

Geological, Petrological and Geochemical Evidence for Progressive Construction of an Arc Crustal Section, Sierra de Valle Fértil, Famatinian Arc, Argentina

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The petrogenesis of calc-alkaline magmatism in the Famatinian arc is investigated in the central Sierra Valle Fértil, a major, lower to middle crustal section of the Early Ordovician active margin of West Gondwana. Large-scale field relationships show a gradual and continuous compositional variation of the plutonic sequence, ranging from olivine-bearing gabbro-norites to hornblende- and biotite-bearing granodiorites. Distinctive lithostratigraphic units are, however, discernible as one compositional type of plutonic rock dominates over mappable areas. These results allow us to identify a continuous plutonic arc stratigraphy that progressively exposes shallower paleo-depths towards the east. At all the exposed levels, calc-alkaline plutonic rocks are volumetrically dominant, interrupted only by granulite-facies migmatites and leucogranites. The migmatites are interpreted to be refractory remnants of supracrustal sedimentary successions, whereas the peraluminous leucogranites have field relationships and chemical and isotopic compositions suggesting that they were produced via anatexis of metasedimentary packages. Mass-balance calculations predict that a parental gabbroic magma after progressive closed-system fractionation would crystallize about 80% of the original mass to yield a granodioritic daughter. Because the crystallizing mineral assemblage comprises hornblende and plagioclase, mass balance suggests a volume of residual amphibole-rich gabbroic rocks much larger than that observed, suggesting that differentiation is significantly driven by open-system processes. Indeed, the combination of field and petrographic observations with bulk-rock geochemistry and petrogenetic modeling demonstrates that

most dioritic and tonalitic rocks are hybrids formed by either (1) bulk assimilation of metasedimentary materials into gabbroic magmas, or (2) multi-stage and complex interactions between gabbroic rocks and metasedimentary-derived leucogranitic melts. The source region of the granodioritic magmas is located at the transition zone between a tonalite-dominated intermediate unit and a granodiorite-dominated silicic unit. Typical granodiorites have a hornblende-bearing mineralogy, metaluminous chemical signature and isotopic compositions [$^{87}\text{Sr}/^{86}\text{Sr}(T) = 0.7075\text{--}0.7100$ and $\epsilon_{\text{Nd}}(T) \sim -5.0$] broadly overlapping those of the tonalites of the intermediate rock unit. These major compositional features of the granodiorites can be best explained if three end-member components contribute to their generation. As field observational data suggest, primitive gabbroic rocks, metaluminous intermediate magmas and anatectic leucogranitic melts mixed to produce the calc-alkaline granodiorites; however, the exact petrological process generating the granodioritic magmas is unclear because the mafic end-member may have been incorporated as mafic inclusions in the intermediate magmas or as syn-magmatic dikes, or both. The polygenetic nature of the intermediate to silicic plutonic rocks, along with the preponderance of parental gabbroic rocks at the inferred base of the plutonic column, suggests an upward growth of the intermediate to silicic crust that involved the complete reconstitution of the pre-existing crustal configuration. The main implication of this study is that intermediate and silicic plutonic rocks in the Valle Fértil section formed within a crustal column in which the mass transfer and heat

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input of mantle-derived magmas promoted fusion of fertile metasedimentary rocks and favored mixing of gabbroic or dioritic magmas with crustal granitic melts. Our results lend support to models asserting that the thermal and material budget of arc magmatism is primarily governed by the rate at which mafic magmas ascend from their mantle sources and intrude repeatedly into the crust.

KEY WORDS: *igneous petrology; granitoids; open system; batholith construction; active continental margin*

INTRODUCTION

The petrogenesis of calc-alkaline magmatic suites at destructive ocean–ocean or ocean–continent plate margins has received a great deal of attention over the years (Moore, 1959; Bateman & Eaton, 1967; Pitcher *et al.*, 1985; Anderson, 1990; among many others). The evolution of subduction zone magmatism is generally of great significance in understanding the evolution of the continental crust of the Earth (Taylor & McLennan, 1985; Rudnick, 1995; Rudnick & Gao, 2003) and is also critical to societal issues such as volcanic hazards and the discovery of a large range of economic deposits [for a comprehensive review see Stern (2002)].

One of the major limitations in uniquely resolving the range of processes contributing to generating diversity in arc magma compositions is the rare and incomplete preservation of arc crustal sections exposing deep-seated plutonic levels (DeBari, 1997; Miller & Snoke, 2009). The incomplete view of arcs through exposures, heavily biased toward the upper parts of the system, makes most geochemical and petrogenetic models difficult to test and uniquely resolve. Among the most important issues at most arcs are: (1) the composition of the primitive magmas derived from the mantle; (2) the amount of water carried by these magmas; (3) the processes that generate the compositional diversity and silica enrichment in particular; (4) the overall mass balance of mafic vs silicic magmas and the composition and extent of magmatic residues.

The early Ordovician arc of northwestern Argentina, northeastern Chile, southwestern Bolivia and Peru is a relatively short-lived but extensive linear belt of calc-alkaline plutonic and volcanic rocks that formed along the proto-margin of South America (Pankhurst & Rapela, 1998; Lucassen & Franz, 2005; Chew *et al.*, 2007) via subduction of the Iapetus Ocean. Subduction and related arc magmatism ended at least along the southern parts (in present-day coordinates) by the collision of the Cuyania continental microplate. Famatinian arc rocks are exposed as basement rocks throughout the Altiplano–Puna region as well as in the Sierras Pampeanas (Fig. 1).

The Sierra Valle Fértil basement uplift of the Sierras Pampeanas is of particular importance because it exposes

a tilted and relatively undeformed section of the middle to deep crust of the Famatinian arc (Otamendi *et al.*, 2009a; Ducea *et al.*, 2010). This study presents new geochemical and isotopic data on representative rocks of the Valle Fértil paleo-arc section in an effort to distinguish the most significant petrogenetic processes that constructed the crustal plutonic column. We use new and previously published data to show that arc diversity is generated within the middle crust of the upper plate where hydrous, mantle-derived mafic magmas interacted extensively with a thick sequence of continental margin metasedimentary rocks at depths of 15–27 km.

GEOLOGICAL SETTING

Famatinian arc

The geological setting of the Famatinian arc has been presented in detail elsewhere (e.g. Pankhurst *et al.*, 1998; Coira *et al.*, 1999; Lucassen & Franz, 2005; Otamendi *et al.*, 2009a; Ducea *et al.*, 2010). Below, we highlight some of the key aspects of the regional geology that are most relevant to this study.

The Famatinian arc, including its extension into the present-day Puna plateau, extends between latitudes 24° and 39°S along the active foreland of the central Andes (Fig. 1). This early Ordovician arc was part of the active margin of Gondwana, which evolved during a long-lived period between the latest Neoproterozoic and late Paleozoic. Today, large segments of accretionary orogens formed on the Gondwanan margins are found from Australia through the Pacific border of Antarctica to the northern Andes (Cawood, 2005; Chew *et al.*, 2007).

At about latitude 28°S the Famatinian arc displays a pronounced change in the exposed paleo-depth. The deepest plutonic levels are dominantly granodioritic and tonalitic batholiths exposed along a roughly north–south-trending belt that extends from latitude 28° to 33°S (Fig. 1), coinciding with an area where the Nazca plate is at present subducting at a relatively shallow angle under South America (Barazangi & Isacks, 1976). To the north of latitude 28°S, within the Puna plateau, early Ordovician volcanic complexes expose the extrusive cover of the arc formed coevally with sedimentary sequences (Mannheim & Miller, 1996; Toselli *et al.*, 1996; Bahlburg, 1998; Coira *et al.*, 1999; Cisterna *et al.*, 2004; Kleine *et al.*, 2004).

From the southern Puna to the central Sierras de Famatina and La Rioja (Fig. 1), the early Ordovician Famatinian batholiths are emplaced into a metasedimentary crustal sequence. In the Sierras de Famatina and La Rioja, muscovite, biotite, sillimanite, andalusite and cordierite coexist in some of the metasedimentary rocks (Saal *et al.*, 1996; Dahlquist *et al.*, 2005), suggesting pressures lower than 4 kbar, corresponding to middle-crustal depths (<15 km). At middle-crustal levels, the batholith host-rocks are siliciclastic successions, mostly from a

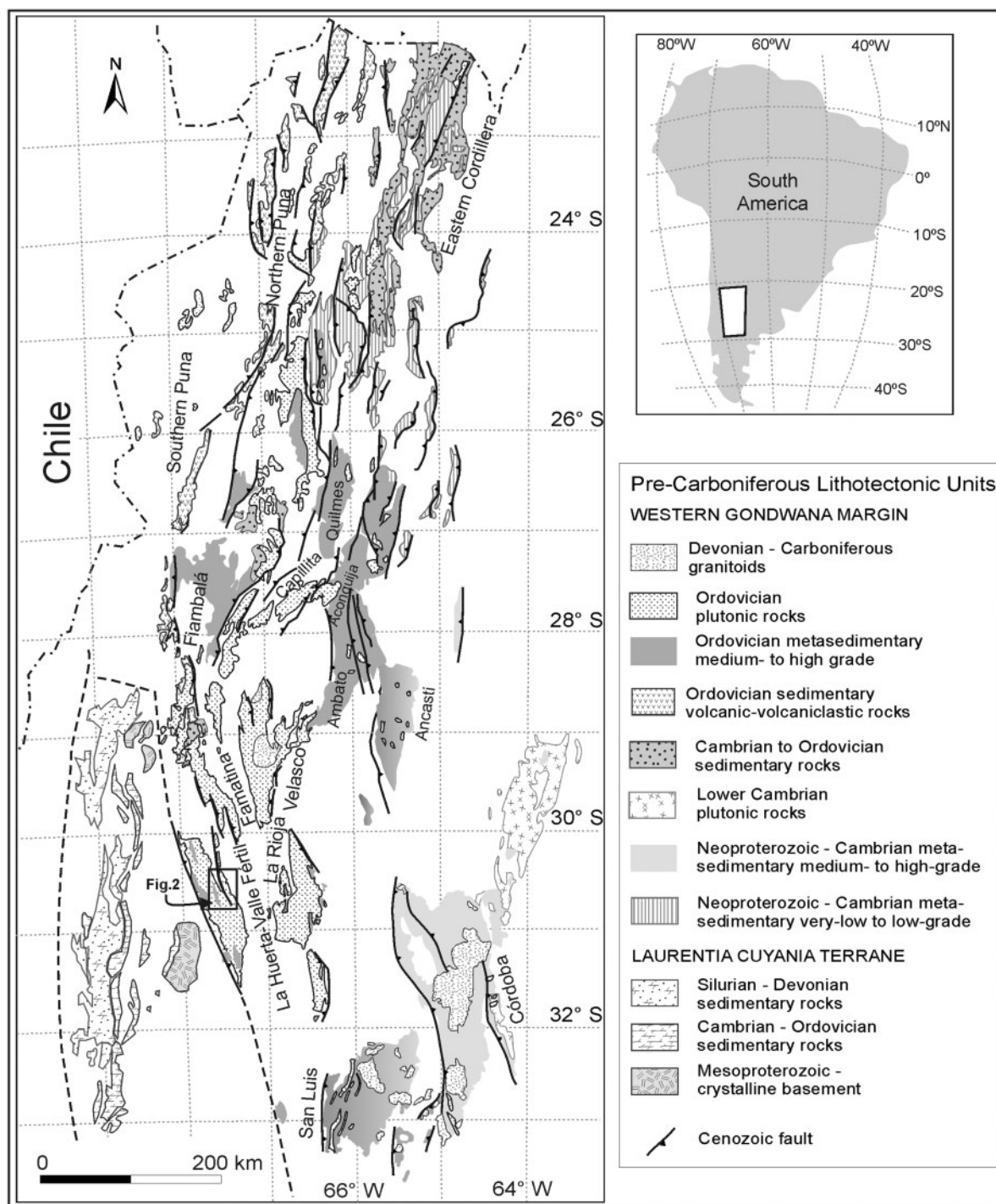


Fig. 1. Geological map showing the present-day extent of the Early Ordovician arc and its location with respect to the western Gondwana landmasses, and the Laurentia-derived Cuyania terrane. This sketch map is modified after Bahlburg (1998), Coira *et al.* (1999), and Astini & Dávila (2004). Inset shows the location of the map in South America.

marine environment, formed between the early Cambrian and early Ordovician (Mángano & Buatois, 1996; Aceñolaza *et al.*, 2000; Collo *et al.*, 2009). The mafic batholith roots of early Ordovician age are at paleo-depths of

exposure deeper than 20 km (>6 kbar of pressure) in the Sierras Valle Fértil-Huerta range (Otamendi *et al.*, 2008; Ducea *et al.*, 2010; Gallien *et al.*, 2010). Widespread partial to almost complete melting of pelitic and greywacke

host-rocks took place at more than 15 km depth (>5 kbar) (Otamendi *et al.*, 2009a). Below these depths, mafic igneous rocks dominate the plutonic sequence and were intruded into deeply buried metasedimentary rocks.

Because the Famatinian arc crust shows a clear along-strike gradient in the paleo-depth of exposure, we can observe the pre-Ordovician crustal configuration into which the entire plutonic–volcanic arc sequence was emplaced. The oldest outcropping stratigraphic unit along the Famatinian arc is a quartz-rich turbidite succession that was deposited from Middle to Late Cambrian times (Clemens & Miller, 1996; Astini & Dávila, 2004). This shallow marine succession does not preserve diagnostic fossils, but its pattern of detrital zircon ages suggests a maximum depositional age of Middle Cambrian (Collo *et al.*, 2009; Adams *et al.*, 2010). Middle to Upper Cambrian turbidite successions extend over a large region of northwestern Argentina and, where their base is observed, they unconformably overlie folded Late Neoproterozoic to Early Cambrian turbidite sequences (Aceñolaza *et al.*, 2000; Piñán Llamas & Simpson, 2006). Sedimentation in shallow to deep marine environments continued from the latest Cambrian to early Ordovician (Tremadoc) and followed the first appearance of early Ordovician volcanism in the late Tremadoc (Mángano & Buatois, 1996; Mannheim & Miller, 1996).

During the earliest Paleozoic the oceanic plate descending beneath the West Gondwana margin carried a continental microplate that had previously rifted from Laurentia (Thomas & Astini, 1996). When this continental microplate (Cuyania) docked along the Gondwana margin about 460 Myr ago, subduction-related magmatism ceased along the Famatinian segment of the Ordovician arc (Astini & Dávila, 2004). The ensuing collisional event closed the suture between the Famatinian arc plutonic crust and the Cuyania terrane (Mulcahy *et al.*, 2011). By the Carboniferous, the Gondwana margin had migrated hundreds of kilometers outboard. The Famatinian plutonic crust was unconformably overlain by sedimentary sequences deposited between the late Carboniferous and middle Tertiary. Magmatism was restricted to minor basaltic volcanism along major normal faults during Triassic extension. Shallow subduction of the Nazca plate beneath South America at the latitude of this study area commenced about 15 Myr ago, perhaps because of the subduction of the Juan Fernández Ridge (Yáñez *et al.*, 2001). As a result, the deformation front migrated inland and coupling between the two plates increased (Jordan & Allmendinger, 1986); the Sierra Valle Fértil and other segments of the Famatinian arc were uplifted during this period as basement-cored anticlines.

Sierra de Valle Fértil section

The transition between lower- and upper-crustal levels is exposed in the Sierra Valle Fértil within the western belt

of the exhumed Famatinian arc. Integrating field relationships from hand-specimen through outcrop to regional scales, several previous studies have allowed the position of the rock units at the time the magmatism was active to be inferred (Mirrè, 1976; Vujovich *et al.*, 1996; Otamendi *et al.*, 2009a; Gallien *et al.*, 2010). The shallowest part of the exposed section corresponds to its eastern boundary, whereas deeper levels of the Ordovician crust are exposed to the west (Otamendi *et al.*, 2008, 2009a). The array of the lithological units crudely correlates with the variation in inferred paleo-depth, exhibiting progressively increasing silicic compositions eastwards (Fig. 2). At the deepest paleo-depths in the western Valle Fértil section, steeply dipping layering in igneous cumulates reflects overall tilting of more than 60° from their inferred original paleo-horizontal orientations. In sharp contrast, paleomagnetic data reveal that the eastern granodioritic batholiths experienced unroofing with imperceptible tilting (Rapalini & Pinotti, personal communication).

The deep-seated plutonic section of the early Ordovician arc is almost entirely igneous with subordinate screens of migmatitic metasedimentary rocks that were metamorphosed and partially melted during magma intrusion. Consequently, there are no stratigraphic relationships and mapping in the area is based entirely on thermobarometry studies of the metamorphic rocks and grouping of broad rock types. Because of the clear preponderance of one rock type across large field areas, the igneous crust is divided into five lithostratigraphic units, which are informally named mafic, intermediate, intermediate transitional, silicic, and metasedimentary.

The mafic lithological unit represents the deepest exposed crustal levels and is dominated by amphibole gabbro-norite and orthopyroxene–amphibole–biotite diorite. Whole-rock compositions reveal that almost all of the gabbroic rocks are cumulates; however, the original igneous textures have undergone varying degrees of re-crystallization. Only olivine-bearing layered bodies, presumably representing magma chamber deposits at upper levels of the mafic unit, preserve igneous features from millimeter to whole body length scales (Otamendi *et al.*, 2010a). The general framework of the mafic unit is one of numerous sills that intruded previously solidified gabbroic rocks and pre-existing crust, of which only highly residual granulite-facies metasedimentary remnants are found. Orthopyroxene-bearing and amphibole-rich quartz diorites are the second dominant rock type of the mafic unit and appear either as veins interlayered with the gabbroic rocks or as discrete bodies several hundreds of meters in length.

A cross-sectional view through the Valle Fértil region suggests that a tonalite-dominated intermediate unit grew overlying the mafic unit. The boundary zone between these units is transitional and appears as sequences of

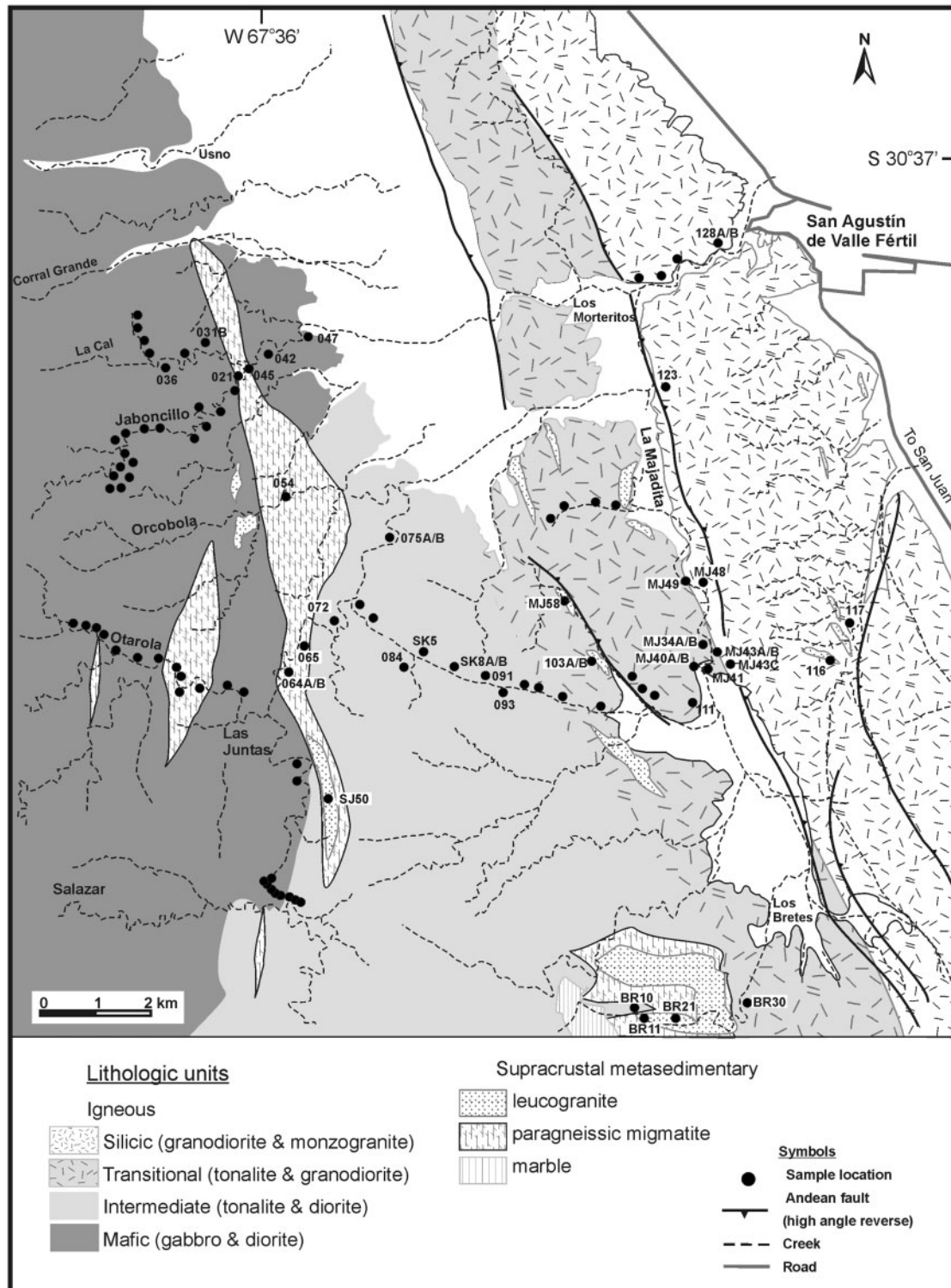


Fig. 2. Geological map of the central section of the Sierra de Valle Fértil showing the sample locations. Sample numbers are keyed to descriptions in Table 1. Other sample locations (after Otamendi *et al.*, 2009a, 2009b, 2010a, 2010b) are also shown because many of these specimens are used in our discussion.

mixed mafic and metasedimentary migmatites that are complexly interstratified with mafic and intermediate plutonic rocks (Otamendi *et al.*, 2009b). The intermediate unit consists of a heterogeneous suite of amphibole-rich and biotite-bearing tonalites. The dominant tonalites contain significant proportions (up to 50%) of gabbroic components that appear in many shapes and sizes. Large (up to 2 m) blocks of gabbro with sharp edges and rounded shapes are scattered throughout the tonalites. Mafic enclave swarms are channel-shaped and comprise several tens of tightly packed, small (5–25 cm), rounded, mafic enclaves. These display gradational, mingled contacts with their tonalitic host. Typical dikes vary between 1 cm and 1 m in thickness, are several tens of meters in length and display irregular edges at outcrop scales (Fig. 3a). Mafic dikes show massive fined-grained texture, but typical chilled margins are mostly absent. The dikes show convex-outward lobes and are internally composite (Fig. 3b). Mafic dikes meander through tonalitic rocks. Dikes with these characteristics are interpreted to have intruded when the host had a fraction of melt to allow ductile deformation of the mafic dike (e.g. Wiebe, 1993; Turnbull *et al.*, 2010). By contrast, tonalites are also intruded by mafic dikes with sharp boundaries that display tabular shapes and cross cut the magmatic foliation (Otamendi *et al.*, 2009a). The latter mafic dikes intruded when the tonalites had crystallized.

At the inferred upper levels of the intermediate unit, the tonalites that dominate this lithological unit are commingled with leucogranite (Fig. 3c). At a regional scale, a mappable belt characterized by a mixture of tonalitic rocks and leucogranites occurs as a boundary zone separating the intermediate unit and the silicic unit; therefore, in our mapping this is defined as the 'intermediate transitional unit' (Fig. 2). Within this transitional zone, coarse-grained amphibole- and biotite-bearing tonalites grade eastwards to a mixture of biotite tonalites and leucogranites. One distinctive petrological characteristic of the transitional intermediate unit is the coexistence of tonalites and leucogranites commingled from hand specimen to map scales (Fig. 3d and e). The tonalites and leucogranites are mingled rather than mixed, perhaps because of the large difference in the densities and viscosities of the magmas from which the tonalite and leucogranite crystallized (e.g. Furman & Spera, 1985; Frost & Mahood, 1987). Mafic inclusions are entrained within, and carried by the tonalitic magma. However, because leucogranitic veins neither cross-cut nor mingle with the mafic inclusions, we infer that the mafic inclusions were extensively crystallized when the tonalites began interacting with the leucogranites (Fig. 3e). In the upper parts of the transitional intermediate unit, the tonalites display an increase in the modal abundance of quartz and biotite, and include lens-shaped bodies of granodiorite. The boundary between the

transitional and silicic units is gradational and we have located it where the granodiorites become the dominant lithology.

The silicic unit comprises numerous plutons that together resemble the typical shallow exposure of a Cordilleran batholith. The silicic unit mostly consists of coarse-grained granodiorites that have either inequigranular or porphyritic textures. The typical granodiorite contains biotite and hornblende. Tonalites occur as irregular bodies inside the granodiorites. Except for the absence of K-feldspar, the petrographic characteristics of the tonalites resemble those of the granodiorites in several respects. The contact between the tonalites and granodiorites is transitional and extends over a few tens of centimeters. Primary foliation in the granodiorites and tonalites is defined by flattened mafic inclusions and the planar orientation of plagioclase, hornblende and biotite crystals (Fig. 3f). Inclusion swarms occur in almost all locations and are generally concordant with the subvertical primary foliation (Castro *et al.*, 2008). Measured perpendicular to the strike of the foliation, some zones with >50% inclusions are several meters thick; single inclusions in these zones are smaller than 1 m and range from fine-grained, homogeneous, amphibole gabbro to complexly mixed hybrid enclaves. The mixed enclaves are identified as such because they consist of fine-grained biotite and amphibole clots coexisting with coarse-grained K-feldspar and quartz. This mixture reflects extensive digestion of the mafic inclusions into the host granodiorites and tonalites (e.g. Turnbull *et al.*, 2010). In many cases field relationships suggest that mafic dikes that intruded into the evolving granodiorites experienced rapid solidification, as they seem to have been brecciated inside the still-moving host magma (Fig. 3g). The same variety of amphibole-rich gabbroic rocks is frequent in the silicic unit as discrete bodies with sizes of up to 1 km² exposed on surfaces. Even less abundant than the gabbroic bodies, metasedimentary rocks occur as blocks tens of meters long scattered throughout the silicic unit.

The metasedimentary unit consists of paragneissic migmatites (metatexite ≫ diatexite). This lithostratigraphic unit occurs as kilometer-long belts intercalated with mafic and intermediate igneous rocks or appears as widespread, meter-long fragments within all four igneous units. Within the metasedimentary unit, typical migmatites are characterized by well-developed layering that results from the alternation of mesosomes and leucosomes. Mesosomes typically include quartz, plagioclase, biotite, garnet, cordierite, sillimanite, and K-feldspar with modal variations that extend to the absence of up to two of these rock-forming phases. Thermobarometric estimates based on this mineral assemblage yielded pressures between 5.2 and 7.5 kbar at temperatures of about 805 ± 35°C for migmatitic packages located at the upper levels of the mafic

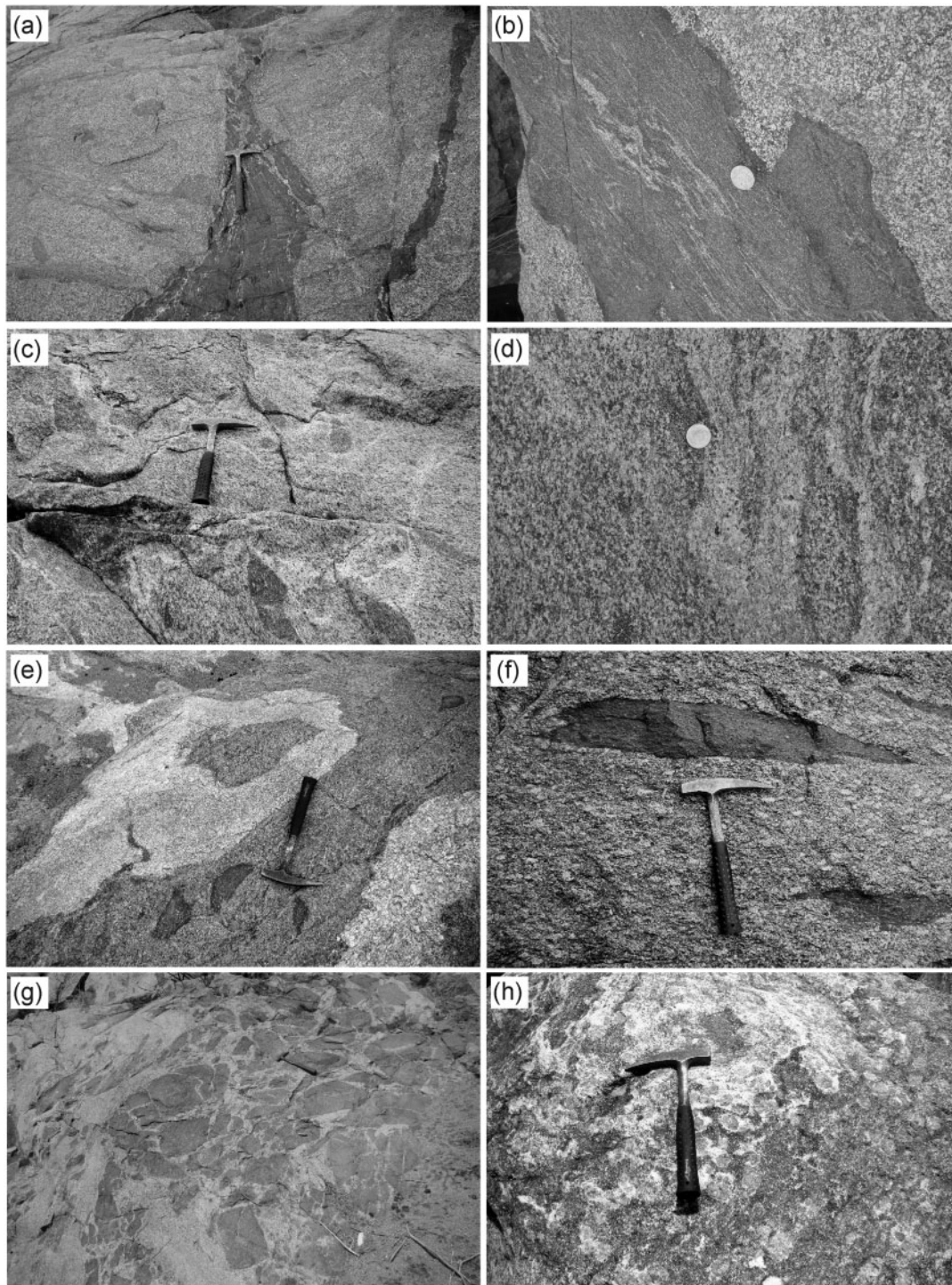


Fig. 3. (a) Mafic dike in quartz diorite from the intermediate unit. The aligned mafic inclusions preserve evidence of magmatic flow in the diorite. The dike has a sinuous pattern suggesting that it was deformed and disrupted by the movement of the host magma. Locally, the magmatic foliation cuts across the diorite dike. (b) Detail of a mafic dike in a tonalitic host-rock showing the cusped boundary and composite nature that are characteristic of syn-magmatic dikes. (c) Example of amphibole-rich tonalite complexly commingled with leucogranitic veins in the intermediate transitional unit. The leucogranite develops a network of connected veins that extensively permeate the tonalite. Most leucogranitic veins do not cut across the mafic inclusions, which were enclosed in the tonalite before the tonalitic and leucogranitic magmas began to interact. (d) Detail of the interlayering of tonalite and leucogranite (white bands). The mineral colour index is the most evident feature that characterizes the contrasting magmas; tonalite contains more than 20% modal hornblende plus biotite, whereas the leucogranite has low modal proportions of mafic minerals. The magmatic foliation of these rocks is close to parallel, suggesting that the two contrasting rocks coexisted as magmas. (e) The transitional intermediate unit, showing the close coexistence of three lithologies: leucogranite, tonalite, and mafic inclusions. (f) Magmatic foliation in a granodiorite from the silicic unit shown by the strong preferred orientation of feldspar megacrysts. (g) Syn-magmatic mafic dike that was broken apart when intruded into granodiorite of the silicic unit. (h) Residual migmatite that occurs as screens in the transitional unit and contains a high modal proportion of garnet and cordierite porphyroblasts. In this case, garnet is surrounded by leucocratic domains of quartz and K-feldspar.

unit (Otamendi *et al.*, 2008; Gallien *et al.*, 2010; Tibaldi *et al.*, 2011). Commonly, a leucogranitic body appears in the core or along one border of the large metasedimentary packages (Fig. 2). The transitional contacts between paragneissic migmatites and leucogranites, as well as the widespread appearance of residual mineral assemblages in the metasedimentary unit (Fig. 3h), suggest that the paragneissic migmatites and leucogranites are of a residue–melt complementary nature. In the central Valle Fértil section, U–Pb zircon geochronology indicates that the migmatites attained peak metamorphic conditions and experienced anatexis during the emplacement of the early Ordovician magmas (Rapela *et al.*, 2001; Ducea *et al.*, 2010).

METHODS

Sampling rationale

The selection of samples for the present study was based on previous information on the field relationships and petrological characteristics of the distinct regional-scale lithostratigraphic units summarized above (see Otamendi *et al.*, 2009a). All the samples were collected on a nearly east–west transect across the Famatinian plutonic section from the central Sierra Valle Fértil. The samples studied cover the full range of lithological diversity, as outlined for each lithostratigraphic unit (Table 1 and Fig. 2). To establish the origin of the intermediate and silicic plutonic rocks, significant emphasis was put on sampling petrological variations throughout lithological units dominated by diorites, tonalites and granodiorites. Further sampling was aimed at characterizing a spectrum of supracrustal sedimentary-derived lithologies, which were metamorphosed and partially melted in spatial and temporal relation with the early Ordovician plutonism (Ducea *et al.*, 2010). In particular, the anatectic leucogranites derived from the metasedimentary packages were sampled within all the major units. This allows identification and comparison of compositional variations among the leucogranites in the lithological units dominated by other plutonic rock types. Mafic inclusions collected from either enclaves or large blocks in the distinct units are indicated in Table 1; the sampled inclusions are listed after the rocks that host them.

Data and methods

We analyzed major and trace elements and Sr–Nd isotopes for around 40 bulk-rock samples; these data are reported in Tables 2 and 3. A set of 24 representative rocks was selected for the determination of their Pb isotope composition (Table 4).

Most of the specimens selected for whole-rock chemical analysis consisted of more than 2 kg of fresh material. For specimens with sample number starting with ‘KS’, fresh samples with no weathered surfaces and/or secondary minerals were powdered in an Al₂O₃-lined crusher. Powder

aliquots were sent to the Geoanalytical Laboratory of the School of Earth and Environmental Sciences, Washington State University, for major and trace element analysis by X-ray fluorescence (XRF) and inductively coupled plasma mass spectrometry (ICP-MS). Procedures and reproducibility have been reported by Johnson *et al.* (1999).

The remaining specimens collected were crushed and pulverized as described by Otamendi *et al.* (2009b). In the latter sample population, major elements and selected trace elements (Ba, Rb, Sr, and Zr) were determined by XRF, whereas trace elements were measured by ICP-MS at the University of Huelva, Spain. The equipment, reproducibility and accuracy of analytical procedures at the University of Huelva have been reported by de la Rosa *et al.* (2001).

The isotopic ratios $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ and the trace element concentrations of Rb, Sr, Sm, and Nd for all the samples were measured by thermal ionization mass spectrometry on whole-rock samples at the University of Arizona following the procedure described in detail elsewhere (Ducea & Saleeby, 1998; Otamendi *et al.*, 2009b).

Washes from the cation column separation were used for separating Pb in Sr-Spec resin (Eichrom, Darien, IL) columns using a protocol developed at the University of Arizona. Samples were loaded in 8 M HNO₃ in the Sr-Spec columns. Pb elution was achieved via 8 M HCl. Lead isotope analysis was carried out by multicollector (MC)-ICP-MS on a GV Instruments (Hudson, NH) system at the University of Arizona (Thibodeau *et al.*, 2007; Drew *et al.*, 2009). Samples were introduced into the instrument by free aspiration with a low-flow concentric nebulizer into a water-cooled chamber. A blank, consisting of 2% HNO₃, was run before each sample. Before analysis, all samples were spiked with a Tl solution to achieve a Pb/Tl ratio of 10. Throughout the experiment, the standard National Bureau of Standards (NBS)-981 was run to monitor the stability of the instrument. All results were Hg-corrected and empirically normalized to Tl by using an exponential law correction. To correct for machine and interlaboratory bias, all results were normalized to values reported by Galer & Abouchami (2004) for the National Bureau of Standards (NBS)-981 standard ($^{206}\text{Pb}/^{204}\text{Pb} = 16.9405$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.4963$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.7219$). Internal error reflects the reproducibility of the measurements on single samples, whereas external errors are derived from long-term reproducibility of the NBS-981 Pb standard and result in part from the mass bias effects within the instrument. In all cases, external error exceeds the internal errors and is reported below. External errors associated with each Pb isotopic ratio are as follows: $^{206}\text{Pb}/^{204}\text{Pb} = 0.028\%$; $^{207}\text{Pb}/^{204}\text{Pb} = 0.028\%$; $^{208}\text{Pb}/^{204}\text{Pb} = 0.031\%$.

Table 1: Summary of petrographic observations

Unit and sample	Number on map ¹	Rock type	Mineralogy ²	Age ³ (Ma)
<i>Mafic</i>				
KS021	021	Hornblende gabbro	Pl-Opx-Hbl	
KS031B	031B	Two-pyroxene gabbro	Opx-Cpx-Hbl-Pl-(Mt)	
KS036	036	Hornblende gabbro	Pl-Opx-Hbl-(Mt)	
KS042	042	Olivine-hornblende gabbro	Ol-Opx-Pl-Hbl	
KS047	047	Diorite	Pl-Hbl-Opx-Bt-Qtz-(Mt)	
KS065	065	Two-pyroxene gabbro	Cpx-Opx-Hbl-Pl	
<i>Intermediate</i>				
KS072	072	Quartz diorite	Pl-Qtz-Hbl-Bt-(Mt)	
KS075A	075A	Diorite	Hbl-Bt-Pl-Qtz	
KS075B	075B	Amphibole gabbro (MI) ⁴	Hbl-Pl-Mt	
KS084	084	Tonalite	Qtz-Pl-Hbl-Bt-(Mt)	472.3 ± 5.8
VFSK5	SK5	Quartz diorite	Hbl-Pl-Qtz-Bt-(Zrn, Ap, Mt)	
VFSK8A	SK8A	Tonalite	Pl-Qtz-Bt-Hbl-(Zrn)	
VFSK8B	SK8B	Quartz diorite (MI)	Hbl-Pl-Qtz-(Mt, Zrn)	
KS093	093	Tonalite	Qtz-Pl-Hbl-(Bt, Mt, Ep)	472.9 ± 5.9
<i>Transitional</i>				
KS111	111	Tonalite	Qtz-Pl-Bt-(Mt, Aln)	
KS103B	103B	Amphibole gabbro (MI)	Hbl-Pl-(Qtz)	
VFBR10	BR10	Tonalite	Qtz-Pl-Hbl-Bt-(Mt, Zrn)	
VFBR11	BR11	Granodiorite	Qtz-Pl-Kfs-Bt-(Mt, Ep)	
VFBR30	BR30	Tonalite	Qtz-Pl-Hbl-Bt-(Mt)	
VFMJ34A	MJ34A	Tonalite	Pl-Qtz-Bt	
VFMJ34B	MJ34B	Tonalite	Hbl-Pl-Qtz-Bt	
VFMJ40A	MJ40A	Tonalite	Pl-Qtz-Bt-(Zrn, Ap, Mt)	
VFMJ40B	MJ40B	Leucogranite	Qtz-Kfs-Pl-Bt-(Ap, Mt, Zrn)	
KS103A	103A	Leucogranite	Qtz-Kfs-Pl-Bt	472.7 ± 10.4
VFMJ58	MJ58	Leucogranite	Qtz-Kfs-Pl-Bt-(Ms, Ap, Mt)	
<i>Silicic</i>				
KS117	117	Granodiorite	Qtz-Kfs-Pl-Bt-(Ep, Ox, Aln)	470.3 ± 7.7
KS128A	128A	Granodiorite	Qtz-Pl-Kfs-Bt-(Ep, Mt)	473.4 ± 8.2
KS128B	128B	Quartz diorite (MI)	Pl-Hbl-Bt-Qtz-(Ap, Mt, Ep)	
VFMJ41	MJ41	Granodiorite	Qtz-Pl-Kfs-Bt-(Sph, Ap, Mt)	
VFMJ43A	MJ43A	Tonalite	Hbl-Pl-Qtz-Bt-(Zrn, Ap, Mt)	
VFMJ43B	MJ43B	Quartz diorite (MI)	Hbl-Pl-Qtz-Bt-(Zrn, Ap, Mt)	
VFMJ43C	MJ43C	Quartz diorite	Pl-Qtz-Bt-Hbl-(Zrn, Ap, Mt)	
VFMJ48A	MJ48	Tonalite	Pl-Qtz-Bt-Hbl-(Ap, Mt)	
VFMJ49	MJ49	Amphibole gabbro (MI)	Hbl-Pl-Bt-(Ap, Zrn)	
KS116	116	Leucogranite	Qtz-Kfs-Pl-Bt	
KS123	123	Leucogranite	Qtz-Pl-Kfs-Bt-(Mt, Ep)	
<i>Metasedimentary</i>				
KS045	045	Metapelite migmatite	Pl-Qtz-Kfs-Bt-Sil-(Mt, Zrn)	
KS054	054	Metapelite migmatite	Qtz-Pl-Bt-Kfs-Crd-Sil-(Mt)	
KS064B	064B	Metapelite migmatite	Qtz-Pl-Bt-Grt-Sil-(Mt)	
KS091	091	Metapelite migmatite	Qtz-Pl-Crd-Bt-Sil-(Zrn)	476.9 ± 11.4
KS064A	064A	Leucogranite	Qtz-Kfs-Pl-Crd-(Bt)	
VFSJ50	SJ50	Leucogranite	Qtz-Kfs-Pl-(Bt, Ms, Mnz)	
VFBR21	BR21	Leucogranite	Qtz-Kfs-Pl-(Bt, Ms, Zrn)	

¹Labels as used in map shown as Fig. 2.²Mineral abbreviations after Kretz (1983), rock-forming minerals listed in order of decreasing abundance from left to right, accessory mineral are in parentheses.³See Ducea *et al.* (2010) for details of U-Th-Pb zircon geochronology.⁴(MI) denotes the specimen is a mafic inclusion hosted by the preceding rock.

Table 2: Whole-rock major element (wt %) and trace element (ppm) composition

Unit:	Mafic						Intermediate		
	KS021	KS031B	KS036	KS042	KS047	KS065	KS072	KS075A	KS075B
SiO ₂	48.35	49.25	47.66	44.59	49.02	48.99	59.56	53.79	49.25
TiO ₂	0.87	0.67	0.45	0.11	1.29	0.87	0.79	0.85	0.99
Al ₂ O ₃	17.51	4.22	19.00	21.97	18.28	15.24	16.53	18.51	14.86
Fe ₂ O ₃ *	12.06	18.22	10.30	7.53	13.79	10.67	8.01	9.63	14.28
MnO	0.19	0.34	0.21	0.12	0.24	0.18	0.13	0.20	0.25
MgO	7.45	20.62	9.79	11.69	4.98	8.59	3.25	4.22	7.12
CaO	11.41	6.22	11.63	13.36	9.27	12.96	5.87	6.96	10.65
Na ₂ O	0.95	0.28	1.01	0.64	2.16	1.65	2.46	2.63	1.44
K ₂ O	0.25	0.11	0.22	0.10	0.69	0.19	1.98	2.34	0.59
P ₂ O ₅	0.12	0.01	0.05	0.01	0.28	0.07	0.17	0.44	0.05
LOI	0.91	0.72	0.44	0.52	0.83	1.01	1.06	0.74	0.82
Total	100.07	100.67	100.76	100.63	100.83	100.42	99.80	100.32	100.30
Sc	40.6	64.6	36.4	14.5	49.3	47.7	33.9	51.8	51.3
Rb	5.1	2.0	1.0	3.2	19.8	1.2	69.9	82.3	4.5
Sr	300.1	19.9	165.1	189.3	215.5	106.0	169.7	185.2	305.3
Y	14.1	12.4	15.8	2.4	13.1	20.3	41.3	54.7	25.2
Zr	38.7	14.3	28.0	6.2	192.5	43.2	177.7	52.2	29.4
Nb	2.1	0.9	2.7	0.3	8.4	2.3	9.9	13.4	3.7
Cs	0.1	0.1	0.0	0.2	0.8	0.0	0.9	1.4	0.1
Ba	134.2	13.4	41.2	21.8	162.6	23.9	405.6	407.1	67.9
La	10.09	2.27	9.50	1.38	16.19	4.30	21.79	19.29	8.37
Ce	22.58	7.50	23.54	2.99	38.96	11.15	51.32	52.76	18.68
Pr	2.98	1.24	3.08	0.37	5.48	1.64	6.80	7.75	2.32
Nd	12.15	6.18	6.05	1.47	24.72	7.55	8.26	13.67	23.55
Sm	2.86	1.94	2.69	0.37	6.10	2.23	2.75	3.63	6.61
Eu	1.02	0.35	0.70	0.21	1.69	0.85	1.56	1.94	0.97
Gd	3.09	2.18	2.80	0.44	6.49	3.06	8.25	10.77	3.53
Tb	0.49	0.40	0.48	0.08	1.10	0.57	1.43	1.84	0.70
Dy	2.87	2.58	3.04	0.49	6.78	3.83	8.62	11.14	4.75
Ho	0.58	0.52	0.61	0.10	1.37	0.82	1.71	2.24	1.04
Er	1.57	1.46	1.76	0.27	3.72	2.27	4.52	5.96	2.98
Tm	0.22	0.21	0.27	0.04	0.53	0.34	0.62	0.81	0.44
Yb	1.39	1.36	1.73	0.25	3.27	2.08	3.54	4.67	2.81
Lu	0.22	0.21	0.27	0.04	0.51	0.32	0.51	0.68	0.44
Hf	1.21	0.60	0.95	0.20	5.07	1.39	4.82	1.69	1.14
Ta	0.10	0.06	0.14	0.03	0.48	0.15	0.36	0.60	0.22
Pb	5.03	0.49	3.92	1.60	6.36	1.31	8.69	10.30	4.91
Th	1.70	0.38	0.15	0.30	1.45	0.15	3.41	1.17	1.47
U	0.34	0.07	0.04	0.08	0.51	0.07	0.26	0.33	0.27

(continued)

BULK-ROCK CHEMISTRY**Overview**

Rock types are established based on petrographic and major element characteristics (Le Maitre *et al.*, 1989).

Gabbro includes a broad spectrum of rock types that show both cumulate and non-cumulate textures, but always have SiO₂ < 52 wt % (see Otamendi *et al.*, 2009a). Some ultramafic cumulate rocks (SiO₂ < 45 wt %) plotted in subsequent variation diagrams were taken from previous

Table 2: Continued

Unit:	Intermediate					Transitional			
	KS084	VFSK5	VFSK8A	VFSK8B	KS093	KS111	KS103A	KS103B	VFBR10
SiO ₂	52.56	52.44	74.96	45.23	58.54	65.62	73.83	49.99	61.51
TiO ₂	1.26	1.25	0.14	1.05	0.86	0.69	0.12	0.92	0.83
Al ₂ O ₃	18.38	18.88	14.32	16.95	16.74	15.31	12.62	16.78	16.09
Fe ₂ O ₃ *	10.63	10.64	1.35	14.52	8.30	6.06	1.18	10.71	7.58
MnO	0.16	0.19	0.02	0.24	0.18	0.09	0.02	0.21	0.15
MgO	4.27	4.23	0.74	7.23	3.67	1.72	0.29	6.53	2.65
CaO	6.57	7.26	5.93	12.12	5.84	4.30	1.34	10.17	5.93
Na ₂ O	2.16	2.45	1.66	1.32	2.49	2.94	1.64	2.17	2.65
K ₂ O	2.62	2.35	0.50	0.40	2.17	1.95	6.46	1.20	1.44
P ₂ O ₅	0.25	0.29	0.08	0.09	0.14	0.19	0.04	0.19	0.21
LOI	0.67	0.88	0.69	1.00	0.93	0.84	1.24	1.06	0.70
Total	99.53	100.85	100.37	100.16	99.86	99.70	98.77	99.93	99.74
Sc	29.7	41.6	1.7	58.3	44.4	14.0	1.1	33.4	35.9
Rb	120.5	94.4	27.9	5.5	139.3	83.8	84.8	33.6	51.7
Sr	186.7	210.4	150.1	139.1	164.8	149.7	150.7	252.1	203.9
Y	24.4	41.8	1.9	18.5	37.6	16.1	2.8	22.7	14.1
Zr	311.7	37.3	12.0	37.2	171.8	194.6	104.4	77.7	103.0
Nb	10.5	11.6	1.9	2.7	12.0	10.6	1.2	5.5	8.5
Cs	1.8	1.7	0.7	0.0	1.3	2.1	0.5	1.1	0.7
Ba	493.5	371.4	91.5	55.1	347.5	187.1	1127.4	227.5	379.6
La	15.91	16.90	5.54	7.76	42.14	30.38	4.85	19.74	17.68
Ce	36.34	41.48	8.48	19.21	86.45	61.42	7.71	43.64	36.37
Pr	4.76	6.72	0.74	2.74	9.86	7.16	0.80	5.66	4.01
Nd	18.55	24.83	2.46	11.68	37.91	26.65	2.87	21.84	17.82
Sm	4.62	3.83	0.37	3.16	8.49	5.16	0.56	4.68	4.46
Eu	1.41	1.94	0.29	0.93	1.53	1.09	1.01	1.49	1.26
Gd	5.21	9.24	0.37	3.52	7.98	4.34	0.46	4.79	3.06
Tb	0.84	1.51	0.05	0.57	1.31	0.66	0.07	0.76	0.55
Dy	5.04	8.78	0.30	3.55	7.91	3.63	0.42	4.51	3.31
Ho	1.00	2.13	0.04	0.86	1.55	0.70	0.09	0.92	0.60
Er	2.61	5.19	0.16	2.19	4.08	1.80	0.30	2.49	1.51
Tm	0.37	0.71	0.02	0.29	0.56	0.25	0.05	0.35	0.18
Yb	2.23	4.05	0.13	1.89	3.23	1.46	0.37	2.27	1.24
Lu	0.34	0.56	0.02	0.26	0.46	0.24	0.08	0.36	0.18
Hf	8.14	1.18	0.37	1.42	4.74	5.47	3.44	2.40	2.52
Ta	0.46	3.40	2.30	2.64	0.50	0.58	0.06	0.33	0.45
Pb	9.02	20.22	15.85	12.80	11.07	17.26	25.59	7.69	11.23
Th	1.06	1.12	0.96	1.11	7.04	9.40	0.79	3.59	0.94
U	0.32	0.22	0.30	0.20	0.33	0.47	0.41	0.59	0.45

(continued)

studies (Otamendi *et al.*, 2010a). The classification of the intermediate and silicic plutonic rock types correlates well with the SiO₂ contents of the rocks as the great majority of them fall in the following compositional ranges: quartz

diorites 52 < SiO₂ < 56 wt %; tonalites 55 < SiO₂ < 67 wt %; granodiorites–monzogranites 65 < SiO₂ < 71 wt % (on an anhydrous basis). Leucogranites are characterized by low color indices (<5) and SiO₂ > 72 wt %.

Table 2: Continued

Unit:	Transitional							Silicic	
	VFBR11	VFBR30	VFMJ-34A	VFMJ-34B	VFMJ-40A	VFMJ-40B	VFMJ58	VFMJ-43B	VFMJ-43A
SiO ₂	69.67	61.85	66.09	46.00	66.94	73.81	71.84	48.95	49.35
TiO ₂	0.38	0.81	0.64	1.27	0.63	0.12	0.06	0.91	1.19
Al ₂ O ₃	15.01	16.18	15.74	16.18	15.68	14.20	15.51	17.15	17.05
Fe ₂ O ₃ *	2.90	6.56	5.67	14.70	5.32	0.82	0.45	11.25	11.07
MnO	0.03	0.13	0.12	0.26	0.09	0.02	0.01	0.20	0.21
MgO	1.15	2.88	1.75	6.34	2.00	0.30	0.57	6.52	6.21
CaO	2.82	5.79	4.73	10.47	3.86	1.22	1.37	10.63	10.23
Na ₂ O	2.97	2.81	2.91	2.40	2.99	1.80	2.30	2.03	2.25
K ₂ O	3.95	1.99	2.46	1.09	2.32	8.00	7.80	1.21	1.46
P ₂ O ₅	0.07	0.13	0.18	0.39	0.15	0.22	0.06	0.15	0.39
LOI	0.80	0.70	0.40	0.77	0.81	0.46	0.34	1.15	0.87
Total	99.75	99.83	100.68	99.87	100.78	100.98	100.30	100.14	100.28
Sc	21.8	37.3	n.a.	n.a.	19.5	3.3	1.1	35.7	47.6
Rb	101.2	81.9	n.a.	n.a.	126.1	165.0	133.6	29.0	51.5
Sr	160.7	209.5	n.a.	n.a.	158.6	169.3	141.1	283.6	311.7
Y	13.0	20.1	n.a.	n.a.	15.0	15.9	8.9	26.5	40.0
Zr	115.1	94.2	n.a.	n.a.	41.2	16.3	32.0	54.9	46.5
Nb	9.9	10.8	n.a.	n.a.	14.6	2.7	1.3	7.3	12.3
Cs	0.9	0.7	n.a.	n.a.	1.6	0.8	0.5	0.7	1.6
Ba	669.6	365.5	n.a.	n.a.	227.3	1202.0	754.8	135.8	246.7
La	74.56	18.79	n.a.	n.a.	20.65	13.40	22.62	15.22	32.05
Ce	153.33	43.46	n.a.	n.a.	39.81	27.87	44.38	37.63	69.24
Pr	17.41	5.21	n.a.	n.a.	5.02	3.74	5.70	5.65	11.12
Nd	29.67	27.00	n.a.	n.a.	11.07	10.86	19.81	23.09	39.80
Sm	7.51	5.63	n.a.	n.a.	2.76	2.88	3.89	5.61	8.27
Eu	1.30	0.96	n.a.	n.a.	1.15	1.65	1.58	1.37	2.40
Gd	6.78	4.16	n.a.	n.a.	3.84	3.88	3.22	5.48	9.59
Tb	0.86	0.68	n.a.	n.a.	0.57	0.61	0.40	0.87	1.45
Dy	3.48	4.41	n.a.	n.a.	3.17	3.38	2.03	5.11	8.12
Ho	0.54	0.86	n.a.	n.a.	0.69	0.74	0.37	1.26	1.97
Er	1.50	2.55	n.a.	n.a.	1.57	1.67	0.83	3.25	4.80
Tm	0.14	0.32	n.a.	n.a.	0.17	0.18	0.08	0.44	0.69
Yb	0.90	2.31	n.a.	n.a.	0.96	1.09	0.62	2.73	4.04
Lu	0.10	0.26	n.a.	n.a.	0.11	0.12	0.07	0.39	0.59
Hf	3.01	2.67	n.a.	n.a.	1.30	0.64	1.68	2.15	2.27
Ta	0.70	1.39	n.a.	n.a.	3.17	2.50	1.42	0.93	1.33
Pb	32.13	11.32	n.a.	n.a.	19.57	45.02	51.98	8.82	11.33
Th	6.95	1.61	n.a.	n.a.	4.28	2.81	15.60	2.54	2.86
U	1.06	0.2	n.a.	n.a.	0.34	0.39	0.76	0.61	0.73

(continued)

Leucotonalitic veins and dikes in the mafic migmatites, as described by Otamendi *et al.* (2009b), are not included in all the diagrams, but will be used in some of the discussion.

Gabbroic rocks appear in all of the lithostratigraphic units; a petrographic distinction, however, is that within the transitional and silicic units the gabbroic rocks contain calcic-amphibole as the sole ferromagnesian phase

Table 2: *Continued*

Unit:	Silicic								
	KS128B	VFMJ-43C	VFMJ41	KS117	KS128A	VFMJ-48A	VFMJ49	KS116	KS123
SiO ₂	48.81	57.56	68.44	65.41	68.51	66.20	49.12	76.54	71.72
TiO ₂	1.56	0.61	0.57	0.63	0.52	0.61	0.42	0.03	0.25
Al ₂ O ₃	17.29	20.25	14.99	15.15	14.63	16.00	16.61	12.09	13.59
Fe ₂ O ₃ *	13.80	6.66	5.09	6.02	4.62	5.61	7.74	0.44	2.17
MnO	0.31	0.09	0.09	0.12	0.09	0.10	0.18	0.01	0.04
MgO	4.32	2.01	1.49	1.79	1.30	1.60	9.45	0.08	0.71
CaO	9.16	7.67	4.27	4.65	4.18	5.03	11.83	1.18	2.05
Na ₂ O	2.97	3.66	3.16	2.84	2.96	3.14	2.12	1.84	2.15
K ₂ O	1.49	1.56	1.75	2.27	1.71	1.96	0.86	5.80	5.48
P ₂ O ₅	0.26	0.24	0.15	0.16	0.12	0.16	0.06	0.02	0.05
LOI	0.80	0.61	0.95	0.47	0.75	0.47	1.75	1.07	1.15
Total	100.77	100.92	100.93	99.52	99.38	100.88	100.14	99.11	99.35
Sc	41.1	23.1	22.7	22.6	9.1	18.7	47.7	0.8	3.6
Rb	40.3	74.1	92.0	77.2	82.6	83.0	48.5	114.6	118.5
Sr	204.1	345.5	151.3	165.3	130.1	202.2	277.3	124.8	140.0
Y	38.4	18.6	25.0	35.9	18.8	25.5	47.7	5.5	7.6
Zr	165.2	53.9	58.2	166.1	136.9	48.4	48.1	59.7	110.1
Nb	9.9	7.3	13.3	10.4	8.0	9.8	13.8	0.7	3.6
Cs	1.6	3.6	3.3	1.3	3.3	2.4	1.7	0.5	1.2
Ba	168.1	236.0	223.2	469.4	227.4	352.8	214.7	873.9	1116.1
La	28.64	23.40	38.71	33.95	26.87	33.06	34.02	9.61	35.13
Ce	66.83	42.87	70.48	70.20	51.58	59.60	75.41	18.21	67.56
Pr	8.66	5.38	9.37	8.14	5.88	7.88	12.55	2.02	6.94
Nd	22.71	19.40	25.95	19.44	15.51	23.38	50.34	7.06	13.08
Sm	5.10	3.92	4.53	4.36	2.82	5.81	11.28	1.29	1.82
Eu	2.31	1.49	1.30	1.39	1.02	1.47	2.52	1.03	0.79
Gd	7.74	3.81	5.77	6.95	3.62	5.05	10.72	1.21	2.07
Tb	1.26	0.57	0.89	1.20	0.59	0.78	1.63	0.19	0.28
Dy	7.73	3.43	5.17	7.47	3.58	4.61	9.45	1.06	1.50
Ho	1.55	0.84	1.21	1.50	0.72	1.14	2.32	0.21	0.29
Er	4.18	2.04	2.85	3.97	2.02	2.96	5.87	0.52	0.80
Tm	0.58	0.28	0.38	0.54	0.30	0.42	0.88	0.07	0.12
Yb	3.65	1.62	2.11	3.08	1.98	2.61	5.25	0.44	0.74
Lu	0.56	0.22	0.28	0.44	0.32	0.36	0.78	0.08	0.12
Hf	4.38	1.77	1.59	4.58	4.00	1.32	2.22	2.50	3.33
Ta	0.53	3.95	2.27	0.47	1.09	1.89	3.73	0.08	0.18
Pb	11.21	18.14	13.50	10.38	13.47	22.27	11.45	31.30	22.35
Th	1.36	2.52	8.60	5.64	9.99	8.04	5.40	4.42	9.35
U	0.67	0.59	1.03	0.44	2.60	0.52	1.22	0.33	0.30

(continued)

(Table 1). Diorites and tonalites appear in the intermediate, transitional and silicic units spanning a wide compositional range from 52 to 65 wt % SiO₂; however, whereas diorites with SiO₂ < 57 wt % dominate the

intermediate unit, tonalites are more abundant in the transitional and silicic units. The appearance of granodiorites and monzogranites is restricted to the transitional and silicic units.

Table 2: Continued

Unit:	Metasedimentary				Leucogranite		
	KS045	KS054	KS064B	KS091	KS064A	VFSJ50	VFBR21
SiO ₂	55.43	63.35	60.42	73.80	75.57	76.12	77.23
TiO ₂	1.38	1.03	1.21	0.72	0.07	0.09	0.06
Al ₂ O ₃	18.41	16.42	17.06	11.69	13.02	12.81	12.49
Fe ₂ O ₃ *	11.27	8.70	9.43	4.81	1.18	0.68	0.50
MnO	0.19	0.18	0.27	0.06	0.05	0.01	0.03
MgO	4.48	3.11	4.67	1.61	0.54	0.08	0.03
CaO	0.87	1.49	1.07	0.93	1.18	1.10	1.25
Na ₂ O	1.60	2.23	1.35	1.83	2.18	2.76	2.56
K ₂ O	4.86	2.64	3.07	3.01	4.63	5.47	5.17
P ₂ O ₅	0.05	0.06	0.05	0.13	0.08	0.01	0.02
LOI	1.40	1.06	1.65	1.21	0.83	0.60	0.50
Total	99.94	100.27	100.25	99.80	99.33	99.73	99.84
Sc	23.4	17.6	21.6	11.0	1.4	12.3	15.1
Rb	92.8	69.1	105.2	96.1	84.4	132.2	87.6
Sr	164.4	132.9	108.3	132.0	229.6	107.6	68.3
Y	17.1	31.1	35.8	36.1	8.3	3.9	10.0
Zr	252.3	230.3	245.6	315.3	47.7	88.1	162.6
Nb	12.9	15.4	18.2	12.9	0.8	0.3	0.2
Cs	0.7	1.2	1.9	0.8	0.5	0.5	0.2
Ba	929.2	636.9	547.3	481.9	1125.3	728.4	488.8
La	79.46	59.09	68.37	42.23	18.93	17.14	15.19
Ce	166.35	124.28	149.76	85.67	36.93	35.39	27.84
Pr	18.54	14.13	16.40	10.31	3.88	3.86	2.59
Nd	41.74	44.37	42.50	39.65	14.05	16.55	7.20
Sm	6.77	8.36	8.06	7.62	2.82	3.44	1.67
Eu	2.52	1.82	1.87	1.60	2.15	1.19	0.77
Gd	8.46	8.67	10.12	7.22	2.42	2.10	1.36
Tb	1.00	1.28	1.50	1.19	0.35	0.32	0.19
Dy	4.51	6.86	8.02	7.00	1.84	1.27	1.54
Ho	0.74	1.27	1.45	1.40	0.34	0.22	0.40
Er	1.67	3.21	3.63	3.86	0.85	0.63	1.72
Tm	0.20	0.42	0.49	0.57	0.12	0.07	0.29
Yb	1.25	2.40	2.81	3.55	0.71	0.32	2.09
Lu	0.22	0.37	0.44	0.55	0.11	0.04	0.43
Hf	7.37	6.86	7.17	9.03	1.39	2.42	4.25
Ta	0.39	0.80	0.71	0.78	0.07	0.59	0.38
Pb	26.00	19.78	18.06	20.38	47.62	33.93	30.66
Th	26.68	23.41	24.48	15.64	5.80	9.28	3.68
U	0.47	1.05	1.88	5.33	4.27	0.47	0.36

*Total Fe given as Fe₂O₃.

n.a. indicates that the analytical datum was not analysed; LOI, loss on ignition.

Leucogranites show *in situ* relationships within the meta-sedimentary migmatites and appear as veins or dikes in the intermediate, transitional and silicic units. True leucogranites are mostly absent in the mafic unit. Except for

the leucogranites, gabbroic sills and dikes are the only rocks with igneous plutonic textures that have been found intruded into the migmatite packages from the metasedimentary unit.

Table 3: Sr and Nd isotope and trace element data

	Rb (ppm)	Sr (ppm)	⁸⁷ Rb/ ⁸⁶ Sr	⁸⁷ Sr/ ⁸⁶ Sr	σ %	⁸⁷ Sr/ ⁸⁶ Sr (475 Ma)	Sm (ppm)	Nd (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	σ %	¹⁴³ Nd/ ¹⁴⁴ Nd (475 Ma)	ε _{Nd} (475 Ma)
Mafic Unit													
KS021	5.11	300.13	0.0490	0.706825	0.001680	0.706495	2.86	12.15	0.142041	0.512323	0.000015	0.511882	-2.8
KS031B	2.03	19.94	0.2932	0.711415	0.000170	0.709437							
KS036	0.98	165.11	0.0171	0.709174	0.000213	0.709059	2.69	6.05	0.268592	0.512729	0.000025	0.511895	-2.6
KS042	3.15	189.29	0.0478	0.708606	0.000212	0.708283	0.37	1.47	0.152177	0.512223	0.000015	0.511802	-4.4
KS047	19.83	215.49	0.2647	0.712647	0.000384	0.710862	6.10	24.72	0.149259	0.512228	0.000015	0.511764	-5.1
KS065	1.22	106.06	0.0329	0.705446	0.000296	0.705224	2.23	7.55	0.178681	0.512626	0.000025	0.512071	0.9
Intermediate Unit													
KS072	69.88	169.66	1.1869	0.726737	0.004840	0.718731	2.75	8.26	0.201112	0.512285	0.000012	0.511660	-7.2
KS075A	82.29	185.19	1.2797	0.722425	0.002000	0.713793	3.63	13.67	0.160431	0.512308	0.000008	0.511810	-4.2
KS075B	4.48	305.30	0.0422	0.709624	0.000199	0.709339	6.61	23.55	0.169591	0.512249	0.000031	0.511722	-6.0
KS084	120.51	186.73	1.8592	0.726570	0.004000	0.714030	4.62	18.55	0.150641	0.512199	0.000027	0.511731	-5.8
VFSK5	94.35	210.42	1.2910	0.721061	0.001800	0.712353	3.83	24.83	0.093205	0.512007	0.000010	0.511717	-6.0
KS093	139.33	164.77	2.4361	0.726279	0.000625	0.709848	8.49	37.91	0.135400	0.512146	0.000029	0.511725	-5.9
Transitional Unit													
KS111	83.84	149.67	1.6132	0.723158	0.003330	0.712277	5.16	26.65	0.117093	0.512089	0.000010	0.511725	-5.9
KS103A	84.84	150.70	1.6212	0.722168	0.000185	0.711233	0.56	2.87	0.119071	0.512097	0.000018	0.511727	-5.9
KS103B	33.63	252.12	0.3836	0.709550	0.000156	0.706962	4.68	21.84	0.129386	0.512273	0.000012	0.511871	-3.0
VJBR10	51.68	203.91	0.7294	0.715467	0.001800	0.710547	4.46	17.82	0.151104	0.512125	0.001500	0.511656	-7.3
VFBR11	101.24	160.66	1.8160	0.730654	0.002100	0.718405	7.51	29.67	0.153079	0.512058	0.001100	0.511582	-8.7
VFBR30	81.87	209.47	1.1250	0.717057	0.001100	0.709469	5.63	26.99	0.126127	0.512172	0.001200	0.511780	-4.8
VFMJ40A	126.14	158.58	2.2917	0.726666	0.002200	0.711209	2.76	11.07	0.150741	0.512170	0.002000	0.511702	-6.4
VFMJ40B	164.97	169.28	2.8086	0.730281	0.001500	0.711337	2.88	10.86	0.160108	0.512245	0.001100	0.511748	-5.5
Silicic Unit													
KS117	77.20	165.63	1.3417	0.718219	0.000016	0.709169	4.36	19.44	0.135660	0.512193	0.000009	0.511772	-5.0
KS128A	82.57	130.09	1.8270	0.719868	0.000013	0.707545	2.82	15.51	0.109805	0.512104	0.000018	0.511763	-5.2
KS128B	40.30	204.09	0.5682	0.714455	0.000432	0.710623	5.10	22.71	0.135776	0.512224	0.000008	0.511802	-4.4
VFMJ40D	157.60	171.82	2.6436	0.730986	0.001300	0.713155	2.46	12.04	0.123647	0.512106	0.001900	0.511722	-6.0
VFMJ41	91.89	151.29	1.7488	0.721716	0.001500	0.709920	4.53	25.95	0.105601	0.512099	0.001400	0.511771	-5.0
VFMJ43A	47.78	269.09	0.5107	0.710931	0.001600	0.707486	8.27	39.80	0.125556	0.512241	0.001500	0.511851	-3.4
VFMJ48A	82.98	202.18	1.1813	0.717403	0.001200	0.709435	5.81	23.38	0.150196	0.512213	0.003000	0.511746	-5.5
VFMJ49	31.28	191.59	0.4696	0.710763	0.002000	0.707596	0.90	3.44	0.157573	0.512323	0.002500	0.511833	-3.8
KS116							1.29	7.06	0.110899	0.512069	0.000015	0.511724	-5.9
KS123	118.50	140.03	2.4378	0.727370	0.000140	0.710927	1.82	13.08	0.084266	0.512033	0.000010	0.511771	-5.0
Metasedimentary Unit													
<i>Migmatite</i>													
KS045	92.79	164.39	1.6264	0.727802	0.000024	0.716832	6.77	41.74	0.097996	0.511983	0.000017	0.511679	-6.8
KS054							8.36	44.37	0.113927	0.512026	0.000010	0.511672	-6.9
KS064B	105.19	108.29	2.8023	0.740721	0.000189	0.721820	8.06	42.50	0.114614	0.512013	0.000018	0.511657	-7.2
KS091	96.08	132.02	2.0974	0.730505	0.000019	0.716358	7.62	39.65	0.116221	0.512045	0.000011	0.511684	-6.7
<i>Leucogranite</i>													
KS064A	84.38	229.61	1.0560	0.725600	0.000017	0.718477							
VFSJ50	132.15	107.63	3.5397	0.734196	0.001800	0.710321	3.44	16.55	0.125698	0.512134	0.000800	0.511744	-5.5
VFBR21	87.61	68.26	3.7021	0.738337	0.002000	0.713366	1.67	7.20	0.140114	0.512031	0.001200	0.511596	-8.4

ε_{Nd} values are calculated as deviations from a chondritic uniform reservoir in parts per 10⁴, using present-day values of ¹⁴³Nd/¹⁴⁴Nd = 0.512638 and ¹⁴⁷Sm/¹⁴⁴Nd = 0.1967 (Wasserburg *et al.*, 1981; Faure, 1986). Ages of rocks are from an average of the Famatinian plutonism (475 Ma) as estimated after Ducea *et al.* (2010).

Table 4: Pb isotope data

Sample	$^{206}\text{Pb}/^{204}\text{Pb}$	2 σ	$^{207}\text{Pb}/^{204}\text{Pb}$	2 σ	$^{208}\text{Pb}/^{204}\text{Pb}$	2 σ	$^{206}\text{Pb}/^{204}\text{Pb}$ (475 Ma)	$^{207}\text{Pb}/^{204}\text{Pb}$ (475 Ma)	$^{208}\text{Pb}/^{204}\text{Pb}$ (475 Ma)
<i>Mafic unit</i>									
KS021	18.510	0.0102	15.653	0.0107	38.432	0.0267	18.18	15.63	37.90
KS031B	18.842	0.0085	15.699	0.0101	39.147	0.0300	18.14	15.66	37.91
KS036	18.214	0.0102	15.635	0.0107	38.070	0.0267	18.16	15.63	38.01
KS042	18.225	0.0025	15.642	0.0024	38.102	0.0094	17.98	15.63	37.81
KS047	18.624	0.0102	15.731	0.0107	38.558	0.0267	18.23	15.71	38.20
KS065	18.459	0.0085	15.666	0.0107	38.379	0.0300	18.20	15.65	38.20
<i>Intermediate unit</i>									
KS075B	18.489	0.0102	15.675	0.0107	38.575	0.0267	18.22	15.66	38.10
KS072	18.357	0.0102	15.688	0.0107	39.952	0.0267	18.21	15.68	39.32
KS075A	18.338	0.0102	15.660	0.0107	38.214	0.0267	18.18	15.65	38.04
KS084	18.303	0.0102	15.640	0.0107	38.201	0.0267	18.13	15.63	38.02
KS093	18.267	0.0102	15.642	0.0107	38.112	0.0267	18.12	15.63	37.12
<i>Transitional unit</i>									
KS103B	18.827	0.0102	15.669	0.0107	38.768	0.0267	18.45	15.65	38.03
KS111	18.477	0.0102	15.657	0.0107	39.600	0.0267	18.34	15.65	38.73
KS103A	18.320	0.0102	15.655	0.0107	39.251	0.0267	18.24	15.65	39.20
<i>Silicic unit</i>									
KS128B	18.670	0.0102	15.676	0.0107	38.911	0.0267	18.38	15.66	38.72
KS117	18.478	0.0102	15.658	0.0107	39.092	0.0267	18.27	15.65	38.23
KS128A	19.822	0.0102	15.797	0.0107	39.374	0.0267	18.85	15.74	38.17
KS123	18.307	0.0102	15.646	0.0107	38.644	0.0267	18.24	15.64	37.99
KS116	18.276	0.0102	15.654	0.0107	38.273	0.0267	18.23	15.65	38.05
<i>Metasedimentary unit</i>									
KS045	18.289	0.0102	15.658	0.0107	39.855	0.0267	18.20	15.65	38.22
KS054	18.434	0.0102	15.672	0.0107	39.778	0.0267	18.17	15.66	37.89
KS064B	18.747	0.0102	15.701	0.0107	40.039	0.0267	18.23	15.67	37.85
KS091	18.784	0.0102	15.709	0.0107	39.392	0.0267	17.49	15.64	38.16
KS064A	18.254	0.0102	15.663	0.0107	38.170	0.0267	17.82	15.64	37.98

Initial Pb isotopic ratios are calculated at 475 Ma using trace element contents presented in Table 2.

Rocks with igneous plutonic textures are divided into two groups. The predominant group spans the full compositional range of the gabbro–diorite–tonalite–granodiorite–monzogranite series, which is the plutonic equivalent of the basalt–andesite–dacite–rhyolite volcanic suite. The volumetrically subordinate group consists of peraluminous felsic leucogranites that occur as *in situ* migmatite leucosomes (KS064A), para-autochthonous bodies intruded into metasedimentary migmatites (VFSJ50, VFBR21), and lens-shaped meter-scale veins or dikes in the transitional and silicic units (VFMJ40B, KS103A, VFMJ58, KS116, KS123).

Major and trace elements

Overall, the bulk-rock composition of the igneous plutonic suite varies continuously from 41 to 71 wt % SiO_2 .

Non-cumulate plutonic rocks within the Valle Fértil section define a typical sub-alkaline trend, with most rocks falling in the intermediate-K field (Fig. 4a). Within the compositional range of the non-cumulate plutonic rocks CaO , MgO , TiO_2 , and FeO^* decrease, whereas Na_2O and K_2O increase with increasing SiO_2 . However, although in the FeO^*/MgO vs SiO_2 diagram the data show considerable scatter, it is evident that the variation trend crosses the boundary between tholeiitic and calc-alkaline series (Fig. 4b).

At the low- SiO_2 ($\text{SiO}_2 < 48$ wt %) end of the spectrum, most of the rocks are igneous cumulates dominated by either olivine + pyroxene $\pm$ calcic amphibole or plagioclase-rich assemblages that plot very close to the tie line connecting olivine and plagioclase in a CaO vs MgO diagram (Fig. 4c). The cumulate rocks deviate from the main trend in Harker variation diagrams, because their compositions

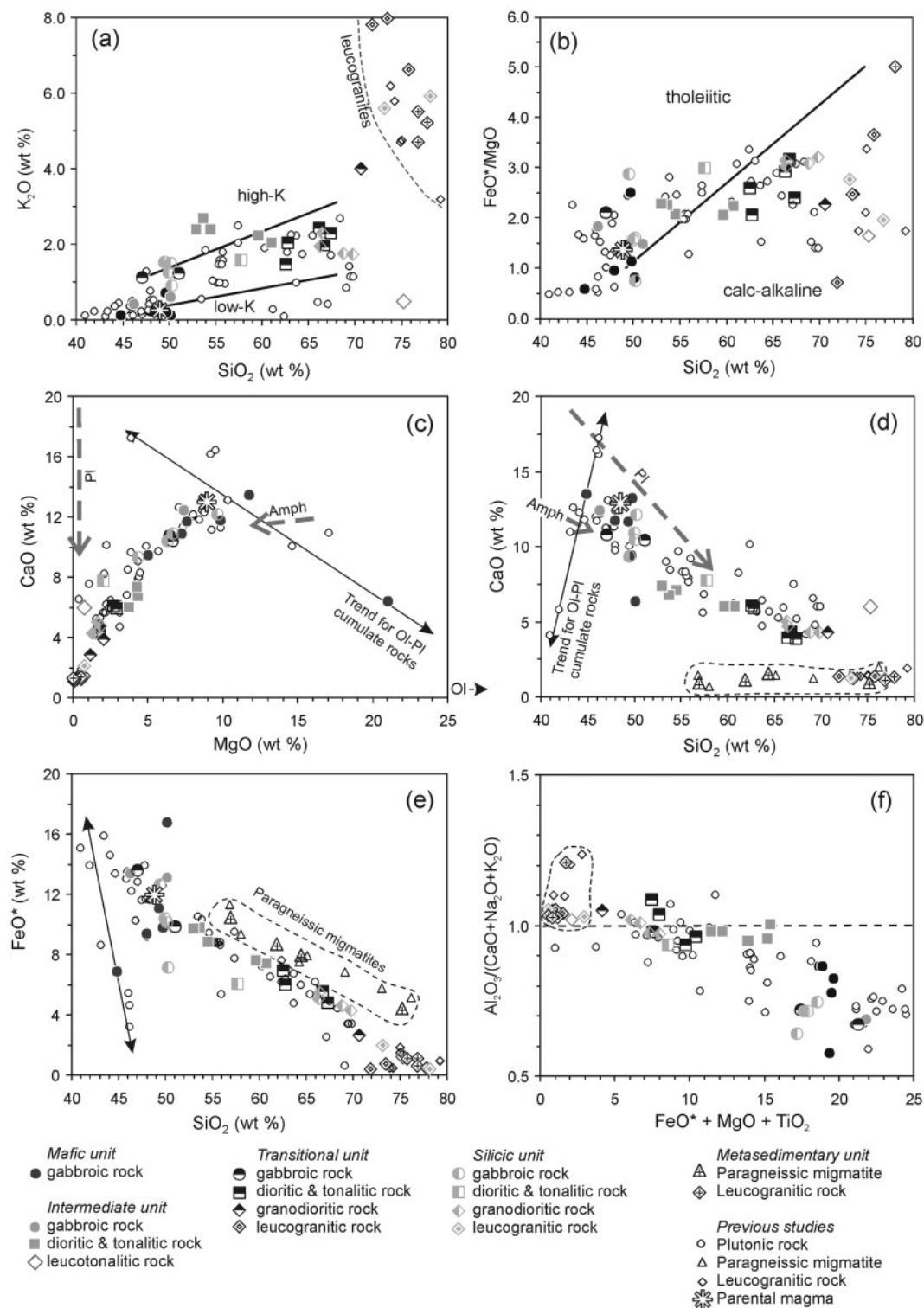


Fig. 4. Whole-rock major element abundances. (a) K_2O vs SiO_2 diagram with classification boundaries after Le Maitre *et al.* (1989). (b) FeO^*/MgO vs SiO_2 diagram showing the tholeiitic vs calc-alkaline boundary from Miyashiro (1974). (c) CaO vs MgO diagram. (d, e) CaO and FeO^* vs SiO_2 . (f) Co-variation of $Al_2O_3/(CaO+Na_2O+K_2O)$ on a molar basis vs $FeO^* + MgO + TiO_2$ in weight per cent. The thin continuous double-arrow lines reflect the compositional trends observed in mafic-ultramafic cumulate bodies (Otamendi *et al.*, 2010a). The thick dashed single-arrow lines show the compositional variation of plagioclase (Pl) and amphibole (Amph) in gabbroic to granodioritic rocks. Parental magma composition and data sources from previous studies are from Otamendi *et al.* (2009a, 2009b, 2010b).

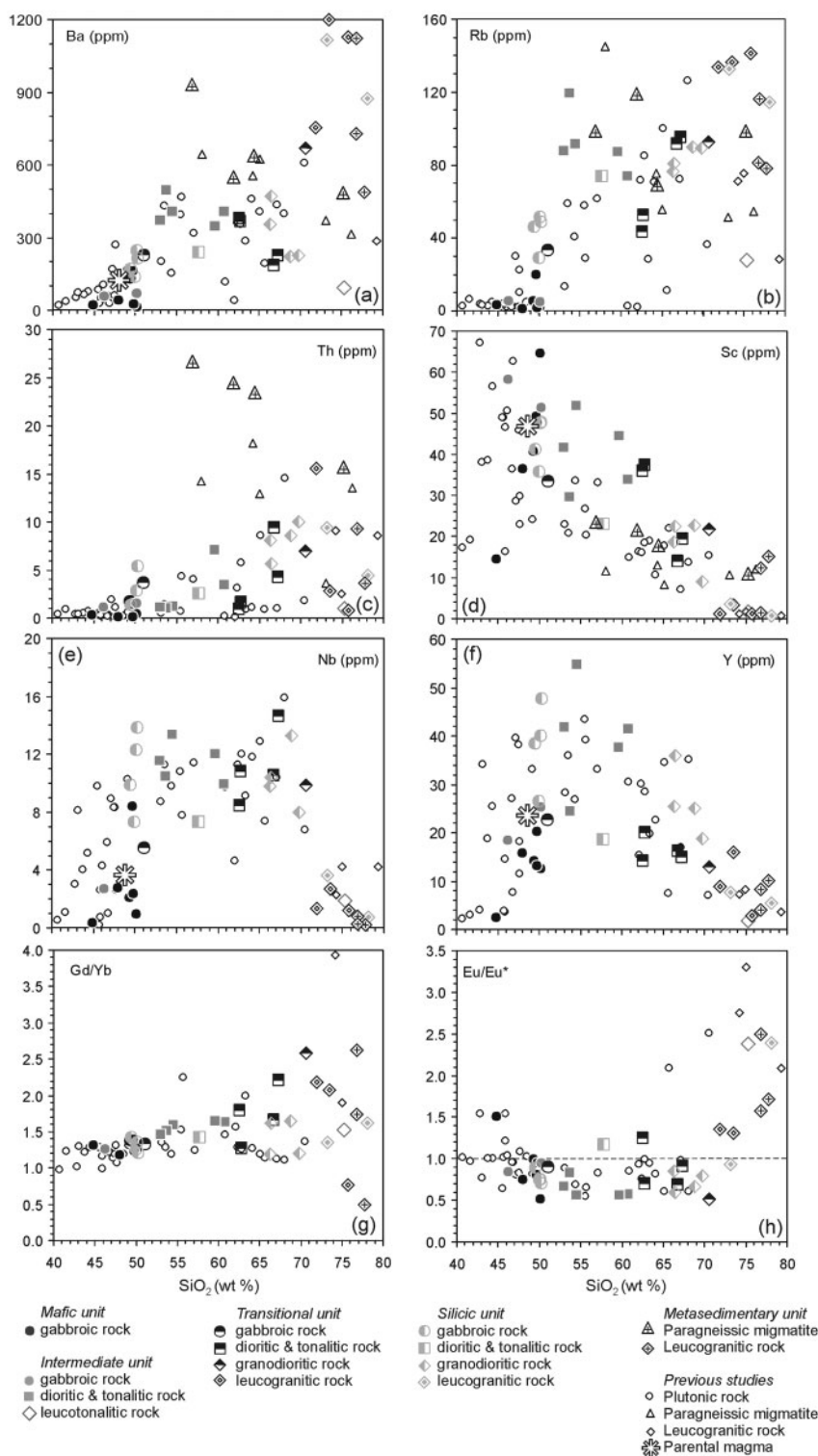


Fig. 5. (a–g) Variations of Ba, Rb, Th, Sc, Nb and Y (ppm) and Gd/Yb and Eu/Eu* vs SiO_2 in the Valle Fértil central section. In the Gd/Yb and Eu diagrams the abundance of the rare earth elements is normalized to CI chondrite (Anders & Grevesse, 1989). The abundance of trace elements in the parental magma was inferred by Otamendi *et al.* (2009a, 2010b).

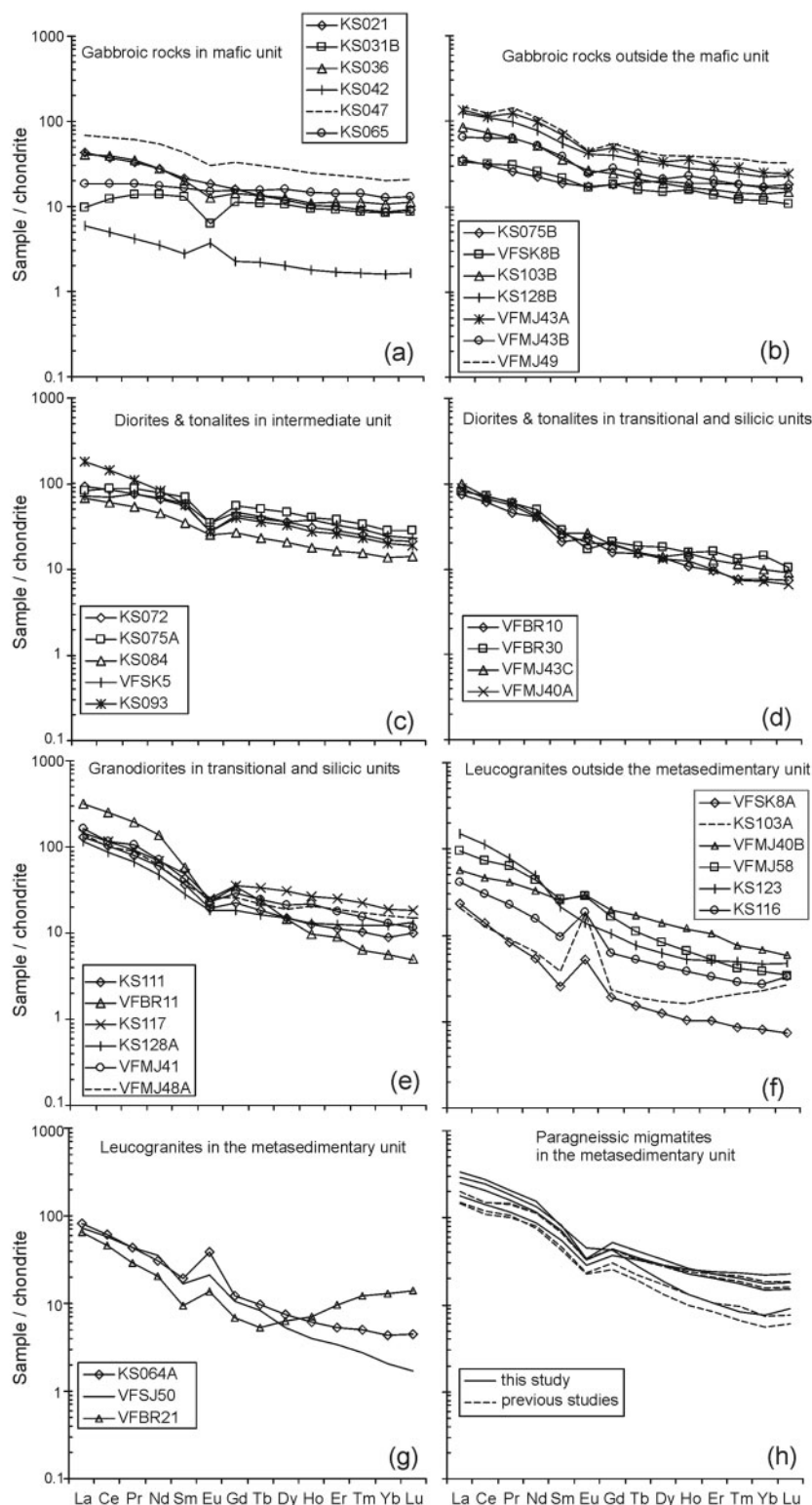


Fig. 6. Whole-rock REE patterns normalized to CI chondrite. (a, b) Gabbroic rocks from the mafic and the other lithostratigraphic units, respectively. (c) Dioritic and tonalitic rocks characteristic of the intermediate unit. (d) Dioritic and tonalitic rocks in the transitional and silicic units. (e) All the granodiorites collected for this study. (f) Leucogranites occurring as bodies or dikes within rock units dominated by plutonic rocks. (g) One leucogranite specimen separated from a stromatic migmatite leucosome (KS064A) and two leucogranites taken from large bodies in the metasedimentary packages. (h) All the paragneissic migmatites collected for this study; some representative specimens from Otamendi *et al.* (2009a) are distinguished by dashed lines.

are not a simple back-projection of well-defined trends of decreasing MgO (not shown), FeO* and CaO with increasing SiO₂ as defined by the plutonic suite (Fig. 4d and 4e). The parental magma composition for the plutonic calc-alkaline series is mafic (SiO₂ ~ 48–49 wt %), high-Mg (MgO ~ 8–9 wt %), moderately high-Al (Al₂O₃ ~ 15–16 wt %) and hydrous (Otamendi *et al.*, 2009a, 2010a). In several major oxide co-variation diagrams, the most plausible composition for the primitive parental magma falls at lower SiO₂ and higher MgO and CaO contents than the intersection between the non-cumulate and cumulate compositional trends (Fig. 4c–e). This feature of the major element data may reflect a change of the early crystallizing mineral assemblage that drives magma differentiation. Whereas the chemical diversity of the cumulate rocks can be caused by the mechanical separation between olivine + pyroxenes and plagioclase-rich assemblages, the paucity of olivine and clinopyroxene in the non-cumulate rocks requires removal of amphibole rather than olivine + pyroxenes to drive the compositional trend from mafic through intermediate to silicic rocks (Otamendi *et al.*, 2010a). Compared with the gabbroic rocks (SiO₂ < 52 wt %) from the mafic unit, the gabbroic rocks in the transitional and silicic units have higher SiO₂, K₂O and total alkalis contents.

In general, the trace element abundances in the rocks of each lithostratigraphic unit are variable but appear largely dependent on rock type. Gabbroic rocks from all of the rock units have higher abundances of compatible trace elements (Ni, Cr and Sc), but lower concentrations of Ba, Rb, Th, Cs and light rare earth elements (LREE: La–Sm) than the tonalitic and granodioritic rock types (Table 2, also see Fig. 5). Moreover, in Harker type variation diagrams trace elements display trends that are broadly consistent with their behavior in cogenetic igneous series. The co-variation of Rb and Ba vs SiO₂ mimics the compositional trend of K₂O vs SiO₂. However, even though Rb and Ba are incompatible with respect to the mineral assemblages of the mafic–ultramafic cumulates, the degree of incompatibility of Ba and Rb is not constant throughout the plutonic suite. The concentration of Ba and Rb in the diorites and tonalites is twice that of the gabbroic rocks, but similar to that of the granodiorites, clearly exhibiting a decrease of incompatibility with increasing SiO₂ (Fig. 5a and b). In contrast to Rb and Ba, the abundance of Sr does not exhibit a clear trend with increasing SiO₂ (not shown); however, the anomalously high Sr abundances in some gabbroic rocks clearly reflect the accumulation of plagioclase. The variation trend of Th with increasing SiO₂ characterized by scatter and enrichment is representative of other highly incompatible lithophile elements, such as Cs, U and Pb (Fig. 5c).

Because Ni and Cr are strongly sequestered by olivine and spinel during cumulate rock formation, Sc reflects the

changes of compatible elements from the gabbroic to the more evolved rocks better than Ni and Cr (Fig. 5d), essentially because Sc is incorporated into pyroxene, amphibole and magnetite with moderate to high partition coefficients (i.e. K_d values of 1–5; Dunn & Sen, 1994; Nielsen *et al.*, 1994). Within the scatter of data, the abundance of Sc progressively decreases from the gabbros through diorites and tonalites to granodiorites; most plutonic rocks have Sc abundances lower than the parent magma, reflecting the fractionation of amphibole ± orthopyroxene ± magnetite throughout the igneous evolutionary series.

The behavior of the high field strength elements (HFSE), such as Nb (Fig. 5e), Ta, Hf and Zr, resembles that observed for very large highly incompatible elements (i.e. Th and U) in several respects. However, one singular feature of the trace element variations in the plutonic series is the inflection of the Y vs SiO₂ trend at SiO₂ ~ 55 wt % (Fig. 5f). With some scatter of data, this inflection is also observed in a sub-group of heavy rare earth elements (HREE: Tb–Lu) and would cause middle REE (MREE) to HREE ratios (e.g. Gd_N/Yb_N in chondrite-normalized abundances) to increase with increasing SiO₂. However, Gd_N/Yb_N ratios progressively spread over a wider range of values with increasing SiO₂ (Fig. 5g). With the notable exception of plagioclase-cumulate mafic rocks and a few tonalites, all of the rest of the plutonic rocks show weak to strong Eu negative anomalies [Fig. 5h; Eu anomaly = Eu/Eu* where Eu* = $\sqrt{(\text{Sm} \times \text{Gd})}$], reflecting plagioclase involvement in their petrogenesis.

The primary feature of the REE data is that the shape of the chondrite-normalized (N) REE patterns systematically varies from nearly flat in the gabbros to very steep in the granodiorites (Fig. 6). One gabbroic sample (KS065) with a non-cumulate texture is distinguished by a nearly flat REE pattern and no Eu anomaly, making it a good candidate for constraining the primitive magma composition. Hence, considering KS065 as a reference, the other gabbroic rocks in the mafic unit appear to be either cumulates or slightly differentiated gabbroic rocks (Fig. 6a). Whereas an overall depleted REE pattern with a pronounced positive Eu anomaly is characteristic of plagioclase-rich cumulates (KS042), a two-pyroxene cumulate rock (KS031B) has La_N lower than Sm_N and a marked negative Eu anomaly. In contrast, differentiated gabbroic rocks have steeper REE patterns mainly because they have higher LREE contents than the most primitive gabbro (i.e. KS065).

Seven gabbroic rocks collected from the intermediate, transitional and silicic units display progressive REE enrichment ($60 < \sum \text{REE} < 200$) and fractionated REE patterns (La_N/Yb_N = 2–6). The REE patterns of these gabbroic rocks span the range between gabbros in the mafic unit and granodiorites in the silicic unit, perhaps reflecting LREE incorporation from the host intermediate

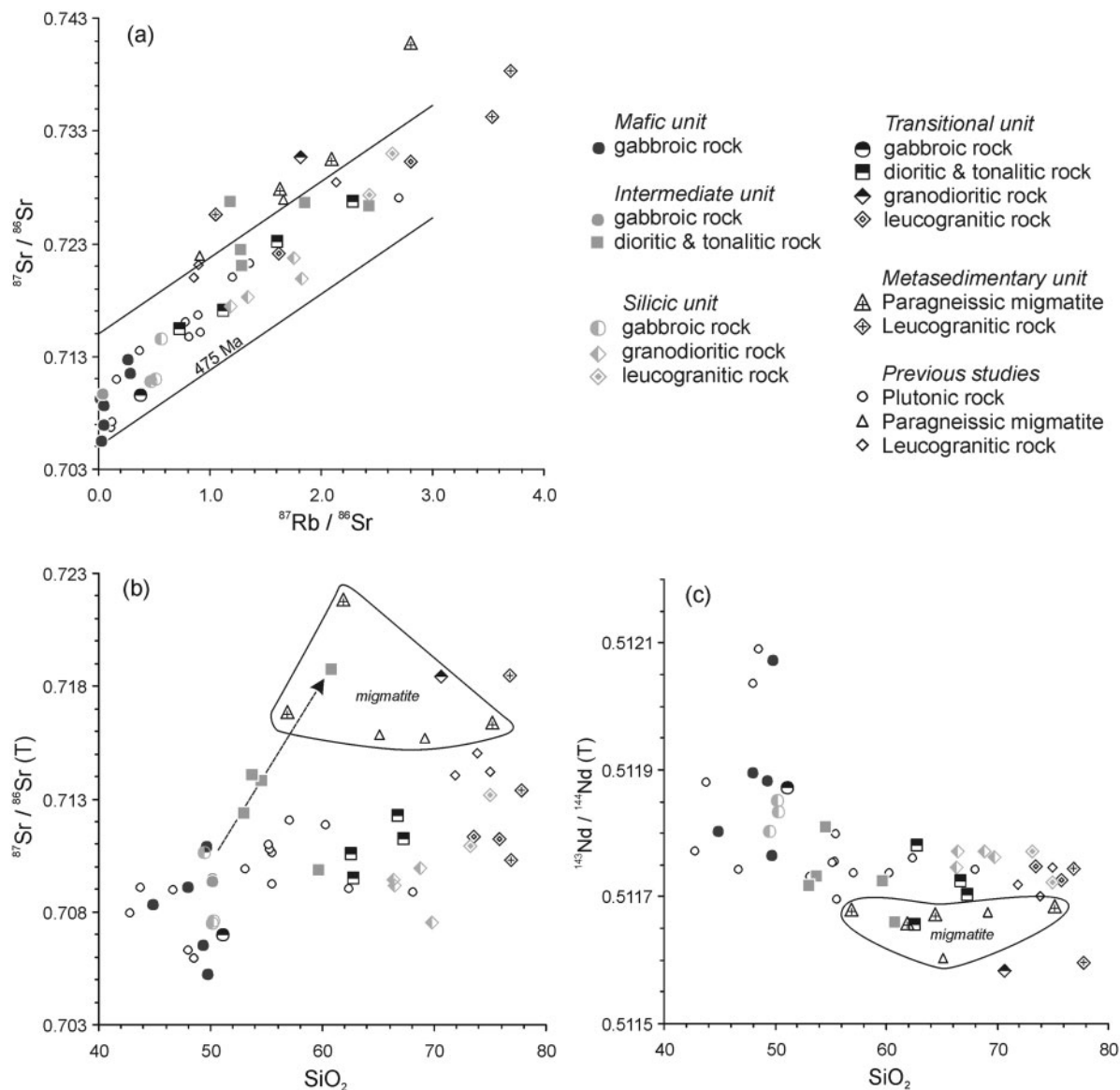


Fig. 7. (a) Variation of $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{87}\text{Rb}/^{86}\text{Sr}$ showing two Rb–Sr isochrons calculated at 475 Ma. One isochron was computed using an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.705 that is typical of the primitive gabbroic rocks, whereas the other was arbitrarily initiated at $^{87}\text{Sr}/^{86}\text{Sr} = 0.715$ to reflect the lowest possible value of a paragneissic migmatite. Most plutonic rocks fall between these two theoretical isochrons. (b, c) Variations of initial $^{87}\text{Sr}/^{86}\text{Sr}$ (b) and $^{143}\text{Nd}/^{144}\text{Nd}$ (c) vs SiO_2 . The arrow in (b) was arbitrarily drawn to highlight the compositional trend defined by the diorites and tonalites of the intermediate unit that have more radiogenic Sr isotope compositions than other petrographically similar rocks. In all the diagrams, Sr and Nd isotope compositions for the Valle Fértil crustal section are initial isotopic ratios recalculated at $t = 475$ Ma (e.g. Ducea *et al.*, 2010). The distinction of plutonic, paragneissic migmatites and leucogranites follows the same criteria as used in previous diagrams (see text for further explanation).

to silicic magmas into an REE budget originally controlled by cumulate formation (Fig. 6b).

Within the intermediate unit, most of the diorites and tonalites analyzed in this study span a limited range in both absolute and relative abundances of REE (Fig. 6c). The diorites and tonalites show variable fractionation patterns between LREE and HREE ($\text{La}_N/\text{Yb}_N = 2.9\text{--}9.0$) with pronounced negative Eu anomalies ($\text{Eu}/\text{Eu}^* \sim 0.6$).

One diorite (KS084) has lower REE contents and a less marked Eu anomaly than other petrographically similar rocks (Fig. 6c). Four tonalites collected from the transitional and silicic units have REE patterns similar to those of the tonalites that dominate the intermediate unit. However, tonalites in the transitional and silicic units are distinct from typical tonalites in the intermediate unit in that they lack Eu anomalies (Fig. 6c and d).

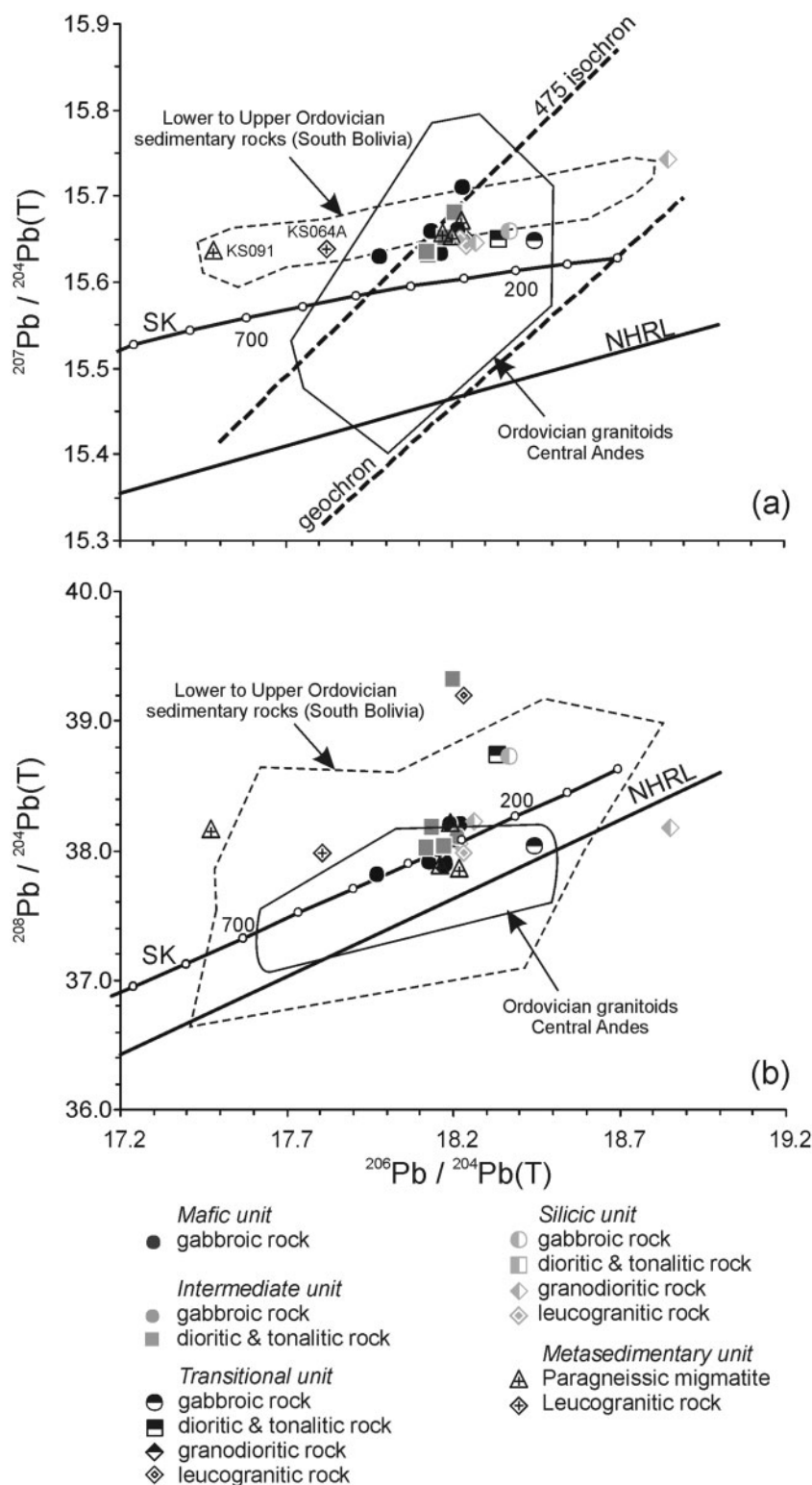


Fig. 8. Age-corrected ($t = 475$ Ma) Pb isotope ratios for the igneous and metasedimentary rocks from the central Valle Fértil section. (a) $^{207}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$, showing the Stacey & Kramers (1975) curve (SK) of common Pb growth for the average Earth, the Northern Hemisphere Reference Line (NHRL) after Hart (1984), and compositional fields for Ordovician granitoids from the central Andean basement (Lucassen *et al.*, 2002), and Lower to Upper Ordovician sedimentary rocks from southern Bolivia (Egenhoff & Lucassen, 2003). The isochron and geochron were calculated following the procedure of Stacey & Kramers (1975) and Faure & Mensing (2005). (b) Projection of the same rocks, reference lines and compositional fields as in (a) on the $^{208}\text{Pb}/^{204}\text{Pb}$ vs $^{206}\text{Pb}/^{204}\text{Pb}$ variation diagram.

All the granodiorites display smooth and straight steeply sloped REE patterns ($\text{La}_N/\text{Yb}_N = 8\text{--}13$) with variable negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.6\text{--}0.9$) (Fig. 6e). Notably, the granodiorites have broadly similar MREE to HREE abundances (measured as $\text{Gd}_N/\text{Yb}_N = 1.2\text{--}1.5$). One monzogranite (VFBR11) in the transitional unit has distinctly higher LREE and lower HREE contents, resulting in a more steeply sloped pattern ($\text{La}_N/\text{Yb}_N = 57$).

Whole-rock compositions of the leucogranites and paragneissic migmatites

Irrespective of location within the lithostratigraphic framework, leucogranites have a silica range restricted to values higher than 71.5 wt %, and molecular $\text{Al}_2\text{O}_3/(\text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})$ ratios higher than unity (Fig. 4f). Leucogranites have low concentrations of CaO, FeO^* , MgO and TiO_2 (Fig. 4c–e). Major element abundances are not smoothly correlated with silica content. Leucogranites plot on the high-silica side of the FeO^* , MgO, TiO_2 and CaO vs SiO_2 trends defined by the plutonic rocks spanning the range from gabbro to monzogranite (Fig. 4). Some leucogranites are K_2O -enriched ($\text{K}_2\text{O} > 6$ wt %; Fig. 4a) and have distinctively elevated Ba abundances commonly associated with marked positive Eu anomalies (Fig. 5a and h). Compared with the paragneissic migmatites, the leucogranites have generally higher SiO_2 and K_2O abundances, but lower MgO, FeO^* , TiO_2 and CaO contents.

In Harker variation diagrams the trace element contents of the leucogranites are scattered but broadly correspond to the high-silica extension of the global array displayed by the plutonic rocks (Fig. 5). In contrast to the coherent behavior shown by most of the trace elements, the leucogranites appear to be markedly depleted in Nb with respect to the concentration expected for a composition derived from the most evolved plutonic granodiorites (Fig. 5e).

Outside the metasedimentary unit, leucogranites show a wide variety of REE patterns (Fig. 6f). One leucogranite (KS123) from the silicic unit has a straight fractionated REE pattern characterized by high LREE contents and no Eu anomaly. This signature is considered typical of non-differentiated anatectic melts (Sawyer, 1987). Two other leucogranites (VFMJ40B and VFMJ58) display similar REE patterns with high LREE to HREE ratios ($\text{La}_N/\text{Yb}_N = 15\text{--}21$) and weak positive Eu anomalies ($\text{Eu}/\text{Eu}^* \sim 1.25$). In sharp contrast, the other three leucogranites have REE patterns characterized by marked positive Eu anomalies ($\text{Eu}/\text{Eu}^* = 2.4\text{--}6.1$) associated with a general depletion of the other REE (Fig. 6f). In the latter cases, the size of the Eu anomaly reflects an overall depletion of LREE, MREE and HREE rather than variability of the absolute Eu content.

All the leucogranites collected within the metasedimentary unit have REE patterns with pronounced positive Eu

anomalies (Fig. 6g). Evidently, the shape of the REE pattern of one leucogranite (VFBR21) is controlled by the melting of garnet in the source, which explains the strong enrichment in the HREE and the positive slope from MREE to HREE pattern ($\text{Dy}_N/\text{Yb}_N < 1$).

Paragneissic migmatites show very uniformly shaped chondrite-normalized REE patterns characterized by moderate fractionation ($\text{La}/\text{Yb} = 10\text{--}45$), well-defined negative Eu anomalies ($\text{Eu}/\text{Eu}^* = 0.51\text{--}0.77$), and HREE contents between seven and 22 times chondrite (Fig. 6h). These REE features are typical of siliciclastic sedimentary rocks; hence, the overall shape of the REE patterns of paragneissic migmatites is similar to that of post-Archean average shales (Taylor & McLennan, 1985). Most of the steepness of the REE patterns of the paragneissic migmatites reflects LREE (La/Sm) fractionation, because the patterns of REE heavier than Eu are nearly flat.

Radiogenic isotopes

Two groups of gabbroic rocks have been identified based on their Sr and Nd isotopic compositions. Some gabbros are distinctly grouped as they have initial Sr and Nd isotopic ratios similar to those of the Bulk Earth (Table 3). This group of gabbros, including KS065 from this study, combine low $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ (~ 0.7052) and relatively high $\epsilon_{\text{Nd}}(\text{T})$ of around +1, so that they are the most isotopically primitive rocks from the plutonic sequence. By contrast, most of the gabbroic rocks have $\epsilon_{\text{Nd}}(\text{T})$ values spanning a wide compositional range from -2 to -5 that crudely correlate with increasing $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ from 0.7065 to 0.7108 (as shown and discussed below). Therefore, many gabbroic rocks that have apparently primitive whole-rock compositions are unlikely to be derived from a single, uniform mantle source; moreover, the $\epsilon_{\text{Nd}}(\text{T})$ values of the most primitive gabbros indicate crustal contamination or isotopic input from an enriched lithospheric mantle source.

Considering all the rock units, the diorites and tonalites span a large Sr and Nd isotope range, with $0.7095 < ^{87}\text{Sr}/^{86}\text{Sr}(\text{T}) < 0.7140$ and $0.51166 < ^{143}\text{Nd}/^{144}\text{Nd}(\text{T}) < 0.51181$, whereas the granodiorites have a more restricted range of $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ from 0.7075 to 0.7100 and $^{143}\text{Nd}/^{144}\text{Nd}(\text{T})$ of about 0.51175. The isotopic composition of the granodiorites overlaps with that of the diorites, but is notably less radiogenic in Sr than that of the tonalites (Table 3).

Paragneissic migmatites and leucogranites derived from migmatites within the metasedimentary unit have the most radiogenic $^{87}\text{Sr}/^{86}\text{Sr}$ and unradiogenic $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic compositions, with $^{87}\text{Sr}/^{86}\text{Sr}(\text{T}) > 0.715$ and $\epsilon_{\text{Nd}}(\text{T}) < -6.0$ (see also Otamendi *et al.*, 2010b). Notably, the leucogranites collected from the transitional and silicic units have Sr isotopic compositions distinctly less radiogenic than those of the leucogranites from the metasedimentary unit.

Table 5: Stepwise crystal fractionation models calculated through least-squares multiple regressions

Model (1), gabbroic to diorite							
Component:	Daughter	Amph ²	Pl (An90)	Mt	Parent	Calculated ³	
Sample:	VFSK5 ¹				KS021		r^2
SiO ₂	53.96	46.26	44.15	0.00	49.96	49.60	0.13
Al ₂ O ₃	19.43	13.08	36.22	0.00	18.09	19.02	0.85
FeO*	9.85	10.77	0.09	99.00	11.21	11.21	0.00
MgO	4.46	15.87	0.00	0.00	7.70	8.62	0.85
CaO	7.57	11.94	18.45	0.00	11.79	10.81	0.96
Na ₂ O	2.52	1.45	1.05	0.00	0.98	1.89	0.82
K ₂ O	2.35	0.64	0.04	0.00	0.26	1.34	1.18
Coefficient	0.458	0.415	0.130	0.023		$\sum r^2 =$	4.80
Model (2), diorite to low-Si tonalite							
Component:	Daughter	Amph	Pl (An66)	Mt	Parent	Calculated	
Sample:	KS093				VFSK5		r^2
SiO ₂	60.40	46.31	51.61	0.00	53.96	54.23	0.07
Al ₂ O ₃	17.27	12.00	31.05	0.00	19.43	19.02	0.17
FeO*	7.71	14.90	0.07	99.00	9.85	9.85	0.00
MgO	3.79	12.79	0.00	0.00	4.46	3.93	0.28
CaO	6.03	11.63	13.48	0.00	7.57	8.17	0.37
Na ₂ O	2.57	1.22	3.75	0.00	2.25	2.56	0.00
K ₂ O	2.24	1.00	0.20	0.00	2.35	1.53	0.7
Coefficient	0.608	0.127	0.225	0.033		$\sum r^2 =$	1.27
Model (3), low-Si tonalite to normal tonalite							
Component:	Daughter	Amph	Pl(An62)	Mt	Parent	Calculated	
Sample:	VFBR30				KS093		r^2
SiO ₂	63.50	45.17	51.97	0.00	60.40	60.47	0.00
Al ₂ O ₃	16.61	12.89	30.76	0.00	17.27	16.87	0.16
FeO*	6.06	15.38	0.10	99.00	7.71	7.71	0.00
MgO	2.96	12.15	0.00	0.00	3.79	3.38	0.16
CaO	5.94	11.63	12.72	0.00	6.03	6.63	0.36
Na ₂ O	2.88	1.75	4.34	0.00	2.57	2.84	0.07
K ₂ O	2.04	1.03	0.10	0.00	2.24	1.83	0.17
Coefficient	0.856	0.070	0.057	0.015		$\sum r^2 =$	0.93
Model (4), tonalite to granodiorite							
Component:	Daughter	Amph	Pl(An53)	Bt	Parent	Calculated	
Sample:	KS117				VFBR30		r^2
SiO ₂	67.07	44.39	54.40	39.24	63.50	63.48	0.00
Al ₂ O ₃	15.53	12.51	29.20	18.82	16.61	16.60	0.00
FeO*	5.56	17.25	0.09	19.99	6.06	6.28	0.05
MgO	1.84	10.59	0.00	11.31	2.96	2.58	0.14
CaO	4.77	12.26	10.92	0.00	5.94	6.01	0.00
Na ₂ O	2.91	1.66	5.25	0.08	2.88	2.99	0.01
K ₂ O	2.33	1.33	0.14	10.56	2.04	2.11	0.00
Coefficient	0.801	0.093	0.095	0.010		$\sum r^2 =$	0.21

¹All compositions normalized to 100% anhydrous as required by multiple regression methods.²Mineral composition is after our published and unpublished data. Text provides details on sources of published data.³Best-fit calculated major element composition of assumed parent component. r^2 is sum of the square of the residuals.

*Total Fe given as FeO.

Table 6: *Mixing of crustal melts and residual mineral assemblages calculated through least-squares multiple regressions*

Model (5), mixing of crustal melt with residual mineral assemblage of amphibole gabbroic rocks							
Component:	Contaminant	Amph ²	Pl (An80)	Mt	Diorite ³	Calculated ⁴	
Sample:	KS103A ¹				VFSK5		r ²
SiO ₂	75.93	46.26	48.25	0.00	53.96	54.05	0.01
Al ₂ O ₃	12.98	13.08	33.32	0.00	19.43	18.79	0.40
FeO*	1.09	10.77	0.00	99.00	9.85	9.85	0.00
MgO	0.30	15.87	0.00	0.00	4.46	3.92	0.29
CaO	1.36	11.94	16.13	0.00	7.57	8.70	1.29
Na ₂ O	1.70	1.45	2.20	0.00	2.52	1.68	0.70
K ₂ O	6.64	0.64	0.10	0.00	2.35	2.54	0.04
Coefficient	0.355 ⁵	0.241	0.332	0.069		Σr ² =	2.73

Model (6), mixing of crustal melt with residual mineral assemblage of orthopyroxene gabbroic rocks							
Component:	Contaminant	Opx	Pl (An80)	Mt	Diorite	Calculated	
Sample:	KS103A				VFSK5		r ²
SiO ₂	75.93	52.44	48.25	0.00	53.96	54.08	0.01
Al ₂ O ₃	12.98	1.46	33.32	0.00	19.43	19.23	0.04
FeO*	1.09	24.54	0.00	99.00	9.85	9.86	0.00
MgO	0.30	21.11	0.00	0.00	4.46	4.27	0.04
CaO	1.36	0.45	16.13	0.00	7.57	7.86	0.09
Na ₂ O	1.70	0.02	2.20	0.00	2.52	1.49	1.05
K ₂ O	6.64	0.00	0.10	0.00	2.35	1.94	0.17
Coefficient	0.284	0.198	0.458	0.047		Σr ² =	1.39

¹All compositions normalized to 100% anhydrous with seven major oxides.

²Sources for mineral compositions are provided in text and Table 5.

³Composition of high-K diorite (VFSK5) modeled mixing crustal melt (KS103A) with residual mineral assemblages corresponding to gabbroic rocks.

⁴Best-fit calculated major element composition of assumed resultant component.

⁵Fraction of crustal melt in rock calculated by multiple regression model.

r² is sum of the square of the residuals.

The variations of measured $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{87}\text{Rb}/^{86}\text{Sr}$ exhibit a strongly covariant array associated with a systematic trend of isotopic enrichment that ranges from the gabbroic rocks to the leucogranites (Fig. 7a). Collectively, the isotope data suggest a trend toward more radiogenic initial Sr isotope ratios in compositionally more evolved rocks, as illustrated by the general array of increasing $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ with increasing SiO_2 . However, compositional trends with distinct slopes can be distinguished within the general arrays of $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ vs SiO_2 (Fig. 7b). Whereas typical diorites and tonalites from the intermediate unit show a strong increase in initial Sr isotope ratios

over a narrow SiO_2 range, most of the other plutonic rocks have Sr isotope ratios that vary between 0.708 and 0.713, but do not correlate with increasing SiO_2 abundance. The variation of $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ vs SiO_2 also shows that some tonalites are significantly more radiogenic than other plutonic rocks that are geochemically more evolved (Fig. 7b).

Except for the most primitive gabbroic rocks, there is no clear correlation between $^{143}\text{Nd}/^{144}\text{Nd}(\text{T})$ and SiO_2 (Fig. 7c). Gabbroic rocks with $\text{SiO}_2 < 50$ wt % have a wide range of Nd-isotope ratios from 0.5121 to 0.5117; however, with few exceptions, the gabbroic rocks have higher

$^{143}\text{Nd}/^{144}\text{Nd}(\text{T})$ than tonalites and granodiorites. At the other extreme of the SiO_2 spectrum, the leucogranites have relatively unradiogenic Nd isotope ratios between 0.5118 and 0.5117. The initial Nd isotope ratios for most diorites, tonalites and granodiorites lie within the same range as the leucogranitic rocks (Fig. 7c). Remarkably, the $^{143}\text{Nd}/^{144}\text{Nd}(\text{T})$ values of the paragneissic migmatites are the lowest and vary over a restricted range of 0.51167 ± 0.00002 .

In general, the plutonic rocks from the Early Ordovician Valle Fértil section appear to be characterized by depleted Pb isotopic compositions of age-corrected $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ combined with relatively enriched $^{207}\text{Pb}/^{204}\text{Pb}(\text{T})$ and $^{208}\text{Pb}/^{204}\text{Pb}(\text{T})$ values, most of them lying above the average crustal growth curve of Stacey & Kramers (1975). Early Ordovician plutonic rocks also appear to be displaced well above the Northern Hemisphere Reference Line (NHRL) in Pb isotope space (e.g. Hart, 1984).

In the $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ vs $^{207}\text{Pb}/^{204}\text{Pb}(\text{T})$ diagram (Fig. 8a), most rocks from the mafic and intermediate units lie close to and align roughly parallel to a 475 Ma reference isochron, thus reflecting a single-stage common Pb isotopic history. Also in this plot, nearly all the plutonic rocks define a small domain that occurs inside the compositional field of Ordovician granitoids from northwestern Argentina (e.g. Lucassen *et al.*, 2002).

An increase in the Pb isotopic ratios is observed from mafic to silicic rock units, spanning a $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ range from 17.98 to 18.85. In this trend one granodiorite (KS128A) is distinctively more radiogenic than the other plutonic rocks, an observation supported by the significantly higher U abundance in this granodiorite compared with the other plutonic rocks. The Pb isotope systematics for this granodiorite are possibly affected by the anomalously high concentration of one accessory U-bearing mineral of the epidote group (also see Table 1).

On the $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ vs $^{207}\text{Pb}/^{204}\text{Pb}(\text{T})$ plot (Fig. 8a), the paragneissic migmatites define a trend that perfectly overlaps the field of the Early–Middle Ordovician sedimentary sequence from southern Bolivia (e.g. Egenhoff & Lucassen, 2003). Like the Cambrian–Ordovician sedimentary sequences form the central Andean basement (i.e. northern Argentina, northeastern Chile and southern Bolivia), the paragneissic migmatites seem to be derived from Gondwanan sources that are distinctively enriched in $^{207}\text{Pb}/^{204}\text{Pb}$ at any given $^{206}\text{Pb}/^{204}\text{Pb}$ compared with Laurentian sources (see Kay *et al.*, 1996; Tosdal, 1996). One paragneissic migmatite (KS091) retains some common Pb signature from an older source characterized by relatively high $^{207}\text{Pb}/^{204}\text{Pb}(\text{T})$ and low $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ composition. A leucogranite (KS064A) from the metasedimentary unit also reflects, in part, Pb inheritance from an older source. In contrast, all the other paragneissic migmatites and leucogranites overlap the $^{207}\text{Pb}/^{204}\text{Pb}(\text{T})$ vs $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$

compositional field defined by most of the gabbroic and dioritic rocks (Fig. 8a).

In the $^{208}\text{Pb}/^{204}\text{Pb}(\text{T})$ vs $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ diagram, the plutonic rocks define a subvertical array scattering towards higher $^{208}\text{Pb}/^{204}\text{Pb}(\text{T})$ than the average crustal growth curve of Stacey & Kramers (1975) (Fig. 8b). However, the data for the plutonic rocks mostly plot within the broad compositional field of Ordovician granitoids from northwestern Argentina. Some gabbroic and tonalitic rocks extend towards higher $^{208}\text{Pb}/^{204}\text{Pb}$ values, displaying an isotopic variation that may result from the high Th/U ratios in these plutonic rocks.

Within the $^{208}\text{Pb}/^{204}\text{Pb}(\text{T})$ vs $^{206}\text{Pb}/^{204}\text{Pb}(\text{T})$ plot, the paragneissic migmatites and leucogranites define a large range of $^{208}\text{Pb}/^{204}\text{Pb}(\text{T})$ variation with a spread that encompasses the field defined by Ordovician sedimentary rocks from northern Argentina and southern Bolivia (Lucassen *et al.*, 2002; Egenhoff & Lucassen, 2003).

Mass-balance constraints on differentiation processes

Mass-balance models were performed using least-squares multiple regression and combine whole-rock major element chemistry (Table 2) with mineral composition data from previous studies (Otamendi *et al.*, 2009a, 2009b, 2010a). The mass-balance models constrain the extent to which closed-system fractionation and open-system mixing can drive the diversification of the plutonic rocks.

Table 5 shows a set of models (1–4) using first a primitive gabbroic rock (KS021) as the starting magma composition and then progressively more ‘evolved’ parent melts required to generate a typical granodiorite (i.e. KS117). Because the evolutionary model sequence requires a typical gabbro to be representative of the parental magma, the model cannot explain the origin of the gabbroic rocks, which are presumed to be derived from juvenile magmas (Otamendi *et al.*, 2010a). The most relevant outcomes of the mass-balance calculations are as follows: (1) the poor statistical fit of model 1 (as expressed by $\sum R^2 \gg 1$; e.g. Le Maitre, 1979) illustrates the impossibility of deriving typical diorites from any gabbroic parent magma by closed-system differentiation; (2) upon accumulation, the residual crystallizing assemblage for all the steps will yield different combinations of amphibole, plagioclase and magnetite residual assemblages; (3) if each step were to exhaust the respective parent, the end result would be residual or cumulate gabbroic rocks and granodioritic magmas amounting to a ratio of about 8:2.

As inferred from the poor fits of the major element balances, typical diorites are unlikely to be the closed-system differentiation derivatives from gabbroic parental magmas. The main limitation is the impossibility of crystal fractionation involving any observable mineral assemblage to account for the increase in K_2O (and Na_2O to a lesser extent) by the amount measured in parent–daughter pairs

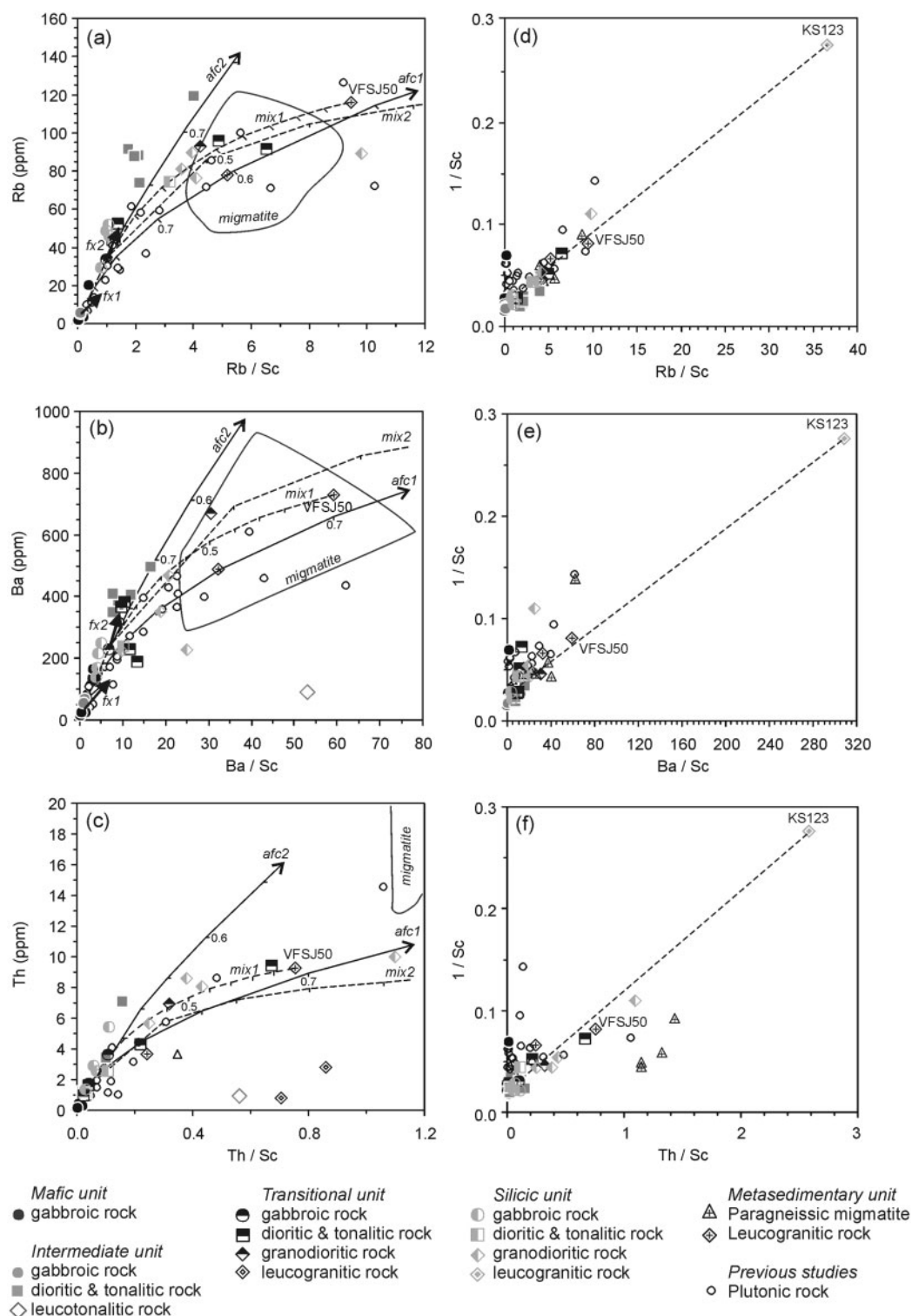


Fig. 9. (a–c) Variation of incompatible elements vs trace element ratios of selected incompatible elements to Sc for plutonic rocks, metasedimentary migmatites and leucogranites. Two simple fractional crystallization models (*fx*) illustrate the trace element changes driven by closed-system differentiation. One model (*fx1*) uses as starting composition a gabbroic composition (KS021) and the fractionating phase assemblage computed through mass balance (i.e. model 1 in Table 5). Another differentiation model (*fx2*) uses as parental magma a gabbroic specimen from the transitional unit (KS103B) and a fractionating assemblage of amphibole, plagioclase and magnetite (see text for details). Two model AFC and mixing curves are shown with a primitive gabbroic rock as the starting composition or juvenile mixing end-member. One curve (*mix1* and *afc1*) uses specimen (VFSJ50) from a leucogranitic body in the metasedimentary unit as the crustal component, whereas the other (*mix2* and *afc2*) uses a leucogranitic dike (KS123) from the silicic unit as the supracrustal contaminant. Tick marks on the *afc* curves represent the fraction of original magma remaining. Tick marks on mixing lines are at 0.1 end-member fractions unless omitted for clarity. (d–f) 1/Sc vs incompatible element/Sc ratios, which are companion plots to test the mixing hypotheses inferred from elemental variations following the method of Langmuir *et al.* (1978).

within the range of SiO_2 from gabbro to diorite. In fact, alkaline oxides plus lime contribute to a large extent (60%) to the divergence between measured and calculated daughter compositions (i.e. model 1 in Table 5). Adding olivine, clinopyroxene, and orthopyroxene as a single phase or combination of phases to the Amph–Pl–Mt three-phase fractionating assemblage gives unacceptable statistical fits.

Alternatively, differences in the measured and calculated composition of the diorites are probably due to variable amounts of mixing between the residual assemblage of the parental gabbroic magmas and leucogranitic crustal melts. Least-squares mass balance involving metasedimentary-derived melts and residual phases from crystallization of the gabbroic magmas provide a test for the role of mixing in the generation of the diorites (also see Feeley *et al.*, 2002). Table 6 illustrates representative calculations involving leucogranite as the crustal melt and an Amph–Pl–Mt residual assemblage. The results are statistically improved with respect to crystal fractionation and support the possibility that diorite compositions can be produced by addition of leucogranite melt to a residual gabbroic assemblage. Inclusion of orthopyroxene as a residual phase in the gabbroic rocks gives an even better fit; however, in contrast to our petrographic observations it also requires the absence of amphibole in the residual assemblage (Table 6). The problem of how to achieve a mass-balance model perfectly matching the geochemistry of the diorites may reflect variable degrees of mixing between gabbroic residual assemblages and crustal melts following limited fractional crystallization of mafic magmas.

Closed-system crystal fractionation as the sole process driving lithological diversification also creates a major mass-balance problem. Starting with a mafic ($\text{SiO}_2 \sim 49$ wt %) parental magma, accumulating a residual assemblage to construct the diorites upon mixing with crustal melts, and then differentiating by closed-system fractional crystallization, the maximum amount of silicic magma that may be produced is about 20% of the original mafic precursor (Tables 5 and 6). This raises the question: where is the huge volume of gabbroic cumulates formed as a by-product of closed-system differentiation? A large volume of unseen residual cumulates, which must have extended about 60 km into the lower crust and uppermost lithospheric mantle, is required to generate a minimum of 15 km thick silicic upper crust. The presence of chemically depleted hornblende-dominated gabbroic rocks throughout the intermediate and silicic plutonic section may indicate that mafic cumulates also formed at lower and middle crustal levels (Fig. 2, Tables 1 and 2). However, assuming that arc crust uniquely forms from differentiation of mantle-derived melts may result in an unrealistically high volume of residual mafic rocks. If crustal rocks contributed to the production of the intermediate and silicic plutonic rocks, the amount of residual cumulates representing the

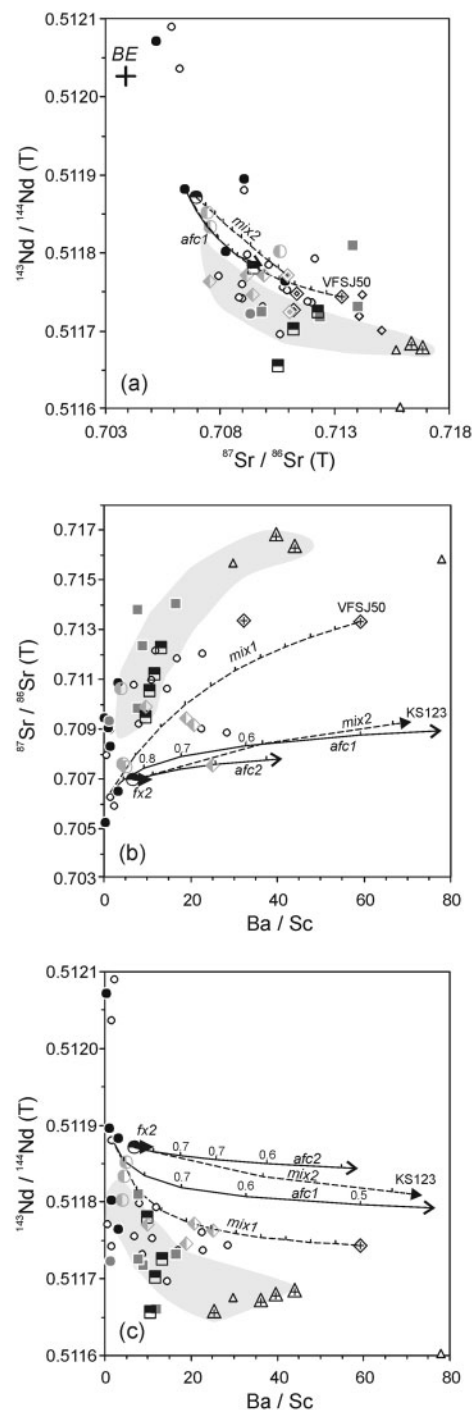


Fig. 10. (a) Variation of $^{143}\text{Nd}/^{144}\text{Nd}$ vs $^{87}\text{Sr}/^{86}\text{Sr}$ (at 475 Ma) for plutonic rocks, metasedimentary migmatites and leucogranites. BE denotes the Bulk Earth composition at 475 Ma after Faure (1986). (b) $^{87}\text{Sr}/^{86}\text{Sr}$ at 475 Ma vs Ba/Sc for the same lithologies as (a). (c) $^{143}\text{Nd}/^{144}\text{Nd}$ at 475 Ma vs Ba/Sc. Assimilation–fractional crystallization (afc) and mixing (mix) modelling was performed following the same rationale as described in Fig. 9. In these diagrams the shaded field is arbitrarily drawn to show that many dioritic and tonalitic rocks plot within a hyperbolic band between plutonic mafic and metasedimentary end-members. These intermediate plutonic rocks cannot be replicated by the computed afc curves or bulk mixing lines.

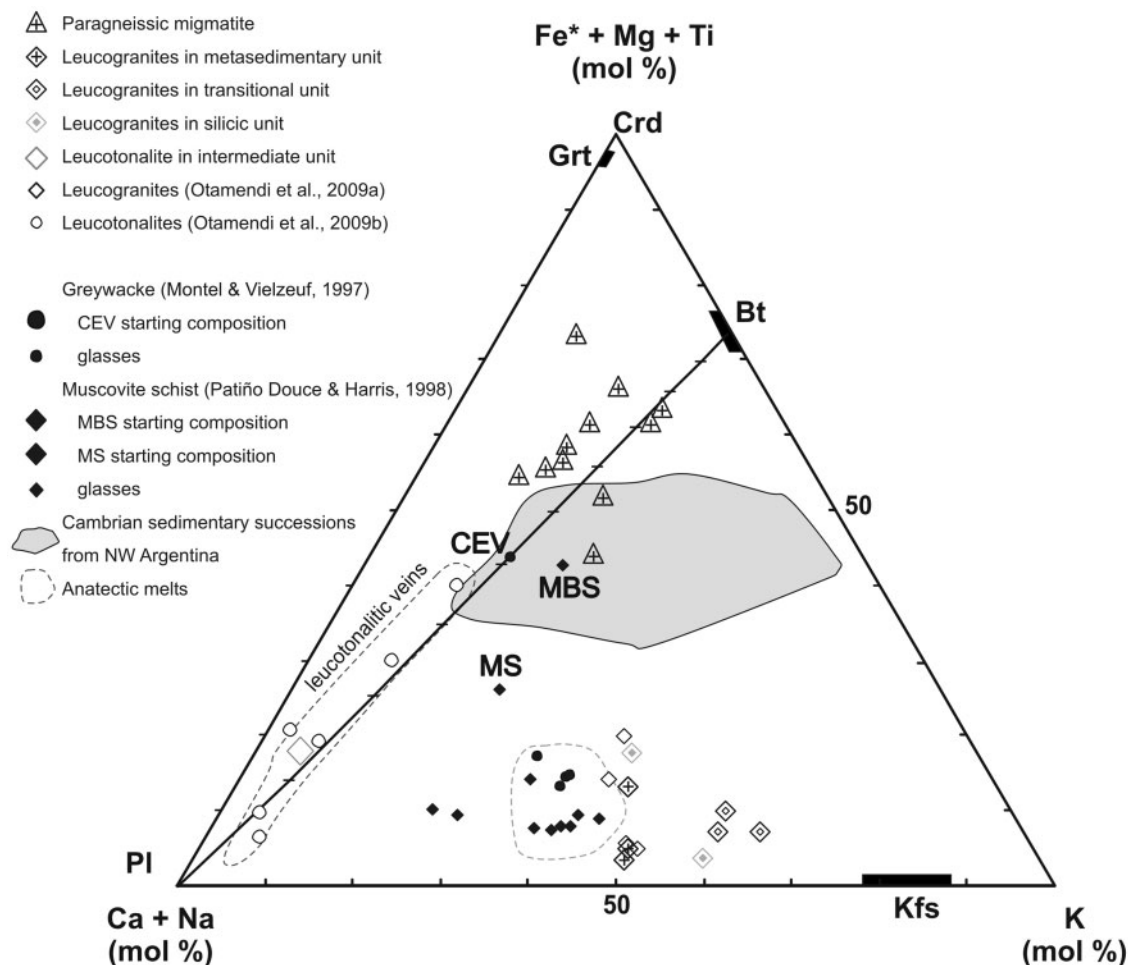


Fig. 11. $(\text{Fe}^* + \text{Mg} + \text{Ti})$ – $(\text{Ca} + \text{Na})$ – K ternary diagram (after Solar & Brown, 2001) for leucogranites and granulite-facies metasedimentary migmatites. For comparison, also plotted are the compositions of starting materials and glasses equilibrated at middle crustal pressures (5–8 kbar) from the experimental studies of Montel & Vielzeuf (1997) and Patiño Douce & Harris (1998). The composition of Cambrian sedimentary successions from northwestern Argentina, which are the most plausible protoliths to the Early Ordovician metasedimentary sequences, is taken from Zimmerman & Bahlburg (2003) and Zimmerman (2005). Leucotonalitic veins in the mafic migmatites from the same section are from Otamendi *et al.* (2009b).

complement to the tonalites and granodiorites would be much lower than that inferred from closed-system mass balance.

Discrimination of differentiation processes through trace element and radiogenic isotope systematics

To quantify the processes potentially responsible for generating the dioritic, tonalitic and granodioritic rocks of the intermediate and silicic units, we evaluated several differentiation models using trace element systematics and often-used equations [for a general presentation see Rollinson (1993)].

The enrichment of an incompatible trace element in a derivative melt relative to its concentration in the parental

melt (C_i/C_o) during crystal fractionation increases exponentially with decreasing residual melt fraction (f). This behavior is valid for either equilibrium or Rayleigh fractionation models, and at a realistic remaining melt fraction ($f \geq 0.4$; i.e. up to 60% crystallization; see Table 5) the difference in the estimated values between these two models is irrelevant when compared with the range of values that result from using the full spectrum of published mineral/melt partition coefficients (Rollinson, 1993). Even considering extremely low bulk partition coefficients ($D \sim 0.01$), which govern the behavior of strongly incompatible trace elements, the maximum enrichment of C_i/C_o would be of about 2.5 at $f = 0.4$. This simple argument may be used to infer that dioritic or tonalitic rocks in the intermediate unit cannot be derived from typical gabbros in the mafic

unit, because the highly incompatible lithophile elements (K, Ba and Rb) in the latter rocks are on average 4–15 times lower than in the former (Figs 4a and 5a, b). An alternative explanation is to consider crystal fractionation at very low fractions of residual melts ($f \ll 0.4$); however, this would conflict with the variations of elements such as Sc, Sr and Eu. Although Sc, Sr and Eu are moderately compatible in the gabbroic mineral assemblage, they are not depleted in the diorites and tonalites relative to the gabbroic rocks, as one would expect if fractionation occurred at high degrees of crystallization and the gabbroic rocks were the sole parents to the diorites or tonalites.

To illustrate these arguments, the arrows labeled fx using either gabbroic ($fx1$) or dioritic ($fx2$) starting compositions in Fig. 9a and b show calculated Rayleigh fractional crystallization trends. The models use early crystallizing mineral assemblages computed through mass balance, which are 41% hornblende, 13% plagioclase, 2% magnetite, and 46% residual melt for $fx1$; and 13% hornblende, 23% plagioclase and 3% magnetite, and 61% residual melt for $fx2$. These simple calculations graphically illustrate the incapacity of closed-system differentiation to replicate the trace element variation, therefore reflecting the need for considering more complex, multi-source processes.

The method of DePaolo (1981) is used as a first-order approximation to model the effects of an assimilation–fractional crystallization process (AFC). We compute AFC models using the same starting composition and fractionating assemblages as discussed above, which are combined with distinct crustal contaminants. Whereas one AFC model (curve $afc1$) assumes that the leucogranites (VFSJ50) from the metasedimentary unit are the crustal contaminant assimilated during the early stages of fractional crystallization (model $fx1$), another model (curve $afc2$) considers that leucogranites collected from the intermediate and silicic units (KSI23) are the most likely contaminants during crystal fractionation in the latter units (model $fx2$). We prefer to use leucogranites as the assimilate rather than metasedimentary migmatites for two reasons. Field evidence shows that migmatites are partly melted before being assimilated into the evolving magmas (Fig. 3h), and as computed the AFC models are easily compared with equivalent mixing ('mix') models (Fig. 9a–c). A conservative r value (ratio of rate of assimilation to crystallization) of 0.5 is used in all the AFC models because, even though assimilation at deep crustal depths may be more effective in terms of rates of assimilation to crystallization (e.g. Reiners *et al.*, 1995), the computed AFC curves have a similar projection for r between 0.5 and 0.8.

The modeled AFC curves are successful in accounting for the relative variation of incompatible (Ba, Rb and Th) and compatible (Sc) trace elements. The AFC modeling roughly replicates both the spread of data variation and the hyperbolic data array of the dioritic to granodioritic

rocks. The most important difference between the two assimilation models is that the incompatible elements (Rb, Ba, and Th) increase more rapidly in model $afc2$, mainly because the trace element ratios Rb/Sc, Ba/Sc and Th/Sc are higher in the leucogranites from the silicic units than in those leucogranites interlayered in the metasedimentary unit. This creates a controversial observation, for the reason that the $afc2$ model better replicates the composition of the diorites and tonalites from the intermediate unit; however, the input compositions for modeling $afc2$ are taken from rocks in the more evolved transitional and silicic units.

The shape of the data array in element-ratio variation diagrams might also be interpreted in terms of two-component mixing hyperbolae (e.g. Langmuir *et al.*, 1978). For comparison of mixing hypotheses with the AFC models, we computed two possible mixing arrays using as end-members the same compositions used as starting parental magma and contaminant in the AFC models. The relevant result is that a clear distinction between AFC and mixing cannot be made, making it difficult to assess the contribution of mixing of contrasting magmas relative to crystal-fractionation coupled with assimilation. However, Fig. 9a–c illustrates that the model results best satisfy the compositional data for tonalitic and granodioritic rocks when either in AFC models the degree of crystallization is lower than 50% or in the mixing models the fraction of the gabbroic end-member is higher than 50%. Regardless of which process dominated differentiation, the modeled results of the two alternative processes imply that more than one-half of the dioritic, tonalitic and granodioritic rocks is made up of the gabbroic component.

Figure 9d–f shows companion diagrams in which the values of $1/Sc$ are plotted against ratios of incompatible element to Sc; in these diagrams the mixing relationships are sustained by the linear variation of the data (Langmuir *et al.*, 1978; Feeley *et al.*, 2002; Schiano *et al.*, 2010). The compositions of the diorites, tonalites and granodiorites broadly fall on straight lines for which gabbroic rocks and leucogranites always appear as the end-member compositions. Consistent with elemental modeling, the diorites, tonalites and granodiorites lie closer to the gabbroic end-member than to the leucogranites, therefore reflecting the preponderance of the juvenile component in most of them.

To gain further insights into the nature of the crustal contaminants and to discriminate between AFC and mixing processes, additional modeling that combines Sr and Nd isotopic ratios with trace elements was performed. Because it is not easy to quantify self-diffusion isotopic exchange among reservoirs that had contrasting compositions before being physically connected by magmatic processes, modeling that includes isotopes is a gross oversimplification (Leshner, 1990). This is particularly

problematic as the isotopic compositions of leucogranites mingled with tonalites and granodiorites are different from those of leucogranites interlayered with metasedimentary migmatites, in that the latter are more radiogenic in Sr and less radiogenic in Nd than the former (Fig. 7), implying that the radiogenic isotope ratios of the leucogranites from the intermediate and silicic units may have changed as a result of isotopic exchange with coexisting plutonic rocks (Leshner, 1994).

The results of the most representative calculations with Sr- and Nd-isotope ratios are shown in Fig. 10, in which model curves were computed using the same conceptual framework and input parameters as those used for the contrasting model curves in Fig. 9. These models cannot produce the wide range of $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ values observed for the dioritic, tonalitic and granodioritic rocks. The difference is, in part, due to the fact that many intermediate and silicic rocks plot on a diffuse trend that lies at lower $^{143}\text{Nd}/^{144}\text{Nd}$ than that predicted by the model (Fig. 10a). To reduce the discrepancy, a different crustal contaminant is required for either AFC or the mixing curves to pass through the data. As illustrated by the shaded field in Fig. 10, Sr- and Nd-isotope ratios suggest that the bulk migmatites acted as either the contaminant in the AFC process or the crustal end-member for mixing models. The covariation of isotope ratios with Ba/Sc shows that no single model can reproduce the full compositional spectrum of the intermediate and silicic plutonic rocks (Fig. 10b and c). Because open-system processes make Ba/Sc ratios increase more rapidly compared with closed-system differentiation, the combination of isotopic ratios and trace element abundances suggests some sort of open-system process, because this variation trend cannot be produced by closed-system crystal fractionation or partial melting processes.

DISCUSSION

Origin of leucogranites

Leucogranites have SiO_2 contents higher than 71 wt % and K_2O -enriched compositions compared with the other plutonic rocks of the calc-alkaline series in the Valle Fértil section (Fig. 4). The leucogranites are also weakly to strongly peraluminous and strongly felsic with low abundance of transition metals (measured as $\text{TiO}_2 + \text{FeO}^* + \text{MgO}$) (Fig. 4f), reflecting a major element composition similar to that of melts experimentally generated from pelitic and greywacke starting materials (Vielzeuf & Holloway, 1988; Patiño Douce & Johnston, 1991; Montel & Vielzeuf, 1997). Geochronological data support the hypothesis that the leucogranites were derived by anatexis of metasedimentary migmatites, because the U/Pb zircon crystallization age of 472.7 ± 10 Ma for one leucogranite in the transitional unit coincides with the age of high-grade metamorphism of the migmatites (Ducea *et al.*, 2010).

The Valle Fértil section contains both leucogranitic leucosomes that show in-source relations with metasedimentary rocks and leucogranitic veins and bodies that are not attached to their source. Thus, the central question is whether the leucosomes and leucogranites in the metasedimentary unit are the in-source counterpart to the leucogranites in the other units.

The best preserved sedimentary protoliths to the granulite-facies migmatites of the Valle Fértil section are found in Late Neoproterozoic to Late Cambrian turbiditic sedimentary successions that are exposed over large regions of northwestern Argentina (Bahlburg, 1998; Zimmermann & Bahlburg, 2003; Zimmermann, 2005). The starting materials used in the experimental studies of Montel & Vielzeuf (1997) and Patiño Douce & Harris (1998) lie within the compositional field of early Paleozoic turbidites from northwestern Argentina (Fig. 11), and provide constraints on the compositional changes experienced by partially melted migmatites.

The fact that the migmatites of the Valle Fértil section have lost granitic melt is clearly appreciated by their displacement toward the $\text{Ti} + \text{Fe}^* + \text{Mg}$ apex in the ternary plot $(\text{Ti} + \text{Fe}^* + \text{Mg}) - (\text{Ca} + \text{Na}) - \text{K}$ with respect to the sedimentary and very low grade turbiditic successions from northwestern Argentina (e.g. Solar & Brown, 2001).

The composition of the leucosomes and leucogranites in the metasedimentary migmatites is similar to that of glasses produced by biotite dehydration melting from pelite and greywacke starting materials (Fig. 11), suggesting that some of the leucogranites may be near-primary anatectic melts. Most leucogranites, however, are enriched in K with respect to experimentally produced melts (Fig. 11). Potassium enrichment in the leucogranites may be attributed to K-feldspar accumulation upon cooling and movement (Sawyer, 1987). An alternative explanation for K enrichment in the leucogranites is the generation of K-feldspar as a peritectic product of the incongruent melting reaction (e.g. Carrington & Watt, 1995), and subsequent entrainment of part of this K-feldspar that remains suspended in the highly viscous anatectic magma during extraction. Identifying the actual process driving K enrichment may not be feasible, as the two alternative processes explain the strong enrichment in Ba and the positive Eu anomaly and would display similar petrographic features.

Crystallization of peritectic garnet gives rise to HREE depletion in leucogranites. Although the growth and shielding of zircon in the mesosomes during partial melting may contribute to HREE depletion in the anatectic magmas (Watt & Harley, 1993), the strong depletion of HREE associated with high Gd/Yb ratios is very probably governed by garnet rather than zircon. A few leucogranites with low LREE abundance ($\text{La}_N < 50$) may reflect crystallization of monazite, which would result from

apatite-bearing precursors at the initial stages of partial melting (e.g. Wolf & London, 1995). The phase changes that accompany partial melting of metasedimentary protoliths explain the REE patterns, which are typical of anatectic products. As first noted by Saywer (1987) for anatectic magmas, the largest positive Eu anomalies appear in those specimens with the lowest total REE contents, whereas differentiated samples with the highest REE contents have weak to marked negative Eu anomalies (Fig. 6f and g). Noticeably, most leucogranites in all the rock units share these compositional characteristics, which is the strongest supporting evidence for the field observation that all the leucogranites are crustal magmas derived from metasedimentary sources.

A singular aspect is that the $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ and $^{143}\text{Nd}/^{144}\text{Nd}(\text{T})$ isotopic ratios for the great majority of the leucogranites do not lie within the isotopic field of the migmatites from the metasedimentary unit (Fig. 7). This observation contradicts field, petrographic and geochemical data, which consistently suggest that the leucogranites were derived from a sedimentary protolith with the migmatites representing the residual complement. Although a few leucogranitic veins or bodies display Sr- and Nd-isotopic ratios similar to those of the migmatites, many other leucogranites have lower $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ and higher $^{143}\text{Nd}/^{144}\text{Nd}(\text{T})$ ratios than the metasedimentary migmatites (Fig. 7c and d). These isotopic differences, although minor, are somewhat systematic and may reflect either distinct initial isotopic compositions inherited from heterogeneities in the metasedimentary protoliths or, as discussed above, isotopic exchange with contrasting plutonic rocks, or both.

Origin of diorites, tonalites and granodiorites

The magmas from which the gabbros crystallized had already inherited supracrustal isotopic signatures while maintaining their mafic whole-rock chemistry. This interpretation is based on the high initial $^{87}\text{Sr}/^{86}\text{Sr}(\text{T}) > 0.706$ and low initial $\epsilon_{\text{Nd}}(\text{T}) < -2$ values of the gabbroic rocks, with some of them being as low as an $\epsilon_{\text{Nd}}(\text{T})$ of -5 (e.g. Taylor, 1980; DePaolo, 1981). Crustal isotopic signatures in gabbroic rocks make the relative isotopic contribution of mantle and crustal sources to the intermediate plutonic rocks less distinct (Fig. 7).

An explanation for the origin of dioritic and tonalitic magmas in the Famatinian arc has previously been proposed (Otamendi *et al.*, 2009b). Essentially, because the model for the origin of these intermediate magmas was mainly based on observations of small outcrops, our conclusion that the dioritic and tonalitic magmas were generated by the interaction between gabbroic magmas and melts sourced from sedimentary packages did not preclude the possibility that some of the intermediate magmas formed by either crystal fractionation or partial melting of

gabbroic materials. Covering a much larger scale, our present study shows that the diorites and low- SiO_2 tonalites ($52 < \text{SiO}_2 < 62$ wt %) of the intermediate unit result from the incorporation of (supra)crustal sedimentary lithologies with high $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ ratios into gabbroic rocks and/or magmas; however, based only on geochemical data, it is difficult to ascertain whether the gabbroic materials assimilated bulk metasedimentary rocks or reacted with anatectic leucogranitic melts. Notably, a distinction can be made among plutonic rocks of the intermediate unit, as there is a group of diorites and low- SiO_2 tonalites that depart from the main compositional variation trend (Fig. 4a). Some diorites and tonalites are distinct because their elemental and isotopic compositions differ from those of the rest of the petrographically similar rocks. The diorites and low- SiO_2 tonalites in question lie in the high-K field of the K_2O vs SiO_2 diagram and are isotopically distinct compared with other plutonic rocks that are chemically more evolved (Fig. 7). The simplest interpretation of the covariation between $\epsilon_{\text{Nd}}(\text{T})$ and $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ displayed by the high- K_2O diorites and tonalites is that they contain contributions from two end-member components, one characterized by high $\epsilon_{\text{Nd}}(\text{T})$ and low $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ and the other by low $\epsilon_{\text{Nd}}(\text{T})$ and high $^{87}\text{Sr}/^{86}\text{Sr}$. The Sr and Nd isotope composition of these diorites leaves little doubt about their hybrid nature. However, because these diorites fall well outside the main compositional trend in several critical chemical parameters, they cannot represent the intermediate state between parental gabbroic rocks and derivative intermediate plutonic rocks with normal K_2O abundances.

To some extent, the diorites and tonalites falling in the medium field of the K_2O vs SiO_2 diagram can be derived by magmatic differentiation, mainly by hornblende + plagioclase $\pm$ Fe-Ti oxide fractional crystallization, from mafic (gabbroic) igneous precursors (Table 5). However, the least-squares fractionation models calculated also require that differentiation operates in several discrete steps. Even progressing as a stepwise process, the cumulative result of simple fractional crystallization is inadequate to raise the concentration of incompatible trace elements in the gabbroic parental magmas to those of the derivative tonalitic magmas; to account for the minimum fourfold increase in the concentration of K, Rb, Ba and Th more than 70% of aggregate crystal fractionation is needed (Figs 5 and 9). In such cases, the advance of differentiation would be too large to be consistent with the variation of moderately compatible trace elements (e.g. Sc and Sr).

As inferred from mass-balance calculations, trace element systematics and isotopic ratios, if crystal fractionation in mafic (gabbroic) magma chambers contributed to the formation of typical tonalitic magmas, this must have been accompanied by crustal assimilation or mixing with crustal melts (Table 6; Figs 9 and 10). The preponderance

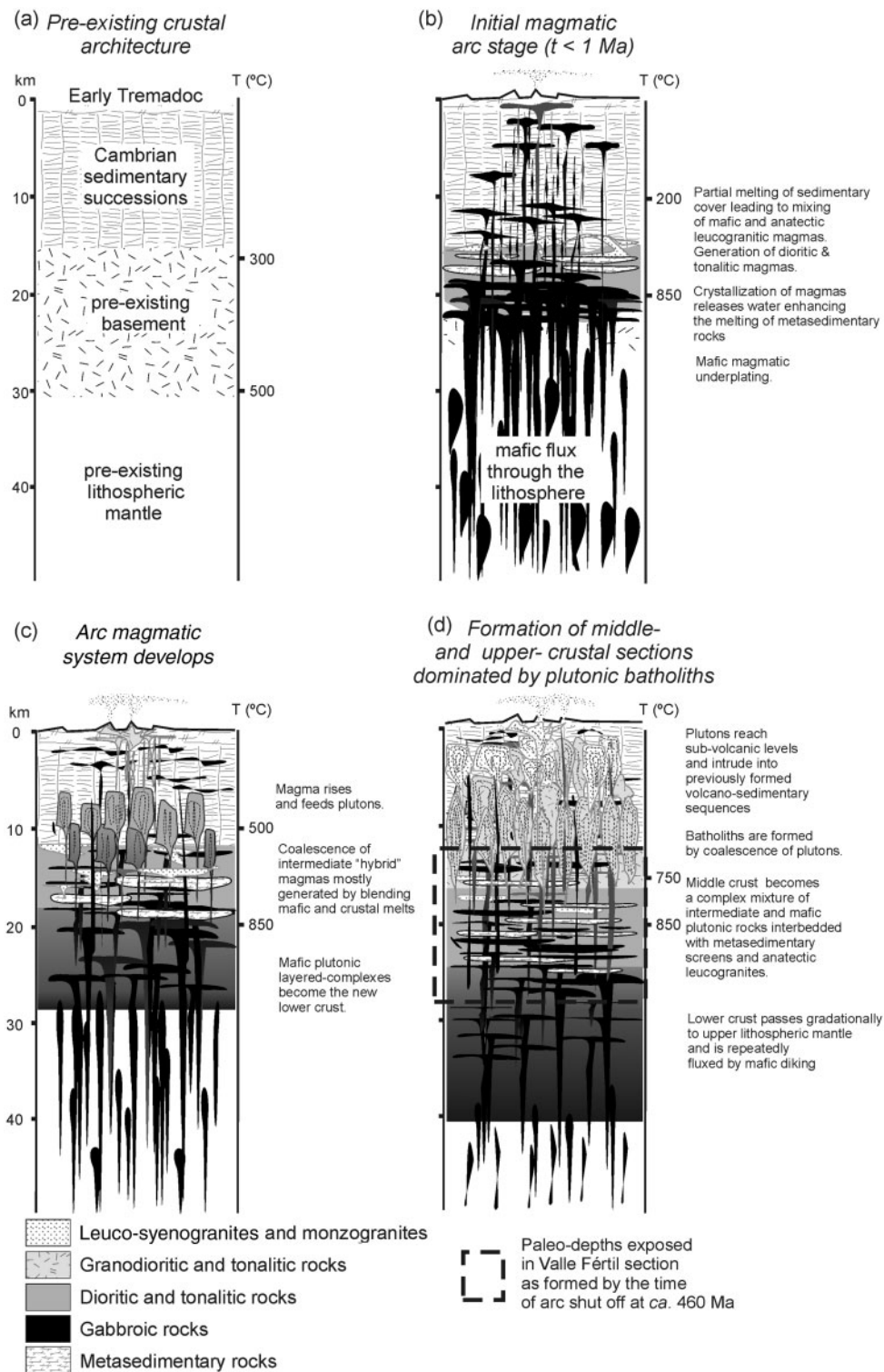


Fig. 12. Four-stage diagram schematically summarizing the evolution of the Famatinian paleo-arc crust as inferred from the Valle Fertil section. In (a) the position of isotherms reflects the geothermal gradient for a surface heat flow of 60 mW m^{-2} computed by Chapman & Furlong (1992). In (b) the isotherms are those computed by Annen & Sparks (2002) for injecting basaltic magmas at 20 km depth for a total duration of 8×10^5 years. The rate of magmatic addition is constant at a 50 m thick sill every 10 kyr, and basaltic magmas are at 1300°C . In (b) and (c) representative temperatures and pressures are after Otamendi *et al.* (2010b). The model of deep batholithic structure conserves the main features outlined in the studies of Hildreth (1981), Ducea (2002), Saleeby *et al.* (2003) and Miller *et al.* (2009).

of a crustal component is compositionally more evident for the tonalites in the transitional unit, where the process of magma mingling is observed in the field (Fig. 3d and e). The alternative that typical tonalites formed after crystal fractionation in dioritic magma chambers, which had already incorporated a crustal component into them, cannot be ruled out on the basis of either mass-balance modeling or compositional variation.

An unavoidable field observation is that granodioritic magmas do not form after partial melting of either a single rock type or complex rock sequences. Hence, the problem is how to generate large volumes of the granodioritic magmas that led to the formation of plutonic rocks characterized by the coexistence of amphibole and biotite, a metaluminous signature, and relatively alkali-enriched ($\text{Na}_2\text{O} + \text{K}_2\text{O} > 4.7$ wt % and $\text{Na}_2\text{O}/\text{K}_2\text{O} > 1$) composition (e.g. Patiño Douce, 1999).

The elemental and isotopic compositional characteristics of the granodiorites can be best explained if three end-member components contribute to their generation. Two end-members are the tonalites and leucogranites, which are blended together to form homogeneous granodiorites. This hypothesis is based on field observations that show the advance of hybridization between tonalites and leucogranites at different stages of development (Fig. 3c and d). The third component is mafic and appears as gabbroic and dioritic enclaves, amphibole-rich gabbroic bodies, and chilled gabbroic dikes and sills.

The granodioritic magmas were produced through: (1) crystal fractionation in evolving tonalitic magma chambers, which were rejuvenated by intrusion of syn-plutonic mafic dikes and already had a supracrustal component incorporated into them; (2) assimilation (or bulk melting) of exposed metasedimentary rocks into intermediate magmas chambers that carried variably assimilated fragments of their precursor mafic rocks; or (3) intrusion of anatectic leucogranites forming a network of dikes into still hot intermediate plutonic rocks, resulting in intimate mixing of all tonalitic and leucogranitic components. Although the field observations support the last process, for this to be effective either the intermediate rocks should have been at, or above, their solidus temperature or the invading leucogranitic melts should have been more voluminous than the intermediate rocks (Sparks & Marshall, 1986).

To a first approximation the variation of trace elements and Sr and Nd isotope ratios in the tonalites and granodiorites is best explained by progressive mixing processes involving mantle-derived magmas and metasedimentary material. However, the major petrological process cannot be one of simple binary mixing. Element–element and isotope–element variations in diagrams such as Figs 9 and 10 illustrate that, in general, the chemically evolved granodiorites do not have the highest $^{87}\text{Sr}/^{86}\text{Sr}(\text{T})$ and lowest

$\epsilon_{\text{Nd}}(\text{T})$, as one would expect if they had originated from magmas richer in the metasedimentary-derived end-member.

There are intermediate and silicic plutonic rocks that in terms of petrography and whole-rock major-element composition should be given the same name; however, their generation may involve different petrological processes. Clearly petrographically similar rocks contain distinct proportions of mantle-derived juvenile and crustal components. This petrological heterogeneity might not be appreciated while studying plutonic rocks emplaced at shallow crustal depths, as the collection and ascent of relatively large magma masses to upper crustal levels would mix and blend the crystals and melts derived from different plutonic and/or crustal precursors (Miller *et al.*, 1988; Beard *et al.*, 2005; Davidson *et al.*, 2007).

Construction of the Famatinian arc and implications for arc plutonic crust

The configuration of the crust at the time of initiation of a subduction-related magmatic system plays an important role in controlling the petrological products of the magmatic system. The Early Ordovician Famatinian batholiths were generated and emplaced into a sedimentary environment fed by continental detritus and deposited into marine basins (Mángano & Buatois, 1996; Bahlburg, 1998; Zimmermann & Bahlburg, 2003; Astini & Dávila, 2004). The nature and composition of the crustal basement underlying these basins remains unclear. The basins formed along the West Gondwana margin and probably reflect the transition between Paleozoic oceanic lithosphere to the west (in present coordinates) and South American Proterozoic lithosphere to the east (Piñán Llamas & Simpson, 2006). Two critical uncertainties are the total thickness of the crust and the thickness of the sedimentary cover that was deposited between the latest Neoproterozoic and the earliest Ordovician. The stratigraphic record provides the most compelling evidence to reduce these uncertainties. It shows that the Famatinian volcanism erupted close to sea level (Mángano & Buatois, 1996).

Simple isostatic calculations show that for the top of the lithosphere to be near sea level either a continental (granitic) basement or an oceanic (basaltic) crust can accommodate between 5 and 10 km of sedimentary loading if the total thickness of the crust is less than 30 km (e.g. Cloos, 1993). For the Famatinian case, a 30 km thick crust composed of 15 km of accreted sedimentary sequences overlying a crystalline basement can explain the existence of metasedimentary packages buried at lower crustal depths (20 km) by the time mafic magmatic activity began to dominate the arc lower crust (Otamendi *et al.*, 2008). Considering this pre-arc crustal structure, we propose a multiple-step model for generating the Famatinian arc crust (Fig. 12). At the deepest exposed levels of the arc,

the Valle Fértil plutonic section is dominated by mafic rocks, but it does not expose the lithospheric mantle–crust boundary (Otamendi *et al.*, 2009a). Field evidence indicates that the magmatic products leaving behind mantle residues and reaching the seismic Moho are mostly basaltic melts.

During the initial stages of arc magmatism, mafic magmas would have been emplaced near the basement–cover boundary, because the contrasting densities across this interface act as a filter for hydrous mafic magmas. Although the density of mafic magmas is very sensitive to the fraction of dissolved water in the melt, the densities of typical mafic arc melts range between 2.45 and 2.55 g cm⁻³ (Sisson *et al.*, 1996). The first rising mafic magmas to come across the base of the sedimentary cover could have stalled as they became neutrally buoyant with respect to the pre-existing country rocks. The mafic magmas that crossed the lower levels of the sedimentary cover were either emplaced as sill-like bodies or erupted at the surface (Fig. 12b). Clear indications that mafic magmas were able to penetrate into the sedimentary cover appear in the upper part of the Famatinian paleo-arc crust, where basaltic lava flows, fine-grained gabbroic sills, dikes and laccoliths are found interbedded with early Ordovician sedimentary sequences (Mannheim, 1993; Mannheim & Miller, 1996). The relative volume of gabbroic and dioritic magmas that reached the upper crust was, however, a minor proportion of that injected into deep-seated crustal domains as sills and dikes. The early additions of mafic magma have one major consequence: the geothermal gradient must have increased dramatically, leading to the progressive heating of sedimentary packages close to and above their solidus temperature (Huppert & Sparks, 1988; Bergantz, 1989). An interval of at least 1 Myr of crustal heating is required to produce sufficient partial melting of the pre-existing sedimentary protoliths (Annen & Sparks, 2002; Annen *et al.*, 2006).

Evidence for the processes occurring during the early stages of arc magmatism is also found at the deepest exposed levels of the Valle Fértil section (Otamendi *et al.*, 2009a, 2009b). As shown above, the radiogenic isotopic composition of the gabbros and diorites in the mafic unit requires that supracrustal rocks exist at the stagnation depths of the early mafic magmas. The main neutral buoyancy levels of the early mafic magmas reached the bottom of the pre-arc sedimentary packages (Fig. 12b). Field observations suggest that the sedimentary rocks were rapidly heated to partial melting temperatures, rather than being bulk assimilated into the mafic magmas. The main process generating intermediate magmas was initially promoted by the emplacement of primitive mafic magmas supplying heat, and upon crystallization, releasing aqueous fluids. The dramatic increase of the thermal gradient caused metasedimentary rocks at lower and middle crustal levels

to melt. After undergoing partial melting, the sedimentary packages released anatectic melts that invaded and reacted with the crystallizing mafic magmas. In turn, invading mafic magmas that penetrated into the metasedimentary packages were entrained in, and commingled with anatectic felsic melts (Otamendi *et al.*, 2010b). The end result of the complex interaction between rapidly melted metasedimentary sequences and hot mafic magmas was the formation of a network of dioritic to tonalitic veins and bodies throughout the mafic lower crustal levels.

Dioritic and tonalitic veins in the gabbroic rocks were predominantly derived from juvenile igneous components but they also bear a crustal component (Fig. 10). Recurrent tapping of mafic magma chambers residing in the lower crust sustained the thermal and material budget to respectively promote anatexis of fertile crustal protoliths and provide the mass flux for blending juvenile and crustal magmas (e.g. Davidson *et al.*, 1988; Hildreth & Moorbath, 1988; Feeley *et al.*, 2002). Horizontally sheeted, partially molten intermediate bodies enhanced the ascent of subsequent magmas to higher crustal levels where larger volumes of magma became thoroughly hybridized, resulting in gradational contacts (Fig. 12b). Hybrid intermediate magma chambers would have been the main outcome of this general process. Where the mafic juvenile component was predominant, the overall magmatic process generated intermediate magmas, which upon accumulation would homogenize and ascend to upper crustal levels (Otamendi *et al.*, 2009b).

Conditions for generating metaluminous amphibole-bearing granodiorites and monzogranites were attained when the middle crust became to a large extent dominated by intermediate plutonic rocks; calc-alkaline silicic rocks resulted from either fractional crystallization in intermediate magma chambers or hybridization between intermediate plutonic rocks and anatectic leucogranites (Fig. 12c). When the thermal conditions at mid-crustal depths (*c.* 15–20 km) reached temperatures above 800°C, the rate of magma generation became high enough to form a large volume of magma, which ascended driven by buoyancy forces.

The production of intermediate to silicic magmas is envisaged to have developed sequentially in batches as melt in volumetrically variable magma chambers accumulated and ascended. Because the arc crust evolved with time, the generation of intermediate and silicic plutonic units caused the crust to thicken. Given that intermediate and silicic rocks have low-density quartz–plagioclase assemblages, progressively less mafic magma was capable of crossing the boundary between crustal levels dominated by quartz-absent mafic rocks (mafic unit) and quartz–plagioclase-dominated rocks (intermediate unit). This is due to the density-filter effect, which predicts that the predominance of quartz-feldspathic plutonic rocks at middle

crustal levels would force primitive mafic magmas to stagnate at progressively greater depths (Leeman, 1983; Sisson *et al.*, 1996). The temporal change in the composition of the dominant plutonism from the early entrance of mafic magmas to the large volume of intermediate and silicic magmas with a progressively larger crustal component was thought to be a first-order feature of Cordilleran magmatism (e.g. Pitcher *et al.*, 1985). The preponderance of plutonic rocks in the Valle Fértil crustal section is typical and characteristic of deep-seated Cordilleran magmatic systems (Miller *et al.*, 2009). Furthermore, the overall composition of the plutonic rocks dominating the crust becoming progressively more mafic with depth is also a feature of arc crustal sections (e.g. Christensen & Mooney, 1995; Ducea *et al.*, 2003).

After a few million years (between 1.5 and 5.0 Myr) of magmatic evolution, the rate of production of intermediate to silicic magmas, mainly at mid-crust depths, is large enough for the magmas to coalesce, ascend and feed magma chambers (Fig. 12d; also see Annen & Sparks, 2002). As suggested by Miller *et al.* (1988, 2009) and Miller & Paterson (2001), magma transfer to upper crustal levels is probably accommodated by a network of channels and/or as multiple pulses of narrow, elongate, visco-elastic diapirs. Consistent with this hypothesis, at upper crustal depths in the Famatinian arc, tonalites and granodiorites form elongate, vertically sheeted plutons oriented parallel to the major elongation direction of the arc (e.g. Pankhurst *et al.*, 1998, 2000; Lucassen & Franz, 2005).

Field evidence for the progressive construction of the Famatinian plutonic–volcanic system is clear even at a regional scale. At a mature stage of the evolution of the central segment of the Famatinian arc, monzogranitic magmas stalled in shallow magma chambers and constructed large plutons. The roofs of these shallow plutons were emplaced at the base of the Middle Cambrian sedimentary succession (Collo & Astini, 2008). Zircon geochronology in the Valle Fértil section shows that the building of a completely new section of arc crust encompassed a time span of about 10 Myr (Ducea *et al.*, 2010). During this time interval, and by the time the arc was shut down, the quartz-bearing intermediate and silicic plutonic sequence had grown enough to dominate the middle to upper levels of the arc crust (Otamendi *et al.*, 2010b).

CONCLUSIONS

Production of hydrous mafic magmas in the mantle wedge above subducting slabs is the main thermal and magmatic input at destructive plate margins (Gill, 1981; Davies & Stevenson, 1992; Gaetani & Grove, 1998; Ulmer, 2001). A deep crustal section dominated by amphibole-rich gabbroic rocks directly underlying an intermediate plutonic crust, like that observed in this study, supports this hypothesis and provides constraints for the process involved in

the construction of intermediate and silicic crust. The most significant findings from this study are as follows.

- (1) Some dioritic and tonalitic rocks may reflect the most evolved magmas derived from fractionation of hydrous primitive mafic magmas or partial melting of gabbroic rocks. However, the principal process involved in generating large volumes of dioritic and tonalitic magmas includes contamination from a supracrustal metasedimentary source.
- (2) Formation of dioritic and tonalitic magmas takes place to a sufficiently large extent to allow the separation and rise of distinct magma batches. While the intermediate magmas are en route to upper crustal levels, they can experience crystal fractionation and reaction-driven assimilation of mafic fragments entrained in them (e.g. Beard *et al.*, 2005). Crystal fractionation and assimilation of mafic inclusions cause opposite variations in elemental compositions; the extent of the latter process may be constrained by the radiogenic isotopic compositions of the magmatic rocks. Moreover, to generate intermediate composition magma chambers with melt fractions higher than the melt escape threshold at middle crustal depths, the entire middle to lower crust must be hot, a condition that is met by the influx of mafic magma at a high rate of intrusion (Miller *et al.*, 1988, 2009; Annen & Sparks, 2002; Annen *et al.*, 2006). This stage of thermal maturity of the magmatic systems is evidenced by mafic dikes and sills intruding into still hot or partially molten intermediate plutonic rocks (Fig. 3a and b).
- (3) A major compositional change occurs when intermediate magmas interact with either metasedimentary rocks or metasedimentary-derived melts. Complex melt–mineral reactions, mingling and mixing between tonalites and leucogranites are observable in the field (Fig. 3) and evident from major element variations (Fig. 4). Granodioritic to monzogranitic rocks display a compositional trend that crosses the boundary between metaluminous and peraluminous fields; therefore, the simplest explanation for the generation of this silicic plutonic suite is reaction between crustal melts (i.e. leucogranites) and rocks or magmas having metaluminous intermediate calc-alkaline compositions (e.g. Patiño Douce, 1999). The preservation over large regions of the tonalitic rocks including leucogranites and paragneissic migmatitic septa is therefore taken to reflect the local existence of rheological and/or thermal barriers preventing the full mixing of these materials.
- (4) The generation of new plutonic arc crust fully reprocesses the pre-existing crust (e.g. Saleeby *et al.*, 2003); however, the total thickness and the internal crustal structure of the plutonic crust are determined by the

total time span of magmatic activity. The arc section studied here was built up by a single arc episode during which the magmatic loci remained nearly stationary within the reference frame of the upper plate (Ducea *et al.*, 2010; Fig. 12). In contrast, many continental batholithic belts appear to have been constructed over several cycles of arc magmatic activity and to have experienced spatial migration of the magmatism; with time these develop more complex internal crustal structures (Ducea, 2002). The latter continental margin arcs would construct thicker crustal sections and extend their dense roots into the mantle (Ducea & Saleeby, 1996).

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REFERENCES

- Aceñolaza, F. G., Miller, H. & Toselli, A. J. (2000). The Pampean and Famatinian cycles—superposed orogenic events in the West Gondwana. *Zeitschrift für Angewandte Geologie, Sonderheft*, SH 1, 337–344.
- Adams, C. J., Miller, H., Aceñolaza, F. G., Toselli, A. J. & Griffin, W. L. (2010). The Pacific Gondwana margin in the late Neoproterozoic–early Paleozoic: Detrital zircon U–Pb ages from metasediments in northwest Argentina reveal their maximum age, provenance and tectonic setting. *Gondwana Research* **19**, 71–83.
- Anders, E. & Grevesse, N. (1989). Abundances of the elements—meteoritic and solar. *Geochimica et Cosmochimica Acta* **53**, 197–214.
- Anderson, J. L. (ed.) (1990). *The Nature and Origin of Cordilleran Magmatism*. Geological Society of America, *Memoirs* **174**, 414 p.
- Annen, C. & Sparks, R. S. J. (2002). Effects of repetitive emplacement of basaltic intrusions on thermal evolution and melt generation in the crust. *Earth and Planetary Science Letters* **203**, 937–955.
- Annen, C., Blundy, J. D. & Sparks, R. S. J. (2006). The genesis of intermediate and silicic magmas in deep crustal hot zones. *Journal of Petrology* **47**, 505–539.
- Astini, R. A. & Dávila, F. M. (2004). Ordovician back arc foreland and Ocoyic thrust belt development on the western Gondwana margin as a response to Precordillera terrane accretion. *Tectonics* **23**, TC4008, doi:10.1029/2003TC001620.
- Bahlburg, H. (1998). The geochemistry and provenance of Ordovician turbidites in the Argentine Puna. In: Pankhurst, R. J. & Rapela, C. W. (eds) *The Proto-Andean Margin of Gondwana*. Geological Society, London, *Special Publications* **142**, 127–142.
- Barazangi, M. & Isacks, B. I. (1976). Spatial distribution of earthquakes and subduction of the Nazca plate beneath South America. *Geology* **4**, 686–692.
- Bateman, P. C. & Eaton, J. P. (1967). Sierra Nevada Batholith. *Science* **158**, 1407–1417.
- Beard, J. S., Ragland, P. C. & Crawford, M. L. (2005). Reactive bulk assimilation: A model for crust–mantle mixing in silicic magmas. *Geology* **33**, 681–684.
- Bergantz, G. W. (1989). Underplating and partial melting: implications for melt generation and extraction. *Science* **245**, 1093–1095.
- Carrington, D. P. & Watt, G. R. (1995). A geochemical and experimental study of the role of K-feldspar during water-undersaturated melting of metapelites. *Chemical Geology* **122**, 59–76.
- Castro, A., Martino, R., Vujovich, G., Otamendi, J., Pinotti, J., Deramo, F., Tibaldi, A. & Viñao, G. (2008). Top-down structures of mafic enclaves within the Valle Fértil magmatic complex (Early Ordovician, San Juan, Argentina). *Geologica Acta* **6**, 217–229.
- Cawood, P. A. (2005). Terra Australis Orogen: Rodinia breakup and development of the Pacific and Iapetus margins of Gondwana during the Neoproterozoic and Paleozoic. *Earth-Science Reviews* **69**, 249–279.
- Chapman, D. S. & Furlong, K. P. (1992). Thermal state of the continental lower crust. In: Fountain, D. M., Arculus, R. & Kay, R. W. (eds) *Continental Lower Crust. Development in Geotectonics* **23**, 179–199.
- Chew, D. M., Schaltegger, U., Kosler, J., Whitehouse, M. J., Gutjahr, M., Spikings, R. A. & Miskovic, A. (2007). U–Pb geochronologic evidence for the evolution of the Gondwanan margin of the north–central Andes. *Geological Society of America Bulletin* **119**, 697–711.
- Christensen, N. I. & Mooney, W. D. (1995). Seismic velocity structure and composition of the continental crust: a global view. *Journal of Geophysical Research* **100**(B7), 9761–9788.
- Cisterna, C. E., Medina, M. E. & Fernández Turiel, J. L. (2004). Evidencias de procesos de fraccionamiento y mezcla de magmas en el Complejo Intrusivo Las Angosturas, Sistema de Famatina. *Revista de la Asociación Geológica Argentina* **59**, 45–56.
- Clemens, K. & Miller, H. (1996). Sedimentología, proveniencia y posición geotectónica de las sedimentitas del Precámbrico y Paleozoico inferior del Sistema de Famatina. In: Aceñolaza, F., Miller, H. & Toselli, A. (eds) *Geología del Sistema de Famatina*. *Münchener Geologische Hefte* **19A**, 31–50.
- Cloos, M. (1993). Lithospheric buoyancy and collisional orogenesis: subduction of ocean plateaus, continental margins, island arcs, spreading ridges, and seamounts. *Geological Society of America Bulletin* **105**, 715–737.
- Coira, B., Kay, S. M., Pérez, B., Woll, B., Hanning, M. & Flores, P. (1999). Magmatic sources and tectonic setting of Gondwana margin Ordovician magmas, northern Puna of Argentina and Chile. In: Ramos, V. & Keppie, J. (eds) *Laurentia–Gondwana Connections before Pangea*. Geological Society of America, *Special Papers* **336**, 145–170.
- Collo, G. & Astini, R. (2008). La Formación Achavil: una nueva unidad de bajo grado metamórfico en la evolución Cámbrica superior del Famatina. *Revista de la Asociación Geológica Argentina* **63**, 344–362.

- Collo, G., Astini, R., Cawood, P. A., Buchan, C. & Pimentel, M. (2009). U–Pb detrital zircon ages and Sm–Nd isotopic features in low-grade metasedimentary rocks of the Famatina belt: implications for late Neoproterozoic–early Palaeozoic evolution of the proto-Andean margin of Gondwana. *Journal of the Geological Society, London* **166**, 303–319.
- Dahlquist, J. A., Rapela, C. W. & Baldo, E. G. (2005). Petrogenesis of cordierite-bearing S-type granitoids in Sierra de Chepes, Famatinian orogen, Argentina. *Journal of South America Earth Sciences* **20**, 231–251.
- Davidson, J. P., Ferguson, K. M., Colucci, T. & Dungan, M. A. (1988). The origin and evolution of magmas from San Pedro–Pellado Volcanic Complex, S. Chile: multicomponent sources and open system evolution. *Contributions to Mineralogy and Petrology* **100**, 429–445.
- Davidson, J. P., Morgan, D. J., Charlier, B. L. A., Harlou, R. & Hora, J. M. (2007). Microsampling and isotopic analysis of igneous rocks: implications for the study of magmatic systems. *Annual Review of Earth and Planetary Sciences* **35**, 273–311.
- Davies, J. H. & Stevenson, D. J. (1992). Physical models of source region of subduction zone volcanics. *Journal of Geophysical Research* **97**(B2), 2037–2070.
- DeBari, S. (1997). Evolution of magmas in continental and oceanic arcs: the role of the lower crust. *Canadian Mineralogist* **35**, 507–519.
- de la Rosa, J. D., Chacón, H., Sánchez de la Campa, A., Carrasco, R. & Nieto, J. M. (2001). Metodología y análisis de elementos trazas-REE mediante ICP-MS del standard SARM 1 (granito) y SARM 4 (norita). In: Lago, M., Arranz, E. & Galé, C. (eds) *Libro de Actas del III Congreso Ibérico de Geoquímica*. Zaragoza: Instituto Tecnológico de Aragón, pp. 435–438.
- DePaolo, D. J. (1981). Trace element and isotopic effects of combined wall rock assimilation and fractional crystallization. *Earth and Planetary Science Letters* **53**, 189–202.
- Drew, S. T., Ducea, M. N. & Schoenbohm, L. M. (2009). Mafic volcanism on the Puna Plateau, NW Argentina: Implications for lithospheric composition and evolution with an emphasis on lithospheric foundering. *Lithosphere* **1**, 305–318.
- Ducea, M. N. (2002). Constraints on the bulk composition and root foundering rates of continental arcs: a California arc perspective. *Journal of Geophysical Research* **107**(B11), 2304, doi:10.1029/2001JB000643.
- Ducea, M. N. & Saleeby, J. B. (1996). Buoyancy sources for a large, unrooted mountain range, the Sierra Nevada, California: evidence from xenolith thermobarometry. *Journal of Geophysical Research* **101**, 8226–8244.
- Ducea, M. N. & Saleeby, J. B. (1998). The age and origin of a thick mafic ultramafic root from beneath the Sierra Nevada batholith. *Contributions to Mineralogy and Petrology* **133**, 169–185.
- Ducea, M. N., Kidder, S. & Zandt, G. (2003). Arc composition at mid-crustal depths: Insights from the Coast Ridge Belt, Santa Lucia Mountains, California. *Geophysical Research Letters* **30**, 1703, doi:10.1029/2002GL016297.
- Ducea, M. N., Otamendi, J. E., Bergantz, G., Stair, K., Valencia, V. & Gehrels, G. (2010). Timing constraints on building an intermediate plutonic arc crustal section: U–Pb zircon geochronology of the Sierra Valle Fértil, Famatinian Arc, Argentina. *Tectonics* **29**, TC4002, doi:10.1029/2009TC002615.
- Dunn, T. & Sen, C. (1994). Mineral/matrix partition coefficients for orthopyroxene, plagioclase, and olivine in basaltic to andesitic systems: A combined analytical and experimental study. *Geochimica et Cosmochimica Acta* **58**, 717–733.
- Egenhoff, S. O. & Lucassen, F. (2003). Chemical and isotopic composition of Lower to Upper Ordovician Sedimentary Rocks (Central Andes/South Bolivia): implications for their source. *Journal of Geology* **111**, 487–497.
- Faure, G. (1986). *Principles of Isotope Geochemistry*. New York: John Wiley, 464 p.
- Faure, G. & Mensing, T. M. (2005). *Isotopes: Principles and Applications*. New York: John Wiley, 928 p.
- Feeley, T. C., Cosca, M. A. & Lindsay, C. R. (2002). Petrogenesis and implications of calc-alkaline cryptic hybrid magmas from Washburn Volcano, Absaroka Volcanic province, USA. *Journal of Petrology* **43**, 663–703.
- Frost, T. P. & Mahood, G. A. (1987). Field, chemical and physical constraints on mafic–felsic magma interaction in the Lamarck Granodiorite, Sierra Nevada, California. *Geological Society of America Bulletin* **99**, 272–291.
- Furman, T. & Spera, F. J. (1985). Co-mingling of acid and basic magma with implications for the origin of mafic I-type xenoliths: field and petrochemical relations of an unusual dyke complex at Eagle Lake, Sequoia National Park, California, U.S.A. *Journal of Volcanology and Geothermal Research* **24**, 151–178.
- Gaetani, G. A. & Grove, T. L. (1998). The influence of water on melting of mantle peridotite. *Contributions to Mineralogy and Petrology* **131**, 323–346.
- Galer, S. J. G. & Abouchami, W. (2004). Mass bias correction laws suitable for MC-ICP–MS measurement (abstract), in 14th Annual Goldschmidt conference. *Geochimica et Cosmochimica Acta* **68**, A542.
- Gallien, F., Mogessie, A., Bjerg, E., Delpino, S., Castro de Machuca, B., Thöni, M. & Klötzli, U. (2010). Timing and rate of granulite facies metamorphism and cooling from multi-mineral chronology on migmatitic gneisses, Sierras de La Huerta and Valle Fértil, NW Argentina. *Lithos* **114**, 229–252.
- Gill, J. B. (1981). *Orogenic Andesites and Plate Tectonics*. New York: Springer, 390 p.
- Hart, S. R. (1984). A large-scale isotope anomaly in the Southern Hemisphere mantle. *Nature* **309**, 753–757.
- Hildreth, W. (1981). Gradients in silicic magma chambers: implications for lithospheric magmatism. *Journal of Geophysical Research* **86**(B11), 10153–10192.
- Hildreth, W. & Moorbath, S. (1988). Crustal contributions to arc magmatism in the Andes of Central Chile. *Contributions to Mineralogy and Petrology* **98**, 455–489.
- Huppert, H. E. & Sparks, R. S. J. (1988). The generation of granitic magmas by intrusion of basalt into continental crust. *Journal of Petrology* **29**, 599–624.
- Johnson, D. M., Hooper, P. R. & Conrey, R. M. (1999). XRF analysis of rocks and minerals for major and trace elements on a single low dilution Li-tetraborate fused bead. *Advances in X-Ray Analysis* **41**, 843–867.
- Jordan, T. E. & Allmendinger, R. W. (1986). The Sierras Pampeanas of Argentina: a modern analogue of Rocky Mountain foreland deformation. *American Journal of Science* **286**, 737–764.
- Kay, S. M., Orrell, S. & Abbruzzi, J. M. (1996). Zircon and whole rock Nd–Pb isotopic evidence for a Grenville age and a Laurentian origin for the basement of the Precordillera in Argentina. *Journal of Geology* **104**, 637–648.
- Kleine, T., Mezger, K., Zimmermann, U., Münker, C. & Bahlburg, H. (2004). Crustal evolution along the Early Ordovician Proto-Andean margin of Gondwana: trace element and isotope evidence from the Complejo Igneo Pocitos (Northwestern Argentina). *Journal of Geology* **112**, 503–520.
- Kretz, R. (1983). Symbols for rock-forming minerals. *American Mineralogist* **68**, 277–279.

- Langmuir, C. H., Vocke, R. D., Hanson, G. N. & Hart, S. R. (1978). A general mixing equation with applications to Icelandic basalts. *Earth and Planetary Science Letters* **37**, 380–392.
- Leeman, W. P. (1983). The influence of crustal structure on compositions of subduction-related magmas. *Journal of Volcanology and Geothermal Research* **18**, 561–588.
- Le Maitre, R. W. (1979). A new generalized petrological mixing model. *Contributions to Mineralogy and Petrology* **71**, 133–137.
- Le Maitre, R. W., Bateman, P., Dudek, A., Keller, J., Lameyre, J., Le Bas, M. J., Sabine, P. A., Schmid, R., Sørensen, S., Streckeisen, A., Woolley, A. R. & Zanettin, B. (1989). *A Classification of Igneous Rocks and Glossary of Terms*. Oxford: Blackwell Scientific, 193 p.
- Leshner, C. E. (1990). Decoupling of chemical and isotopic exchange during magma mixing. *Nature* **344**, 235–237.
- Leshner, C. E. (1994). Kinetics of Sr and Nd exchange in silicate liquids: theory, experiments, and applications to uphill diffusion, isotopic equilibration, and irreversible mixing of magmas. *Journal of Geophysical Research* **99**, 9585–9604.
- Lucassen, F. & Franz, G. (2005). The early Palaeozoic Orogen in the Central Andes: a non-collisional orogen comparable to the Cenozoic high plateau? In: Vaughan, A., Leat, P. T. & Pankhurst, R. (eds) *Terrane Processes at the Margins of Gondwana Geological Society, London, Special Publications* **246**, 257–373.
- Lucassen, F., Harmon, R., Franz, G., Romer, R. L., Becchio, R. & Siegel, W. (2002). Lead evolution of the Pre-Mesozoic crust in the Central Andes (18–27°): progressive homogenization of Pb. *Chemical Geology* **186**, 183–197.
- Mángano, M. G. & Buatois, L. A. (1996). Shallow marine event sedimentation in a volcanic arc-related setting: the Ordovician Suri Formation, Famatina Range, northwest Argentina. *Sedimentary Geology* **105**, 63–90.
- Mannheim, R. (1993). Genese der Vulkanite und Subvulkanite des alt-paläozoischen Famatina-Systems, NW-Argentinien, und seine geodynamische Entwicklung. *Münchner Geologische Hefte* **7**, 1–155.
- Mannheim, R. & Miller, H. (1996). Las rocas volcánicas y subvolcánicas copaleozoicas del Sistema de Famatina. In: Aceñolaza, F., Miller, H. & Toselli, A. (eds) *Geología del Sistema de Famatina. Münchner Geologische Hefte* **19A**, 159–186.
- Miller, C. F., Watson, E. B. & Harrison, T. M. (1988). Perspectives on the source, segregation and transport of granitoid magmas. *Transactions of the Royal Society of Edinburgh: Earth Sciences* **79**, 135–156.
- Miller, R. B. & Paterson, S. R. (2001). Influence of lithological heterogeneity, mechanical anisotropy, and magmatism on the rheology of an arc, North Cascades, Washington. *Tectonophysics* **342**, 351–370.
- Miller, R. B. & Snoke, A. W. (2009). The utility of crustal cross sections in the analysis of orogenic processes in contrasting tectonic settings. In: Miller, R. B. & Snoke, A. W. (eds) *Crustal Cross Sections from the Western North American Cordillera and Elsewhere: Implications for Tectonic and Petrologic Processes. Geological Society of America, Special Papers* **456**, 1–38.
- Miller, R. B., Paterson, S. R. & Matzel, J. P. (2009). Plutonism at different crustal levels: Insights from the ~5–40 km (paleodepth) North Cascades crustal section, Washington. In: Miller, R. B. & Snoke, A. W. (eds) *Crustal Cross Sections from the Western North American Cordillera and Elsewhere: Implications for Tectonic and Petrologic Processes. Geological Society of America, Special Papers* **456**, 125–149.
- Mirré, J. C. (1976). *Descripción Geológica de la Hoja 19e, Valle Fértil, Provincias de San Juan y La Rioja*. Buenos Aires: Servicio Geológico Nacional, pp. 1–70.
- Miyashiro, A. (1974). Volcanic rock series in island arcs and active continental margins. *American Journal of Science* **274**, 321–355.
- Montel, J.-M. & Vielzeuf, D. (1997). Partial melting of metagreywackes, part II, compositions of minerals and melts. *Contributions to Mineralogy and Petrology* **128**, 176–196.
- Moore, J. G. (1959). The quartz diorite boundary line in the Western United States. *Journal of Geology* **67**, 198–210.
- Mulcahy, S. R., Roeske, S. M., McClelland, W. C., Jourdan, F., Iriondo, A., Renne, P. R., Vervoort, J. D. & Vujovich, G. I. (2011). Structural evolution of a composite middle to lower crustal section: The Sierra de Pie de Palo, northwest Argentina. *Tectonics* **30**, TC1005, doi:10.1029/2009TC002656.
- Nielsen, R. L., Forsythe, L. M., Gallagher, W. E. & Fisk, M. R. (1994). Major and trace element magnetite–melt partitioning. *Chemical Geology* **117**, 167–191.
- Otamendi, J. E., Pinotti, L. P., Basei, M. A. S. & Tibaldi, A. M. (2010b). Evaluation of petrogenetic models for intermediate and silicic plutonic rocks from the Sierra de Valle Fértil–La Huerta, Argentina: petrologic constraints on the origin of igneous rocks in the Ordovician Famatinian–Puna paleoarc. *Journal of South American Earth Sciences* **30**, 29–45.
- Otamendi, J. E., Tibaldi, A. M., Vujovich, G. I. & Viñao, G. A. (2008). Metamorphic evolution of migmatites from the deep Famatinian arc crust exposed in Sierras Valle Fértil–La Huerta, San Juan, Argentina. *Journal of South American Earth Sciences* **25**, 313–335.
- Otamendi, J. E., Cristofolini, E., Tibaldi, A. M., Quevedo, F. & Baliani, I. (2010a). Petrology of mafic and ultramafic layered rocks from the Jaboncillo Valley, Sierra de Valle Fértil, Argentina: implications for the evolution of magmas in the lower crust of the Famatinian arc. *Journal of South American Earth Sciences* **29**, 685–704.
- Otamendi, J. E., Vujovich, G. I., de la Rosa, J. D., Tibaldi, A. M., Castro, A. & Martino, R. D. (2009a). Geology and petrology of a deep crustal zone from the Famatinian paleo-arc, Sierras Valle Fértil–La Huerta, San Juan, Argentina. *Journal of South American Earth Science* **27**, 258–279.
- Otamendi, J. E., Ducea, M. N., Tibaldi, A. M., Bergantz, G., de la Rosa, J. D. & Vujovich, G. I. (2009b). Generation of tonalitic and dioritic magmas by coupled partial melting of gabbroic and meta-sedimentary rocks within the deep crust of the Famatinian magmatic arc, Argentina. *Journal of Petrology* **50**, 841–873.
- Pankhurst, R. J. & Rapela, C. W. (1998). Introduction. In: Pankhurst, R. J. & Rapela, C. W. (eds) *The Proto-Andean Margin of Gondwana. Geological Society, London, Special Publications* **142**, 1–9.
- Pankhurst, R. J., Rapela, C. W. & Fanning, C. M. (2000). Age and origin of coeval TTG, I- and S-type granites in the Famatinian belt of NW Argentina. *Transactions of the Royal Society of Edinburgh, Earth Sciences* **91**, 151–168.
- Pankhurst, R. J., Rapela, C. W., Saavedra, J., Baldo, E. G., Dahlquist, J., Pascua, I. & Fanning, C. M. (1998). The Famatinian magmatic arc in the central Sierras Pampeanas: an Early to Mid-Ordovician continental arc on the Gondwana margin. In: Pankhurst, R. J. & Rapela, C. W. (eds) *The Proto-Andean Margin of Gondwana. Geological Society, London, Special Publications* **142**, 343–368.
- Patiño Douce, A. E. (1999). What do experiments tell us about the relative contributions of crust and mantle to the origin of granitic magmas? In: Castro, A., Fernández, C. & Vigneresse, J. L. (eds) *Understanding Granites: Integrating New and Classical Techniques Geological Society, London, Special Publications* **168**, 55–75.
- Patiño Douce, A. E. & Harris, N. (1998). Experimental constraints on Himalayan anatexis. *Journal of Petrology* **39**, 689–710.
- Patiño Douce, A. E. & Johnston, A. D. (1991). Phase equilibria and melt productivity in the pelitic system: implications for the origin of peraluminous granitoids and aluminous granulites. *Contributions to Mineralogy and Petrology* **107**, 202–218.

- Piñán Llamas, A. & Simpson, C. (2006). Deformation of Gondwana margin turbidites during the Pampean orogeny, north-central Argentina. *Geological Society of America Bulletin* **118**, 1270–1279.
- Pitcher, W. S., Atherton, M. P., Cobbing, E. J. & Beckinsale, R. D. (eds) (1985). *Migmatism at a Plate Edge—the Peruvian Andes*. London: Blackie.
- Rapela, C. W., Pankhurst, R., Baldo, E., Casquet, C., Galindo, C., Fanning, C. & Saavedra, J. (2001). Ordovician metamorphism in the Sierras Pampeanas: new U–Pb SHRIMP ages in central-east Valle Fértil and the Velasco batholith. *Proceedings III South American Symposium of Isotope Geology*. Pucón (Chile): Sociedad Geológica de Chile, pp. 616–619.
- Reiners, P. W., Nelson, B. K. & Ghiorso, M. S. (1995). Assimilation of felsic crust by basaltic magma: Thermal limits and extents of crustal contamination of mantle-derived magmas. *Geology* **23**, 563–566.
- Rollinson, H. R. (1993). *Using Geochemical Data: Evaluation, Presentation, Interpretation*. London: Longman, 352 p.
- Rudnick, R. L. (1995). Making continental crust. *Nature* **378**, 571–578.
- Rudnick, R. L. & Gao, S. (2003). Composition of the continental crust. In: Rudnick, R. L. (ed.) *The Crust. Treatise on Geochemistry* **3**, 1–64.
- Saal, A., Toselli, A. J. & Rossi, J. N. (1996). Granitoides y rocas básicas de la Sierra de Paganzo. In: Aceñolaza, F., Miller, H. & Toselli, A. (eds) *Geología del Sistema de Famatina. Münchner Geologische Hefte* **19A**, 119–210.
- Saleeby, J. B., Ducea, M. N. & Clemens-Knott, D. (2003). Production and loss of high-density batholithic root, southern Sierra Nevada, California. *Tectonics* **22**, 1064, doi:10.1029/2002TC001374.
- Sawyer, E. W. (1987). The role of partial melting and fractional crystallization in determining discordant migmatite leucosome compositions. *Journal of Petrology* **28**, 445–473.
- Schiano, P., Monzier, M., Eissen, J.-P., Martin, H. & Koga, K. T. (2010). Simple mixing as the major control of the evolution of volcanic suites in the Ecuadorian Andes. *Contributions to Mineralogy and Petrology* **160**, 297–312.
- Sisson, T. W., Grove, T. L. & Coleman, D. S. (1996). Hornblende gabbro sill complex at Onion Valley, California, and a mixing origin for the Sierra Nevada batholith. *Contributions to Mineralogy and Petrology* **126**, 81–108.
- Solar, G. S. & Brown, M. (2001). Petrogenesis of migmatites in Maine, USA: possible source of peraluminous leucogranite in plutons? *Journal of Petrology* **42**, 789–823.
- Sparks, R. S. J. & Marshall, L. A. (1986). Thermal and mechanical constraints on mixing between mafic and silicic magmas. *Journal of Volcanology and Geothermal Research* **29**, 99–124.
- Stacey, J. S. & Kramers, J. D. (1975). Approximation of terrestrial lead isotope evolution by a two-stage model. *Earth and Planetary Science Letters* **26**, 207–221.
- Stern, R. J. (2002). Subduction zones. *Reviews of Geophysics* **40**, 1012, doi:10.1029/2001RG000108, 2002.
- Taylor, H. P., Jr (1980). The effects of assimilation of country rocks by magmas on $^{18}\text{O}/^{16}\text{O}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ systematics in igneous rocks. *Earth and Planetary Science Letters* **47**, 243–254.
- Taylor, S. R. & McLennan, S. M. (1985). *The Continental Crust: its Composition and Evolution*. Oxford: Blackwell Scientific, 312 p.
- Thibodeau, A. M., Killick, D. J., Ruiz, J., Chesley, J. T., Deagan, K., Cruxent, J. M. & Lyman, W. (2007). The strange case of the earliest silver extraction by European colonists in the New World. *Proceedings of the National Academy of Sciences of the USA* **104**, 3663–3666.
- Thomas, W. A. & Astini, R. A. (1996). The Argentine Precordillera: A traveler from the Ouachita embayment of North American Laurentia. *Science* **273**, 752–757.
- Tibaldi, A. M., Álvarez Valero, A. M., Otamendi, J. E. & Cristofolini, E. A. (2011). Formation of paired pelitic and gabbroic migmatites: an empirical investigation of the consistency of geothermometers, geobarometers, and pseudosections. *Lithos* **122**, 57–75.
- Tosdal, R. M. (1996). The Amazon–Laurentian connection as viewed from the Middle Proterozoic rocks in the central Andes, western Bolivia and northern Chile. *Tectonics* **15**, 827–842.
- Toselli, A. J., Durand, F. R., Rossi de Toselli, J. N. & Saavedra, J. (1996). Esquema de evolución geotectónica y magmática Eopaleozoica del sistema de Famatina y sectores de Sierras Pampeanas. *Proceedings XIII Congreso Geológico Argentino*, Vol. 5. Buenos Aires: Asociación Geológica Argentina, pp. 443–462.
- Turnbull, R., Weaver, S., Tulloch, A., Cole, J., Handler, M. & Ireland, T. (2010). Field and geochemical constraints on mafic–felsic interactions, and processes in high-level arc magma chambers: an example from the Halfmoon pluton, New Zealand. *Journal of Petrology* **51**, 1477–1505.
- Ulmer, P. (2001). Partial melting in the mantle wedge—the role of H_2O in the genesis of mantle-derived ‘arc-related’ magmas. *Physics of the Earth and Planetary Interiors* **127**, 215–232.
- Vielzeuf, D. & Holloway, J. R. (1988). Experimental determination of fluid-absent melting relations in the pelitic system: consequences for crustal differentiation. *Contributions to Mineralogy and Petrology* **98**, 257–276.
- Vujovich, G. I., Godeas, M., Marín, G. & Pezzutti, N. (1996). El complejo magmático de la Sierra de La Huerta, provincia de San Juan. *Proceedings XIII Congreso Geológico Argentino*, Vol. 3. Buenos Aires: Asociación Geológica Argentina, pp. 465–475.
- Wasserburg, G. J., Jacobsen, S. B., DePaolo, D. J., McCulloch, M. T. & Wen, T. (1981). Precise determination of Sm/Nd ratios, Sm and Nd isotopic abundances in standard solutions. *Geochimica et Cosmochimica Acta* **45**, 2311–2323.
- Watt, G. R. & Harley, S. L. (1993). Accessory phase controls on the geochemistry of crustal melts and restites produced during water-undersaturated partial melting. *Contributions to Mineralogy and Petrology* **114**, 550–566.
- Wiebe, R. A. (1993). The Pleasant Bay layered gabbro–diorite, Coastal Maine: ponding and crystallization of basaltic injections into a silicic magma chamber. *Journal of Petrology* **34**, 461–489.
- Wolf, M. B. & London, D. (1995). Incongruent dissolution of REE- and Sr-rich apatite in peraluminous granitic liquids: differential apatite, monazite, and xenotime solubilities during anatexis. *American Mineralogist* **80**, 765–775.
- Yáñez, G. A., Ranero, C. R., von Huene, R. & Díaz, J. (2001). Magnetic anomaly interpretation across the southern central Andes (32°–34°S): The role of the Juan Fernández Ridge in the late Tertiary evolution of the margin. *Journal of Geophysical Research* **106**, 6325–6345.
- Zimmermann, U. (2005). Provenance studies of very low- to low-grade metasedimentary rocks of the Puncoviscana complex, Northwest Argentina. In: Vaughan, A. P. M., Leat, P. T. & Pankhurst, R. J. (eds) *Terrane Processes at the Margins of Gondwana. Geological Society, London, Special Publications* **246**, 381–416.
- Zimmermann, U. & Bahlburg, H. (2003). Provenance analysis and tectonic setting of the Ordovician clastic deposits in the southern Puna Basin, NW Argentina. *Sedimentology* **50**, 1079–1104.