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Compositional variation during monogenetic volcano growth and its implications for magma supply to continental volcanic fields

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Abstract: Individual volcanoes of continental monogenetic volcanic fields are generally presumed to erupt single magma batches during brief eruptions. Nevertheless, in two unrelated volcanic fields (the Waipiata volcanic field, New Zealand, and the Miocene–Pliocene volcanic field in western Hungary), we have identified pronounced and systematic compositional differences among products of individual volcanoes. We infer that this indicates a two-stage process of magma supply for these volcanoes. Each volcano records: (1) intrusion of a basanitic parent magma to lower- to mid-crustal levels and its subsequent fractionation to form a tephritic residual melt; (2) subsequent transection of this reservoir by a second batch of basanitic melt, with tephrite rising to the surface at the head of the propagating basanite dyke. Eruption at the surface then yields initial tephrite, typically erupted as pyroclasts, followed by eruption and shallow intrusion of basanite from deeper in the dyke. By analogy with similar tephrite–basanite eruptions along rift zones of intraplate ocean-island volcanoes, we infer that fractionation to tephrite would have required decades to centuries. We conclude that the two studied continental monogenetic volcanic fields demonstrate a consistent history of early magmatic injections that fail to reach the surface, followed by capture and partial eruption of their evolved residues in the course of separate and significantly later injections of basanite that extend to the surface and erupt. This systematic behaviour probably reflects the difficulty of bringing small volumes of dense, primitive magma to the surface from mantle source regions. Ascent through continental crust is aided by the presence in the dyke head of buoyant tephrite captured during transection of the earlier-emplaced melt bodies.

Keywords: Otago, Pannonian Basin, magmas, composition, volcanism.

Monogenetic volcanoes are small and occur as scoria cones, tuff cones and rings, and maars; they form from single, typically brief eruptions (Walker 1993). Monogenetic volcanoes form in two distinct settings: (1) as isolated fields of volcanoes on continental lithosphere, ranging from thinned lithosphere (<30 km) resulting from stretching and extension (e.g. Ethiopia (Barberi & Varet 1970; Ebinger *et al*. 1993) and the Basin and Range province (Brandon & Goles 1995)) to normal or thick lithosphere (e.g. the San Francisco field (Conway *et al*. 1998) and Hopi Buttes (White 1991)); (2) as 'parasitic' vents along the rift zones or flanks of large polygenetic central volcanoes (e.g. Tolbachik (Doubik & Hill 1999), La Palma (Klügel *et al*. 1999) and Hawaii (Moore 1992)).

Some single eruptions forming monogenetic volcanoes atop large central volcanoes are known to have produced petrologically varied magmas (Klügel 1997, 1998; Klügel *et al*. 1999, 2000) that reflect the presence of magma reservoirs within the large volcano. Such variation has not been noted in single small-volume monogenetic volcanoes typical of continental fields, which are thought to lack stable magma storage zones. These volcanoes are formed by more or less direct eruption of magma from the mantle, with each volcano resulting from successful propagation of a small batch of magma to the surface along a new pathway (Spera 1984; Hasenaka & Carmichael 1985, 1987; Connor *et al*. 2000). In this paper, we present evidence for similar, single-eruption, compositional variation in products of two widely separated and unrelated continental monogenetic volcanic fields, and suggest that the 'plumbing systems' and magma supply dynamics for such fields may be more nuanced than hitherto appreciated.

Waipiata and western Hungarian volcanic fields

The Waipiata volcanic field occupies an area of *c*. 5000 km^2 onland in southeastern New Zealand (Fig. 1) and continues offshore over a similar, but very poorly constrained, area. It formed between *c*. 16 and 12 Ma (Coombs *et al*. 1986), on continental crust of >30 km thickness (Koons *et al*. 1999) during a period of mild crustal extension (King 2000). In this study, remnants of 55 volcanoes of the Waipiata volcanic field were examined (Németh 2001), some 75% of which comprise glassy pyroclastic rocks intruded by dykes and sills, and/or overlain by lavas.

An intracontinental Miocene–Pliocene volcanic field in western Hungary developed between 7.56 and 2.3 Ma (Balogh *et al*. 1986) across an area half the size of the Waipiata volcanic field (Fig. 1). There are two distinct parts to the western Hungarian volcanic field (Fig. 1), and these developed together in a tectonic regime that varied during that time from transtensional to transpressional (Fodor 1995). Volcanoes of the western Hungarian volcanic field also consistently comprise basal glassy pyroclastic units overlain by lavas that cap buttes (Németh & Martin 1999).

The pyroclastic rocks of each of the studied parts of the western Hungarian volcanic field contain various proportions of country rock clasts (Fig. 2), and apparently represent vent-filling

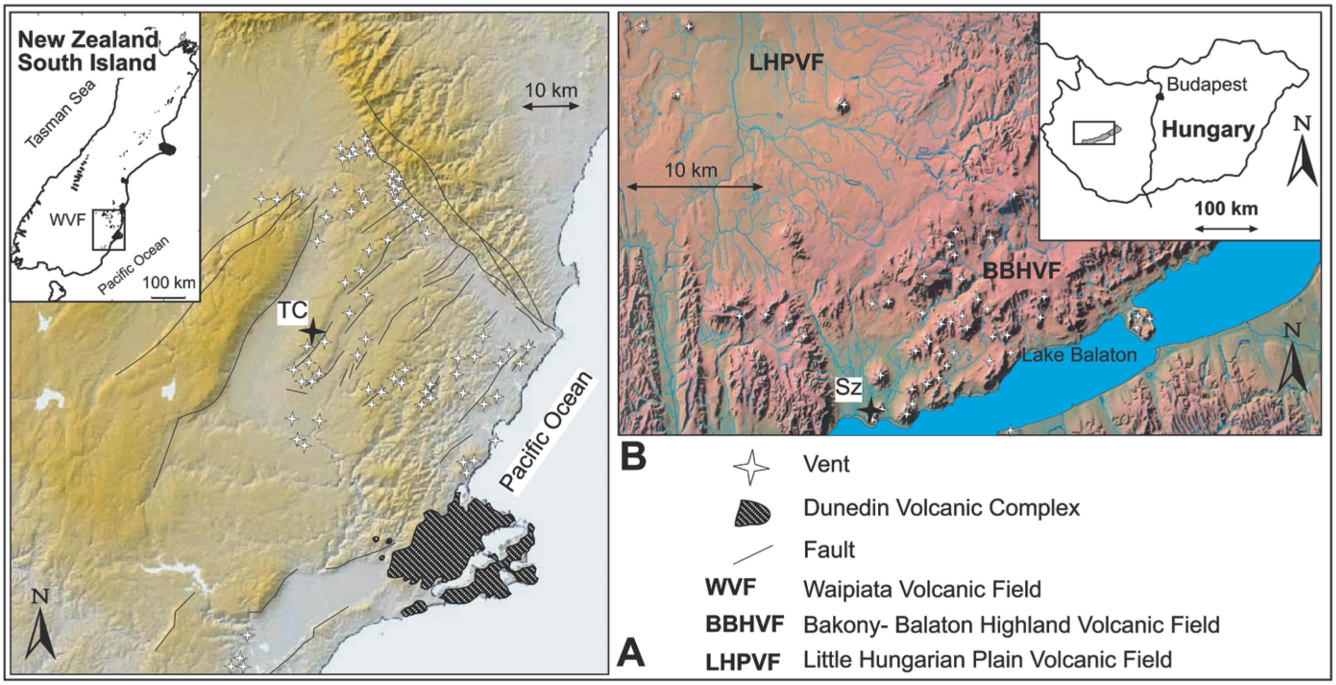

Fig. 1. Overview maps of the two studied volcanic fields. (**a**) Waipiata volcanic field in the South Island of New Zealand; (**b**) small-volume intracontinental volcanic fields in western Hungary. TC, 'The Crater'; Sz, Szigliget.

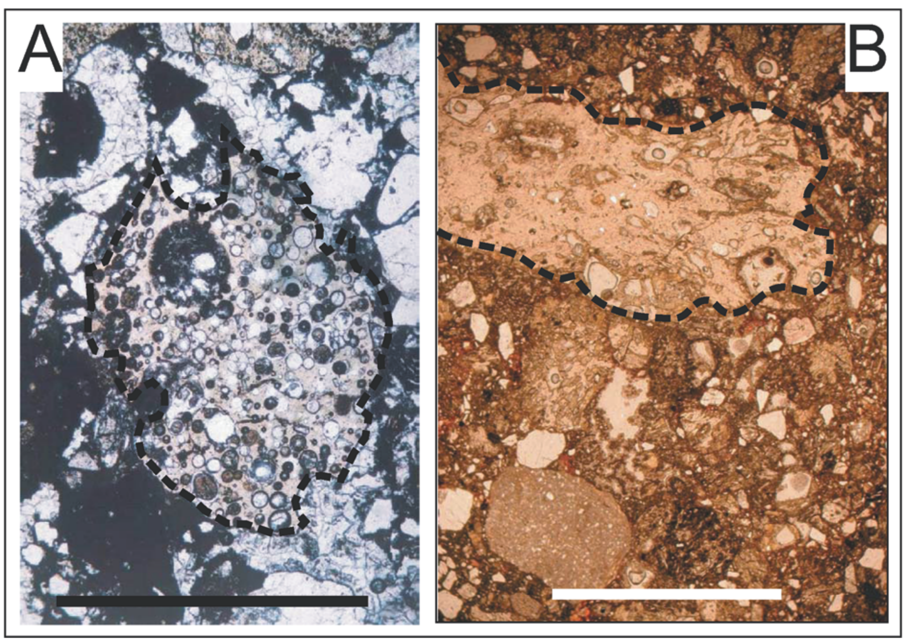

Fig. 2. Photomicrographs of representative volcanic glass shards from the studied volcanic fields. (**a**) 'The Crater'; (**b**) Szigliget. The moderately vesicular texture of the glass shards, and the paucity of microlites, should be noted. Scale bars represent 1 mm.

assemblages and locally preserved well-bedded tuff ring deposits. Dykes and lavas have subplanar to highly irregular, locally peperitic (Martin & Németh 2000), contacts with pyroclastic rocks, suggesting intrusion soon after emplacement of the tuffs and tuff breccias while they were still unconsolidated. The pyroclastic rocks typically have aphyric or sparsely feldspar-phyric juvenile clasts (Fig. 2), whereas the slightly younger dykes and lavas are characterized by abundant pyroxene ± kaersutite phenocrysts. There are abundant deep-seated cumulate fragments, 1–15 cm in size, in the topmost beds of pyroclastic deposits at some of the volcanoes.

Analytical methods and results

Samples of pyroclastic rocks were examined in thin section, and those containing remnants of fresh, isotropic glass were analysed by electron microprobe to obtain glass compositions. The juvenile clasts themselves are glassy or palagonitic, with variably abundant feldspar and pyroxene microlites as well as occasionally some larger plagioclase phenocrysts of andesine composition (An 50%). Vesicularity estimated from thin section ranges from a few percent to perhaps up to 50%, with typical clasts of moderate vesicularity (<25%) (Fig. 2). Vesicles are small, generally circular to elliptical in section (Fig. 2), and separated by thick glass walls; such pyroclasts are generally interpreted to result from phreatomagmatic eruption processes (Heiken 1974; Fisher & Schmincke 1984; White & Houghton 2000).

Results of microprobe analyses of pyroclast glass were compared with existing whole-rock XRF analyses of dykes and lavas from the same volcanoes (Embey-Isztin 1993; Donnelly 1996; Figs 3 and 4, Table 1). Deposits of 'The Crater' typify features of single volcanoes in the Waipiata volcanic field (Németh 2000), and have pyroclastic rocks with less MgO, less FeO_t, and more K_2O than dykes and lavas from the same vent (Fig. 4a and b; Table 1). CIPW norm calculations indicate that the pyroclast glasses are tephrite (regardless of Fe_2O_3 to FeO ratios chosen), and lavas are basanite (Table 1). A similar degree of composi-

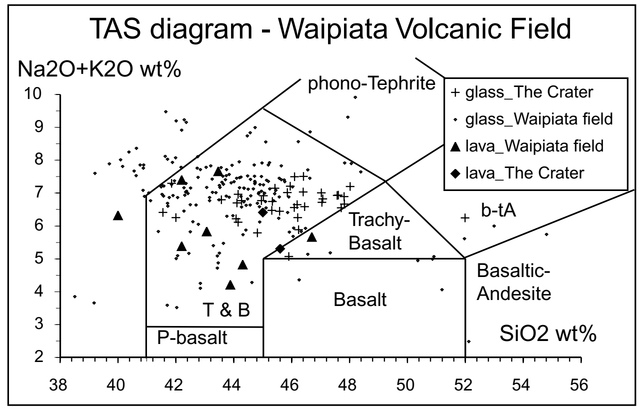

Fig. 3. Total alkali v. silica (TAS) diagram for volcanic glass shards and lava flows from throughout the Waipiata field. Because TAS differences within single volcanoes are modest, and vary across different ranges, the fields for glass shard and lava compositions are not distinct in this diagram.

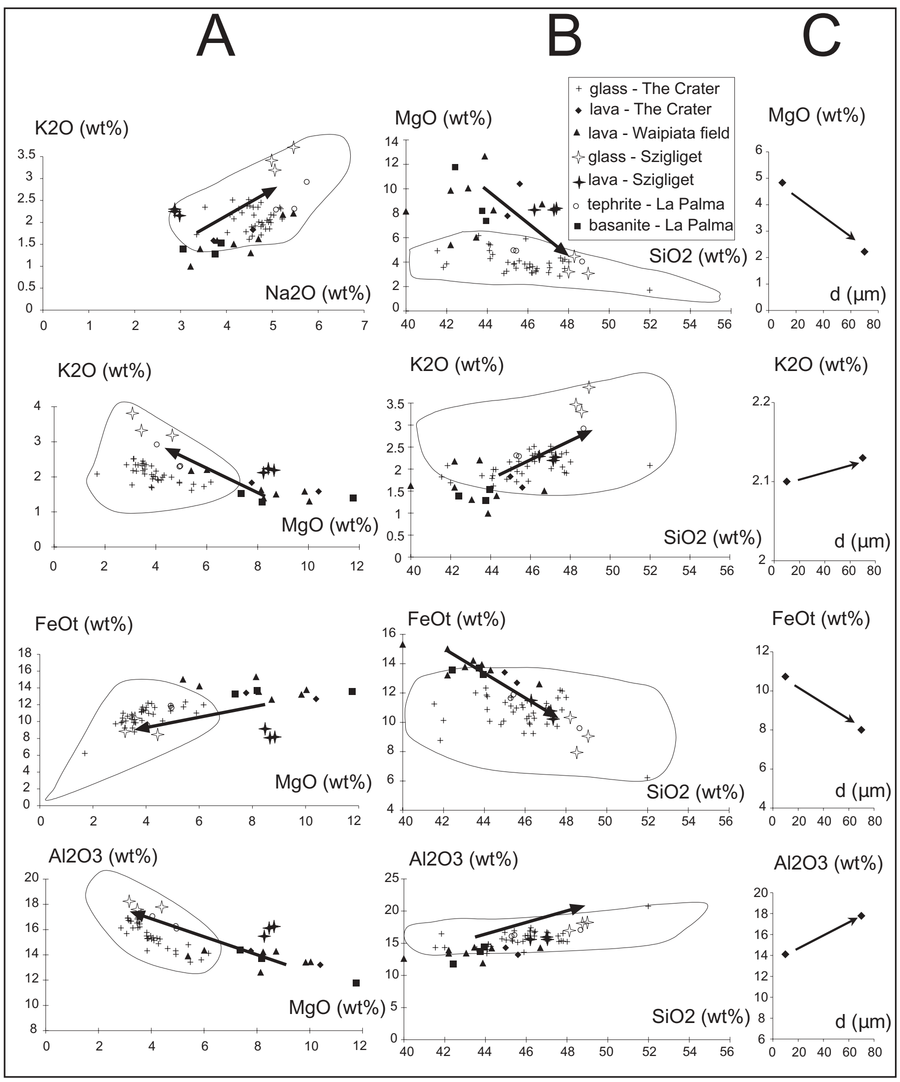

Fig. 4. Major element discrimination diagrams for volcanic glass shards and lava flows from individual volcanoes in the Waipiata volcanic field (WVF) and western Hungary. From the Waipiata volcanic field, data from a single volcanic remnant are presented (glass and lava from 'The Crater'). For comparison with other lava flow compositions, each diagram shows the data field for all Waipiata lava flows (lava - Waipiata field). From western Hungary, data from a similar vent remnant are shown (glass - Szigliget; lava - Szigliget). The outlined field represents all glass data from both the Waipiata field and western Hungarian fields. In (**a**), Na_2O v. K_2O and MgO v. K_2O, FeO_{total} and Al_2O_3 are plotted. (Note the separation of data representing volcanic glass shards v. lava flows, suggesting fractional crystallization with separation of olivine, minor clinopyroxene and magnetite.) In (**b**), representative Harker diagrams (SiO_2 v. MgO, K_2O, FeO_t and Al_2O_3) are plotted, showing crystal fractionation trends between volcanic glass shards and lava flows. In each frame of (**a**) and (**b**), data from the La Palma eruption are also plotted to illustrate the similar fractionation trend for La Palma, the Waipiata volcanic field and western Hungary. (**c**) shows the effect on 'apparent' composition of using larger electron beam diameter (d; abscissa in µm) during electron microprobe measurements of glass shards. It should be noted that increasing beam diameter leads to lower measured values of MgO and FeO_t, thus yielding lower values of normative olivine in CIPW calculation. We used the smaller of the beam diameters, thus conservatively biasing glass measurements to less tephritic compositions.

Table 1. *Composition of volcanic glass shards from 'The Crater', Waipiata volcanic field, compared with XRF data from associated lava flows (first two columns)*

Sample:	CRb-272	CRb2-307	CR2-320	CR15-330	CR23-334	CR22-342	CR8-dyke	CR3-dyke
SiO_2	45.42	47.02	46.41	47.05	47.82	48.02	45	45.6
Al_2O_3	16.94	16.72	16.13	15.3	15.31	15.24	2.46	2.15
TiO_2	2.85	2.7	2.47	3.09	2.93	2.86	14.3	13.2
FeO	10.69	10.98	10.63	11.28	12.13	11.7	13.4	12.69
MnO	0.11	0.13	0.18	0.08	0.16	0.29	0.22	0.19
MgO	3.5	3.14	3.3	3.91	4.22	4.01	7.77	10.4
CaO	7.78	8.33	8.38	9.64	9.24	9.37	8.12	8.16
Na_2O	4.97	4.49	5.17	4.71	4.75	5.03	4.58	3.72
K_2O	2.17	2.51	2.34	2.01	1.91	2.16	1.83	1.58
$Na_2O + K_2O$	7.14	7	7.51	6.71	6.66	7.2	6.41	5.3
Total	94.43	96.02	95.01	97.06	98.48	98.68	99.55	99.38
SiO_2	48.1	48.97	48.85	48.48	48.56	48.66	45	45.6
TiO_2	3.02	2.81	2.6	3.18	2.98	2.9	2.46	2.15
Al_2O_3	17.94	17.41	16.98	15.76	15.55	15.44	14.3	13.2
FeO	11.32	11.44	11.18	11.62	12.32	11.86	13.4	12.69
MnO	0.12	0.14	0.19	0.08	0.16	0.29	0.22	0.19
MgO	3.71	3.27	3.47	4.03	4.29	4.06	7.77	10.4
CaO	8.24	8.68	8.82	9.93	9.38	9.5	8.12	8.16
Na_2O	5.26	4.68	5.44	4.85	4.82	5.1	4.58	3.72
K_2O	2.3	2.61	2.46	2.07	1.94	2.19	1.83	1.58
$Na_2O + K_2O$	7.56	7.29	7.9	6.92	6.76	7.29	6.41	5.3
Total	100	100	100	100	100	100	99.55	99.38
Mg-number	0.43	0.4	0.42	0.46	0.45	0.44	0.57	0.65
ol	4.6	3.53	2.62	1.77	3.78	2.49	16.97	21.64
or	13.58	15.45	14.55	12.24	11.46	12.94	10.81	9.34
ab	19.7	20.31	18.14	18.49	20.75	18.32	16.67	18.7
ne	13.45	10.43	15.11	12.23	10.87	13.44	11.96	6.92
an	18.54	18.8	14.62	15.11	15.04	12.79	13.06	14.65
D.I.	46.74	46.19	47.81	42.96	43.08	44.7	39.45	34.96

Analyses were performed by electron microprobe (JEOL 8600 Superprobe at the Geology Department, Otago University), using 15 kV acceleration voltage, 10–20 µm electron beam diameter on polished thin sections and OXIDE9 standard. First set of data represents raw chemical composition; second set represents data normalized to 100%. CIPW norms were calculated for 0.3 Fe_2O_3 to FeO ratios. Mg-number, magnesium number; ol, normative olivine; or, normative orthoclase; ab, normative albite; ne, normative nepheline; an, normative anorthite; D.I., differentiation index. The total values varied between 94 and 99% on measured volcanic glass shards from lapilli tuffs and tuffs. This variation seems only to manifest in the SiO_2 content of the measured glass shard indicating SiO_2 loss of the glass shard. However the deeper reason of such effect needs further investigation. For the comparison of various data sets a uniform normalization has been applied on each glass shard data. Glass shard data over 97% of total values is generally considered a very good measurement in microprobe studies and perhaps a 100% normalization does not give significantly new results in calculation of CIPW norms.

tional variation between pyroclast glasses of phreatomagmatic pyroclastic rocks and intruding dykes has been recognized from the vent remnant of Szigliget, western Hungary (Németh *et al*. 2000) (Fig. 4; Table 1).

The studied pyroclasts contain modest proportions of microlites and few or no phenocrysts, which allows comparison with whole-rock XRF results. Use of a wide electron beam (>50 µm) captures both glass and local microlite compositions, and yields similar SiO_2 values coupled with increased total alkalis (Fig. 4c) and higher totals overall. We infer that these differences result from capture of a mixture of feldspar and/or clinopyroxene microlites and glass within the larger beam, and these results reveal rocks that are even more clearly tephritic to phonotephritic, supporting our interpretation of a substantial difference in composition between the pyroclastic rocks and subsequently emplaced dykes and lavas. Modelling of major element trends indicates that the tephrite could have been formed from a basanite parent by simple crystal fractionation dominated by removal of 10–20% olivine plus minor (<5%) titanomagnetite and/or clinopyroxene. All of these phases are present in the basanite lavas (Coombs & Reay 1986; Coombs *et al*. 1986), and are plausible fractionating phases to produce the tephrites.

Discussion

The eruption of both tephrite and basanite during single eruptions of a continental monogenetic volcanic field is unexpected, because each volcano and eruption occupies a new site, each volcano is too short lived to develop its own magma chamber in which differentiation might take place, and the time necessary for differentiation of a basanite parent to tephrite far exceeds the duration of an eruption or of magma transport from source to surface (Table 2).

As indicated by Table 2, differentiation of basanite parent magma to tephrite is, however, commonplace in polyphase alkaline volcanoes. In the Eifel volcanic field, the Rothenberg volcano, which had an original edifice volume of some 0.05 km^3 (Houghton & Schmincke 1989) records a number of eruptions that tapped progressively a zoned magma chamber developed at upper-mantle and/or mid-crustal levels (30–20 km; Schmincke 1977a, 1977b; Duda & Schmincke 1978; Schmincke *et al*. 1983). Such repeatedly tapped, zoned, magma columns are believed to form over hundreds (to thousands?) of years (Hawkesworth *et al*. 2000) from a mafic parent magma. In contrast, individual eruptions playing a part in the construction of Rothenberg volcano probably had durations of days to months on the basis of its sedimentary record (Houghton & Schmincke 1986; Houghton & Schmincke 1989).

In 1949, both basanite and tephrite were ejected during a single, 5 week long eruption that formed monogenetic cones and lava flows on the volcanic island of La Palma, Canary Islands (Klügel *et al*. 1999; White & Schmincke 1999). In this case, the tephrite is inferred to have formed from a basanite parent magma that was injected into the island volcano's deep rift zone in the

Table 2. *Summary comparison of the time scales of magma rise speed (time), eruption duration, and magma differentiation time, with notes on the calculation methods and locations*

Time scales	Notes	Reference
Magma rise speed and/or time	*Basis of calculation*	
Few hours to 4 days from lower crust to the surface	La Palma, Canary Island; Fe–Mg interdiffusion at xenolith fractures	Klügel 1998
0.1–10 m s^{-1} from source	Xenolith settling velocity (Stoke's law) applied to varying fluid behaviour	Sparks *et al*. 1977
36 h from mantle to Moho (0.1 m s^{-1}) and 1.5 h from Moho to surface (5 m s^{-1})	Nógrad–Gömör volcanic field, Hungary alkali basalt hosted xenoliths; fluid inclusion study	Szabó & Bodnar 1996
0.1 m s^{-1} from mantle to Moho then 10 m s^{-1} from Moho to surface	CO_2 inclusion study of peridotite lherzolite nodules	Spera 1984
Eruption duration	*Location of eruption and its type*	
c. 1300 days (14 November 1963–1967) with distinct active periods lasting for days from various vent sites	Surtsey, Iceland; surtseyan style subaqueous to emergent	Thorarinsson 1967
10–20 h (1–2 January 1996)	Karymskoye lake, Kamchatka; surtseyan style sub-lacustrine	Belousov & Belousova 2001
10 days (30 March 1977–09 April 1977), 147 h duration for small-volume lava extrusion	Ukinrek, Alaska; two maar-forming phreatomagmatic, terrestrial eruptions	Self *et al*. 1980; Kienle *et al*. 1980
3 days (1965)	Taal, Philippines; phreatomagmatic eruption	Moore *et al*. 1966
100 days (1955) with a more active period in the beginning lasting for days, then activity changed to fumarole steam	Nilahue, Chile; maar-forming eruption with repeated explosive activity	Müller & Veyl 1956
38 days (1949)	La Palma, Canary Islands; phreatomagmatic explosive eruptions and lava flows	Klügel *et al*. 1999, 2000
Magma differentiation time	*Location and basis of calculation*	
Basalt through andesite to dacite, 1–4 ka	Tonga/Kermadec; U–Th–Ra isotope series	Turner *et al*. 2000
Basalt to rhyolites, 5–6 ka		
Basanite to phonolite via 50% crystal fractionation, 10 ka	Tenerife, Canary Islands; U–Th–Ra isotope series	Hawkesworth *et al*. 2000
Phonolite to more evolved phonolite via 50% crystal fractionation, a few hundred years	Tenerife, Canary Islands; U–Th–Ra isotope series	Hawkesworth *et al*. 2000
Hawaiite to mugearite, $<$200 years	Mt Etna, Italy; U–Th–Ra isotope series	Condomines *et al*. 1995
Basanite to phonolite, 100 ka	Laacher See, Germany; U–Th–Ra isotope series	Bourdon *et al*. 1994
Alkali basalt to trachyte, 90 ka	São Miguel, Azores; U–Th–Ra isotope series	Widom *et al*. 1992
Basanite to tephrite, phonotephrite, $>$13 years (small, shallow batches)	La Palma, Canary Islands; historic account (13 years)	Klügel *et al*. 2000
Basanite to tephrite in mantle, a century or more	Tephrites not petrologically derived from historically erupted magmas	
Basanite to phonolite, $<$150 ka	Mt Erebus, Antarctica; U–Th–Ra isotope series	Reagan *et al*. 1992

upper mantle (200–350 MPa, *c*. 15 km, Moho at ocean-island setting) no later than 1936, the time of the most recent preceding eruption (Klügel *et al*. 2000); this eruption also emplaced some magma higher into the volcanic edifice. After at least 13 years of differentiation (probably more) at depth to produce tephrite, a new injection of basanite transected the deep reservoir, following to the surface tephrite magma that collected additional, phonotephritic magma, from small high-level reservoirs en route (Klügel *et al*. 2000).

A key feature of tephrite–basanite successions, then, is that they reflect evolution of a basanite parent magma over time. When basanite and tephrite are erupted together, as at La Palma in 1949, eruption of magmas from different batches, with different crystal fractionation histories, is inferred (Klügel *et al*. 2000). In the Waipiata and western Hungarian volcanic fields, basanite and tephrite occur together in volcanoes that were neither large and long lived enough to have developed their own differentiated magma columns (compare Rothenberg) nor associated with the magma storage system of a larger, underlying volcano (compare La Palma). How, then, can we interpret their co-occurrence?

The simplest explanation is that the general assumption that continental monogenetic volcanic fields lack significant shallow magma storage zones (Connor & Conway 2000) must be in error. We infer for the Waipiata field that each 'successful' eruption that produces a monogenetic volcano represents at least two injections of magma from the source region (mantle), with the earlier of these injections stalling and differentiating at regions such as rheological and/or density boundaries of the lithosphere (e.g. upper-mantle–lower-crustal and/or lower–upper-crustal levels). Each volcano records one eruption, but also carries the residual products of one or more 'failed eruptions' that leave magma pockets ponded in temporary storage places.

The tephrite–basanite association occurs at many volcanoes of the Waipiata and western Hungarian volcanic fields, and we consider two scenarios to explain this (Fig. 5).

In the first scenario, the sites of the Waipiata and western Hungarian volcanic fields may have initially offered unfavourable stress conditions for transport of small magma batches from mantle source regions to the surface, and injected magma was consistently captured and stored at rheological and/or density boundaries of the lithosphere (Walker 1989; Head & Wilson 1992; Rubin 1995; Watanabe *et al*. 1999). Later, stress conditions became more favourable for magma transport, and injected batches consistently reached the surface, transecting bodies of evolved magma along the way.

In this scenario, thermal decay times require that there was an abrupt switch from one stress regime to another, and that all volcanoes within each field would be of similar ages. Although dating within both fields has been limited, there is a range of some millions of years among dated lavas of each field. This is too long to preserve small bodies of melt in the crust, and suggests that the scenario is appropriate to neither the Waipiata nor the western Hungarian field.

In the second scenario, throughout the area of the Waipiata and western Hungarian volcanic fields, magma was injected to form small (kilometre-scale) mid-crustal magma bodies, then

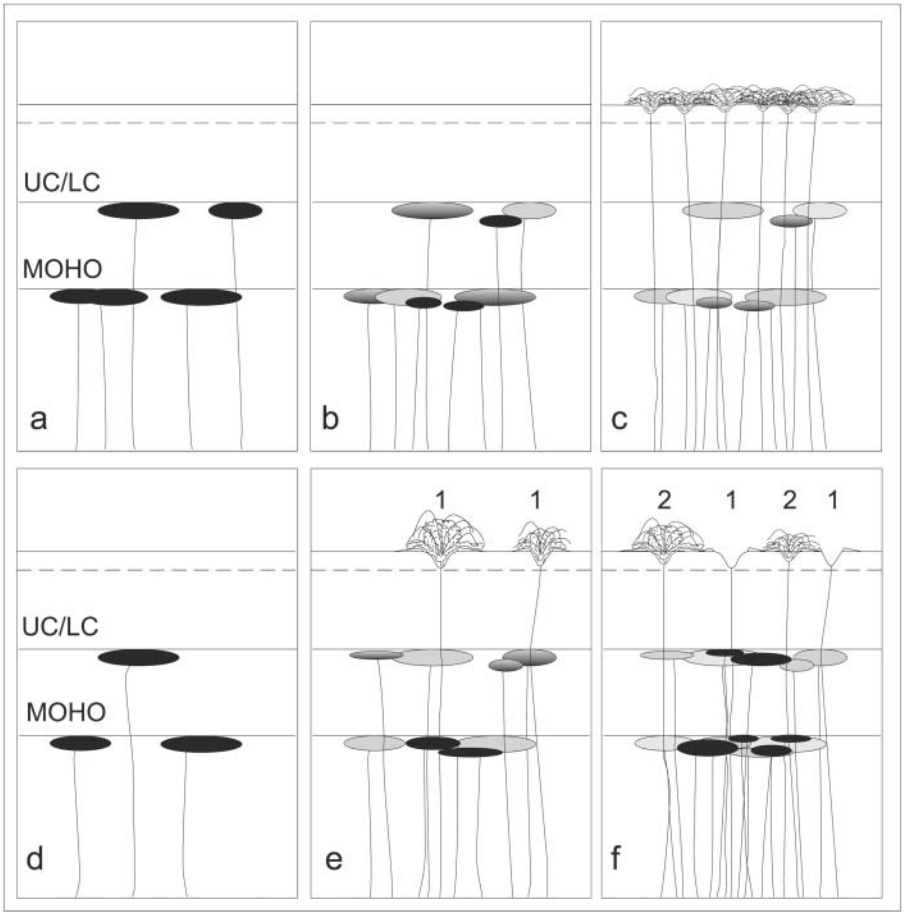

Fig. 5. Simple model for a monogenetic volcanic field magmatic plumbing system. (**a**)–(**c**) illustrate a scenario with initial magmatic underplating, during which magmas are not erupted to the surface because of unfavourable stress regime, and individual basanite melt batches are instead emplaced (**a**) at the level of the Moho, or at other density and/or rheology traps in the crust such as the brittle–ductile boundary between upper and lower crust (UC/LC), where they differentiate over centuries to millennia (**b**). When the stress regime becomes suitable (**c**), a second generation of basanite melt rises to the surface, collecting melt from the evolved crustal accumulations on its way to eruption. However, volcanism at the Waipiata volcanic field was long lived, with volcanoes of many different ages, suggesting that this is not an appropriate scenario. (**d**)–(**f**) show magma repeatedly injected to Moho or upper—lower-crustal levels. At these crustal or base of crust storage sites magma evolves to tephrite or phonotephrite. When a later basanite injection propagates into one of these magma bodies, the storage site is fractured and the more buoyant tephrite advances ahead of the ascending basanite to begin an eruption at the surface. (See text for further discussion.)

injected again from the same or newly formed adjacent source areas. When evolved magma bodies were encountered by later injections, the additional buoyancy provided by the entrained tephritic melts would have favoured propagation of these dykes to the surface. The degassing and expulsion of volatile phases upon mixing of two different magmas (the stalled and the newly intruded) would also have facilitated the propagation of dykes (Rubin 1995), as would the increased hydrostatic pressure developed in the crustal reservoir when encountered by newly intruding basanite dyke(s). In this scenario, basanite magma would rarely reach the surface unless an evolved, tephritic reservoir was encountered at mid-crustal depths, although exceptionally large batches of basanite might extend to the surface without a boost from tephrite buoyancy and/or increased hydrostatic pressure as a result of melt mixing. Diachronous eruptions, as indicated by existing dates, are more easily accommodated in this scenario. It also bears a strong physical resemblance to the model for the 1949 La Palma eruption, with the main distinction being that storage is not localized within the rift zones of a large volcanic system.

Conclusion

Continental monogenetic volcanic fields are subject to the same physical constraints as other volcanic systems. Dense mantle-derived magmas are prone to ponding near their levels of neutral buoyancy, at depths of 25–30 km, at the upper-mantle–continental crust boundary and/or in rheological and density contrast zones between the brittle–ductile transition at mid-crustal levels (Ryan 1987; Walker 1989; Lister 1991; Lister & Kerr 1991; Watanabe *et al*. 1999). Eruption of such magmas in small volumes requires substantial injected volumes of which only a small proportion reaches the surface, and/or specific stress conditions within the transected lithosphere (Watanabe *et al*. 1999). We have identified mechanisms that may help bring small volumes of magma to the surface, and suggest that where these mechanisms operate there should be a systematic presence of early erupted evolved rocks. These are commonly present as pyroclastic deposits rather than lava because their early arrival favours interaction with groundwater to produce phreatomagmatic eruptions, and/or because they have higher volatile contents. We are unable to assess the universality of this behaviour, because pyroclastic deposits have not been sampling targets for most petrological investigations. We find, however, similar compositional bimodality among early pyroclastic and subsequent effusive products associated with two widely spaced and unrelated volcanic fields, each of which is a typical intracontinental, monogenetic field, and each of which was at the time of volcanism exposed to only weak stresses related to nearby transcurrent plate boundaries. We conclude that similar compositional bimodality among eruptive products of intracontinental volcanoes in volcanic fields may be the rule, rather than the exception.

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