Magma mixing in undersaturated alkaline volcanics, Cantal, Massif Central, France

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Abstract

Several examples of magma mixing exist within the undersaturated alkaline magma series of the Tertiary/Quaternary volcanics in the French Massif Central. This study describes magma mixing in the Puy Griou/Griounot area of the Cantal volcano (10–3 Ma). Petrographic evidence for injection of blebs of basic magma into phonolitic host magmas is abundant (cauliform inclusions, liquid–liquid contacts, vesiculation and chilling). Compositions of the inclusions are basic tephrite, whereas the hosts are miaskitic phonolites. Petrographic examination reveals the presence of disequilibrium mineralogical features (e.g. Mg-rich olivine in phonolites) and strong zoning in many clinopyroxenes. Transfer of phenocrysts between basic inclusions and phonolite hosts was common, and can be seen clearly in the wide range of compositions of clinopyroxene. Hornblende, magnetite and olivine were also transferred from inclusions to host.

Sr and Nd isotope data indicate that, unlike most other fractionated magmas of the region, phonolites which show evidence for magma mixing are uncontaminated by the continental crust and have isotopic ratios similar to local primitive basic magmas. This leads to the suggestion that the magma mixing event took place at great depth, rather than being a high-level phenomenon. The phonolites were thus generated by high-pressure fractional crystallisation of an earlier basanitic or tephritic parent, perhaps at upper-mantle depths. This conclusion may explain why some phonolites elsewhere in the world have entrained spinel lherzolite mantle xenoliths.

KEYWORDS: magma mixing, alkaline volcanics, Massif Central, France.

Introduction

IT is now widely accepted that magma mixing is a major igneous process and is responsible for some of the variability observed in volcanic and plutonic rocks. Application of fluid dynamic concepts to magma chamber processes has shown that magma mixing is probably a common occurrence if a fractionating magma is tapped and replenished (Huppert and Sparks, 1980; Huppert et al., 1982). Often magma mixing has been considered to be the trigger for major volcanic eruptions. Field, petrological and geochemical evidence indicates the existence of magma mixing in different magma series: in tholeiitic volcanics (e.g. Wright and Fiske, 1971; Anderson and Wright, 1972; Wright, 1973; Donaldson and Brown, 1977; Dungan and Rhodes, 1978; Mørk, 1984), calc-alkaline volcanics (e.g. Sakuyama, 1979; Gerlach and Grove, 1982) and alkaline volcanics (e.g. Brooks and Printzlau, 1978; Barton et al., 1982; Worner and Wright, 1984; Wolff, 1985). This study presents evidence for magma mixing in undersaturated alkaline volcanics (basaltic tephrite–phonolite) of the French Massif Central.

Characteristic physical features of partially mixed magmas include their macroscopic heterogeneity, e.g. disseminated globules, emulsions, banding, cauliform 'bombs' of one lava type within another of contrasting composition. Further evidence includes the preservation of the appearance of liquid–liquid contacts after chilling, the transfer of phenocrysts from host to inclusion and vice versa and the consequent disequilibrium textural features of resorption and reverse zoning.

Mixing commonly occurs between two magmas of contrasting compositions related by crystal fractionation. In rare cases more than two compositional end-members are involved or one end-member is already a mixed magma (e.g. Wager et al., 1965). In all cases, if complete homogenisation occurs, the resulting mixture will be a hybrid magma in which the only evidence of mixing is mineralogical disequilibrium (e.g. Brooks and

Printzlau, 1978). Such a hybrid magma will have a chemical composition which plots on a tie-line between the two end-members in any two element plot and which will therefore differ in composition from the normal intermediate products of fractional crystallisation, especially if such fractionation trends are strongly curved.

Geological setting

The alkaline volcanic province of the French Massif Central is dominated by two late Tertiary central volcanoes, Cantal (10–3 Ma) and Mont Dore (2.5–0.25 Ma). In these volcanoes, two main magma series have been recognised (Maury and Varet, 1980): an alkali olivine-basalt–trachyandesite–trachyte–rhyolite series which tends to silica oversaturation and an undersaturated basanite-tephrite–phonolite series. A review of igneous activity in the region is given by Downes (1987).

The existence of mixed magmas in the saturated magma series of the Massif Central has been well documented (e.g. Gourgaud and Maury, 1984; Gourgaud and Camus, 1984). Examples of mixing between basalt and hawaiite, hawaiite and trachyandesites and between trachyandesite and rhyolite magmas are well-known from the Mont Dore volcano and the Chaine des Puys. It is considered that hybridisation has given rise to much of the diversity of the trachyandesites of Mont Dore (Gourgaud and Maury, 1984). This study describes magma mixing between members of the undersaturated magma series of Cantal. Similar mixed magmas have been reported from Mont Dore (Mossand, 1983).

The phonolite domes of Puy Griou (1694 m) and Griounot (1514 m) are situated in the centre of the Cantal volcano between the valleys of the Jordanne and Cere rivers. K–Ar age dates indicate that they were extruded 6.5–7.0 Ma ago (Baubron and Demange, 1977), post-dating the collapse of the central caldera (7.5 Ma) and are cut by small basic intrusions (Demange, 1974) which may be related to the final phase of magmatic activity of Cantal in which basaltic and basanitic lava flows were erupted (5–3 Ma ago). Puy Griou and Griounot are composed of porphyritic miaskitic (nonperalkaline) phonolite which is chemically the least-evolved phonolitic composition observed among Cantal volcanics, having both low total alkalis (11.8–12.6 wt. %) and low SiO_2 (57.5 wt. %); Downes, 1983. Although field relationships between the two domes are unclear (Demange, 1974), the phonolites are chemically and mineralogically similar and may form a single unit with two sites of extrusion.

Petrography of Puy Griou and Griounot mixed magmas

Small sub-spherical inclusions of dark grey material within the white weathering phonolite are frequently observed in both domes and are particularly well-developed on the north side of the main peak of Puy Griou. These inclusions range from 1 to 15 cm in diameter and have lobate margins; those found on Griounot contain amygdales with an infilling of calcite (Fig. 1a). The dark inclusions do not show any chilled margin but, in some samples, the phonolitic host chills against the inclusions. In many cases, the inclusions themselves contain blebs of the light silvery-grey phonolitic material with acicular quench crystals (Fig. 1b).

The dark inclusions contain abundant phenocrysts of hornblende, clinopyroxene, Fe–Ti oxide, rare olivine, and a sodalite-group feldspathoid. Glomeroporphyritic aggregates are common.
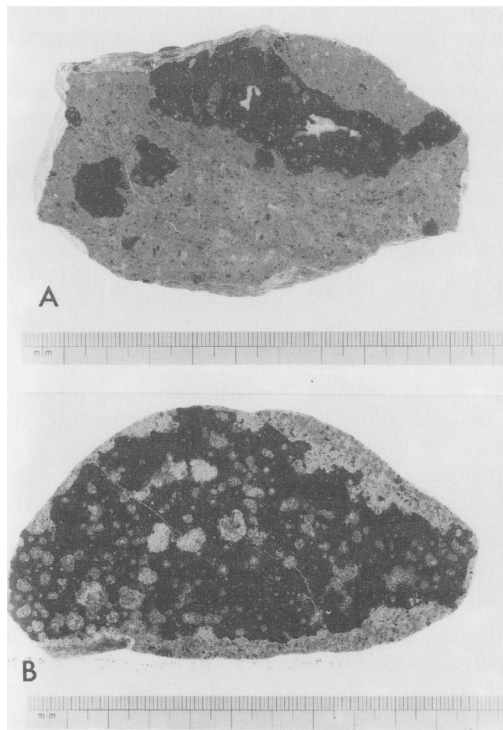

FIG. 1. Photographs of hand-specimens of mixed magmas. A. From Griounot: dark inclusions are basic tephrite, with deformed amygdales containing calcite; light grey host is phonolite. B. From Puy Griou: dark cauliform inclusion of basic tephrite forming an emulsion with phonolite.

Hornblende also occurs as small elongated quench crystals and forms the groundmass with plagioclase. Clinopyroxenes are frequently strongly zoned, and both green and pale brown varieties are observed. An emulsion of rounded blebs of phonolite showing chill textures is often found in the basic inclusions. Phonolite may be flow-banded around the basic inclusions. Rarely, a narrow intermediate zone is developed between the two contrasting compositional types. Fig. 2a and b shows the typical appearance of examples of the basic inclusions in the phonolite host.
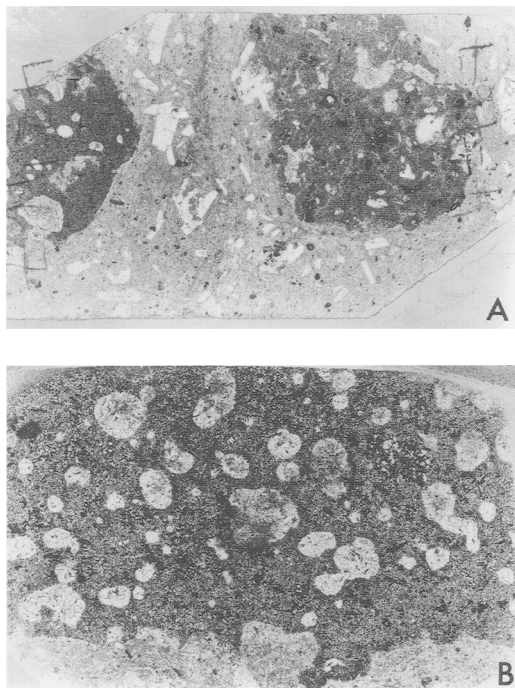

FIG. 2. Thin sections of mixed magmas from Puy Griou. A. Dark areas are basic inclusions; light areas are phonolite host. B. Intimate emulsion between phonolite and basic inclusion.

The phonolite contains sparse phenocrysts of sanidine and bright green clinopyroxene, often in glomeroporphyritic aggregates. Sodic plagioclase occurs as rare phenocrysts. Brown clinopyroxene, olivine, a sodalite-group feldspathoid, hornblende, Fe-Ti oxide, titanite, and biotite are present as microphenocrysts, with apatite and rare zircon as accessory minerals. The groundmass consists of plagioclase and alkali feldspar laths with a strong trachytic texture.

Boundaries between the two compositions are always smooth and rounded. The contacts and the chilling are consistent with the coexistence of two former liquids. Many of the phenocrysts in the phonolite seem to have been out of equilibrium with the surrounding groundmass; they show reaction rims and zoning and sometimes possess dark mantles which are similar in appearance to the groundmass of the basic inclusions.

Bulk geochemistry of end members

Using a defocused electron beam, an estimate was obtained of the chemical composition of the quenched facies of the phonolite host and the basic inclusions of both domes. These results do not include H_2O or P_2O_5 concentrations and hence are only partial analyses. Results are presented in Table 1 and Fig. 3, and are compared with XRF analyses of volcanic rocks of the undersaturated series of Cantal (Downes, 1983, 1987). Iron is reported as wt. % FeO in microprobe data (Table 1), but has been recalculated to Fe_2O_3 for comparison with XRF data (Fig. 3). The microprobe analysis of the Puy Griou host and the bulk XRF analysis of the same extrusion are closely comparable (Table 1), indicating that the analyses are valid. The dark inclusions in both domes are basic tephrite in composition (50–52 wt. % SiO_2), similar to lavas from Suc Gros and Puy Niermont, 8 km to the north. Quenched phonolitic hosts are similar to the least evolved phonolites of the region with between 57 and 59 wt. % SiO_2, and low Na_2O contents.

Comparison of the averages of the two phonolitic domes indicates that they are similar in composition, but that the Puy Griou phonolite is less siliceous than Griounot and has higher concentrations of TiO_2, MgO, CaO, and lower Na_2O and K_2O. The basic tephritic inclusions in both domes are also of broadly similar composition, but again the tephrite of Puy Griou is slightly less differentiated than that of Griounot, having higher concentrations of TiO_2, MgO, CaO and lower SiO_2, Na_2O and K_2O. Fig. 3 shows that the compositions of the phonolites and inclusions fall within the range of volcanic rocks of the undersaturated series of Cantal. Although the chemical trends seen in major element variation diagrams (Fig. 3) are nearly linear, slight departures from linearity exist in MgO–SiO_2, TiO_2–SiO_2 and K_2O–SiO_2 plots. Downes (1984) considered this magmatic series to be related by fractional crystallisation of olivine, clinopyroxene, magnetite, and apatite. Plagioclase and alkali feldspar become important fractionating phases only in the evolved phonolites, which develop strong

TABLE 1. Partial chemical analyses of the quenched facies of the end-members of magma mixing in Puy Griou and Griounot. Analyses 1-15 were obtained using a defocused electron-microprobe beam. XRF analysis (A) is of a basic tephrite from Puy Niermont, Cantal (Downes 1983); XRF analysis (B) is of the Griou phonolite (with no apparent basic inclusions) (Downes 1983). Loss on ignition and P_2O_5 not included; all Fe expressed as FeO.

BASIC END-MEMBER (GRIOU)

	1	2	3	4	Av 1-4	XRF(A)
SiO_2	50.07	49.94	51.86	48.63	50.13	51.84
TiO_2	2.31	2.74	2.35	2.51	2.48	2.07
Al_2O_3	18.86	18.47	18.41	18.30	18.51	17.62
FeO	5.62	7.67	7.22	7.50	7.00	6.97
MnO	0.10	0.18	0.16	0.14	0.15	0.20
MgO	2.83	3.54	3.08	3.37	3.21	2.78
CaO	7.43	8.04	6.41	7.40	7.32	7.18
Na_2O	7.27	6.37	6.77	7.53	6.99	5.27
K_2O	2.34	1.92	2.88	2.49	2.41	3.42
Total	97.17	99.10	99.39	98.35	98.2	97.89

PHONOLITE HOST (GRIOU)

	5	6	7	8	Av 5-8	XRF(B)
SiO_2	54.61	58.74	59.69	56.74	57.45	57.38
TiO_2	0.65	0.40	0.41	0.53	0.50	0.64
Al_2O_3	22.06	18.91	20.00	18.48	19.86	19.24
FeO	1.58	1.43	2.44	2.44	1.97	2.90
MnO	0.03	0.09	0.10	0.04	0.06	0.15
MgO	0.70	1.61	0.95	0.89	1.04	1.42
CaO	1.16	1.77	1.97	1.73	1.66	2.34
Na_2O	9.59	6.13	8.08	6.96	7.69	7.37
K_2O	4.97	5.70	5.13	5.22	5.26	5.27
Total	95.35	94.88	98.82	93.15	95.49	96.80

BASIC END-MEMBER (GRIOUNOT) / PHONOLITE (GRIOUNOT)

	9	10	11	Av (9-11)	12	13	Av(12-13)
SiO_2	51.12	52.77	53.11	52.33	59.90	58.78	59.34
TiO_2	1.65	1.64	1.50	1.60	0.19	0.29	0.24
Al_2O_3	18.80	18.98	19.00	18.93	20.94	20.44	20.69
FeO	5.12	5.43	4.86	5.14	2.05	2.09	2.07
MnO	0.20	0.24	0.18	0.21	0.13	0.03	0.08
MgO	2.45	1.94	1.53	1.97	0.24	0.33	0.29
CaO	5.82	6.00	5.23	5.68	1.06	1.10	1.08
Na_2O	7.47	6.96	7.72	7.38	7.87	7.91	7.89
K_2O	3.68	3.20	3.36	3.41	6.16	6.17	6.17
Total	96.5	97.35	96.78	96.65	98.70	97.26	97.85

negative Eu anomalies. Hornblende, present as sparse phenocrysts in most tephrites, may also be involved in fractional crystallisation, but if so, its removal does not cause a trend to silica saturation. This is in contrast to the coexisting weakly alkaline magma series in the volcano, which shows a strong trend to silica oversaturation and abundant amphibole phenocrysts. Since the compositions of the phonolite host and the basic inclusions are similar to other local undersaturated lavas, it is probable that they have also been formed by fractionation of parental basanitic magmas prior to mixing.

Mineral chemistry of the mixed magmas

Phenocryst phases present in basanites from Cantal are olivine, titansalite and magnetite. Basic tephrites contain titansalite, magnetite, horn-blende, and apatite, while evolved tephrites additionally contain plagioclase phenocrysts (Downes, 1983). Therefore, in terms of mineralogy as well as chemistry, the basic inclusions (containing olivine, hornblende, titansalite, and magnetite) are on the boundary between basanite and tephrite. However, the inclusions differ from normal basic tephrite lavas in their abundance of hornblende, normally present only as sparse phenocrysts in tephrites. Phonolites from Cantal are variable in their mineralogy, depending on their bulk chemistry (Varet, 1968). The Griou/Griounot phonolites appear from their chemical compositions to be among the least evolved non-peralkaline (miaskitic) variety. Their phenocryst mineralogy (ferrosalite, sanidine, plagioclase, sodalite-group feldspathoid, hornblende, Fe–Ti oxide, titanite, and biotite) resembles that of typical miaskitic phonolites in the area (Varet, 1968;
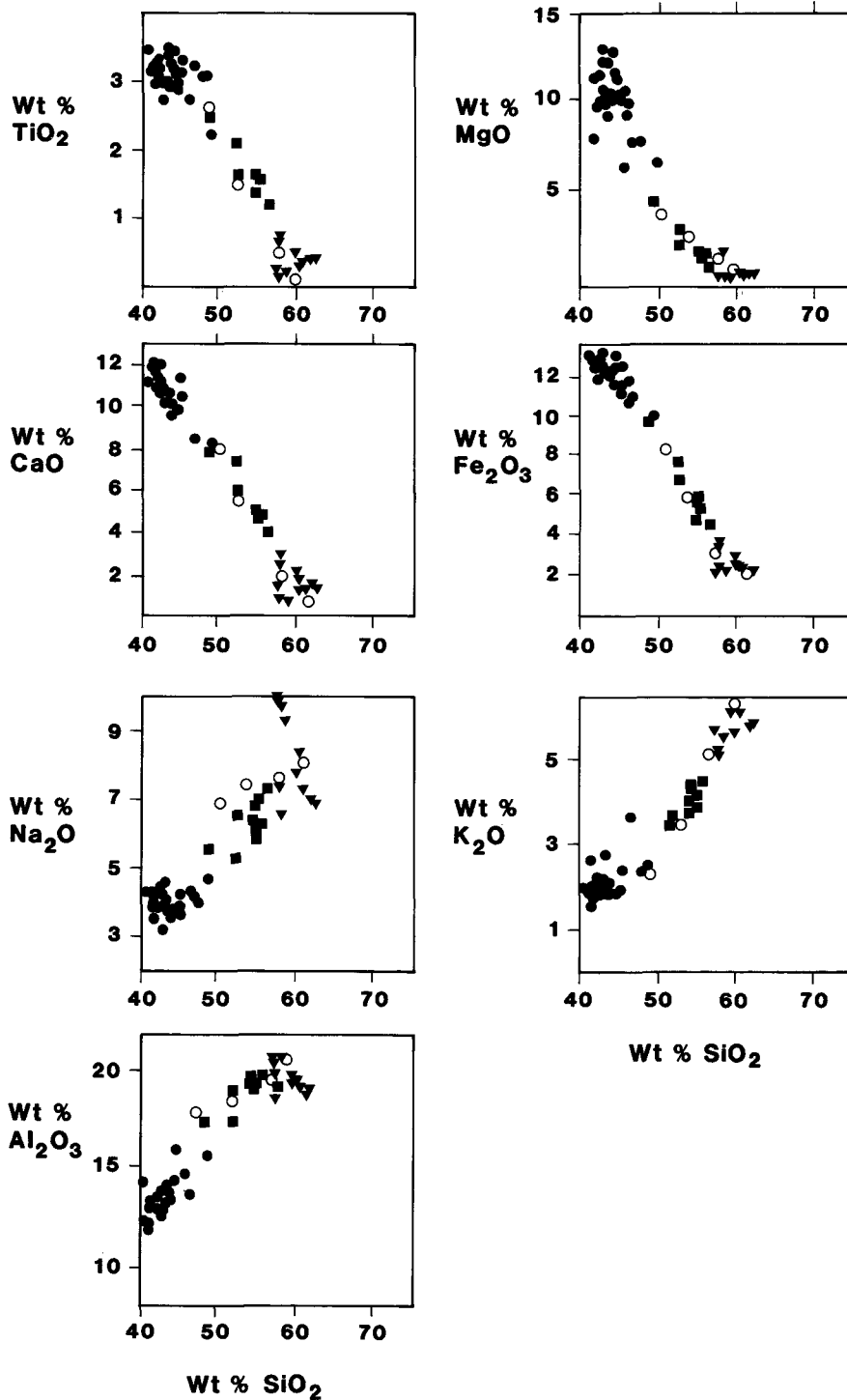
FIG. 3. Bulk geochemistry of end members of mixed magmas from Griou/Griounot (open circles) compared with XRF analyses of undersaturated volcanics from Cantal (Downes, 1983, 1987). Filled circles: basanites; filled squares: tephrites; filled triangles: phonolites.

Downes, 1983). Microprobe analyses of phases from both the phonolite hosts and the basic tephritic inclusions reveal that extensive phenocryst exchange has occurred between the two magma types.

Olivine. Resorbed crystals of olivine are abundant in some samples of the Puy Griou phonolite. These are Fo_{82-83} (Table 2), similar to olivine phenocrysts in the basic inclusions (Fo_{83-84}) and in local basanites (Downes, 1983). Such magnesian olivine could not have been in equilibrium with a phonolitic liquid and it can be inferred that the transfer of phenocrysts across the liquid–liquid contacts between the two magmas was common. Olivine appears to be one of the two minerals which exchanged most commonly between the basic inclusions and the host phonolite, the other being clinopyroxene.

Clinopyroxene. Cantal phonolites contain clinopyroxene phenocrysts which are green ferrosalites with 10–20% of the acmite end-member, while basic tephrites of the volcano contain brown titansalites (Downes, 1983). In the basic inclusions from Griou/Griounot, both green and brown clinopyroxenes are found, often displaying strong zoning from green cores to brown rims. Brown and green clinopyroxenes also occur together in the host phonolite but brown clinopyroxenes are less common than green ones.

Fig. 4 shows the variation in composition of clinopyroxenes from basic inclusions and phonolites. Representative analyses are presented in Table 2. Analyses of the cores of clinopyroxenes from the basic tephrites fall into two groups (Fig. 4); analyses of the brown rims of zoned clinopyroxenes are identical to those of the brown clinopyroxene phenocrysts. The zoning observed in clinopyroxenes from the basic inclusions is from green ferrosalite cores rich in Fe, Si, Na, and Mn to brown titansalite rims with higher concentrations of Ti and Al. Within the brown clinopyroxenes themselves, a slight zonation from cores rich in Fe and Ti to rims with lower Fe and Ti is observed (Fig. 4). It is therefore probable that the abundant brown clinopyroxenes are normal phenocrysts, while the less common clinopyroxenes with green ferrosalite cores were xenocrysts transferred from the phonolite host. The brown titansalite rims achieved equilibrium with the basic tephrite magma into which the clinopyroxenes were mixed.

In the phonolites, the majority of clinopyroxenes are green sodic ferrosalites. They are usually unzoned but occasionally have brown rims and in one case a zone of brown pyroxene was observed between a green core and a green rim. However, the brown rims and zones do *not* have the same
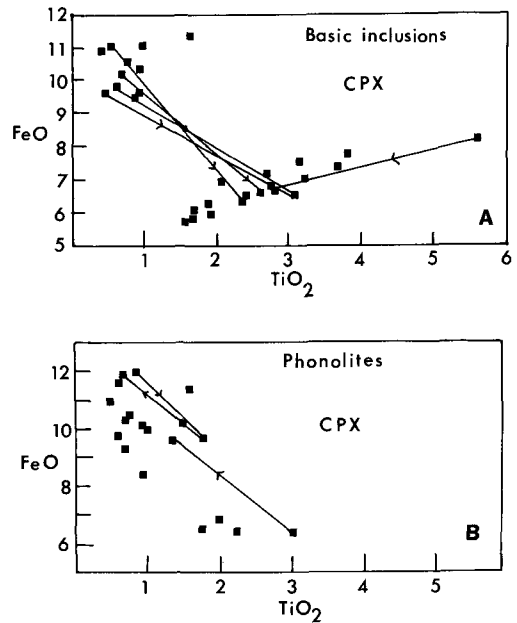

FIG. 4. Compositions of clinopyroxenes from basic tephrite inclusions (A) and phonolite hosts (B) of Puy Griou and Griounot. Two populations are seen: the high FeO, low TiO_2 cpx are ferrosalites, typical of phenocrysts of phonolites; the low FeO, high TiO_2 compositions are titansalites, typical of basanites and tephrites. Arrows show direction of zoning.

composition as the brown clinopyroxenes in the basic inclusions; instead they are only slightly poorer in Fe than the green clinopyroxenes. Analyses of the green clinopyroxenes are similar to those found as green cores in the basic tephrite, but a wider range of compositions is observed. Brown clinopyroxenes of the same composition as the phenocrysts in the basic inclusions are also found in the phonolites. These can have rims similar in composition to the rims on the green clinopyroxene phenocrysts.

Brown clinopyroxenes in phonolites are clearly xenocrysts which have been transferred from the basic inclusions, while the green clinopyroxenes are phenocrysts of the phonolite. The separation of compositions is similar to that seen in the basic inclusions, but the compositional variation in the green sodic ferrosalite phenocrysts is wider and they also show more complex zoning patterns than the ferrosalites transferred to the basic inclusions. Therefore the zoning and the wider variation may post-date the transfer of phenocrysts. From Fig. 4 it is apparent that clinopyroxene phenocrysts were transferred from the host into the inclusions and vice versa.

Table 2. Representative microprobe analyses of minerals in basic tephrite and phonolite from Puy Griou and Griounot samples. Analyses obtained using a Cambridge Instruments Microscan 5 with a LINK EDS system. Operating voltage 20kV, Beam current 100 µA, spot diameter 1 µm.

CLINOPYROXENES

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	51.77	47.39	51.30	47.30	45.68	50.18	47.48	49.94	49.80	51.14	47.82
TiO_2	0.44	2.83	0.68	2.76	3.69	1.92	2.98	1.37	1.76	0.69	1.87
Al_2O_3	1.49	6.90	2.16	6.92	7.88	4.18	6.64	3.97	4.21	2.05	5.53
FeO	9.60	6.73	10.16	6.78	7.37	5.99	6.40	9.57	6.47	10.33	9.65
MnO	0.94	0.19	0.94	0.07	0.04	0.16	0.08	0.50	0.18	0.97	0.25
MgO	10.99	12.44	10.94	12.75	11.62	14.36	12.89	11.06	13.57	11.05	10.56
CaO	22.40	23.11	22.44	23.08	22.84	22.95	22.96	22.47	22.37	21.93	23.24
Na_2O	1.38	0.50	1.35	0.50	0.49	0.11	0.71	1.46	0.78	1.58	0.71
Total	99.01	100.09	99.97	100.16	99.61	99.85	100.14	100.34	99.14	99.74	99.63

Formulae calculated on basis of 6 oxygens

	1	2	3	4	5	6	7	8	9	10	11
Si	1.974	1.768	1.943	1.763	1.720	1.860	1.768	1.880	1.865	1.943	1.835
Al(iv)	0.026	0.232	0.057	0.237	0.280	0.140	0.232	0.120	0.135	0.057	0.165
Al(vi)	0.041	0.072	0.040	0.068	0.070	0.043	0.060	0.056	0.051	0.035	0.085
Ti	0.013	0.079	0.019	0.077	0.105	0.054	0.083	0.039	0.050	0.020	0.054
Fe^{2+}	0.306	0.210	0.322	0.211	0.232	0.186	0.199	0.301	0.203	0.328	0.310
Mn	0.030	0.006	0.030	0.002	0.001	0.005	0.003	0.016	0.006	0.031	0.008
Mg	0.624	0.629	0.618	0.208	0.652	0.793	0.715	0.621	0.757	0.626	0.604
Ca	0.915	0.924	0.911	0.922	0.922	0.911	0.916	0.906	0.898	0.893	0.955
Na	0.103	0.036	0.099	0.036	0.036	0.008	0.051	0.107	0.057	0.116	0.054
En	0.33	0.38	0.33	0.38	0.36	0.42	0.39	0.34	0.41	0.33	0.32
Fs	0.18	0.12	0.19	0.12	0.13	0.10	0.11	0.17	0.11	0.19	0.17
Wo	0.49	0.50	0.48	0.50	0.51	0.48	0.50	0.49	0.48	0.48	0.51

1=core of green cpx in basic inclusion (Griou)
2=rim of green cpx in basic inclusion (Griou)
3=core of green cpx in basic inclusion (Griou)
4=rim of green cpx in basic inclusion (Griou)
5=core of brown cpx in basic inclusion (Griou)
6=rim of brown cpx in basic inclusion (Griou)
7=brown core of cpx in phonolite (Griou)
8=green rim of cpx in phonolite (Griou)
9=core of unzoned brown cpx in phonolite (Griou)
10=core of unzoned green cpx in phonolite (Griou)
11=brown zone within green pyroxene in phonolite (Griounot)

HORNBLENDES

	1	2	3	4
SiO_2	40.29	40.60	38.51	39.31
TiO_2	3.00	3.36	5.60	4.92
Al_2O_3	11.94	11.69	13.65	12.41
FeO	18.17	17.53	11.06	12.92
MnO	1.17	0.93	0.19	0.26
MgO	8.90	9.20	11.94	10.88
CaO	11.28	11.72	12.41	11.77
Na_2O	3.18	2.92	2.67	2.79
K_2O	1.62	1.77	1.13	1.53
Total	99.55	99.71	97.17	96.77

Formulae calculated on basis of 23 oxygens

	1	2	3	4
Si	6.102	6.119	5.774	5.968
Al(iv)	1.898	1.881	2.226	2.032
Al(vi)	0.234	0.196	0.187	0.189
Ti	0.342	0.381	0.631	0.562
Fe^{2+}	2.302	2.210	1.387	1.640
Mn	0.150	0.119	0.024	0.033
Mg	2.009	2.066	2.668	2.462
Ca	1.831	1.893	1.994	1.915
Na	0.934	0.853	0.776	0.821
K	0.313	0.340	0.216	0.296

1=Fe-rich hornblende in phonolite (Griounot)
2=Fe-rich hornblende in tephrite (Griounot)
3=Ti-rich hornblende in phonolite (Griounot)
4=Ti-rich hornblende in tephrite (Griounot)

MAGNETITES

	1	2	3	4
SiO_2	0.40	0.56	0.48	0.83
TiO_2	15.56	13.07	14.88	16.00
Al_2O_3	0.87	0.84	0.89	1.55
FeO	74.75	77.52	75.67	73.32
MnO	3.50	2.05	1.34	2.06
MgO	0.52	0.67	0.67	1.41
Total	95.59	94.73	93.92	96.33

Formulae calculated on basis of 16 oxygens (Fe^{3+} calculated for 12 cations)

	1	2	3	4
Si	0.060	0.085	0.074	0.125
Al	0.155	0.151	0.163	0.275
Ti	1.769	1.494	1.718	1.808
Fe^{3+}	4.186	4.691	4.253	3.860
Fe^{2+}	5.264	5.163	5.464	5.355
Mn	0.448	0.264	0.174	0.262
Mg	0.117	0.152	0.153	0.316

1=Mn-rich magnetite in phonolite (Griou)
2=Mn-poor magnetite in phonolite (Griou)
3=Mn-poor magnetite in tephrite (Griou)
4=Mn-poor magnetite in tephrite (Griou)

OLIVINES

	1	2
SiO_2	39.66	40.39
FeO	15.69	15.20
MnO	0.33	0.23
MgO	43.35	43.95
CaO	0.31	0.33
Total	99.34	100.10

Formulae calculated on basis of 4 oxygens

	1	2
Si	1.006	1.013
Fe^{2+}	0.333	0.319
Mn	0.007	0.005
Mg	1.639	1.642
Ca	0.008	0.009
Fo	83	84

1= phonolite (Griou)
2= basic tephrite (Griou)

Amphibole. Kaersutitic hornblende appears in both the tephritic inclusions and in the phonolite hosts. In the tephrites it occurs both as zoned phenocrysts and as abundant elongated skeletal quench crystals. In the phonolites there are two varieties of hornblende; one is identical in composition to phenocrysts in the basic tephrites and the other is very deep brown and rich in Fe and Mn (Table 2). High Mn contents in minerals is a characteristic feature associated with phonolitic magmas (Downes, 1983). An amphibole of the same composition was found in an associated basic tephrite inclusion (Fig. 5), again indicating that transfer of phenocrysts has occurred. The more common amphiboles are similar in composition to those in tephrites elsewhere in Cantal, i.e. with low Fe and Mn contents (Downes, 1983), and it is probable that these amphiboles occur in the phonolites as transferred xenocrysts.
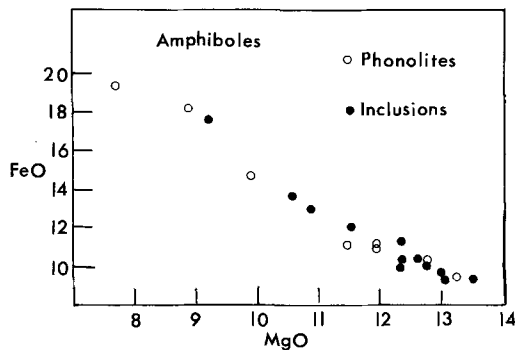
FIG. 5. Amphibole compositions. Total Fe as FeO *vs.* MgO for amphiboles from basic tephrites (filled circles) and phonolites (open circles) from Puy Griou and Griounot. Amphiboles with > 14 wt. % FeO are deep brown in colour and contain higher MnO, indicative of derivation as phenocrysts in the phonolite. Other amphiboles are kaersutitic hornblendes similar to amphiboles found in other Cantal tephrites (Downes, 1983).

Fe–Ti oxides. Magnetite containing 13–16 wt. % TiO_2 occurs in both the phonolite and the basic tephrite with limited compositional variation. Substitution of Al, Mg and Mn is extensive. Phenocrysts in Cantal phonolites have high MnO contents (> 2.5 wt. % MnO) while those from tephrites are less manganiferous (Downes, 1983). This distinction is also seen in magnetites from the mixed magmas of the Griou region (Fig. 6 and Table 2). A transferred magnetite phenocryst of low MnO content was found in the Griou phonolitic host; no example of the reverse transfer has been found.
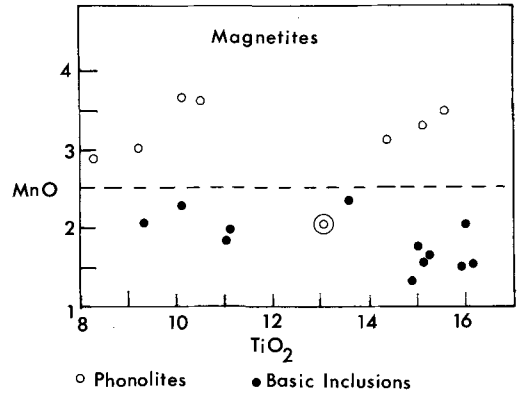

FIG. 6. Wt. % MnO *vs.* wt. % TiO_2 for magnetites from basic inclusions (filled circles) and phonolites (open circles) from Puy Griou and Griounot. Magnetites which are phenocrysts in the phonolites have MnO contents > 2.5 wt. %; those which are phenocrysts in the tephrites have MnO < 2.5 wt. % (Downes, 1983). One magnetite phenocryst in the Griou phonolite (circled) has been transferred from the basic tephrite.

Sr and Nd isotope analyses

Age-corrected $^{87}Sr/^{86}Sr$ and $^{143}Nd/^{144}Nd$ analyses of undersaturated magma series of Cantal (Downes, 1984) are presented in Fig. 7. Results were obtained at the University of Leeds, using techniques described by Downes (1984). A repeat analysis of the Sr isotope ratio of the Puy Griou phonolite has been provided by D. Briot (University of Clermont Ferrand), who also measured the isotopic ratios of the Griounot phonolite. Repeat
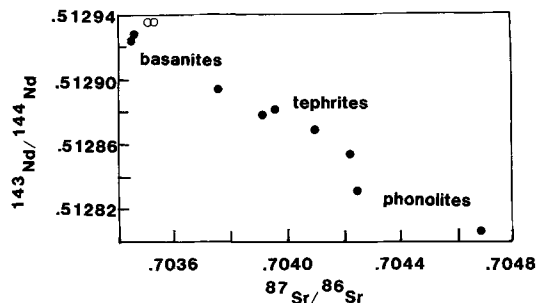

FIG. 7. $^{143}Nd/^{144}Nd$ *vs.* $^{87}Sr/^{86}Sr_i$ for undersaturated volcanics of Cantal (filled circles). Data shown are for basanites, tephrites and phonolites (Downes, 1984). $^{143}Nd/^{144}Nd$ decreases and $^{87}Sr/^{86}Sr_i$ increases with increasing fractionation, indicating progressive assimilation of crustal material. Puy Griou and Griounot phonolites (open circles) have $^{143}Nd/^{144}Nd$ and $^{87}Sr/^{86}Sr$ initial ratios within error of those of uncontaminated basanites from Cantal.

analyses of the $^{87}Sr/^{86}Sr$ and Rb/Sr ratios of the Griounot phonolite were obtained at Royal Holloway and Bedford New College, London. The tephrite inclusions are too intimately emulsified with the phonolite to permit successful separation and analysis.

There is a progressive increase in $^{87}Sr/^{86}Sr$ and decrease in $^{143}Nd/^{144}Nd$ from basanites through tephrites to the majority of phonolites. The Puy Griou phonolite, however, has a high $^{143}Nd/^{144}Nd$ ratio of 0.512935 ± 54 (repeat analysis: 0.512935 ± 14) and a low $^{87}Sr/^{86}Sr$ initial ratio (0.70353 ± 6; repeat analysis 0.70346 ± 2), which are similar to the values for basanites from Cantal. Analyses of the Griounot phonolite reveal almost identical results (initial $^{87}Sr/^{86}Sr$ ratio of 0.703556 ± 20; repeat analysis 0.703489 ± 9; and $^{143}Nd/^{144}Nd = 0.512933 \pm 19$).

Discussion

The macroscopic textures, including the dissemination of the cauliform basic inclusions within the phonolite host and the evidence of chilling, lead to the suggestion that partial magma mixing has occurred in the formation of these rocks. Chemical compositions of the magmas are similar to members of the undersaturated magma series of Cantal. Members of this series are considered to be related to each other by fractional crystallisation and it is probable that the magmas involved in the mixing process were likewise related. Alternatively, the near-linear major element trends could be interpreted as mixing trends and hence the entire magma series could have been formed by mixing. However, there is very little field or petrographic evidence in favour of this hypothesis. The examples of Puy Griou and Griounot contain the only physical evidence of mixing found in the undersaturated series of Cantal. No other phonolites in the area contain similar evidence of mineralogical disequilibrium.

Exchange of phenocrysts has occurred extensively. It is noticeable that all the phenocrysts of the tephrite (olivine, hornblende, titansalite, and Mn-poor magnetite) were transferred to the phonolite, while of the phenocrysts of the phonolite (ferrosalite, Fe-rich hornblende, Mn-rich magnetite, alkali feldspar, plagioclase, feldspathoid, and titanite) only ferrosalite appears to have been transferred to the tephrite in any great quantity. This may indicate that the transfer of phenocrysts occurred before the onset of crystallisation of some phenocryst phases in the phonolite, and hence magma mixing might have taken place at great depth (lower crust or upper mantle). Transfer of phenocrysts from the phonolite to the tephrite might also have been impeded by the higher viscosity of the phonolite magma, or the evidence of such transfer may be poorly preserved because of the tendency towards dissolution of phonolite phenocrysts in the tephrite magma.

The macroscopic features of the samples seem to indicate that the mechanism of transfer of phenocrysts from the basic magma to the phonolite was by the progressive break up and dispersion of the tephrite blebs, and by the formation of emulsions. The cauliform shapes of the basic inclusions could be interpreted as evidence of low-pressure mixing, since ascent from depth might be expected to produce a streaky banded appearance due to shearing. Indeed, the phonolite shows strong flow-banding, which is absent from the tephrite inclusions. The inclusions show only a weak tendency towards elongation (Fig. 1a) and could therefore have been close to solidification by the time the phonolite reached the low-pressure environment. Hence they were solid enough to resist shearing by the flow of phonolite magma. It may be significant in this context that both the inclusions and host show evidence of rapid chilling, but the phonolite displays many areas of extremely feathery crystals of feldspar adjacent to the basic inclusions.

Phonolites have generally been interpreted as low-pressure fractionates of alkali basalt or basanite magmas (e.g. Goles, 1976). This has been proposed as the mechanism of formation of many Massif Central phonolites, since the increase in $^{87}Sr/^{86}Sr$ initial ratios with increasing SiO_2 in the undersaturated series has been interpreted as being due to progressive crustal assimilation linked to fractional crystallisation (Downes, 1984). Such a relationship between silica content and Sr isotopic ratio has been observed in phonolites from many parts of the world (e.g. Marsh, 1987).

Elsewhere in the world there is evidence that some phonolites may have been formed at mantle depths, either by high-pressure fractionation or by direct partial melting of a strongly metasomatised mantle (Irving and Price, 1981). Bailey (1987) argued that the compositional gap frequently observed between basic and evolved alkaline rocks was due to derivation from two separate metasomatised mantle sources, and that such a process could also account for the anomalously large volumes of phonolites and trachytes in some alkaline provinces. Rock (1976) argued that because many phonolites of Monchique and Mt Kenya have very low $^{87}Sr/^{86}Sr$ ratios they were straightforward derivatives of mantle liquids. Price and Compston (1973) reported similar low Sr ratios for phonolites from Dunedin volcano, New Zealand. Irving and Price (1981) summarised earlier work

by Wright (1966) and Price and Green (1972) concerning phonolites which have entrained mantle-derived spinel lherzolite xenoliths and must therefore have been formed at mantle depths within or below the spinel lherzolite stability field.

The main isotopic trend of Cantal volcanics has been attributed to progressive assimilation of continental crust with fractionation (Downes, 1984). The isotopic signatures of the Puy Griou/Griounot mixed magma phonolites are anomalous, being very similar to that of local unfractionated, uncontaminated basanites with low $^{87}Sr/^{86}Sr_i$ and high $^{143}Nd/^{144}Nd$ ratios (Fig. 7). This may indicate that the phonolitic mixed magmas passed through the crust without significant interaction with the wall rocks and consequently escaped contamination. The lack of contamination shown by the Puy Griou phonolite can be explained in several ways, such as particularly rapid ascent or insulation from the crust in unusually large magma chambers. It is equally feasible that the phonolite–tephrite mixing occurred at depths below the base of the crust and subsequently the magma did not pause during ascent. In contrast, other phonolites are formed by fractionation of basanitic and tephritic melts in lower crustal magma chambers where crustal contamination also occurred, and consequently tephrites and phonolites show an isotopic signature offset from that of their mantle-derived parental magmas.

The generation of phonolitic magmas by direct melting of metasomatised mantle (Bailey, 1987) could also explain the anomalous isotopic composition of the Puy Griou/Griounot phonolites. However, this cannot explain why the anomalous isotopic ratios are found only in the phonolites which show petrographic and mineralogical evidence of magma mixing. Other phonolites from the region lie on the crustal contamination trend shown by the lavas of the saturated series (trachyandesite–trachyte–rhyolite) (Downes, 1984), and hence the simplest explanation for their origin is in coupled assimilation and fractionation. The mixed magmas show no evidence of crustal contamination and thus must have a significantly different intratelluric history.

The Puy Griou/Griounot phonolites may therefore have been formed by fractionation of basanite and tephrite magmas in deep magma chambers, perhaps even within the uppermost part of the upper mantle. This was followed by mixing with a small volume of magma similar to their own parental basic tephrite. The ascent of the phonolite magma was subsequently rapid and no further fractionation occurred within the crust. During this time, some of the tephrite blebs disintegrated and released their phenocrysts (particularly olivine and clinopyroxene) into the phonolite magma. If fractionation within the mantle has occurred in other alkaline volcanic regions of the world, it may also account for the presence of rare mantle-derived xenoliths in phonolites and trachytes.

Acknowledgements

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