyrite and pyrite). This mineral association is characteristic of low-grade metamorphism. The fine-grained facies sometimes contain aggregates of chlorite $\pm$ actinolite, perhaps representing secondary products of olivine alteration. In addition, nodule-bearing fine-grained facies contains zoisite aggregates, probably derived from early plagioclase phenocrysts.

A foliation striking parallel to that of the wallrock (S_1) is a relevant structural feature. This tectonic foliation yielded a preferential orientation and the undulating extinction of plagioclase laths, especially in the medium- and coarse-grained facies. In the fine-grained facies, the deformation is sometimes not so marked, and ophitic textures, including clinopyroxene, brown amphibole and ilmenite, as well as zoning in plagioclase, can be observed. In the medium-grained facies from the top half of the sill, a marked increase in deformation and an enrichment in apatite can be observed (up to 3.82% in vol.). The coarse-grained pegmatoid facies show unzoned and deformed plagioclase (albite), an intersertal texture, and a great abundance of ilmenite in some samples. Occasionally, small patches of felsic components can be observed. They are mainly composed of plagioclase and quartz, with a granophyric texture interpreted as being due to a late eutectic crystallization of the magma. Sometimes, in the coarse grained facies centimetre-scale layers occur, with abundant zoisite and albite, possibly representing strongly altered plagioclase cumulates.

5. Mineral chemistry

According to the classification criteria of Morimoto (1988), the clinopyroxene ($\text{En}_{17-47}\text{Fe}_{10-35}\text{Wo}_{40-49}$) is augite. Both the high contents in Ti, Cr and Si and the low contents in Ca and Na are characteristic features of sub-alkaline and non-orogenic tholeiites (Le-Bas, 1962; Leterrier et al., 1982). Across the tabular bodies, the mgN of clinopyroxene does not change significantly, the highest values being found in the bottom-intermediate

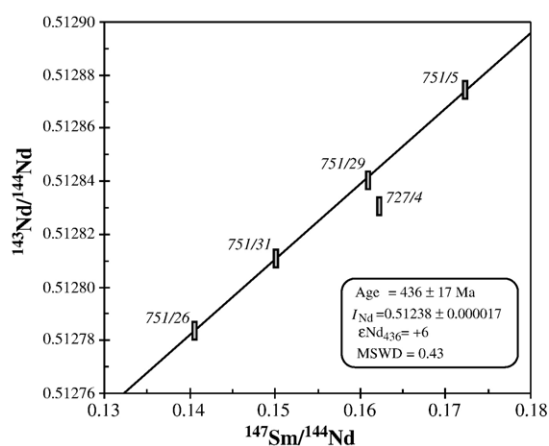


Fig. 2. Sm–Nd whole-rock isochron for the Albuquerque–Villar del Rey–Puebla de Obando sills using samples from the south limb of La Codosera anticlinorium (samples 751/26, 751/31, 751/29 and 751/5).

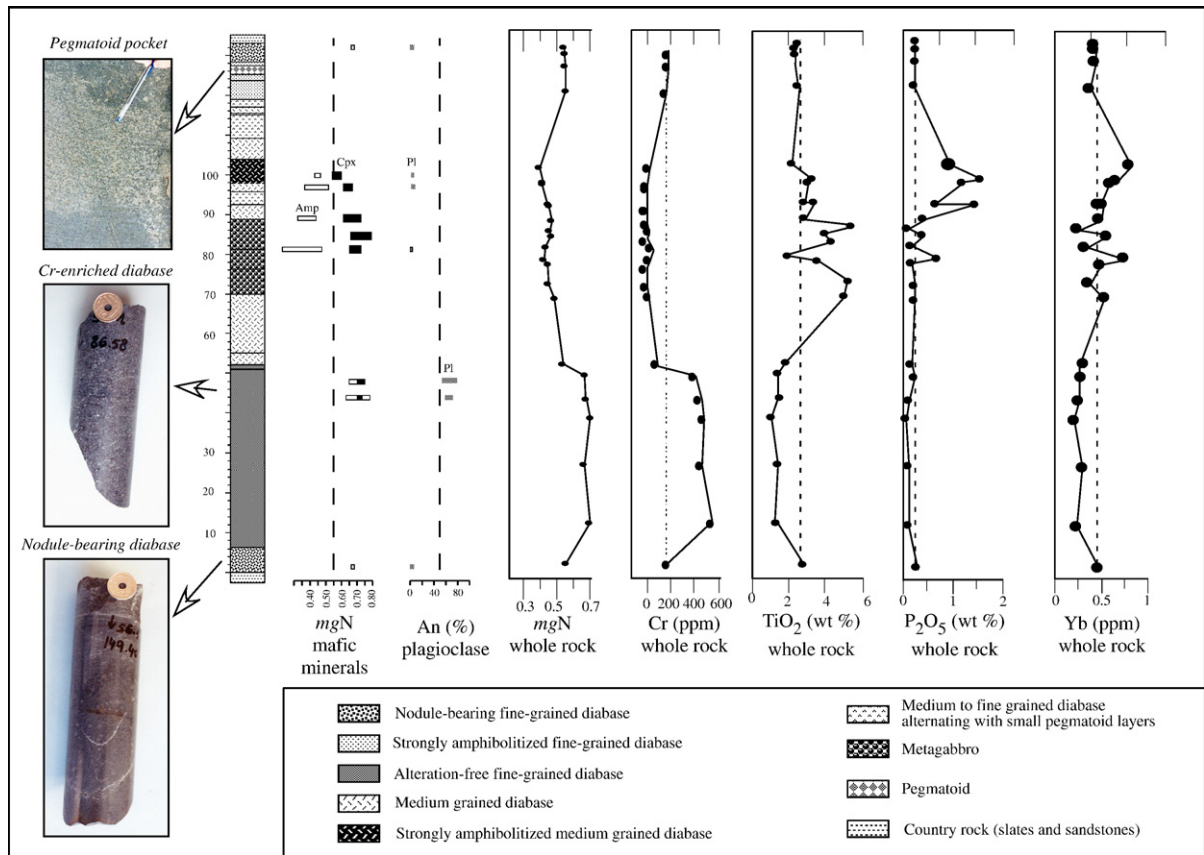


Fig. 3. Representative facies photos, cross section, simplified mineral composition and whole rock profiles with height for sill 7. *mgN*: magnesium number $[Mg/(Mg + Fe)]$ for clinopyroxene (Cpx) and amphibole (Amp); *An%*: $100 \times Ca/(Ca + Na + K)$ in plagioclase; horizontal bars: value ranges; dashed lines connect margin compositions of the sill.

zones and the lowest ones in the central and top-intermediate zones (Fig. 3).

The plagioclase composition ranges between bytownite and albite ($An_{0-79}-Ab_{21-99}-Or_{0-1.3}$). Sills 7, 8, 10 and 14 in Fig. 1c have the highest calcium contents ($>60\%$ An), always located in alteration-free fine-grained facies (Fig. 3). In this case, oscillatory ($An_{75-64-75-60}$) and normal zoning ($An_{65-57-45}$) are observed. A strong hydrothermal alteration prevented the establishment of the original variation in the calcium content in cross-section. According to the classification of Leake et al. (2004), the amphiboles are calcic. The alkali-poor group, with a broad compositional spectrum (tschermakites, magnesiohornblendes, ferrihornblendes, actinolites and ferroactinolites), can be distinguished from the alkali-rich group (pargasites and ferroedenites). Both brown and green hornblende may be enriched or depleted in alkalis, but the titanium content is specific, with a major abundance in brown amphibole crystals. Finally, the brown amphibole and clinopyroxene show a similar *mgN* distribution (Fig. 3).

6. Geothermobarometric evaluations

The liquidus temperature was estimated on the basis of plagioclase and the whole rock composition, using the geothermometers of Kudo and Weill (1970) and Mathez (1973), as well as the methodology of Cotkin and Medaris (1993) for rocks that have not undergone significant subsolidus alteration. Temperature estimations for the chilled margin samples were obtained through the pMELTS program (Ghiorso et al., 2002), assuming a quartz-fayalite-magnetite buffer. Nevertheless, the temperature values obtained from plagioclase, ranging from 1205 to 1103 °C, proved to be similar to those calculated after the pMELTS program (1202–1190 °C) for the chilled margin samples. Temperature values below 900 °C obtained after the apatite saturation model of Harrison and Watson (1984) suggest that apatite saturation was not reached. Lindsley's geothermometer (1983) based on clinopyroxene composition provided a range of 1140–720 °C, the latter value suggesting a subsolidus re-equilibrium. The composition of the brown

Table 2
Selected whole-rock data of LCS sills and their wallrock

Location	Drill hole SG-1										Drill hole SG-3					
Sample	4.02	12.9	18.64	24.5	24.55	35.13	77.3	86.58	111.1	148.8	0.75	0.5	6	13	34.35	38.90
Grain size	M	M	M	M	C	C	F	F	F	FN	FN	F	F	FN	M	M
SiO ₂	45.49	44.20	48.25	46.46	46.94	43.85	47.46	47.11	47.48	48.25	48	48.85	48.63	47.28	51.93	46.57
TiO ₂	3.27	3.39	2.82	4.15	4.17	3.73	1.66	1.62	1.73	2.86	2.69	2.57	2.58	2.74	2.25	3.03
Al ₂ O ₃	13.02	12.56	14.57	14.64	12.33	13.48	14.65	14.64	14.22	14.00	13.95	14.19	13.89	14.09	14.2	13.26
Fe ₂ O ₃	16.38	17.77	12.83	15.26	14.51	15.05	10.22	10.14	10.56	12.34	12.85	12.81	12.48	13.17	13.07	17.01
MnO	0.28	0.27	0.18	0.20	0.21	0.18	0.16	0.16	0.16	0.29	0.18	0.19	0.18	0.17	0.24	0.26
MgO	4.68	5.94	4.60	4.28	5.13	4.91	8.71	8.97	8.84	6.28	7.22	6.32	6.28	6.71	3.4	4.79
CaO	9.51	8.76	9.58	7.74	10.20	14.30	13.04	12.76	12.84	8.02	10.54	9.84	10.19	11.13	7.8	8.22
Na ₂ O	3.51	2.88	4.29	4.38	3.61	1.80	2.07	2.03	2.03	4.56	3.1	3.02	2.98	2.85	4.09	3.27
K ₂ O	0.68	0.48	0.42	0.42	0.20	0.00	0.15	0.13	0.22	0.00	0.14	0.59	0.64	0.23	0.64	0.63
P ₂ O ₅	1.60	1.49	0.51	0.33	0.46	0.22	0.28	0.16	0.16	0.31	0.28	0.26	0.26	0.25	0.99	1.26
LOI	1.48	1.84	1.69	1.95	1.70	1.52	1.76	2.17	1.78	3.46	1.66	1.66	1.66	1.74	1.61	1.86
Total	99.90	99.58	99.72	99.80	99.46	99.03	100.17	99.87	100.04	100.35	99.69	100.3	99.78	100.35	100.21	100.16
Sc	30	24	34	30	42	40	40	42	43	37	36	36	35	39	21	22
V	215	273	384	492	555	498	259	249	262	359	336	321	323	360	133	208
Cr	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	412	460	463	185	184	187	185	169	23	0
Co	32	42	33	40	38	41	41	42	43	37	40	39	39	42	22	35
Ni	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	190	204	197	63	94	91	108	96	0	0
Cu	14	18	33	49	39	b.d.l.	112	112	113	14	99	79	87	46	13	19
Zn	134	151	107	127	113	94	76	79	94	108	97	104	90	98	111	124
Y	81.2	63.9	50.4	44.9	57.8	45.1	24.3	24.9	26.4	43.1	34.8	34.6	35.9	29.9	70.2	56.6
Zr	199	160	239	181	228	183	85	84	97	169	178	182	175	135	233	234
Hf	5.6	5.1	6.7	5.6	7.2	5.8	2.8	2.7	2.9	5.4	4.6	4.8	4.6	3.5	6.1	5.5
Nb	22.2	18.1	12.5	21.2	19.0	14.4	7.1	7.1	7.6	15.9	13.0	12.8	12.9	12.4	24.1	20.0
Ta	1.94	1.64	1.09	1.74	1.61	1.23	0.62	0.62	0.67	1.37	1.10	1.22	1.13	0.92	1.88	1.64
Th	1.81	1.62	2.34	1.58	2.12	1.96	0.72	0.69	0.76	1.60	1.53	1.72	0.94	1.13	2.59	2.04
U	0.52	0.49	0.68	0.45	0.65	0.58	0.22	0.20	0.43	0.46	0.43	0.47	0.28	0.33	0.70	0.64
Cs	2.2	1.3	0.7	1.9	1.9	0.1	0.4	0.7	0.5	0.5	0.2	0.4	0.4	0.2	9.4	2.7
Rb	17	11	6	10	5	2	3	4	3	2	1	14	13	2	19	14
Ba	115	79	53	96	42	23	38	36	40	49	68	119	106	48	175	107
Sr	276	364	183	179	109	378	256	239	247	227	283	280	309	161	331	311
La	27.8	22.1	19.3	15.9	19.1	15.5	7.17	7.06	7.55	14.6	13.8	13.6	14.6	10.8	30.7	23.5
Ce	70.8	56.6	44.6	37.8	46.5	36.7	17.6	17.2	18.5	35.3	33.6	33.0	36.3	27.1	76.8	59.8
Pr	10.0	8.07	5.71	4.97	6.24	4.79	2.39	2.34	2.51	4.68	4.44	4.31	4.69	3.62	10.3	8.18
Nd	52.1	42.2	27.8	24.6	31.4	23.2	12.3	12.0	13.0	23.3	21.1	20.2	21.9	17.5	49.1	40.3
Sm	14.0	11.4	7.33	6.55	8.29	6.21	3.45	3.33	3.74	6.26	5.51	5.47	5.70	4.82	12.6	10.7
Eu	4.52	3.90	2.97	2.38	2.61	2.18	1.35	1.29	1.41	2.06	1.92	1.87	1.89	1.77	4.01	3.56
Gd	15.6	12.8	8.47	7.49	9.68	7.22	3.99	3.93	4.19	7.11	6.12	5.86	6.19	5.21	13.4	11.3
Tb	2.44	1.95	1.37	1.21	1.58	1.21	0.67	0.67	0.73	1.19	1.10	1.07	1.13	0.95	2.32	1.90
Dy	13.7	10.9	8.04	7.24	9.32	7.11	4.06	4.01	4.34	6.94	6.50	6.37	6.62	5.52	13.5	10.7
Ho	2.53	2.03	1.56	1.40	1.78	1.39	0.77	0.76	0.83	1.33	1.26	1.25	1.27	1.08	2.57	2.03
Er	6.82	5.50	4.44	3.99	5.09	4.04	2.20	2.19	2.38	3.84	3.67	3.71	3.81	3.21	7.46	5.75
Tm	0.86	0.68	0.62	0.55	0.70	0.57	0.29	0.29	0.33	0.53	0.52	0.52	0.54	0.44	1.01	0.77
Yb	5.01	4.06	3.66	3.36	4.19	3.44	1.79	1.79	1.92	3.22	3.05	3.09	3.17	2.58	5.90	4.37
Lu	0.67	0.55	0.52	0.49	0.59	0.50	0.26	0.27	0.29	0.48	0.43	0.45	0.46	0.38	0.84	0.61

(continued on next page)

Table 2 (continued)

Location	Drill hole SG-3										Sill 1	Sill 2	Sill 3				Sill 4
Sample	44.60	51.90	56.40	59.85	66.65	70.75	88.95	95.60	104	133.23	750-1	750-2	727-6C	727-6A	727/9	727/10	727/4
Grain size	M	C	C	C	C	C	M	F	F	F	F	F	M	F	FN	FN	M
SiO ₂	49.49	42.06	43.2	52	43.72	44.31	48.62	47.86	47.34	47.25	47.5	47.66	49.03	47.64	47.82	47.67	46.93
TiO ₂	2.84	5.51	4.45	2.20	5.36	5.11	1.98	2.00	1.34	1.43	2.76	2.25	2.03	1.92	1.64	1.66	2.24
Al ₂ O ₃	14.56	12.45	11.74	14.71	12.33	9.64	19.28	16.35	14.47	13.35	13.24	13.76	14.65	14.35	16.16	14.87	13.94
Fe ₂ O ₃	13.11	19.32	18.92	14.19	17.59	16.81	8.62	9.13	9.7	10.5	12.64	12.57	11.72	12.5	11.34	11.44	13.45
MnO	0.21	0.24	0.22	0.14	0.22	0.26	0.13	0.15	0.15	0.16	0.18	0.18	0.19	0.22	0.21	0.20	0.20
MgO	4.25	6.54	5.78	4.17	5.76	6.51	4.11	6.73	9.75	10.28	7.01	6.67	6.26	7.83	7.21	8.14	6.41
CaO	8.96	9.16	10.97	4.73	10.38	13.52	11.11	11.16	13.67	13.1	10.33	10.48	11.51	11.38	10.86	11.94	9.97
Na ₂ O	4.35	2.37	1.98	4.52	2.53	0.82	3.81	3.39	1.81	1.69	2.71	2.45	3.06	2.69	2.48	2.36	3.66
K ₂ O	0.27	0.25	0.41	0.31	0.29	0.11	0.31	0.32	0.23	0.26	0.41	0.44	0.18	0.11	0.58	0.16	0.06
P ₂ O ₅	0.67	0.16	0.2	0.71	0.23	0.26	0.19	0.19	0.11	0.12	0.27	0.24	0.19	0.18	0.15	0.14	0.21
LOI	1.33	2.17	2.37	2.77	2.09	1.54	2.01	2.38	1.59	2.22	2.04	3.3	1.34	1.37	2.03	1.71	2.59
Total	100.04	100.23	100.24	100.44	100.48	98.9	100.17	99.65	100.16	100.36	99.08	100.01	100.16	100.19	100.48	100.29	99.65
Sc	27	46	43	16	46	53	26	36	44	47	41	42	44	39	37	40	40
V	232	609	799	135	694	591	243	244	233	241	353	301	325	285	250	263	336
Cr	0	29	51	25	0	26	93	204	508	547	213	238	29	255	217	266	122
Co	28	57	53	43	57	46	26	33	46	48	42	41	39	47	44	46	44
Ni	0	0	77	0	0	0	57	102	213	192	114	71	46	101	84	101	56
Cu	15	17	77	18	49	11	50	57	126	143	82	100	128	138	197	141	125
Zn	105	140	130	125	119	93	76	68	55	89	104	78	93	100	129	97	108
Y	44.1	21.8	29.5	61.5	31.0	42.1	23.2	22.2	17.4	18.5	34.6	32.3	24.4	22.2	19.7	19.3	34.8
Zr	195	101	128	342	148	184	128	132	79	74	163	145	108	100	83	82	114
Hf	4.9	2.9	3.6	8.5	4.1	4.9	3.0	3.2	1.9	2.0	4.4	3.8	2.8	2.6	2.1	2.2	3.8
Nb	17.2	14.2	10.1	19.4	14.2	17.9	9.5	8.6	4.1	4.6	13.0	9.3	7.7	8.5	6.0	5.8	10.4
Ta	1.39	1.13	0.83	1.71	1.19	1.43	0.77	0.71	0.40	0.42	1.12	0.86	0.72	0.72	0.59	0.57	0.90
Th	1.67	0.55	1.06	4.00	1.27	1.41	0.94	0.90	0.52	0.53	1.61	1.44	0.81	0.74	0.58	0.59	1.17
U	0.49	0.16	0.30	1.13	0.34	0.40	0.27	0.28	0.15	0.15	0.45	0.37	0.23	0.19	0.16	0.16	0.30
Cs	1.1	1.3	10.3	2.8	2.5	1.0	0.3	0.4	0.8	0.4	1.1	1.0	2.7	1.0	4.6	1.9	0.3
Rb	7	5	17	8	5	1	5	9	4	3	8	7	6	2	18	4	1
Ba	91	81	58	87	39	11	54	39	34	33	73	87	82	43	114	34	72
Sr	194	174	226	165	117	357	390	245	230	219	240	233	213	108	263	229	175
La	17.6	6.08	9.01	31.5	11.6	12.8	9.80	8.13	5.13	5.42	13.5	11.6	9.24	7.89	6.94	6.73	10.6
Ce	43.7	16.2	23.3	75.7	28.6	32.9	23.5	20.6	13.0	13.9	33.5	28.6	21.7	19.6	16.8	16.8	25.4
Pr	5.89	2.25	3.22	9.68	3.80	4.55	3.06	2.74	1.79	1.92	4.34	3.79	2.86	2.64	2.23	2.25	3.36
Nd	28.3	11.3	15.8	44.6	18.0	22.4	14.2	13.1	8.99	9.52	20.8	17.9	13.7	12.5	10.8	10.8	17.0
Sm	7.50	3.32	4.43	11.1	4.97	6.42	3.73	3.55	2.79	5.57	4.96	4.96	3.70	3.50	2.94	2.98	4.68
Eu	2.99	1.11	1.60	3.07	1.77	2.09	1.54	1.25	1.05	1.10	1.94	1.79	1.40	1.30	1.20	1.18	1.67
Gd	8.13	3.79	4.93	11.4	5.42	7.15	4.10	3.91	2.93	3.22	6.05	5.44	4.13	3.83	3.33	3.32	5.38
Tb	1.43	0.70	0.92	1.97	1.00	1.32	0.72	0.71	0.54	0.60	1.10	0.99	0.76	0.70	0.61	0.60	0.92
Dy	8.23	4.15	5.43	11.3	5.92	7.91	4.30	4.16	3.17	3.51	6.52	5.97	4.49	4.10	3.59	3.65	5.49
Ho	1.58	0.79	1.05	2.21	1.16	1.55	0.82	0.80	0.62	0.67	1.26	1.17	0.88	0.82	0.69	0.70	1.07
Er	4.59	2.37	3.12	6.48	3.41	4.61	2.44	2.37	1.81	1.99	3.73	3.47	2.57	2.36	2.05	2.05	3.08
Tm	0.63	0.32	0.44	0.91	0.47	0.64	0.34	0.32	0.25	0.27	0.51	0.49	0.35	0.33	0.28	0.28	0.43
Yb	3.66	1.91	2.47	5.44	2.74	3.69	1.97	1.94	1.45	1.57	3.10	2.85	2.10	1.97	1.66	1.70	2.62
Lu	0.52	0.29	0.35	0.77	0.41	0.55	0.29	0.28	0.21	0.23	0.45	0.43	0.31	0.29	0.24	0.25	0.39

Location	Sill 5			Sill 6	Sill 7	Sill 9			Sill 10	Sill 11		Sill 12			Sill 13			Wallrock
Sample	727/5	728/11	728/12	728/7	751/26	751/29	751/30	751/31	751/5	751/10	751-37	728-15	728-17	728-13	728/10	728-20	728-21	Corn-1
Grain size	F	F	M	F	F	F	C	C	M	F	M	FN	FN	M	M	M	C	–
SiO ₂	47.74	48.63	45.46	47.65	47.83	47.09	53.71	53.32	48.63	46.67	48.67	47.79	47.22	48.55	47.74	47.75	47.12	59.35
TiO ₂	2.66	2.28	4.61	1.62	2.56	2.84	2.09	1.63	1.30	2.29	2.00	2.23	1.88	2.30	2.07	1.99	2.14	0.97
Al ₂ O ₃	13.47	13.58	11.98	14.58	13.88	12.86	13.47	13.76	15.17	14.31	15.52	13.77	14.67	14.12	13.72	14.84	13.46	20.04
Fe ₂ O ₃	13.81	11.67	17.38	11.62	12.44	14.10	10.90	13.36	9.79	12.69	11.51	13.26	11.81	12.44	12.27	12.00	13.05	8.72
MnO	0.22	0.18	0.24	0.26	0.19	0.21	0.20	0.23	0.17	0.20	0.20	0.20	0.20	0.18	0.19	0.18	0.21	0.06
MgO	5.82	6.29	5.18	6.87	6.61	5.73	2.34	2.23	7.77	6.84	6.76	6.59	7.22	6.06	6.93	6.40	7.10	1.82
CaO	9.88	9.65	8.55	8.82	9.93	9.87	7.03	5.05	12.68	12.39	11.65	9.49	11.86	10.28	11.71	10.88	12.16	0.31
Na ₂ O	2.92	4.30	3.11	3.92	3.32	2.90	6.13	5.86	2.44	1.35	2.47	3.61	2.72	3.64	2.52	3.33	1.95	0.84
K ₂ O	0.00	0.00	0.25	0.00	0.29	0.22	0.59	0.31	0.14	0.04	0.18	0.19	0.08	0.27	0.00	0.18	0.06	3.15
P ₂ O ₅	0.37	0.25	0.32	0.16	0.28	0.32	0.88	0.91	0.15	0.25	0.25	0.22	0.18	0.21	0.21	0.20	0.20	0.21
LOI	3.31	2.94	2.35	4.00	1.54	2.40	1.25	2.17	0.83	2.62	0.97	2.40	2.22	2.42	2.23	2.39	2.60	4.73
Total	100.12	99.76	99.43	99.44	98.86	98.52	98.57	98.84	99.06	99.65	100.19	99.75	100.05	100.48	99.55	100.14	100.03	100.22
Sc	39	39	40	43	38	39	21	18	43	40	33	37	37	40	43	39	42	21
V	365	312	586	255	334	397	80	65	232	338	230	311	272	321	299	283	312	183
Cr	149	240	b.d.l.	36	262	98	b.d.l.	20	69	272	274	116	240	33	75	57	48	97
Co	43	37	49	42	41	44	17	20	40	44	40	46	44	40	43	41	44	11
Ni	103	83	b.d.l.	58	106	25	b.d.l.	b.d.l.	93	98	100	69	90	58	43	96	62	44
Cu	96	91	34	123	88	96	b.d.l.	b.d.l.	103	107	58	155	123	99	142	124	55	23
Zn	143	113	153	96	108	133	61	103	55	85	84	120	93	90	103	105	71	53
Y	46.4	41.7	51.0	25.7	40.6	51.5	102	104	24.6	40.7	30.7	28.1	22.3	26.5	29.4	23.3	26.2	32.9
Zr	172	174	214	92	161	180	443	466	70	138	153	126	101	124	96	100	110	201
Hf	5.4	5.4	6.8	3.0	4.9	5.7	13.1	13.7	2.3	4.5	4.0	3.3	2.6	3.3	3.1	2.7	3.0	5.5
Nb	14.7	13.4	19.8	8.2	14.6	15.0	35.7	35.8	4.7	11.0	12.2	10.3	8.2	10.2	10.1	9.2	8.7	17.6
Ta	1.22	1.13	1.67	0.71	1.22	1.26	2.83	2.92	0.40	0.91	0.95	0.89	0.71	0.84	0.83	0.73	0.79	1.58
Th	1.86	2.42	2.65	0.69	1.50	1.90	4.80	5.26	0.62	0.99	1.70	0.97	0.74	0.97	0.73	0.76	0.84	18.6
U	0.46	0.63	0.69	0.21	0.45	0.51	1.31	1.56	0.16	0.28	0.43	0.27	0.20	0.27	0.22	0.22	0.24	2.35
Cs	0.6	0.4	1.4	0.1	0.6	0.6	0.2	1.6	2.5	0.1	1.8	0.2	0.4	0.4	0.1	0.2	0.1	4.1
Rb	b.d.l.	b.d.l.	6	b.d.l.	7	7	7	10	7	b.d.l.	3	2	5	4	1	3	1	142
Ba	43	51	61	44	76	53	115	93	30	14	77	94	27	71	62	44	66	365
Sr	119	163	116	181	271	209	365	208	233	291	183	118	180	229	211	280	261	62
La	14.5	15.0	17.9	7.80	13.6	15.2	34.9	39.8	5.61	11.0	12.8	10.9	8.4	10.2	9.37	9.6	7.6	50.6
Ce	34.9	35.5	42.8	18.7	32.8	37.1	86.1	96.4	14.1	27.3	31.0	26.3	20.5	25.1	22.5	23.6	19.6	104
Pr	4.58	4.58	5.58	2.55	4.35	4.99	12.1	12.7	1.95	3.71	4.01	3.47	2.72	3.29	2.96	3.02	2.77	11.4
Nd	23.1	22.3	27.7	12.6	21.3	24.9	60.3	61.8	10.2	18.7	18.9	16.5	13.0	15.6	14.9	14.3	13.9	43.5
Sm	6.20	5.83	7.23	3.53	5.72	6.84	15.8	15.5	3.10	5.10	5.00	4.40	3.54	4.21	4.15	3.87	3.91	8.49
Eu	2.11	2.08	2.42	1.48	2.03	2.39	4.93	5.32	1.19	1.89	1.77	1.57	1.35	1.53	1.57	1.46	1.42	1.66
Gd	7.32	6.75	8.28	4.08	6.75	7.90	17.8	18.0	3.73	6.22	5.31	4.81	3.83	4.69	4.82	4.18	4.33	7.49
Tb	1.21	1.11	1.39	0.69	1.12	1.33	2.94	2.93	0.65	1.06	0.96	0.88	0.69	0.84	0.80	0.75	0.81	1.16
Dy	7.29	6.67	8.18	4.13	6.55	7.87	17.5	17.3	4.03	6.41	5.66	5.19	4.10	4.99	4.82	4.50	4.81	6.48
Ho	1.43	1.29	1.59	0.79	1.25	1.53	3.36	3.37	0.77	1.24	1.12	1.02	0.80	0.96	0.90	0.85	0.94	1.22
Er	4.18	3.76	4.61	2.26	3.60	4.48	9.70	9.81	2.21	3.63	3.34	2.98	2.37	2.88	2.61	2.51	2.74	3.67
Tm	0.59	0.53	0.64	0.31	0.50	0.62	1.35	1.38	0.31	0.51	0.46	0.41	0.32	0.40	0.35	0.35	0.39	0.56
Yb	3.57	3.20	3.89	1.93	3.01	3.77	8.27	8.37	1.86	3.16	2.81	2.45	1.96	2.40	2.15	2.06	2.29	3.38
Lu	0.54	0.46	0.58	0.28	0.44	0.56	1.18	1.25	0.27	0.46	0.41	0.36	0.28	0.35	0.32	0.31	0.33	0.49

F: fine-grained size; FN: fine-grained size groundmass with nodules; M: medium-grained size; C: coarse-grained size; LOI: loss on ignition; b.d.l.: below detection limit.

amphibole together with the geothermometer of Hammarstrom and Zen (1986) allows us to determine the solidus temperature, resulting in a range of 940–820 °C. The latter temperatures are always lower than those calculated after plagioclase and clinopyroxene suggesting that the brown hornblende is a late-stage crystallization phase. Finally, the hornblende-plagioclase geothermometer pair (Plyusnina, 1982) applied to green amphibole and albite provides a temperature range of 480–400 °C, corresponding to low-grade metamorphism conditions.

Pressure conditions were constrained after the barometer of Nimis and Ulmer (1998) using the tholeiitic calibration (TH) as well as the clinopyroxene composition for crystals assumed to be equilibrated at temperatures higher than 1000 °C according to Lindsley's geothermometer. The results obtained range from 280 to 80 MPa (average of 197 MPa and σ -values of ± 60 MPa), indicating that pyroxene crystallization took place under shallow magma emplacement conditions.

7. Geochemistry

7.1. Classification

The above results allow us to infer that subsolidus processes may play an important role in explaining not only the mineral chemistry but also whole-rock chemical characterization; this is why using immobile elements allows us to define a typical tholeiitic evolution from Mg-rich basalt to andesite, according to Jensen and Pyke (1982). Furthermore, the subalkaline nature of the samples studied is evident after the high silica contents of the clinopyroxenes (*graphical information is available in a supplementary data file, see Appendix B*).

7.2. Major elements

Most of the samples are characterized by moderate to high contents in Al_2O_3 , CaO , TiO_2 , Fe_2O_3 , Na_2O and P_2O_5 , and low K_2O contents (Table 2). The trends of magma evolution are marked by a decrease in CaO and an increase in Na_2O and P_2O_5 (Fig. 4). Likewise, the FeO and TiO_2 contents in what we refer to as fine-grained facies and nodule-bearing fine-grained facies show an increase with evolution. A group of samples of medium- and coarse-grained facies shows a strong increase in FeO and TiO_2 , with no significant decrease in MgO when this latter is higher than 5%. In contrast, samples with MgO contents lower than 5% display a slight decrease in both elements, with a relatively strong decrease in MgO . This descending trend is consistent

with ilmenite fractionation since titanomagnetite is absent. Also, a high proportion of ilmenite and, to a lesser extent, Fe-rich clinopyroxene or amphibole accounts for the strong Ti and Fe contents in some samples. The plots of elements such as Ca, Ti, Fe, Na and P of fine-grained and nodule-bearing fine-grained facies fall along straight-line trends, whereas a scattered plot for K is worth noting. Medium and coarse-grained facies show a higher scattering for all the elements than the former ones. Thus, some apatite-enriched samples record extreme P contents, whereas well-defined steep linear trends for Ca and Na appear to be linked to plagioclase or pyroxene enrichment.

7.3. Trace elements

The rocks from the fine-grained facies with scarce alteration are Ni- and Cr-enriched (Fig. 5), suggesting either an olivine-clinopyroxene accumulation and/or a Ni and Cr mobilization for the rest of the facies. A compatible behaviour for these transition elements except for V can be inferred from variation diagrams, showing an incompatible-changing-to-compatible behaviour in the same way as Ti. Fig. 5 shows a scattering in Ba, Sr and Rb, whereas straight linear trends are shown by immobile elements, such as high field strength elements (HFSE) and REE. Good correlations for highly incompatible elements are exemplified by the U and HREE vs. Th diagrams, but the enrichment of apatite in some of the samples causes a relative scattering of MREE and LREE.

Primitive mantle-normalized trace element abundances (Fig. 6) show enrichments in the most incompatible elements, a slightly negative Nb anomaly, a major Th abundance with respect to Ta, and frequent negative P and Zr anomalies. These features are consistent with crustal contamination (e.g. Dupuy and Dostal, 1984) or a metasomatized source. On the other hand, the Ba, Rb and K contents display an incoherent distribution pattern because they behaved as mobile elements during metamorphic or hydrothermal events. The negative Sr anomaly is not uncommon, accounting for a hydrothermal process, whereas a few samples having a slightly positive Sr anomaly can be related to the presence of slightly-altered plagioclase. The strong positive P and Ti anomalies of some samples can be due to the apatite and ilmenite enrichment respectively. The coarse-grained rocks with a negative Ti anomaly would counterbalance the ilmenite enrichment.

The most mafic facies display the lowest REE abundances, whereas the least mafic ones, the coarse-grained facies, have the highest ones. Chilled-margin facies characteristically have intermediate REE abundances,

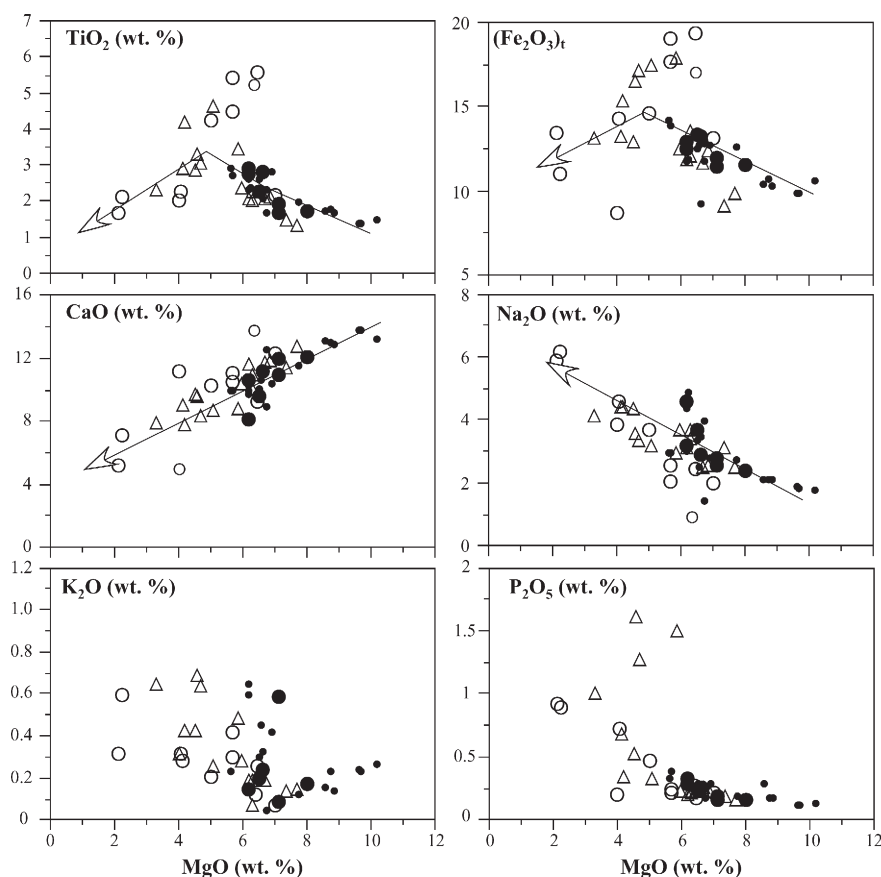


Fig. 4. Major-element variation diagrams. Dots: fine-grained facies; black circles: nodule-bearing fine-grained facies; triangles: medium-grained facies; large empty circles: coarse-grained facies.

similarly to other more than 50-meter thick sills reported (Woodruff et al., 1995; Gibb and Henderson, 2006). These features, along with the subparallel patterns, point to an incompatible behaviour of REE. A slight negative Eu anomaly ($\text{Eu}/\text{Eu}^*: 0.84$) for the most differentiated samples indicates a minor plagioclase fractionation, whereas a slight positive Eu anomaly ($\text{Eu}/\text{Eu}^*: 1.15$) for the most basic ones suggests plagioclase accumulation. It is worth emphasizing that Eu-enriched samples are not the same as the Sr-enriched ones, as is common for moderately to highly hydrothermalized rocks, suggesting a redistribution of Sr during hydrothermal events.

7.4. Vertical variations in whole-rock chemistry of selected sills

The chemical variations of the selected constituents are depicted with respect to height (Fig. 3). It should be noted that the chemical profiles converge at the chilled margins, which show similar compositions. An inward increase in Cr for the fine-grained facies allows us to infer a clinopyroxene accumulation level, confirmed by

observation under the microscope. At the same time, a probable fractionation can be inferred from lower mgN and Cr values for the coarse- and medium-grained size facies. This is also supported by the presence of pegmatoids and vugs filled with quartz and feldspar.

Lu, TiO_2 , Eu and P mirror the patterns of the compatible elements, the medium- and coarse-grained facies showing their highest contents. In other words, the textural changes always correspond to similar chemical changes (Fig. 6). The strong enrichment in P is remarkable at the bottom of the medium-grained facies. Also, a Ti-enrichment appears to be associated with the coarse-grained facies and the medium ones, alternating with pegmatoid layers.

8. Discussion

8.1. Differentiation processes in sill 7: a case of gravitational settling under closed-system conditions

Differentiation processes involved in more than 50-meter thick sill-like mafic bodies are under

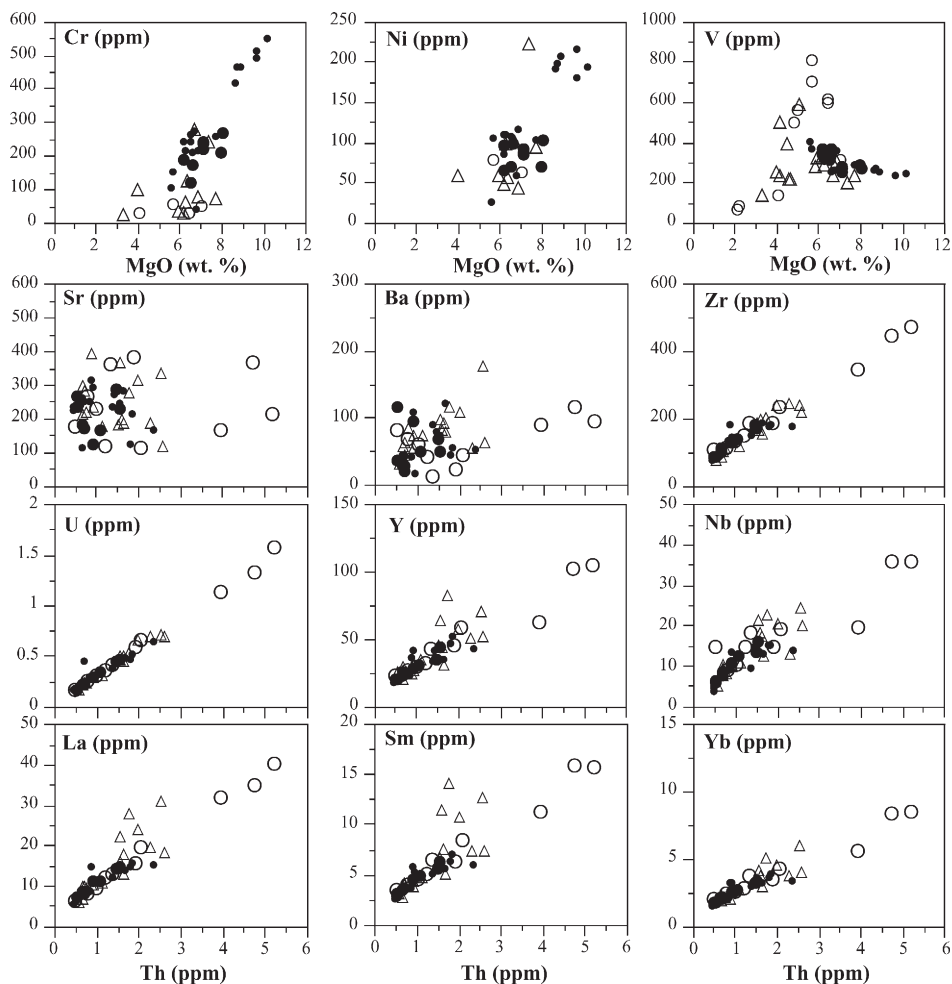


Fig. 5. Variation diagrams for all the samples. Transition elements are plotted vs. MgO; rest of elements vs. Th. Symbols as in Fig. 4.

worldwide revision, mainly as an attempt to explain their complex compositional profiles, consisting of high Cr and MgO basal enrichment, and their high *mgN*. Despite their overlapped compositional typology, taking into account refractory elements basically three profile types have been distinguished: S-, D- and I-shaped profiles (see Latypov, 2003a). Sill number 7 from La Codosera, which we have worked on most intensively, has an S-type affinity profile, showing a higher proportion of differentiated layers than those enriched in Cr and with a high *mgN* (Fig. 3). This type of profile cannot be explained easily and different processes have been suggested, such as a multipulse intrusion (Shirley, 1987; Gibb and Henderson, 2006). Also, a single magma pulse has been invoked in different ways: a) flow differentiation of a magma carrying a large amount of intratelluric phenocrysts (olivine or orthopyroxene) (Marsh, 1996), b) *in situ* crystallization from phenocryst-free magma (Tait and

Jaupart, 1996), c) *in situ* crystallization involving Soret fractionation processes (Latypov, 2003a) and d) gravitational settling of either intratelluric phenocrysts or newly-formed crystals in the chamber (Frenkel' et al., 1989; Ariskin and Barmina, 2006).

A multipulse process should be ruled out because the bulk sill composition, estimated after mass-balance calculations, is similar to that of chilled margins (Fig. 7 a), strongly suggesting that on one hand the latter represent the parental magma and, on the other, that the replenishment process is unlikely to have occurred. A similar composition of the two chilled margins is another aspect to be emphasized, since this is not a common feature for thick sill-like mafic bodies.

With respect to single magma pulse processes, an *in situ* crystallization, as defined by Tait and Jaupart (1996), would not be in agreement with the presence of phenocryst-bearing magmas, exemplified by the nodule-bearing chilled margin facies. Soret diffusion processes

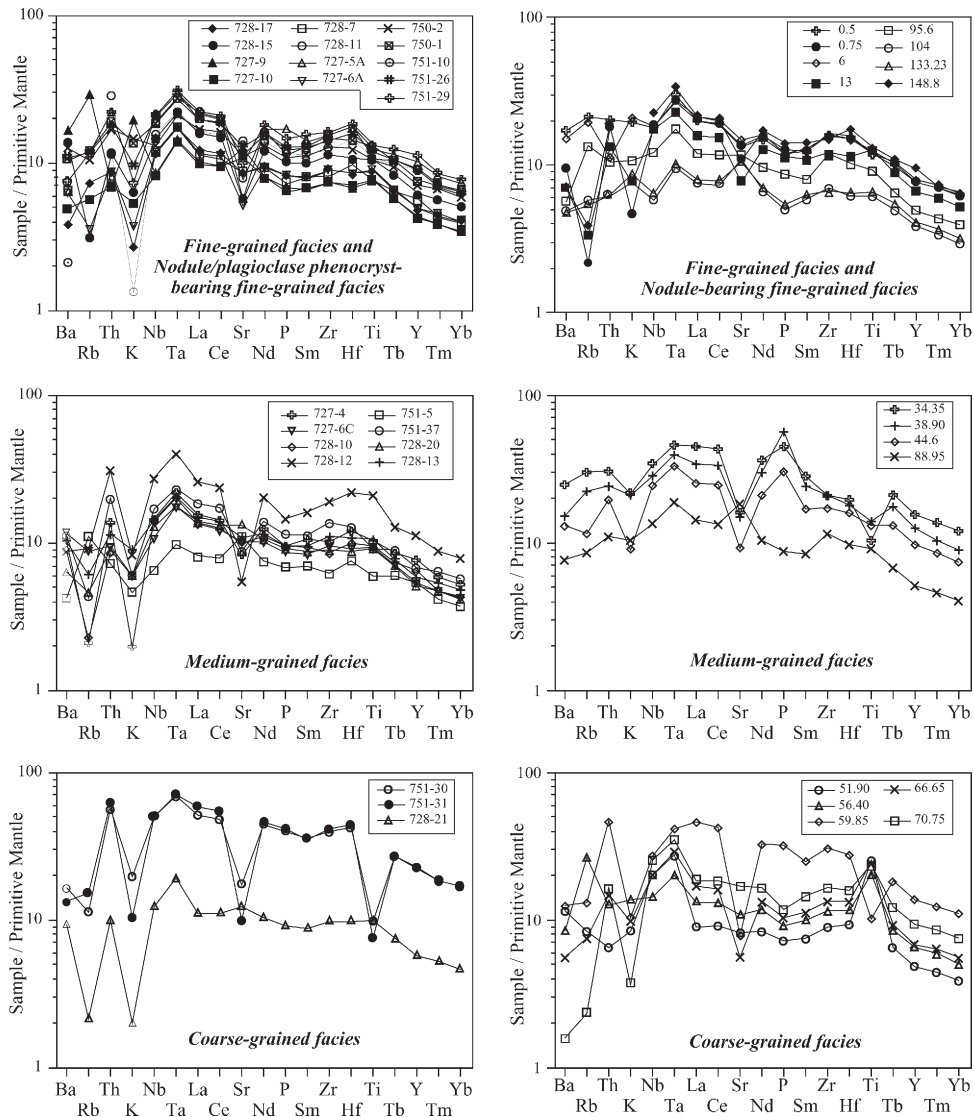


Fig. 6. Primitive mantle-normalized incompatible element plots for selected samples from different sills of the area studied (diagrams on the left column) and from two drill holes through sill number 7 in Fig. 1c (diagrams on right column). Normalising values after Sun and McDonough (1989).

(Latypov, 2003a) can also be ruled out because it would be difficult to reconcile the following two observations: 1) the presence of cumulate textures in Cr-, mgN-enriched layers; and 2) the compositional similarity of chilled margins and the bulk sill composition, which indicates that the parental magma is not the most Cr-enriched sample.

The presence of phenocrysts in marginal layers probably consisting of plagioclase (nodule-bearing chilled margin facies) instead of olivine and pyroxene renders Marsh's model inappropriate.

A gravitational settling model, such as that of Frenkel' et al. (1989) and Ariskin and Barmina (2006), would be supported by: i) the presence of accumulation

levels, ii) the upwards differentiation, and iii) an environment of high amount of supercooling, as is currently accepted for gravitational settling. The formation of the chilled margins and the upward differentiation process are shown in an interpretative two-stage model (Fig. 7b).

The accumulation level is represented by the Cr-enriched facies as shown by: a) the presence of cumulate textures, b) the results of mass-balance calculations and c) the thermometric results. Mass balance modelling indicates that Cr-enriched samples can be reproduced by adding clinopyroxene + olivine + plagioclase (~54%) to the chilled margin compositions (Table 3). Thus, these latter could represent the initial magma, whereas the Cr-enriched samples would be considered to be cumulates.

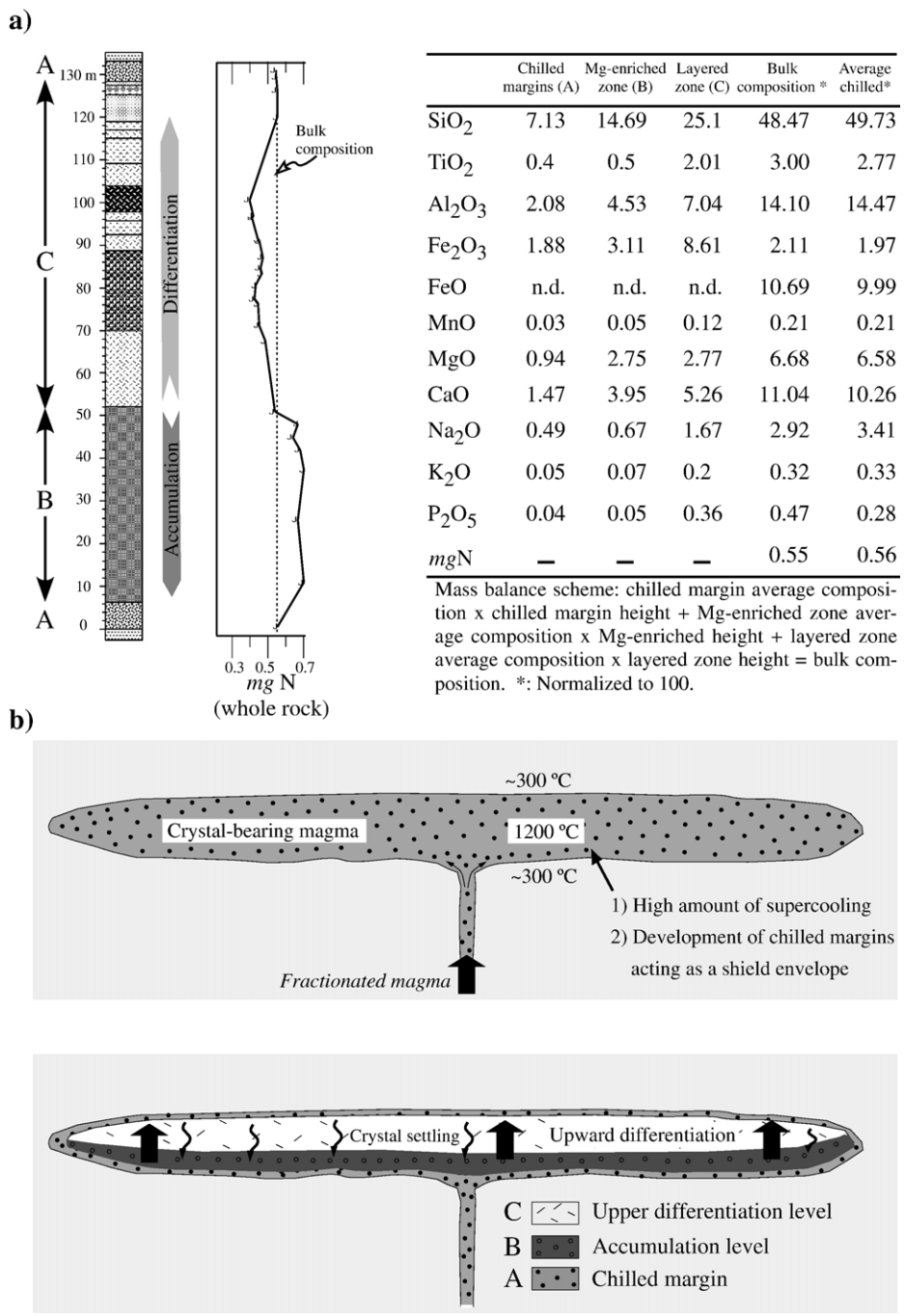


Fig. 7. a) Vertical compositional profile involving mg-number-whole rock-with stratigraphic height and attached table with corresponding compositions for A, B and C zones. Bulk composition was calculated according to mass balance modelling (dotted line). It should be noted that the bulk rock and chilled margin compositions are not dissimilar. (Hatch patterns in the cross section as in Fig. 3). b) Interpretative model illustrating a two-stage sill emplacement. First, a fractionated magma carrying plagioclase phenocrysts would have undergone a high amount of supercooling, leading to a continuous shield of chilled margin (nodule-bearing fine-grained facies). Second, an upwards differentiation process would have been triggered by consequent static-closed-system conditions, giving rise to the formation of the cumulate facies overlying the lower chilled margin level, and the gabbroic and pegmatoid facies upwards. Wallrock temperature according to [Bohrson and Spera \(2001\)](#) for the average of the upper crust.

An important conclusion can be drawn from these observations: the most enriched samples in compatible elements do not necessarily represent the parental magma. The contrasting mineral association from the chilled margin, consisting of plagioclase + groundmass, with respect to that from the overlying accumulation level (plagioclase + clinopyroxene + groundmass), leads us to infer that clinopyroxene saturation was reached just above the bottom chilled margin. A difference in temperature from around 1200–1190 °C for the emplacement of magma, according to pMELTS, to 1100–1040 °C after clinopyroxene crystals in the Cr-enriched level appears to support the late clinopyroxene crystallization, probably at the emplacement level. It is worth noting that the temperature obtained (about 1200 °C) after the melt composition (chilled margin) is similar to that obtained after plagioclase crystals from the cumulate level, accounting for the presence of plagioclase phenocrysts in the chilled margin facies.

ii) Upwards, the residual liquid evolved to a more fractionated composition, the liquid line of descent being reconstructed through the bell-shaped trends in bivariate diagrams. For example, the V–MgO diagram (Fig. 5) permits samples SG1-18.46 and SG3-59.85 from sill 7 to be selected for modelling, both representing the liquid line of descent before and after ilmenite saturation, respectively. A good fit is obtained by fractionating clinopyroxene (44%), olivine (10%), plagioclase (38%) and ilmenite (8%) (Table 3). Finally, the Rayleigh fractional crystallization equation was applied using the latter mineral proportions, although the fit obtained was not acceptable despite the subparallel REE patterns (Fig. 8a), indicating that the

fractionating phase mineral proportions might be correct, but not the F -parameter. Rayleigh fractionation modelling based on immobile elements using the same fractionating phase mineral proportions indicates a residual liquid (F) of about 40% for the most evolved sample (Fig. 8b–f), showing a good fit. This residual liquid proportion is not dissimilar to that obtained according to mass-balance calculations ($F=46\%$) using less altered samples of Cr-enriched diabase (Table 3), but it differs from that estimated after altered gabbroic samples ($F=65\%$).

The results of sill 9 are considered to be equivalent to those of sill 7, since the F value obtained after mass balance modelling with altered samples is lower than that obtained with Rayleigh fractionation (Fig. 8g–h). The decoupling between the mass balance and trace element calculations in our case is related to the hydrothermal activity, since most of the major elements display scattered plotting in bivariate diagrams, whereas immobile trace elements show linear trends as a result of a limited or null redistribution. This is why trace element modelling is preferable for estimating the fraction of residual liquid.

8.2. Differentiation processes for the rest of the sills

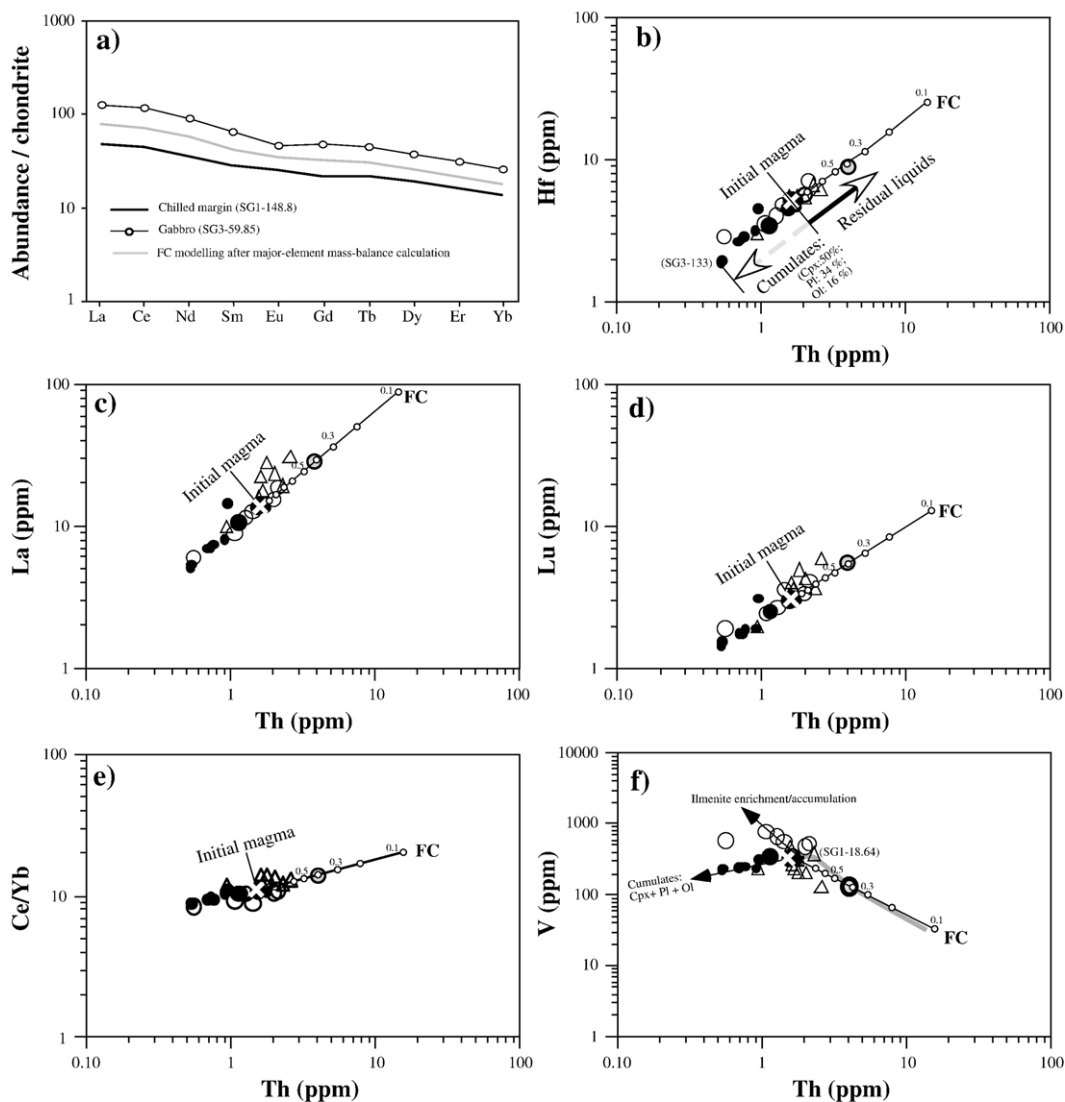
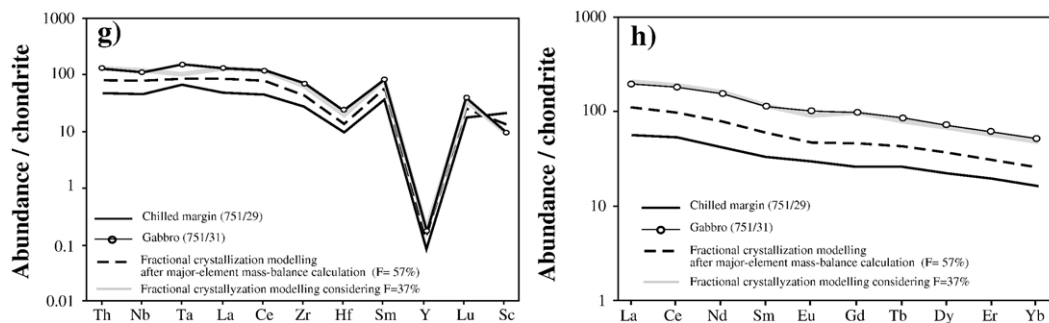
Fractional crystallization seems to have been the most relevant differentiation process, as shown by the following criteria:

- a) *The Sm–Nd isochron.* The Sm–Nd isochron indicates a cogenetic nature for the samples studied (Fig. 2). This points either to a mantle melting process with

Table 3
Multi-stage major element-based least-squares modelling

Modelling	Cumulate minerals (%)				R	Chilled margin (wt.%)
	Cpx	Pl	Ol	Ilm		
<i>Sill 7</i>						
Chilled margin (SG3-0.5)→Cr-enriched diabase (SG1-133)	+27.05	+18.24	+8.66	–	0.21	46
Liquid line of descent	Cumulate minerals (%)				R	F (wt.%)
	Cpx	Pl	Ol	Ilm		
<i>Sill 7</i>						
1. Chilled margin (SG3-0.5)→Gabbro (SG1-18.64)	–36.9	–24.85	–38.23	–	0.08	89
2. Chilled margin (SG3-0.5)→Gabbro (SG3-59.85)	–44.2	–37.71	–10.26	7.82	0.22	65
<i>Sill 9</i>						
3. Chilled margin (751/29)→Gabbro (751/31)	–29.69	–42.38	–19.96	–7.96	0.24	57

R : sum of squared residuals; F : percentage of residual liquid; mineral abbreviations after Kretz (1983); Na₂O, P₂O₅ and MnO not considered in modelling.

Sill 7*Sill 9*

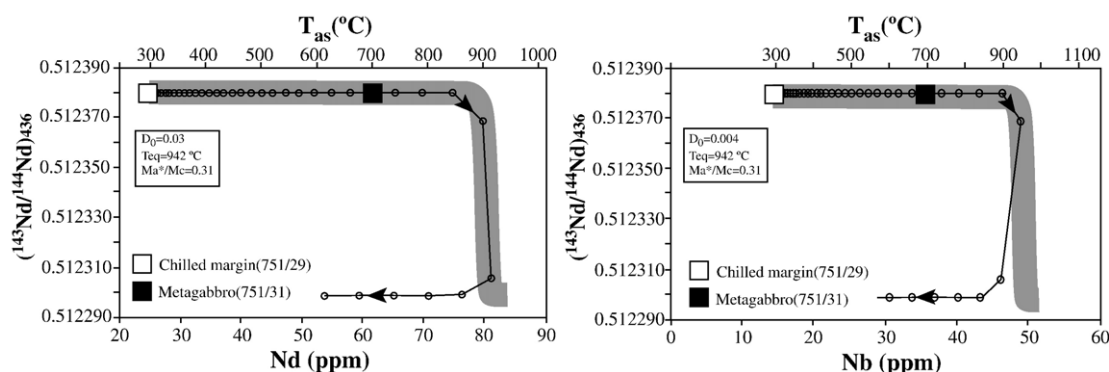


Fig. 9. EC-AFC model using $(^{143}\text{Nd}/^{144}\text{Nd})_t$ vs. Nd and Nb diagrams. Thin line illustrates the evolution of Nd, Nb and Nd-isotope ratios for EC-AFC model of the “standard” upper-crustal case, whereas the heavy grey line represents the variation in wall-rock temperature variation as a consequence of melt heating. In any case, the composition of the most differentiated sample (751–31) indicates a lower temperature than that of the solidus temperature (900 °C) of the wall rock, ruling out an assimilation process. T_{as} : wall rock temperature; T_{eq} : Equilibrium temperature; D_0 : bulk distribution coefficient; Ma^*/Mc : ratio of the mass of anatectic melt to the mass of cumulates.

different degrees of partial melting, or to a closed-system evolution from the same melt (fractional crystallization), ruling out any AFC processes for most of the samples studied. Likewise, a mixing-isochron is not probable owing to the poor correlation in the $^{143}\text{Nd}/^{144}\text{Nd}$ -1/Nd diagram (not included).

- b) $C_L^i - C_L^j$ and $C_L^i/C_L^j - C_L^i$ diagrams. Bivariate plots of highly incompatible (i.e. Th–U, La–Ce, Nb–Ta) or moderately incompatible elements (Lu–Th, Y–Nb) define straight linear trends that intercept close to zero, fitting a regression line excellently in most cases (diagrams not included). This is a characteristic feature of cogenetic samples originated either by fractional crystallization or by partial melting of the same source. There are some exceptions, such as the Zr–Nb regression line, which intercepts at a value of 10 as a result of the different distribution coefficients in clinopyroxene ($D_{Zr}=0.27$ and $D_{Nb}=0.004$) (Kamber and Collerson, 2000), and hence, any mixing process should be ruled out.
- c) *Irregular linear trends*. Geochemical evolutionary trends with inflections in slope may reflect saturation

in a mineral phase, resulting in striking shifts in the residual liquid composition, as is common in continuous crystallization processes. The behaviour of Ti, V and Fe represents further evidence of this, showing an inflection point (Figs. 4 and 5) due to ilmenite saturation, thus pointing to a fractional crystallization process rather than a partial melting process.

- d) *Energy constraints in open-system processes*. Recent Energy-Constrained Open-System process modelling has been performed, allowing a better quantification of assimilation rates in relation to the temperature of both the magma and the country rock according to Spera and Bohrsen (2001) and Bohrsen and Spera (2001). Two consequences can be inferred from these works: a) the onset of the assimilation process depends on the emplacement level, the lower crust providing a suitable scenario for an easier melting process of the country rock, and b) after magma stagnation in the crust, a certain time span is required for the country rock temperature to increase up to its solidus temperature. Assuming that most sills

Fig. 8. Trace-element fractional crystallization graphic modelling for sills 7 and 9. a) REE patterns showing the initial magma (chilled margin), a more differentiated term where the ilmenite is not a fractionating phase (gabbro), and the FC modelling results using only the proportion of fractionating phases and residual liquid estimated based on the mass balance. Note the parallel distribution in all cases. b), c), d), e) and f) log Th vs. log immobile element or immobile element ratio diagrams. Assuming the chilled margin (white crosses) as the initial magma, samples with lower Th contents can be considered to be cumulates, whereas those with higher Th contents would represent different stages of the liquid line of descent, the most Th-enriched sample (SG3-59.85, grey circle) being the most differentiated. The Rayleigh fractionation trend is included for different percentages of residual liquid (small circles on trend), as well as the mineral association and proportion accumulated (see b), and in Th–V the real liquid line of descent (grey line). Chondrite-normalized incompatible element (g) and REE (h) patterns from sill 9, showing the initial magma (chilled margin) and the most differentiated term of the liquid line of descent (the metagabbro 751/31). Two different Rayleigh modelling results have been included: i) heavy dashed line: using only the proportion of fractionating phases and residual liquid according to mass balance calculations, and ii) grey line: using a lower content of residual liquid ($F=37\%$). Partition coefficients, after Nielsen (2005). Symbols, as in Fig. 4. Normalising values after Sun and McDonough (1989) for diagrams “a” and “g”, and after Evensen et al. (1978) for diagram “h”.

studied did not undergo assimilation, the country rock temperature could not have reached the solidus temperature, inhibiting any partial melting processes. In order to check the previous hypothesis, an EC-AFC modelling was performed using thermal and compositional parameters assumed for the upper crust (see [Bohrson and Spera, 2001](#)). Silurian metapelites were used as the wall rock, whereas those of sill 9 were taken as the compositional parameters (isotope Nd signature and trace-element composition) of the magma, including sample 751/29 (chilled margin) as the initial parental magma and 751/31 (meta-gabbro) as the most-differentiated member. The results obtained ([Fig. 9](#)) appear to be in agreement with a closed-system behaviour, at least for this sill, since the Nd (61.8 ppm) and Nb (35.8 ppm) contents of the evolved member, i.e. a meta-gabbro, correspond to a temperature of about 700 °C, which is lower than the temperature required for melting to occur.

8.3. Primary melt composition

The chilled margins from the LCS sills lack some chemical conditions to be considered primary magmas in equilibrium with their mantle source; examples are $mgN \geq 68$, $MgO \geq 11$ wt.%, $Ni > 300$ ppm, and Fe/Mg ratios ranging from 0.4 to 0.8, which is in equilibrium with mantle olivine (Fe_{91}) ([Frey et al., 1978](#); [Albarède, 1992](#)). Hence, the ascending magma *en route* to the emplacement level in the upper crust appears to have undergone some kind of magma fractionation process, either in the magma chamber or in the conduit itself.

The low range of variation in the Al_2O_3/CaO ratio (1.3–1.4) for chilled margins of the LCS sills might be consistent with olivine fractionation after a mantle-derived olivine tholeiitic magma (Al_2O_3/CaO ratio ≈ 1.2), as has been proposed by [Frey et al. \(1978\)](#) for tholeiites and by [Cook et al. \(2005\)](#) for transitional tholeiites. A correction using mass-balance calculations for the assumed fractionated olivine was modelled in order to fit Fe_{91} -bearing equilibrated mantle melts, in agreement with [Albarède \(1992\)](#) and [Aldanmaz et al. \(2006\)](#). Once the proportion of olivine had been estimated and major-element compositions of the chilled margins had been recalculated, some trace-element contents of the chilled margins were also corrected, taking into account the Rayleigh formulation ($C_0 = C_L / F^{(D-1)}$) and assuming the fractionation value to be $F = 1 - X_{O1}$. The results obtained ([Table 4](#)) allow us to approach the bulk composition of the protolith as well as the partial melting degree from the mantle tholeiites.

Table 4

Calculated primary melts and CIPW norms for fractionation-corrected chilled margins

Sill	7	9
Sample	SG0.5*	751/29*
Rock type	Ch.m.	Ch.m.
Added olivine (%)	15.8	19
SiO ₂	48.59	47.86
TiO ₂	2.21	2.40
Al ₂ O ₃	12.19	10.85
Fe ₂ O ₃	1.90	2.09
FeO	9.67	10.65
MnO	0.18	0.21
MgO	13.46	14.69
CaO	8.47	8.36
Na ₂ O	2.59	2.45
K ₂ O	0.51	0.19
P ₂ O ₅	0.22	0.27
mgN	0.71	0.71
Al ₂ O ₃ /CaO	1.44	1.30
<i>CIPW norm</i>		
Pl	42.0	38.8
Or	3.0	1.1
Di	16.4	17.4
Hy	11.8	13.9
Ol	19.3	20.6
Il	4.2	4.6
Mt	2.8	3.0
Ap	0.5	0.6
<i>Trace elements (ppm)</i>		
Nb	10.8	12.2
Zr	153	146
Y	29	42
La	11.45	12.31
Ce	27.79	30.05
Nd	17.01	20.17
Sm	4.61	5.54
Eu	1.57	1.94
Gd	4.94	6.40
Tb	0.90	1.08
Dy	5.37	6.38
Er	3.13	3.64
Tm	0.44	0.50
Yb	2.61	3.07
Lu	0.38	0.46

Ch.m.: chilled margin.

8.4. Source constraints

The constancy of the highly incompatible element ratios (e.g. La/Yb, La/Sm) of the chilled-margin samples prevents a suitable mantle source to be inferred using the dynamic melting inversion (DMI) method proposed by [Zou and Zindler \(1996\)](#). Hence, other kinds of methods must be used, involving different protoliths.

In this work, the whole rock composition of the diabase sills shows contradictory features: on one hand,

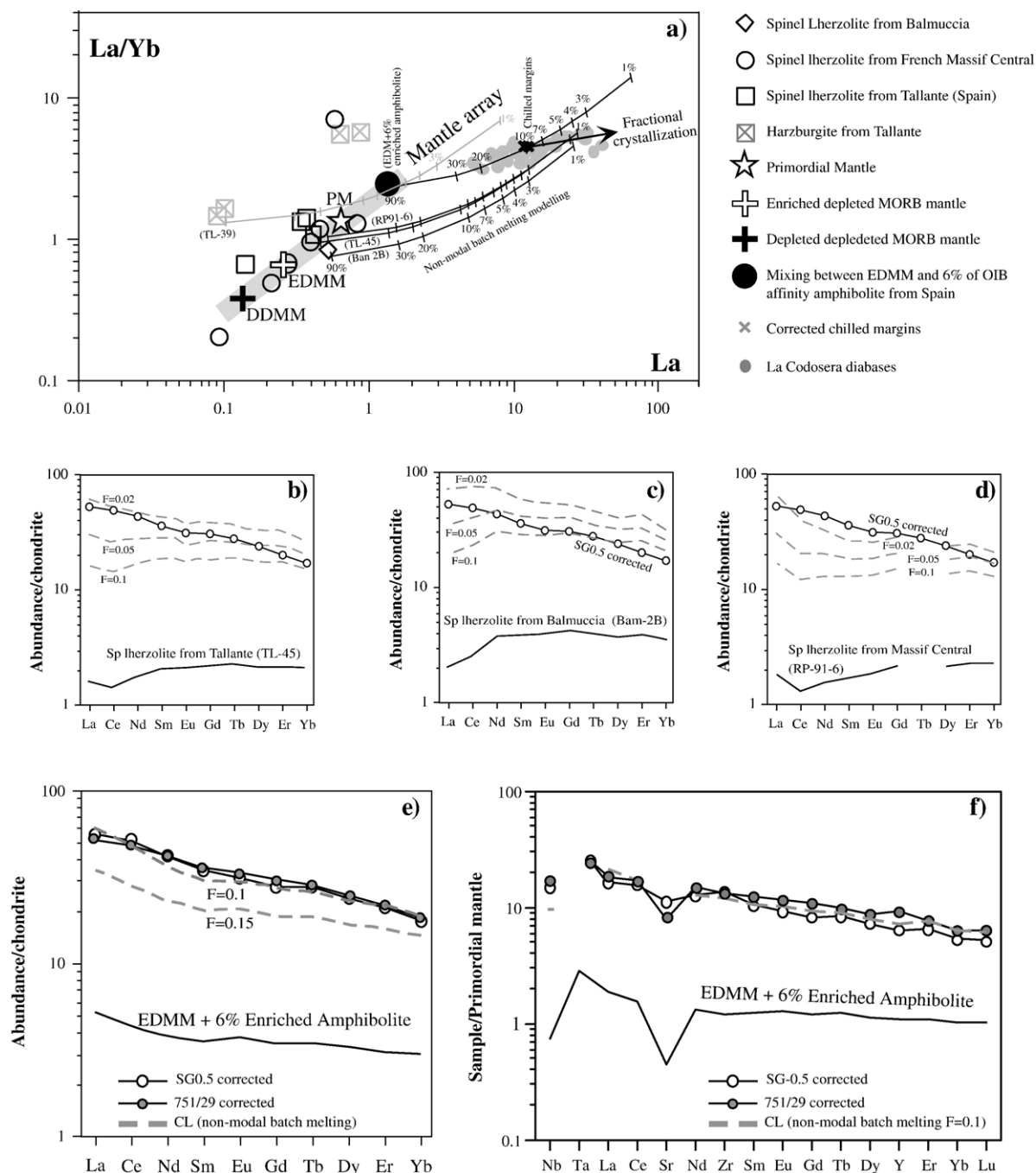


Fig. 10. Non-modal batch-melting modelling using different mantle protoliths. a) La/Yb–La diagram showing melt curves for harzburgites and spinel-lherzolites from Cenozoic European Nb trough xenoliths, as well as for a hybrid protolith consisting of EDM plus 6 wt. of OIB-affinity amphibolite from the Central Iberian Zone (see text). Tick marks on each curve correspond to the degrees of partial melting. Mode of spinel-lherzolites from Tallante is $ol_{0.58} + opx_{0.24} + cpx_{0.11} + sp_{0.02} + pl_{0.05}$ (Beccaluva et al., 2004); mode of Tallante harzburgites is $ol_{0.75} + opx_{0.22} + cpx_{0.03}$ (Beccaluva et al., 2004); mode for the rest of protoliths is $ol_{0.554} + opx_{0.252} + cpx_{0.173} + sp_{0.021}$ (Shaw, 2000), whereas melt mode was taken as the same for all the cases according to Shaw (2000), as follows: $ol_{0.20} + opx_{0.30} + cpx_{0.50}$. Mineral/melt partition coefficients are after Shaw (2000), Kelemen et al. (1993) and Fugimaki et al. (1984). b), c), d), e) and f) Plots of REE and multicomponent patterns for partial melts, quoted protoliths and chilled margin compositions (sills 7 and 9) after correction for crystallization. Normalising values after Sun and McDonough (1989) for incompatible elements, and after Evensen et al. (1978) for REE. EDM: Enriched Depleted MORB Mantle after Workman and Hart (2005); DDM: Depleted Depleted MORB Mantle after Workman and Hart (2005); PM: Primordial Mantle after Sun and McDonough (1989); spinel-lherzolite and harzburgite from Tallante after Beccaluva et al. (2004); spinel-lherzolite from Balmuccia after Hartmann and Wedepohl (1993); spinel-lherzolite from French Massif Central after Zangana et al. (1999).

a mantle component accounted for by a high Nd isotope signature ($\epsilon_{\text{Nd}}^{436} = +6$) and, on the other, a crustal-affinity component, as inferred from a moderate degree of LREE enrichment relative to HREE, as well as from an Nb negative anomaly.

The Nd isotope values of the diabbases studied are intermediate between a depleted mantle (DM) and a primordial mantle (PM) or an enriched mantle (probably EMI). Concerning the Nb trough, different mechanisms have been invoked: i) carbonatite melts that modified the protolith (Ionov et al., 1993); ii) rutile as a residual phase in the source (Kelemen et al., 1993; Rivalenti et al., 1998); iii) chromatographic fractionation produced by alkaline melts (Bedini et al., 1997; Ionov et al., 2002a and b); iv) hydrous fluids from subducting slabs (Downes, 2001), and v) different Nb behaviours in fluids (Hartmann and Wedepohl, 1993). Protoliths that have been metasomatized by carbonatite melts would be expected to show strong Nb, Zr and Hf depletions and low values in the Ti/Eu ratio (<1000 , Downes, 2001), which is not the case. The second possibility can be ruled out owing to the low Nb/Ta ratio (Foley et al., 2002). In addition, the lack of negative anomalies in Ta, Zr and Hf, as well as high Th/Nb ratios makes the involvement of hydrous fluids from subducting slabs to be unlikely (Downes, 2001).

The contrasting behaviours of Nb and Ta is difficult to account for in terms of either fluid involvement, owing to their similar low degree of mobility (Bevins et al., 1984; Rollinson, 1993), or crustal contamination, because potential contaminants, such as slates, have a marked negative Ta anomaly.

Percolation and metasomatism due to a porous flow of alkaline melts/fluids in a depleted mantle would have produced an enrichment in LREE, a depletion in ϵ_{Nd} , and an Nb trough (Ionov et al., 2002a,b). Amphibole seems to be an essential Nb trough-forming mineral phase (Ionov et al., 2002a; Foley et al., 2002), although Nb trough-bearing amphibole-free spinel lherzolites and harzburgites have been reported (Hartmann and Wedepohl, 1993; Zangana et al., 1999; Beccaluva et al., 2004). Thus, clinopyroxene could be the key, arising from its own Nb-depletion.

Additionally, the low Gd/Yb ratio (around 0.9 for all the samples) and the relatively low pressure conditions of magma segregation (1.2 to 1.3 GPa), using the empirical relationship after Albarède (1992), rule out a garnet-bearing metasomatized mantle source. Therefore, spinel lherzolites and harzburgites with Nb troughs would be the most suitable protoliths.

A batch-melting modelling involving selected European spinel-bearing lherzolites and harzburgites having

an Nb trough and lacking a negative Ta anomaly indicates that the calculated melt compositions display a strong depletion in La; accordingly, an unrealistically low partial melting degree would be necessary to fit the corrected chilled margin compositions (Fig. 10). Also, a very low partial melting degree ($\sim 2\%$) might be inferred after chondrite-normalized REE diagrams, depicting patterns poorly fitted to those of corrected chilled margin compositions (Fig. 10b, c, d). Thus, a more REE-enriched mantle source having a higher La/Yb ratio would be required, i.e., a metasomatized mantle source, instead of the one inferred after Cenozoic mantle xenoliths.

An enriched mantle source also can be inferred from the pervasive presence of Early Paleozoic OIB-affinity amphibolitic rocks from the innermost domain of the Central Iberian Zone (“Orthogneiss Domain”), as has been recently reported (López-Plaza et al., *in press*). A significant Nb-trough-inherited mantle plume contribution appears to be plausible since the REE-enriched amphibolitic rocks themselves display the negative Nb anomaly and relatively high ϵ_{Nd} positive values, ranging from +7 to +3.8. A certain contribution by a deep mantle plume in relation to the amphibolitic source would be expected to have occurred owing to the relatively short distance from the “orthogneiss domain” region. A mixing modelling using a depleted mantle component and one of the REE-enriched amphibolites having a within-mantle array isotope signature was assessed. The proportions of both components were determined according to the best fitting for the Nd-isotope signature ($\epsilon_{\text{Nd}}^{436} = +6$) of LCS sills, resulting in 6 wt.% of amphibolitic component and the rest for the depleted mantle component (Fig. 10e and f). The trends of the melts thus calculated after the hybrid mantle source include the chilled margin compositions after correction for crystallization, indicating 10% of partial melting (Fig. 10a, e and f), in agreement with reported values for E-MORB affinity magmas (Shervais, 2004) and for Q-normative basalts from New Zealand (Cook et al., 2005).

9. Conclusions

Ordovician-Silurian mafic sheet-like bodies from the Central Iberian Zone include Mg-rich basalts, Fe-rich basalts and andesites. They show an asymmetrical pattern, consisting of fine-grained diabbases at the bottom, pegmatoids at the top, and metagabbros in the middle of the body.

Since the chilled margin composition is similar to the bulk composition of the sill, a single magma pulse can be inferred. In detail, geochemical modelling using the chilled margin as the parental magma is consistent with

the Cr-rich accumulation level at the bottom and upwards-evolving residual liquids, accounting for the S-type profile. Closed-system behaviour during the fractional crystallization process is also supported by the Sm/Nd isochrone and the large magma/host rock temperature contrast, ruling out any significant contamination at upper crustal level.

A certain amount of olivine fractionation is required to explain the composition of chilled margin (low Cr, Ni, V, Sc and mg-number values).

The relative LILE enrichment, Sm/Nd isotope signature, and Nb trough are geochemical features that point to a metasomatised mantle as the fertile protolith. A small amount of OIB-mantle component (6%) represented by pre-Ordovician LREE-enriched amphibolites outcropping throughout the deepest domain of the Central Iberian Zone was added to a depleted end-member mantle component, resulting in a suitable hybrid mantle source for 10% of partial melting.

Acknowledgements

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Appendix A. (Materials and methods)

Fifty-five fresh samples were selected for the whole-rock analysis and twenty-nine for electron microprobe analyses.

Minerals were analysed with a JEOL microprobe at the “Centro de Microscopía Luis Bru, Universidad Complutense de Madrid”, Spain. Operating conditions were 15 kV and 20 nA, with a counting time of 15 s.

Whole-rock analyses were carried out at the “ACTLABS” in Canada, using the Code 4Lithoresearch package. A 1 g sample was digested with aqua regia and diluted to 250 ml volumetrically. Appropriate international reference materials for the metals of interest were digested at the same time. The samples and standards were analysed on an inductively coupled plasma-atomic emission spectroscopy device (ICP-AES) for major elements, whereas trace elements and REE were determined by ICP-MS.

Five samples were also selected for the study of Sm–Nd isotopes. Sm–Nd analyses were done at the University of the Basque Country, Spain, using a Finnigan MAT262 with a $^{143}\text{Nd}/^{144}\text{Nd}=0.5118435\pm$

$13 (2\sigma, n=18)$ value for La Jolla standard. A $^{150}\text{Nd}/^{149}\text{Sm}$ mixed isotope tracer was used. Chemical procedures for Sm and Nd extractions are described in Pin et al. (1994) and Pin and Santos (1997). Nd and Sm were loaded onto a Ta filament with $\text{H}_3\text{PO}_4\text{--HNO}_3$ (Platzner, 1997) and measured by multicollection in Faraday cups. Blanks for Sm and Nd were negligible. The isochrons were regressed using the Isoplot/Ex program of Ludwig (1999). ϵ_{Nd} values were calculated using CHUR parameters: $^{143}\text{Nd}/^{144}\text{Nd}=0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd}=0.1967$ (DePaolo and Wasserburg, 1976). Decay constants used were: $1.42 \times 10^{-11} \text{ yr}^{-1}$ (^{87}Rb) (Steiger and Jäger, 1977) and $6.54 \times 10^{-12} \text{ yr}^{-1}$ (^{147}Sm) (Lugmair and Marti, 1978).

Appendix B. Supplementary data

Supplementary data associated with this article can be found, in the online version, at [doi:10.1016/j.lithos.2006.12.006](https://doi.org/10.1016/j.lithos.2006.12.006).

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