«Low P» and «High P» alkaline magmas from Mt. Etna: their origin and evolution from the study of mineralogy and chemistry of lavas and of olivine-trapped inclusions

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ABSTRACT. — Phenocryst assemblages, chemical features and magmatic inclusions in Fo-rich olivine crystals were investigated in members from the earliest alkaline successions of Mt. Etna: the «Ancient Alkaline Centres» Unit (AAC) and the «Trifoglietto» Unit (TU). The study confirms the occurrence of two distinct groups in the AAC sequence according to their P_2O_5 contents, namely the «high P» and «low P» lavas, and supports the idea of genetic links between the «high P» and the TU series.

Apatite occurs as micro-phenocryst and even enclosed in the most Mg-rich olivine crystals in the «high P» AAC and TU hawaiites. Alkali basalts were found among «low P» lavas with Cr-spinel, either as isolated micro-phenocrysts or enclosed in Cr-diopside and Fo-rich (Fo_{87}) olivine, more magnesian than in «high P» hawaiites (Fo_{82}). Clinopyroxene mg# values in these rocks are strictly related to those of olivine, suggesting their co-crystallization.

The density of olivine-trapped CO_2 fluid inclusions, found in «low P» alkali basalts, indicates their crystallization at pressures up to around 0.55 GPa, which are consistent with the coexisting Cr-diopside compositions.

Melt inclusions (mg# 69) in the Mg-rich olivines (Fo_{87}) of the «low P» alkali basalts are widely scattered (from «high P» to «low P») in their compositions, whereas they are more homogeneous, with lower incompatible elements concentrations, in less magnesian olivines.

In the most magnesian olivines of «high P» hawaiites, from AAC (Fo_{82}) and TU (Fo_{73}), only «high P» tephritic and phonotephritic melt inclusions (mg# 60%) occur respectively, suggesting that they could have derived by differentiation of a common parent magma.

Furthermore, it is possible that «high P» and «low P» magmas may be originated by different degrees of partial melting of a heterogeneous mantle. Magma mixing and fractional crystallization processes may then account for all the varieties of both the two groups.

RIASSUNTO. — Sono state svolte indagini sulle associazioni di fenocristalli ed il chimismo, nonché su inclusioni magmatiche in cristalli di olivina ricchi in forsterite, di termini attribuiti alle più antiche manifestazioni vulcaniche, affioranti nel basso versante orientale Etneo: i Centri Alcalini Antichi (CAA) e l'Unità del Trfoglietto (UT). Lo studio ha confermato la presenza di due gruppi nella successione dei CAA, distinti sulla base del contenuto in P_2O_5, definiti rispettivamente come lave «ad alto P» ed «a basso P», e ha portato dati a sostegno dell'ipotesi dell'esistenza di stretti rapporti

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nella genesi dei prodotti «ad alto P» (CAA) e quelli dell'Unità del Trifoglietto.

Nelle hawaiiti «ad alto P» dei CAA ed in quelle dell'UT l'apatite si ritrova in microfenocristalli, inclusi anche negli individui olivinici più ricchi in Mg. Basalti a tendenza alcalina sono stati riscontrati tra i termini «a basso P»: essi contengono Cr-spinello in microfenocristalli o come incluso in Cr-diopside e olivina Fo_{87}, più ricca in Mg di quella delle lave «ad alto P» (Fo_{82}). In questi basalti i valori mg# del clinopirosseno sono strettamente legati a quelli dell'olivina, ciò che è in accordo con una loro cristallizzazione contemporanea.

La densità di inclusioni fluide a CO_2 in olivina di alcali basalto «a basso P» indica una pressione di cristallizzazione fino a 0,55 Gpa, coerente con la composizione del coesistente Cr-diopside.

Le inclusioni di «fuso» (mg# 69) in individui di olivina Fo_{87} di basalti «a basso P» mostrano composizioni molto variabili (da povere a ricche in P), mentre sono più omogenee, con concentrazioni di elementi incompatibili più basse, nell'olivina più povera in Mg. Le inclusioni di «fuso» nell'olivina più magnesifera delle hawaiiti «ad alto P», sia dei CAA che dell'UT, hanno rispettivamente composizioni da tefritiche a fonotefritiche «ad alto P», che indicano una loro possibile derivazione per frazionamento da uno stesso magma.

Inoltre si suggerisce che i magmi «ad alto P» ed «a basso P» abbiano avuto origine per diverso grado di fusione parziale, di un mantello eterogeneo. Processi di mescolamento di fusi e di cristallizzazione frazionata possono poi rendere conto dell'origine dei diversi termini di lave riscontrati nella successione esaminata.

KEY WORDS: *Mount Etna, ancient alkali basalts and hawaiites, Phosphorus, heterogeneous mantle, melt and fluid inclusions.*

INTRODUCTION

The study of drilled and field samples, from the eastern piedmont belt of Mount Etna, improved the knowledge on the features and relationship of the most ancient volcanic units (Cristofolini *et al.*, 1991; Finocchiaro, 1995). In the volcanic sequence belonging to the «Ancient Alkaline Centres» (AAC) two different types of lavas were distinguished related to «high P» and «low P» parent magmas respectively. The former type, exclusively found in the eastern and south-eastern sector of the volcano, is characterized by relatively high P ($P_2O_5 > 0.8$ wt%), Ti, K and Rb contents than the latter. Apatite microphenocrysts occur also in the most basic «high P» varieties, whereas they may be observed only a in the differentiated members ($SiO_2 > 52$ wt%) of the «low P» group.

A later «Trifoglietto» Unit (TU) volcanic sequence is exclusively found in the eastern sector of the volcano (Cristofolini *et al.*, 1991), and might then be considered as «comagmatic» to the AAC «high P» group, due to its chemical and mineralogical features: the same incompatible elements content (P, K, Rb) and mineral assemblage (*ol + px + pl +ap + amph + mt*; Finocchiaro, 1995).

Melt and fluid inclusions trapped in Fo-rich olivine from the less differentiated TU and AAC rocks were studied, together with phenocryst assemblages and chemical characters of lavas in order to gain a better understanding of the mantle sources involved in the generation of the different types of magmas.

GEOLOGICAL BACKGROUND

Mt. Etna, located close to the eastern coast of Sicily, is the largest active volcano in Europe. Its activity should be considered within the geological framework of eastern Sicily, since it is located at the front of the south-verging overthrust pile of the Apennine-Maghrebian chain where it is intersected by regional tectonic structures, marking the boundary between the thickened continental crust domain of Sicily and the thinned continental crust of the Jonian Sea (Barberi *et al.*, 1973; Lentini, 1982; Lanzafame *et al.*, 1997). The integration of geological, morpho-structural and petrological data showed that the Etnean volcanic sequence is related to the activity of several distinct centres that followed each other in time (Tanguy, 1980; Cristofolini *et al.*, 1981; Romano, 1982).
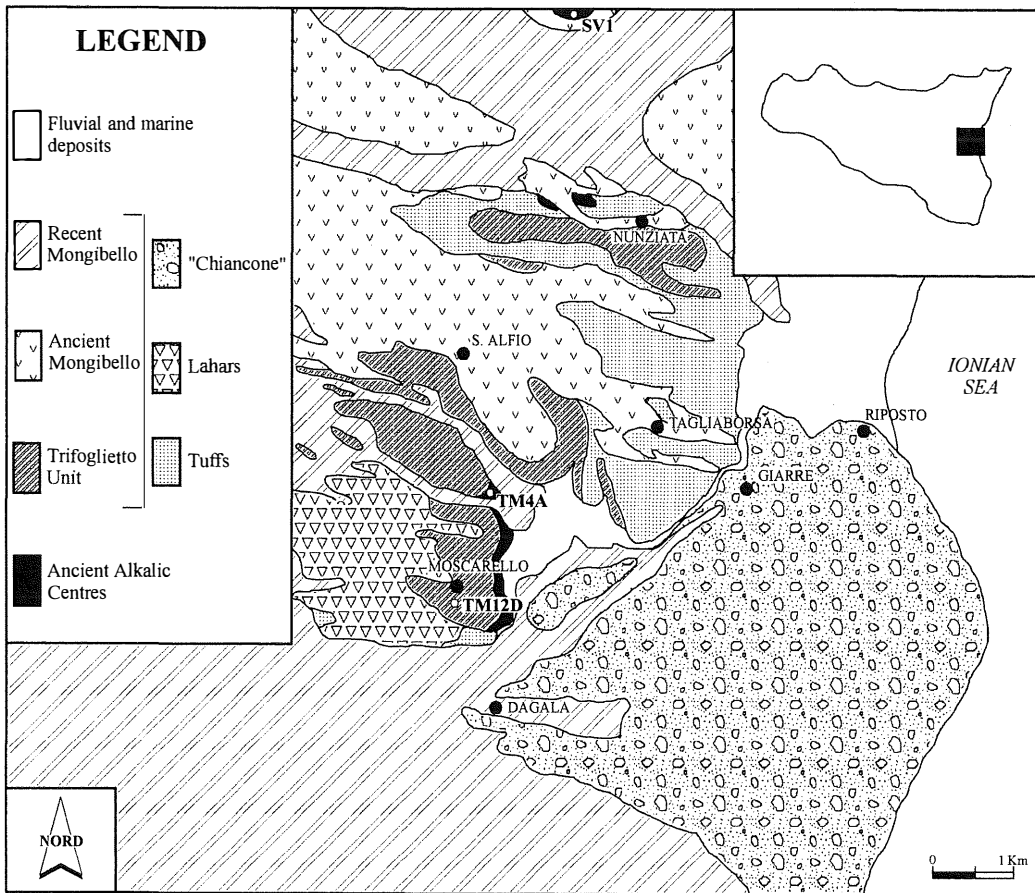
Fig. 1 – Geological map of the eastern slope of Mount Etna volcano. TM, SV: sample locations.

The most ancient lavas of Mt. Etna, tholeiitic to transitional basalts, crop out at the base of the volcanic sequence. On the grounds of K-Ar dating (Gillot *et al.*, 1994) tholeiitic eruptions may be related to two main stages: products of the former, dated at 0.5 ± 0.04 Ma BP, are found at the south-eastern edge of the present-day Etna edifice, and of the latter, dated between 0.33 and 0.27 Ma BP at the base of the south-western sector.

Magmas evolved later towards transitional to alkaline series compositions: the appearance of the most ancient alkaline magmas, dated at no less than 0.17 ± 0.01 Ma BP, raises the question whether the Etnean volcanic activity was continuous or showed a time gap between the eruption of the tholeiites and alkaline series volcanics (Gillot *et al.*, 1994).

The «Ancient Alkaline Centres» (AAC), ranging from basalts to basic mugearites, represent the oldest part of the alkaline sequence. Afterwards, several distinct centres followed each other in time. These, defined on the whole as «Trifoglietto» Unit (TU), are characterised by differentiated volcanics, from hawaiites to mugearites, found only in the eastern sector of the volcanic edifice. The particularly explosive activity of these centres gave rise to a noteworthy amount of tephra along with lavas.

According to different authors (Vv. AA., 1979; Rasà *et al.*, 1982), the most ancient volcanic sequences are best exposed along the lower eastern flanks of the volcano. In particular, lavas from an ancient volcanic sequence at the Timpa di Moscarello fault scarp, and basalts from neighbour areas, among the most ancient Etnean alkaline products (Cristofolini and Spadea, 1975; Spadea, 1972), are the object of this study (fig. 1).

PETROGRAPHY AND CHEMISTRY

Lavas from the Timpa di Moscarello

About eighty samples from the Timpa di Moscarello escarpment were examined: they show porphyritic textures with P.I. (Porphyritic Index) varying from 15 to 30%. The common

mineral association (*pl+cpx+ol+mt*) of Etnean lavas is present, both as Phenocrysts and in the groundmass. In accordance to Finocchiaro (1995), in all of the TU and in some of the AAC samples, apatite and kaersutite may be observed as phenocryst phases.

All of the samples were also analysed for their major and minor elements. They fall in the Na-alkaline field, the AAC products ranging from alkali basalts to mugearites, whereas the TU products are more differentiated, from hawaiites to benmoreites, as clearly shown by the TAS diagram (Le Bas *et al.*, 1986; Le Maitre, 1989) (fig. 2).

According to Cristofolini *et al.* (1991) and Finocchiaro (1995) it is possible to distinguish two different types of lavas within the AAC sequence: one of them is characterized by relatively high P contents ($P_2O_5 > 0.8$ wt%;
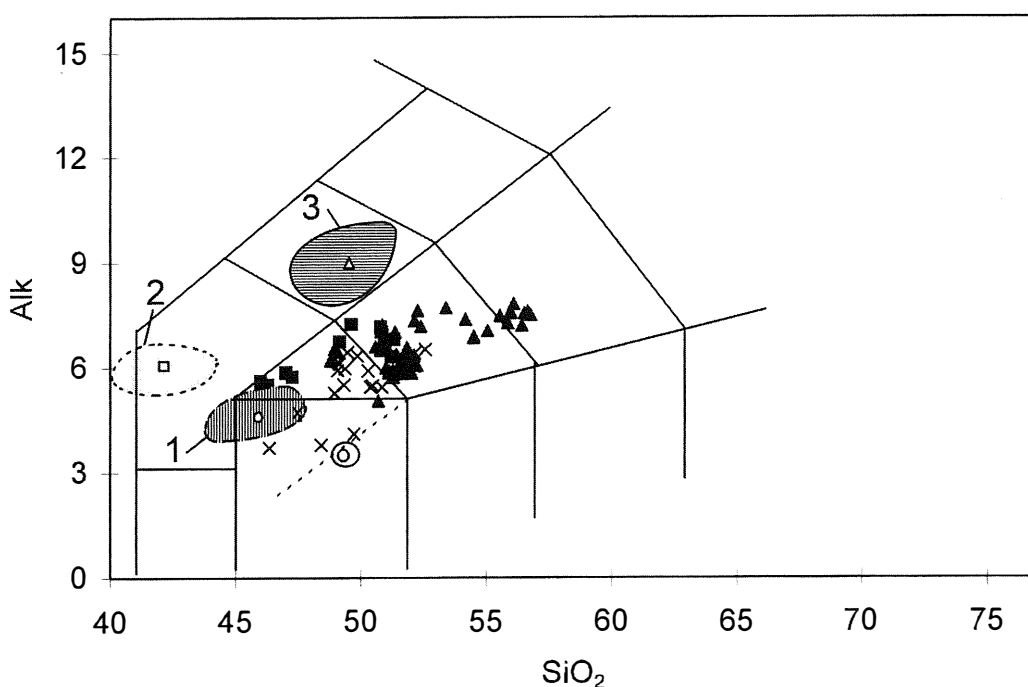

Fig. 2 – TAS (Le Bas *et al.*, 1986). Symbols: Closed square: «Ancient Alkaline Centres» (AAC) - «high P» lavas; Cross: «Ancient Alkaline Centres» (AAC) - «low P» lavas; Closed triangle: «Trifoglietto» Unit (TU). 1: Olivine-trapped inclusions field from the S. Venera alkali basalt (AAC - «low P»); Open circle: average compositions; Double circle: outlier (10b, Table 1); 2: Olivine-trapped inclusions field from the AAC - «high P» hawaiite, Open square: average compositions; 3: Olivine-trapped inclusions field from TU hawaiites; Open triangle: average compositions.
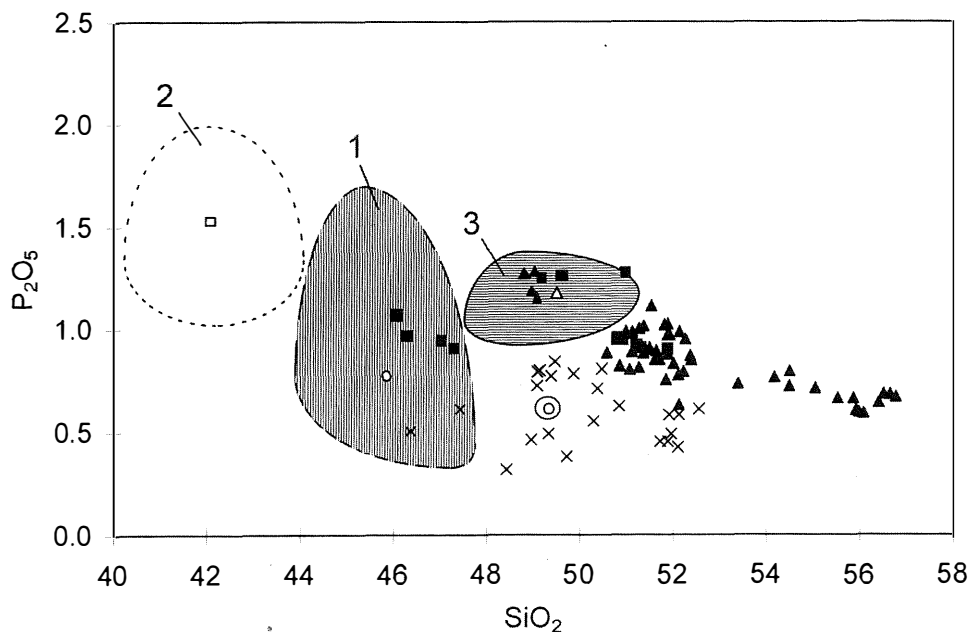
Fig. 3 – P_2O_5 versus SiO_2. Symbols as in fig. 2. For discussion see text.

«high P» lavas) and the other by lower contents of the same element («low P» lavas). The less differentiated members of the AAC «high P» group and of TU, from the Timpa di Moscarello fault scarp, are hawaiites with mg# $[Mg/(Mg+Fe_{tot}) \times 100]$ between 46 and 45. They are porphyritic in structure (P.I.~ 20%), with olivine, clinopyroxene, plagioclase, magnetite, in order of decreasing abundance, and microphenocrystic apatite. Furthermore according to the same authors, the TU volcanic sequence shows chemical characteristics related to the ACC «high P» group (fig. 3).

Alkali basalts from Santa Venera and neigh-bouring areas

The alkali basalt from Santa Venera (sample SV1, Table 1), an ancient lava of the AAC «low P» group, rests on the Quaternary sedimentary substratum (Spadea, 1972). It is a porphyritic basalt with about 30% of phenocrysts of clinopyroxene, olivine, and subordinate plagloclase, and a fine-grained intergranular-textured groundmass composed of the same minerals plus magnetite; some megacrystic olivine and diopside enclosing spinel are also found, sometimes in aggregates. This alkali basalt is one of the most mafic Etnean products (mg# = 65). Similar rocks crop out at Monte Nardello (Spadea, 1972), dated at about 0.12 Ma BP (Condomines *et al.*, 1982), and close to Nunziata (MN1, NUN1).

MINERAL ASSEMBLAGES

Besides the petrographic investigations, microprobe analyses were made on minerals in eleven samples from the Timpa di Moscarello and nearby outcrops (SV 1, MN 1, NUN 1).

Olivine

In the AAC «high P» and TU hawaiites, olivine is generally idiomorphic, and its size attains 0.5 mm, with Fo varying from 77 to 82

TABLE 1

Composition of bulk rocks and melt inclusions trapped in olivine crystals.

Bulk rock	SV1	Inclusion	1	2	3	4a	4b	5	6	7a	7b	8a	8b	9a	9b	10a		ac(N21)	std	10b
		Fo	86.16	86.04	86.73	86.86		86.24	85.87	86.10		84.58		86.17		86.86				
		T°m	1268	1258	1260	1270		1261	1260	1273		1262		1273		1265		ac(N21)	std	
SiO_2	48.43		45.33	46.77	47.58	46.78	46.09	45.91	44.36	46.16	45.62	45.82	45.63	46.89	46.16	45.31		45.85	0.86	49.33
TiO_2	1.23		1.83	1.62	1.71	1.91	1.88	1.38	1.90	1.49	1.58	1.64	1.68	1.67	1.83	1.60		1.66	0.16	1.18
Al_2O_3	17.88		19.06	19.49	20.26	16.70	19.31	18.55	19.28	18.87	18.83	20.03	19.85	19.84	19.96	19.25		19.15	0.81	17.81
FeO*	8.22		6.43	6.67	5.69	6.69	6.45	7.21	6.69	6.61	6.11	7.32	6.72	5.51	5.02	5.78		6.47	0.62	5.21
MnO	0.14		0.12	0.15	0.15	0.01	0.13	0.17	0.14	0.13	0.10	0.12	0.08	0.18	0.01	0.16		0.12	0.06	0.04
MgO	8.56		6.63	7.27	6.11	7.57	7.19	8.24	7.14	6.58	6.91	6.77	6.56	5.48	5.77	6.38		7.03	0.83	6.38
CaO	11.18		15.51	13.00	13.03	14.87	13.52	13.26	15.17	14.78	15.25	13.18	14.14	15.20	15.40	14.81		14.42	0.84	15.85
Na_2O	3.12		3.51	4.06	4.17	3.93	3.68	3.81	3.58	3.77	3.72	3.95	4.04	4.03	4.08	3.99		3.82	0.24	3.14
K_2O	0.66		0.64	0.59	0.92	1.02	0.80	0.92	0.72	0.37	0.45	0.72	0.78	0.44	0.53	1.03		0.70	0.19	0.35
P_2O_5	0.32		0.92	0.31	0.36	0.52	0.94	0.50	0.98	1.20	1.40	0.40	0.45	0.74	1.12	1.67		0.75	0.36	0.61
mg#	65		65	66	66	67	67	67	66	64	67	62	64	64	67	66		66	1.47	69

Bulk rock	TM4A	Inclusion	11	12	13	14	15	16	17a	17b	ac(N14)	std	Bulk rock	TM12D	Inclusion	19	20	21	ac(N5)	std
		Fo	80.99	80.99	80.99	77.69	77.23	80.99	81.83						Fo	70.57	69.97	72.80		
		T°m	1268	1259	1248	1250	1268	1259	1247		ac(N14)	std			T°m	1240	1247	1244	ac(N5)	std
SiO_2	49.17		42.20	41.97	41.72	41.44	44.14	41.72	42.04	40.33	42.05	0.95		49.05		47.36	50.28	48.89	49.46	1.44
TiO_2	1.85		2.33	2.17	2.43	2.16	2.18	2.48	2.36	2.59	2.36	0.15		1.79		2.05	2.06	2.14	2.05	0.10
Al_2O_3	17.66		17.93	17.80	17.57	16.48	16.31	20.35	19.12	20.65	18.24	1.32		17.29		17.59	16.95	20.13	18.41	1.22
FeO*	9.23		10.78	9.56	10.62	13.70	11.76	7.64	9.55	9.61	10.54	1.53		9.54		11.99	10.07	8.66	9.17	1.89
MnO	0.16		0.13	0.11	0.11	0.20	0.15	0.18	0.18	0.16	0.14	0.03		0.16		0.25	0.23	0.15	0.19	0.05
MgO	4.29		7.96	7.44	7.75	8.79	7.30	6.06	7.81	7.24	7.40	0.80		4.63		4.91	4.27	3.84	4.02	0.62
CaO	9.05		11.81	12.93	12.29	10.26	10.75	13.45	11.77	12.49	11.78	0.98		9.37		5.95	6.33	7.04	6.58	0.55
Na_2O	4.9		4.34	4.60	4.37	4.30	4.70	4.91	4.46	4.23	4.53	0.30		4.84		6.8	5.99	5.43	6.39	0.66
K_2O	1.84		1.50	1.40	1.47	1.32	1.38	1.55	1.60	1.34	1.44	0.10		1.63		2.17	2.63	2.32	2.54	0.35
P_2O_5	1.24		1.01	1.97	1.66	1.33	1.31	1.64	1.11	1.32	1.51	0.27		1.27		0.92	1.14	1.37	1.16	0.16
mg#	45		57	58	57	53	53	59	59	57	56	2.39		46		42	43	44	44	1.66

Fo: host olivine composition
T°m: Temperature of melting
FeO*: total iron
ac: averaged composition of N inclusions

std: standard deviation
10b: outlier
mg#: $Mg/(Mg+Fe_{tot})*100$

mol% in the AAC rocks, and from 66 to 78 mol% in the TU ones (Table 1).

In the alkali basalt from Santa Venera («low P» group) olivine is one of the most abundant phases. It is usually idiomorphic, and reaches up to 10 mm across; its Fo content varies from 87 to 74 mol% (Table 1). In detail high Mg olivine crystals (Fo>80%) sometime exhibit well developed kink bands, like those observed by Sachs and Scribano (1985) in olivine crystals from a hawaiite from an ancient alkaline sequence of Mt. Etna. This kind of deformation is relatively common for olivine grown at depth, and was recently described in olivine crystals from Piton de la Fournaise (Bureau *et al*., 1998).

Clinopyroxene

In the AAC «high P» and TU hawaites, phenocrystals are similar to the clinopyroxene commonly found in Etnean alkalic volcanics, generally defined as augite; on the grounds of the proportions of «Quadrilateral» (Quad) and «Others» components, according to Morimoto (1988) it is here classified as an an aluminian ferrian diopside (Al-Fe^{3+} Di) next to the boundary of the augite field. It is generally euhedral, up to 2-3 mm across, with its mg# ranging from 75 to 82, in the AAC hawaiites and from 73 to 80 in the TU ones (Table 2).

Clinopyroxene, the most abuondant phenocryst phase in the alkali basalt from Santa

TABLE 2

Selected compositions of pyroxenes.

Sample	SV1A						TM4A			TM12D			
Px	Cr-Al Di			Al-Fe^{3+} Di									
SiO$_2$	51.87	52.74	52.24	51.22	48.75	46.76	48.50	49.50	49.00	50.30	49.81	46.86	47.46
TiO$_2$	0.55	0.31	0.36	0.97	1.80	1.43	1.20	1.34	2.01	1.29	1.22	1.99	2.09
Al$_2$O$_3$	3.89	2.69	3.02	4.21	5.91	5.96	6.58	5.05	4.49	3.60	4.28	5.90	5.02
Cr$_2$O$_3$	0.22	0.54	0.93	0.68	0.02	0.08	0.24	0.12	0.02	0.00	0.06	0.00	0.03
Fe$_2$O$_3$	2.27	0.07	0.92	2.49	3.45	6.05	2.21	3.85	4.58	3.65	2.57	3.60	4.32
FeO	2.88	3.76	2.91	4.01	4.11	2.77	4.21	·2.47	4.08	3.84	4.37	4.59	4.62
MnO	0.11	0.04	0.05	0.17	0.16	0.25	0.12	0.10	0.23	0.13	0.13	0.10	0.18
MgO	16.02	16.54	16.48	15.07	13.73	13.36	13.81	14.32	13.15	14.24	14.35	12.80	12.63
CaO	23.23	22.58	22.58	22.25	22.32	21.68	22.34	23.70	23.36	23.27	22.74	22.10	22.36
Na$_2$O	0.26	0.24	0.33	0.56	0.55	0.56	0.33	0.42	0.61	0.46	0.32	0.43	0.57
Total	101.30	99.50	99.82	101.63	100.81	98.88	99.53	100.88	101.53	100.79	99.84	98.38	99.30
Si	1.88	1.93	1.91	1.86	1.79	1.76	1.80	1.81	1.80	1.85	1.85	1.78	1.79
Ti	0.02	0.01	0.01	0.03	0.05	0.04	0.03	0.04	0.06	0.04	0.03	0.06	0.06
Al	0.17	0.12	0.13	0.18	0.26	0.26	0.29	0.22	0.20	0.16	0.19	0.26	0.22
Cr	0.01	0.02	0.03	0.02	0.00	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.00
Fe^{3+}	0.06	0.00	0.03	0.07	0.10	0.17	0.06	0.11	0.13	0.10	0.07	0.10	0.12
Fe^{2+}	0.09	0.12	0.09	0.12	0.13	0.09	0.13	0.08	0.13	0.12	0.14	0.15	0.15
Mn	0.00	0.00	0.00	0.01	0.01	0.01	0.00	0.00	0.01	0.00	0.00	0.00	0.01
Mg	0.86	0.90	0.90	0.82	0.75	0.75	0.76	0.78	0.72	0.78	0.79	0.72	0.71
Ca	0.90	0.89	0.89	0.87	0.88	0.88	0.89	0.93	0.92	0.92	0.90	0.90	0.90
Na	0.02	0.02	0.02	0.04	0.04	0.04	0.02	0.03	0.04	0.03	0.02	0.03	0.04
Wo	45.72	44.86	45.00	44.71	45.00	44.74	44.72	47.56	47.85	47.37	45.61	45.96	46.58
En	49.14	48.65	50.00	48.23	46.86	49.34	47.20	47.56	44.17	45.61	46.20	44.72	44.10
Fs	5.14	6.49	5.00	7.06	8.13	5.92	8.07	4.88	7.98	7.02	8.19	9.32	9.32
mg#	86	88	88	81	77	76	81	81	74	79	80	75	73

Cr-Al Di: Chromian aluminian diopside
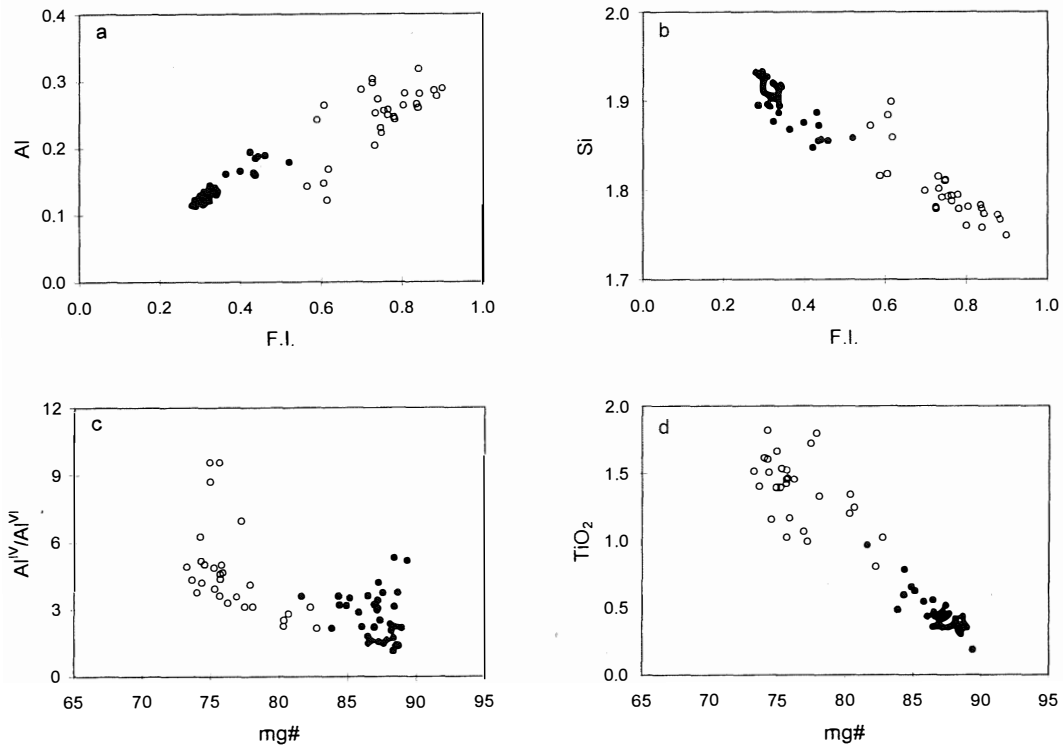Al-Fe^{3+} Di: Aluminian ferrian diopside

Fig. 4a,b – Al and Si versus F. I. respectively. c, d: Al^{IV}/Al^{VI} and TiO_2 versus mg# respectively. Symbols: Closed circle: Cr-Diopside; Open circle: Al-Fe^{3+} Diopside.

Venera, is usually euhedral, up to 0.5-10 mm in size. Its compositional range is very wide: two different pyroxenes may be recognised, as a chromian aluminian diopside (Cr-Al Di) joins to the more commonly found Al-Fe^{3+} Di. Phenocrystic Cr-Al Di, colourless with lower Al and Ti contents than Al-Fe^{3+} Di, often exhibits rounded edges with pale-brown homoaxial overgrowths with compositions similar to the Al-Fe^{3+} Di phenocrysts. mg# ranges from 89 to 80 for Cr-Al Di, and from 82 to 74 for Al-Fe^{3+} Di (Table 2); F.I. $[(Al+Fe_{tot}+Ti+Na)/(Cr+Mg)]$ (fig. 4a, b) is lower in Cr-Al Di than in Al-Fe^{3+} Di and is positively correlated to Al and negatively to Si. Figs. 4c,d show increasing Al^{IV}/Al^{VI} values and increasing TiO_2 with decreasing mg# respectively. Although compositions at the clinopyroxene edges were not investigated,

reference data (cf. Cristofolini et al., 1981) show that the Al^{IV}/Al^{VI} ratio and total Al gradually drop there with differentiation, because of the increased silica activity, and reach values comparable with the ones found in Cr-Al Di.

Spinel group

Various members of the spinel group were found in the analyzed samples:

Cr-spinel mainly enclosed in Cr-diopside and olivine (Fo$_{87-86}$) and very rarely as microphenocrysts only in the alkali basalt from Santa Venera, with spinel proper, as isolated micro-phenocrysts (Table 3);

microphenocrystic and groundmass Ti-magnetite very common in all of the analysed rocks (Table 4).

TABLE 3

Selected compositions of Cr-spinel and spinel.

Sample	SV1						
	T			M-P			
Fo	86.44	85.89	87.08	86.90			
TiO_2	1.04	1.07	1.02	0.84	0.77	0.33	0.37
Al_2O_3	28.73	22.89	23.85	20.71	53.71	58.89	57.88
FeO*	24.80	25.83	23.70	22.05	27.65	21.03	20.69
Cr_2O_3	30.64	36.65	35.68	37.40	0.13	0.06	0.10
MnO	0.15	0.31	0.18	0.13	0.23	0.06	0.14
MgO	13.70	12.04	13.17	12.63	16.07	19.03	18.74
Total	99.08	98.81	97.66	93.78	98.56	99.40	97.92
Al	2.60	2.15	2.23	2.04	4.40	4.60	4.59
Ti	0.06	0.06	0.06	0.05	0.04	0.02	0.02
Fe	1.59	1.72	1.58	1.54	1.61	1.17	1.17
Cr	1.86	2.31	2.24	2.46	0.01	0.00	0.01
Mn	0.01	0.02	0.01	0.01	0.01	0.00	0.01
Mg	1.57	1.43	1.56	1.57	1.67	1.88	1.88
Cr#	0.42	0.52	0.50	0.55	0.00	0.00	0.00

T: Trapped inclusion FeO*: total iron
M-P: Micro-Phenocryst Cr#: Cr/(Cr+Al)
Fo: Host olivine composition

TABLE 4

Selected compositions of Ti-magnetite.

Sample	SV1		TM4A			TM12D	
TiO_2	11.82	12.14	10.99	13.37	11.78	12.95	12.75
Al_2O_3	6.25	6.16	4.80	5.36	7.38	5.99	5.47
FeO*	72.28	71.68	75.58	72.34	71.94	68.35	69.70
Cr_2O_3	0.45	0.20	0.02	0.04	0.01	0.03	0.06
MnO	0.37	0.40	0.53	0.77	0.59	0.53	0.34
MgO	4.76	4.60	2.32	3.37	3.89	5.15	4.98
Total	95.93	95.18	94.24	45.24	95.59	92.98	93.31
Al	0.76	0.75	0.61	0.66	0.89	0.74	0.68
Ti	0.91	0.94	0.90	1.05	0.91	1.02	1.01
Fe	6.21	6.20	6.85	6.30	6.18	5.97	6.13
Cr	0.04	0.02	0.00	0.00	0.00	0.00	0.01
Mn	0.03	0.04	0.05	0.07	0.05	0.05	0.03
Mg	0.73	0.71	0.38	0.52	0.60	0.80	0.78

FeO*: total iron

The Cr-diopside and olivine hosted spinel compositions show a continuous range of Al_2O_3 (21-29 wt%), with low and constant TiO_2 contents (0.84-1.07 wt%) and Cr# (Cr/(Cr+Al) around 0.42 and 0.55. These compositions are alike those of the spinel in the Paternò transitional basalt (Al_2O_3 content around 20 wt% and narrow range of TiO_2 (0.84-1.07 wt%); Kamenetsky and Clocchiatti, 1996).

The spinel micro-phenocrysts show higher Al_2O_3 contents (54-59 wt%) with lower TiO_2 (0.33-0.77 wt%) and Cr_2O_3 (0.06-0.13 wt%) than the Cr-spinel.

Apatite

Apatite occurs as micro-phenocryst, as well as olivine-hosted inclusions, even in the Fo richest olivine crystals (Table 5), only in AAC «high P» and TU hawaiites.

All of the analysed apatites are fluorapatites with F contents ranging from 1.70 to 2.60 wt% and Cl/F ratio higher in micro-phenocrysts

(0.3-0.5) than in acicular olivine trapped crystals in (0.2-0.3), showing values similar to those already measured in apatites from other etnean hawaiites (Metrich, 1990).

The studied apatites are relatively homogeneous, P_2O_5 contents ranging between 40 and 43 wt%, CaO between 52 and 55 wt% respectively, and (Na_2O+SiO_2) between 0.25 and 0.45 wt%.

OLIVINE-TRAPPED FLUID AND MELT INCLUSIONS

Inclusions hosted in olivine crystals were investigated in three of the analyzed samples (SV1, TM4A, TM12D), taken as representative of the least differentiated rocks of the three previously defined groups.

Results of several studies were already published on fluid and melt olivine-hosted inclusions from Mt. Etna (Clocchiatti and Metrich, 1984; Metrich and Clocchiatti, 1989; Kamenetsky et al., 1986; Frezzotti et al., 1991;

TABLE 5

Selected compositions of apatite.

Sample	TM4A			TM12D			
	M-P		T		M-P		T
Fo			80.99	80.05			70.57
CaO	55.18	55.08	54.48	54.37	54.15	50.63	54.21
P_2O_5	43.10	42.14	42.23	43.04	41.48	37.92	42.56
SiO_2	0.23	0.22	0.29	0.24	0.32	5.88	0.18
Na_2O	0.10	0.09	0.04	0.07	0.12	0.16	0.13
La_2O_3	0.05	0.05	0.09	0.05	0.18	0.00	0.00
Ce_2O_3	0.20	0.28	0.00	0.37	0.31	0.10	0.39
F	1.70	1.78	2.12	2.15	1.93	2.12	2.58
Cl	0.55	0.86	0.42	0.63	0.86	1.12	0.71
Cl/F	0.32	0.49	0.20	0.29	0.45	0.53	0.28

T: Trapped inclusion
M-P: Micro-Phenocryst
Fo: Host olivine composition

Clocchiatti *et al.*, 1992; Metrich *et al.*, 1996); Kamenetsky and Clocchiatti, 1996).

Investigations on fluid inclusions in historical hawaiites show their origin at pressures higher than 0.1 GPa. More recent papers on tholeiites, and also on alkali basalts from Mt. Maletto (flank eruptions younger than 14 ka), indicate a crystal growth at pressures up to 0.6 GPa, on the ground of the density of primary CO_2 inclusions in olivine: the CO_2 entrapment depth is consistent with the presence of a deep reservoir beneath Mt. Etna suggested also by petrological evidence from clinopyroxene-bearing nodules, megacrysts and phenocrysts in some Etnean lavas (Scribano, 1989; Nimis *et al.*, 1996), geochemical and geophysical data (Sharp *et al.*, 1980; Him *et al.*, 1997, Tanguy *et al.* 1997).

Several studies showed the existence of olivine-hosted melt inclusions with different incompatible elements contents in the same rock samples and even crystals, and therefore suggested that cystals, grew from magmas generated by melting of a heteroqeneous source. Particularly, in tholeiites, melt inclusions were found with a continuous variation of compositions from high-K and P, Ne-normative to low-K and P ($P_2O_5 \approx 0.2$ wt%), Q-normative. This range in melt compositions cannot be accounted for by crystal fractionation alone and was probably generated during fractional melting of a heterogeneous source (Kamenetsky and Clocchiatti, 1996; Clocchiatti *et al.*, 1998).

The melting process, leading to residual harzburgite, was probably joined to penetration of fluids enriched in the most incompatible elements and volatiles, temporally and genetically linked to the mantle depletion in the basaltic component, caused by continuous melt extraction; these metasomatizing fluids induced significant chemical changes in residual mantle (refertilization of depleted mantle; Elthon, 1992). The shift of the Etnean magmas to alkali basalt compositions is then the witness of an increasing involvement of the metasomatically enriched component in melt generation (Kamenetsky and Clocchiatti, 1996).

Fluid inclusions

Olivine-hosted fluid inclusions were found only in the alkali basalt from Santa Venera. The triple point temperature is close to −56.6°C, indicating that CO_2 is the main fluid component. Primary and pseudosecondary inclusions (Roedder, 1984) can be observed: the former ones are randomly distributed in host olivine with size ranges from < 10 µm to 40 µm; the latter ones, concentrated along healed fractures, are less than 10 µm across. Pseudosecondary fluid inclusions are generally biphase (CO_2 liquid + CO_2 vapour), whereas primary fluid inclusions are monophase (CO_2 liquid) at room temperature.

Homogeneization occurs exclusively into the liquid phase at temperatures spanning from +7° to +29°C (fig. 5), with a CO_2 density indicating a maximum entrapment pressure of 0.45-0.55 GPa (Mäder and Berman, 1991), in agreement with pressures estimated by Nimis *et al.* (1996) for the low-Ti-augite megacrysts, and for the clinopyroxene, similar to Cr-Al Di, from nodules defined as wehrlites by Scribano (1989). A residing magma depth of 15-20 km, recently shown also by the study on fluid inclusions in olivine from Aci Castello tholeiites (0.5 Ma BP) and Mt. Maletto alkali basalts (< 14 ka BP) (Frezzotti *et al.*, 1991; Kamenetsky and Clocchiatti, 1996), agrees with preliminary seismic data on the presence of a deep magma chamber beneath Etna (Sharp *et al.*, 1980).

Melt Inclusions

Olivine-hosted melt inclusions from Mt. Etna display the broadest record of magma evolution prior than eruption (Clocchiatti and Métrich, 1984; Métrich and Clocchiatti, 1989). In this paper exclusively melt inclusions in olivine crystals were studied. They are either round-shaped or irregular and their size varies from 40 to 200 µm, and are in general isolated or arranged along fracture planes. All of the studied inclusions were wholly crystallized, into amphibole, apatite and opaque minerals (Table 6).
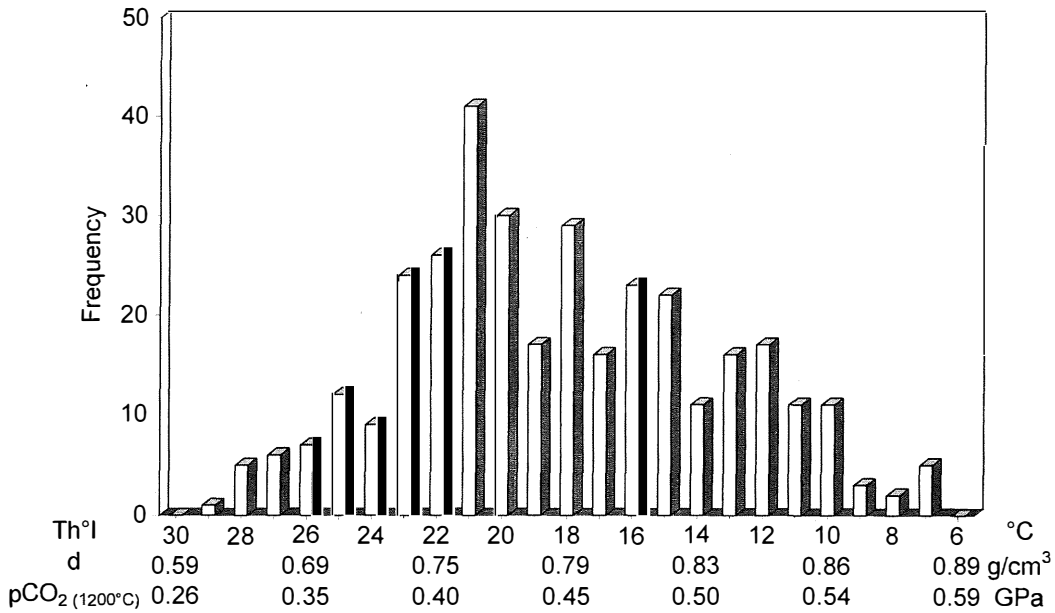
Fig. 5 – Olivine-trapped fluid inclusions from the S. Venera alkali basalt (AAC - «Low - P»): Temperatures of homogenization into liquid phase (T°*hl*), corresponding density (d) and estimated pressure (GPa) assuming a trapping temperature of 1200°C (Mäder and Berman, 1991).

TABLE 6

Selected compositions of daughter minerals in melt inclusions.

	Amphibole		Magnetite		Apatite
SiO_2	40.80	TiO_2	16.20	SiO_2	0.30
TiO_2	3.37	Al_2O_3	2.24	CaO	52.92
Al_2O_3	13.16	FeO*	72.90	Na_2O	0.24
FeO*	10.44	Cr_2O_3	0.14	K_2O	0.07
MnO	0.08	MnO	0.89	P_2O_5	40.68
MgO	14.07	MgO	2.09	Cl	1.96
CaO	12.19			F	n.a.
Na_2O	2.55				
K_2O	0.80				

FeO*: total iron   n.a. : not analyzed

Several processes are responsible for changes of the original compositions within melt inclusions, of which two are the main ones: i) growth of «daughter» mineral phases because of slow cooling after trapping (Clocchiatti and Massare, 1985); ii) wall reaction and postentrapment diffusion between melt inclusions and host mineral. The former process is reversible during homogenization experiments; the latter one depends on the cooling history and diffusion rates of elements involved. In particular, melt inclusions in olivine usually suffer Mg-Fe^{2+} exchange with the host crystal, but other major elements and their ratios are not affected. The original composition of trapped melt may then be approached for homogenized melt inclusions and used for geochemical characterization.

Melting temperatures

Homogenizing experiments were carried out at LPS-Saclay using an optical heating stage in He atmosphere purified with Zr at 800°C (Sobolev *et al.*, 1980). This method involves an initial preparation of the samples by cutting and polishing both sides of the host crystals. Temperatures, measured with a Pt-$Pt_{90}Rh_{10}$ thermocouple with an accuracy of 5°C and checked by measuring the melting temperature of gold in each run, were increased at rates of 0.9° to 0.01°C/sec, below and above 1000°C respectively. Melting temperatures (T°m) (Roedder, 1984) were read when the last of the daughter crystals was resorbed. Samples were then quenched in 1-2 sec, mounted and polished. Melting temperatures for the studied melt inclusions are correlated positively with Fo of the host olivine (fig. 6), and vary from 1240° to 1270° C. They are higher than those measured by Kamenetsky and Clocchiatti (1996), that might be due to reaction with host crystal walls and volatile loss during the original crystallization and the heating experiment (Nielsen *et al.*, 1998).

Melt compositions

Chemical analyses, on polished homogeneized olivine-trapped inclusions, were carried out at the University P. et M. Curie (Paris VI), by means of a CAMEBAX electron microprobe system, operating at 10 nA with a counting time of 10 sec and a beam of 10 μm. The composition of each inclusion was calculated averaging data from 3 to 15 points (Table 1).

The calculated KD value ($(Fe^{2+}/Mg)_{olivine}/(Fe^{2+}/Mg)_{melt}$ varies from 0.32 to 0.53), indicating that the equilibrium condition between host olivine and melt inclusions was not always reached during homogenization experiments (Roeder and Emslie, 1970; Nielsen *et al.*, 1998). This leads one to think that melting of the olivine inclusion wall occurred during the experiments: melt compositions were then recalculated by subtracting an amount of hosting olivine such as the calculated KD value results around 0.3 ± 0.03) (Nielsen *et al.*, 1998): an average amount of 7% olivine was subtracted.

Alkali basalt from S. Venera

The studied melt inclusions, trapped in olivine crystals with a very narrow Fo range (84.5-87 mol%), show mg# values in the range from 62 to 69, that may be considered consistent with near primary magma compositions (Wilson, 1989).

In the TAS diagram (Le Bas *et al.*, 1986; Le Maitre, 1989) (fig. 2) most of the analyzed inclusions fall within the alkali basalt field, slightly extending into the basanite and hawaiite fields. One of the inclusions (10b, Table 1) is considered as representing a clearly distinct subalkaline melt composition and is plotted separately.

Melt inclusions in the most magnesian olivines (Fo_{87}), showing a wide K_2O range (0.35-1.03 wt%) and P_2O_5 (0.61-1.67 wt%; figs. 6, 7), may coexist even within the same crystal (Table 1, 10a, 10b). The scatter in composition, especially for P_2O_5, may be considered real, since standard deviation is very low in most of the analyses (fig. 7). The observed variation is similar to that already observed for melt inclusions in olivine from the Aci Castello and Adrano tholeiites (Clocchiatti
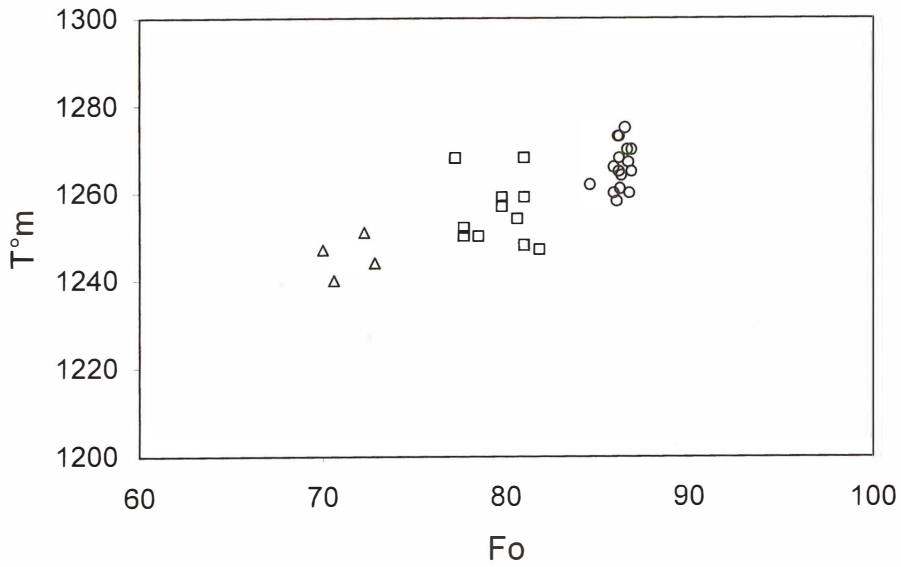
Fig. 6 – Relationship between the composition of host olivine and the temperature of melting (T_m°C) of the last daughter mineral phases in melt inclusions. Symbols: open circle olivine-trapped inclusions from the alkali basalt (AAC - «low P»); open square - AAC «high P» hawaiite; open triangle - TU hawaiite.
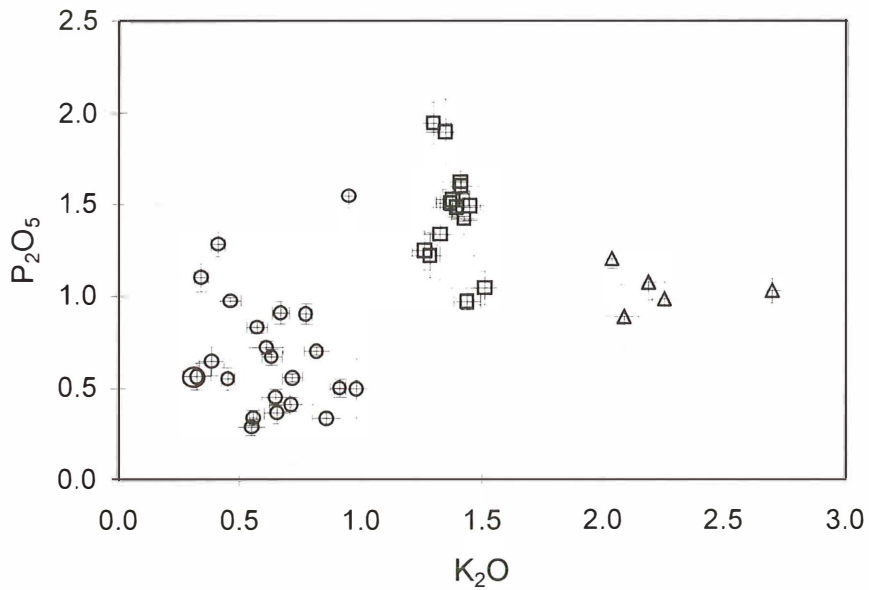

Fig. 7 – P_2O_5 versus K_2O for melt inclusions; bars: standard deviation. Symbols as in fig. 6; Double circle: outlier.

et al., 1992; Kamenetsky and Clocchiatti, 1996): high P_2O_5 contents are commonly related to higher Al_2O_3 and lower SiO_2 and CaO concentrations, sometimes to higher Na_2O and K_2O.

It is noteworthy that in the most evolved melt inclusions (mg# 64-62 in the less magnesian olivines Fo_{86-84}) the incompatible elements concentration (K, Na, P) are lower and less scattered; in particular P_2O_5 tends to the lowest values (Table 1, 8a, 8b; fig. 8).

«high P» hawaiites

Melt inclusions are trapped in olivine crystals with Fo_{81-77} for the AAC – «high P» hawaiites and Fo_{73-70} for the TU ones, and all of them plot in the TAS tephrite to phonoterphrite fields (Le Bas *et al.*, 1985; Le Maitre, 1989) (fig. 2).

Their mg# values, from 42 to 60, indicate that these melts may not be considered as primary magmas (Wilson, 1989): only melt inclusions with the highest mg# values might be considered next to primary compositions, and show incompatible element contents that may be compared with the ones of «high P» melt inclusions in olivine from the S. Venera Alkali basalt (Table 1; figs. 8, 9).

DISCUSSION AND CONCLUSIONS

Alkali basalt from S. Venera

The S. Venera alkali basalt shows very singular characters: as shown in fig. 4*a-b*, clinopyroxene F.I. values are correlated positively with Al and negatively with Si, these data suggest, according to earlier papers (Scribano, 1989; Nimis *et al.*, 1996), that Cr-Al Di (lower F.I. values) represents an early separate from a «primary» magma, whereas Al-Fe^{3+} Di crystallized from a more evolved magma (higher F.I. values). At the same time, the composition of the spinel hosted in Cr-Al Di and olivine (Cr# 0.42-0.55) matches the one found in the Paternò basalt (Kamenetsky and Clocchiatti, 1996).

The most striking feature is that melt inclusions widely differ in their compositions, especially for their little compatible elements (K_2O and P_2O_5, fig. 7) contents even in the most primitive Fo_{87} olivine in accord with earlier papers (Clocchiatti *et al.*, 1992; Kamenetsky and Clocchiatti, 1996). Their overall compositions and K_2O/P_2O_5 values cover a wide range from alkali to transitional basalts (figs. 2, 3, 7 and 8), suggesting that these olivine crystals grew from distinct magmas, generated by partial melting of a heterogeneous source (Clocchiatti *et al.*, 1998). The presence of high-mg# Cr-Al Di (mg# 80-89) associated with high-Fo olivine is an evidence of their co-crystallization at an early stage.

The deformation of high-Mg olivine (Fo>80 %mol) crystals, according to the experimental results (Kohlstedt and Goetze, 1976), suggests that they formed at high pressure. In particular, recent studies (Bureau *et al.*, 1998), on scoria and crystal-rich lapilli samples from Piton de la Fournaise volcano, showed that olivine kink banding is related to crystal deformation due to reservoir replenishment by new magma after they crystallized 15 km deep or more. A first stage of high pressure crystallization is also suggested by CO_2 inclusion data (fig. 5) that show entrapment at up to 0.55 GPa, corresponding to a depth of at least 15 km, in accord with what obtained for Aci Castello and Mt. Maletto fluid inclusions (Kamenetsky and Clocchiatti, 1996), and supported by petrological evidence from clinopyroxene-bearing aggregates and megacrystals (Sachs and Scribano, 1985; Scribano, 1989) and with geochemical data (Tanguy *et al.*, 1997). Seismic data (Sharp *et al.*, 1980; Hirn *et al.*, 1997) suggest the presence of large volumes of partial melt residing beneath the volcano at depths down to 15-20 km, where possibly distinct (Na-alkaline to transitional) basic parental magmas, risen from the upper mantle and residing in a complex fissure system, might evolve by crystal fractionation and mixing.

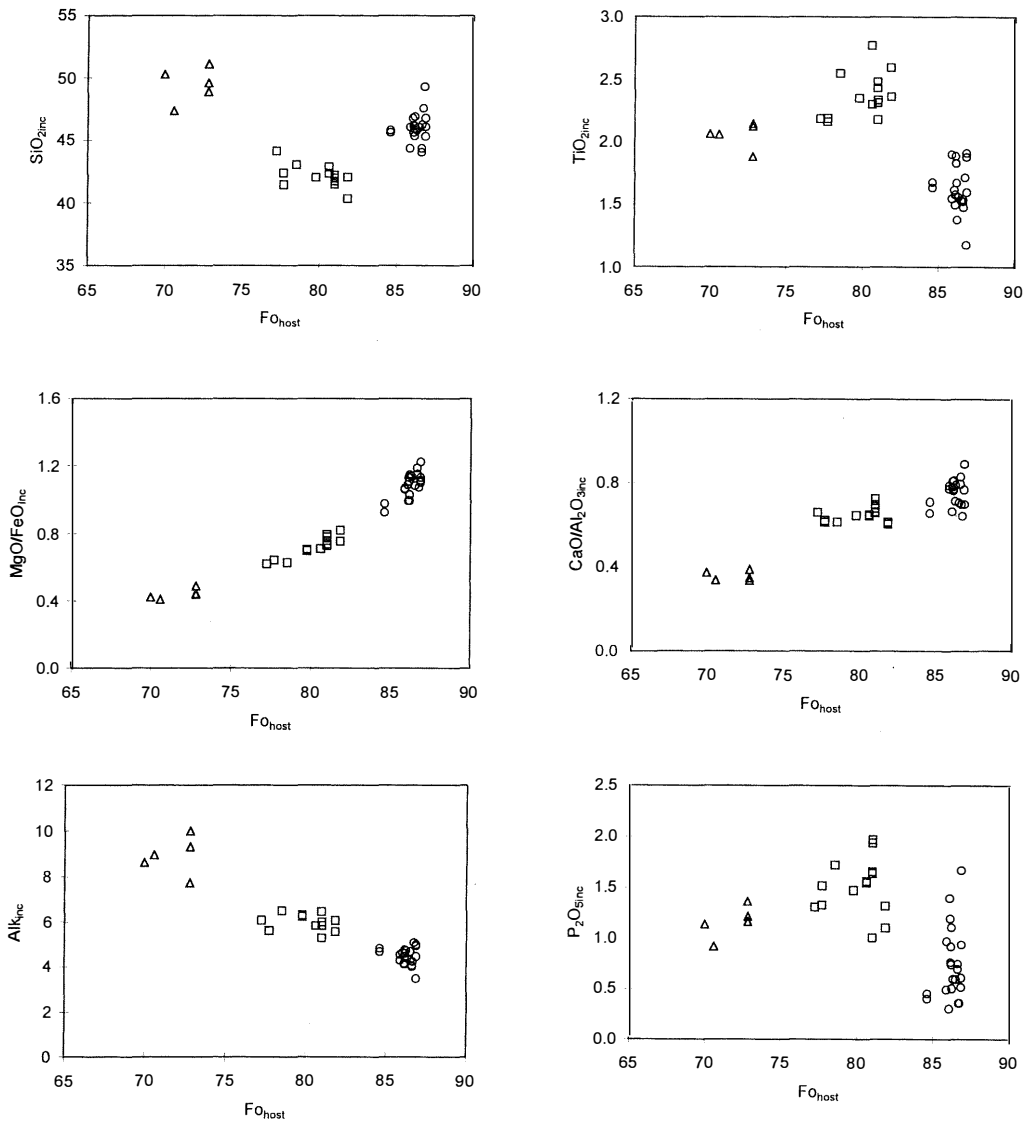Most of the fluid inclusions however were

Fig. 8 – Melt inclusion compositions; see text for discussion. Symbols as in fig. 6.

enclosed at 0.4 Gpa and less (fig. 5), giving evidence that the hosting crystals formed at an early stage of high pressure crystallization from a P-rich alkalic magma, and later at higher levels (0.4 - 0.3 GPa) from a distinct melt. Melt inclusion data suggest that at this later stage the magma was mainly transitional and chacterized by relatively low K and P contents (figs. 2, 3 and 7).

Clinopyroxene evolves towards increasingly higher Al_2O_3 and lower SiO_2 concentrations (Al-Fe^{3+} Di), typical of basalt magmas with relatively low silica activity (Kushiro, 1968; Gupta et al., 1973).
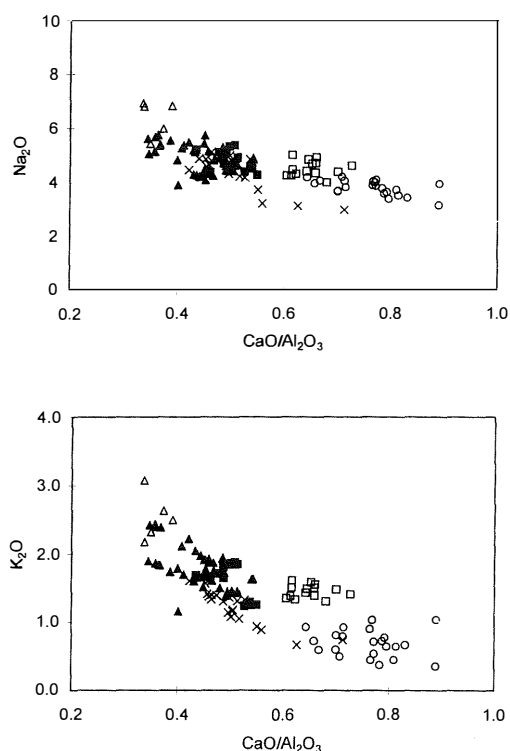
Fig. 9 – CaO /Al$_2$O$_3$ vs. Na$_2$O and K$_2$O in inclusions and in lavas. For symbols see figs. 2 and 6.

With reference to zoned pyroxene compositions, experimental results on simplified systems (Gupta *et al.*, 1973) and several studies about natural clinopyroxenes from alkaline igneous rocks (Kushiro, 1968; Thompson, 1974; Wass, 1979) suggest that pressure changes during fractional crystallization may account for variations in the clinopyroxene chemistry from Cr-diopside to Al-diopside, similar to the ones found in the S. Venera basalt. According to Wass (1979) increasing Al^{IV}/Al^{VI} values and increasing TiO$_2$ with decreasing mg# (fig. 4c-d) indicate the general trend from high to low pressure clinopyroxenes.

The overall data show then that the phenocrysts in the S. Venera basalt formed under polybaric conditions from magmas originated from heterogeneous sources.

«high P» hawaiites

The compositional variations of tephrite to phonotephrite melt inclusions in olivine crystals of these rocks may be interpreted as mainly depending on crystal fractionation, as shown in fig. 8, where their oxide concentrations are plotted versus Fo content in host olivine. Even if some of the oxides (SiO$_2$, TiO$_2$ and P$_2$O$_5$) show a rather regular variation, an abrupt change between compositions of melts hosted in olivines from the AAC and TU hawaiites respectively is apparent for some other ones (e.g. K$_2$O), and for the CaO/Al$_2$O$_3$ ratio. The correlation Of SiO$_2$ to Fo is negative, whereas the ones of TiO$_2$ and P$_2$O$_5$ are slightly positive; these correlations suggest a continuous fractionation of kaersutite (?) and apatite respectively, joined by olivine. The early appearance of apatite is also confirmed by its presence as inclusions even within the most magnesian olivines. Although the FeO, MgO contents are not the original ones, due to host olivine wall partial melting during homogneization experiments (see fig. 8), inclusions show a marked drop of these and of the CaO/Al$_2$O$_3$ values, at the
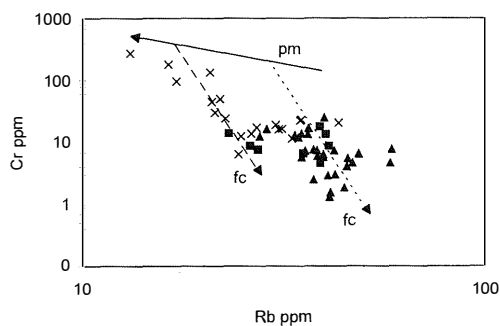

Fig. 10 – Cr versus Rb in the analyzed lavas (Cocherie, 1986); pm: partial melting, fc: fractional crystallization; arrows point toward increasing degrees of melting and differentiation respectively. Symbols as in fig. 2. The least differentiated lavas of AAC are aligned along a fractionation path intersecting the pm vector at a higher degree of melting than most of the «high P» - AAC and TU hawaiites. Some lavas plot joining the two fc lines, and suggest then that mixing occurred.

transition from the AAC to TU hawaiites, whereas K_2O (and possibly Na_2O) increase: if the regular decrease of the MgO/FeO_{tot} ratio is considered, these compositional changes between the two groups might be accounted for by a greatly increased proportion of Ca-bearing mafic minerals (clinopyroxene) among the fractionating phases.

It is then possible to conclude that melt inclusions (fig. 2, fields 2 and 3) from the AAC and TU hawaiites appear as originated by differentiation of tephrite-basanite magmas with similar compositions. In particular, the magma evolution recorded by the inclusions appears as related to continuous fractional crystallization of olivine and minor other phases (apatite), along the whole range of compositions, with clinopyroxene appearing at the transition of the melts, represented by AAC olivine inclusions to the TU ones.

According to Sun and Nesbitt (1977) and Maaloe and Aoki (1977) the mantle is generally P-depleted (200 ppm) with reference to the chondritic value (900 ppm). Only magma produced by very low degrees of partial melting could then be saturated in apatite: assuming a P bulk partition coefficient around 0.03 (Anderson and Greenland, 1969), magma produced by a source-melting of 2%, or higher, contains less than 1 wt% P_2O_5. With a lower bulk partition coefficient (0.01; Green, 1964) it would be possible to obtain alkalic magmas with P_2O_5 around 2 wt%, only at 1% partial melting (Watson 1980).

The phonotephrite melt inclusions in the TM4A hawaiite show P_2O_5 contents ranging between 1 and 2 wt% (Table 1): recalling that alkali basalt (higher partial melting degree) inclusions in SV1 contain up to 1.6 % P_2O_5, these concentrations may then be accounted for by an origin of «high P» magmas through partial melting of a P-enriched mantle source: according to Tanguy *et al.* (1997) the shift from tholeiites towards alkali basalts is related to decreased melting degrees of a larger mantle volume, that was likely composed of a metasomatized peridotite.

Relationship between melt inclusions and rocks in the «low P» and «high P» successions

Compositions of olivine-trapped melt inclusions and rocks are compared in figs. 3 and 9, where P_2O_5 concentrations are plotted versus SiO_2, and those of incompatible elements (Na_2O, K_2O) versus CaO/Al_2O_3 values respectively.

The highest mg# melt inclusions (17a,b), trapped in olivines (Fo_{82}) from the «high P» hawaiite, show P_2O_5 and MgO contents, and CaO/Al_2O_3 values that may be compared to the ones of the P-rich melt inclusion (10a) from olivine (Fo_{87}) crystals of the S. Venera alkali basalt (Table 1), but have lower SiO_2 and higher TiO_2, FeO, and alkalies (fig. 8). The transition from melt compositions hosted in olivine of the «low P» basalts to the ones of the «high P» lavas does not appear possible by simple fractional crystallization processes: in the TAS diagram (fig. 2) the averages of olivine-trapped inclusions from the «high P» AAC hawaiite and from the S. Venera alkali basalt («low P» – AAC), and the inclusion 10b (outlier from the S. Venera alkali basalt) are alined following a trend with SiO_2 negatively correlated to the alkali content, from phnolite to transitional compositions. This might be consistent with an origin of the, primitive magmas by different degrees of partial melting; the wide scatter of P_2O_5 contents, particularly in the S. Venera basalt melt inclusions, might however Support the hypothesis of magmas originated from heterogeneous sources (see 8.1; fig. 6).

There is then strong support for the hypothesis that the parents of the two groups («high P» and «low P» lavas) were originally distinct, and may have generated by different degrees of partial melting of heterogeneous sources.

Mixing at different stages of differentiation of the various magmas represented by the melt inclusions and further fractional crystallization process at crustal levels may account for the evolution of the different members of the two lava groups (figs. 2, 3 and 10):
the «high P» levels (of TU and AAC), lying

on a common variation trend (fig. 2 and 3), may be derived by fractional crystallization from a common original magma generated by mixing of phonolite and alkali basalt melts with high P_2O_5 contents, represented by the melt inclusions in olivine (fig. 2 and 3, fields 1, 2) from the S. Venera basalt (SV1) and the «high P» hawaiite (TM4A);

the «low P» lavas in the AAC sequence should have derived from transitional to alkaline magmas, present as «low P» melt inclusions, in olivine crystals of the S. Venera basalt (fig. 2 and 3, field 1).

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