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Origin of gabbro pegmatite in the Smartville intrusive complex, northern Sierra Nevada, California

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Abstract

Gabbro pegmatites are common in cumulate olivine gabbros in the Smartville complex, northern California. The pegmatites occur as pods and segregations and consist primarily of calcic bytownite or anorthite and clinopyroxene, with minor orthopyroxene and olivine. Amphibole is common both as a late magmatic phase and as a subsolidus alteration of magmatic amphibole and pyroxene. Many of the pegmatites are zoned, with anorthositic rims and pyroxenitic cores. Fine-grained, idiomorphic granular gabbro in some pegmatite cores may be analogous to the quench aplites reported from many granitic pegmatites.

The igneous mineralogy and the mineral chemistry of five pegmatite pods are indistinguishable from that of their host gabbros. This suggests that these pegmatites formed in place and in equilibrium with their gabbro host. One pegmatite intrusive dike is more mafic than its host gabbro, suggesting that this pegmatite melt did not form in place. The presence of gabbro "aplites," the zoning of the pegmatites, and the high temperatures calculated from coexisting pyroxenes support the idea that these pegmatites formed by crystallization of a melt rather than by subsolidus replacement. High fluid pressures are suggested by the compositions of coexisting olivine and plagioclase, by the presence of late magmatic amphibole, by the presence of fluid inclusions, and by the restriction of subsolidus hydrous alteration assemblages largely to the cores of the gabbro pegmatites.

We propose that the pegmatites formed from a mafic intercumulus melt in the presence of a fluid phase. The evolution of the fluid phase may have been enhanced by the drop in confining pressure that accompanied re-emplacement of the cumulate gabbros as crystal mushes. The disruption of the cumulate pile may have enhanced the migration and accumulation of the intercumulus melt.

Introduction

Gabbro pegmatites are a common, but little studied, constituent of mafic intrusive rocks. They have been reported from ophiolites (Moores, 1969; Evarts, 1977; Lindsley-Griffin, 1977; Allen, 1975; Wilson, 1959), layered mafic-ultramafic complexes (Hess, 1960; Taylor and Forester, 1979), Alaska-type ultramafic rocks (Irvine, 1974; James, 1971), gabbroic plutons in orogenic areas (Snoke et al., 1981; Smith et al., 1983; Lovering and Durrell, 1959; this study), and plutonic inclusions in calc-alkaline arc volcanic rocks (Stern, 1979; Conrad and Kay, 1984). The pegmatites range from kilometer-scale plutonic intrusions (Lindsley-Griffin, 1977) to segregations of a few centimeter-sized crystals within a normal gabbro (this report). They commonly occur as pipes, sills, dikes, and lenticular pods.

Most pegmatites reported in the literature contain calcic bytownite or anorthite, but their mafic mineralogy is more diverse. In some pegmatites, amphibole is the principal mafic constituent (Irvine, 1974); in others, pyroxene predominates (Hess, 1960). Pegmatitic ultramafic rocks consisting of various proportions of olivine, pyroxene, and magnetite were reported by Cameron and Desborough (1964) and Viljoen and Scoon (1985). Zoning toward more felsic compositions is rare, but has been reported from some gabbroic pegmatites in the northern Sierra Nevada (Lovering and Durrell, 1959).

The presence of pegmatite is commonly considered an indication that a fluid phase evolved in the magma (Jahns and Burnham, 1969). If one accepts this premise, the presence of gabbro pegmatite in gabbros from diverse geologic environments might suggest that the evolution of a fluid phase is a fundamental part of the crystallization and cooling history of these rocks. In spite of these implications for rocks generally considered to be poor in volatile species, few detailed studies of gabbro pegmatites have been reported (cf. Cameron and Desborough, 1964; Viljoen and Scoon, 1985). In particular, little is known of the geochemical relationship between gabbro pegmatites and their host rocks. This paper reports the textures, mineralogy, and mineral chemistry of several small gabbro pegmatites and their olivine gabbro host rocks from the

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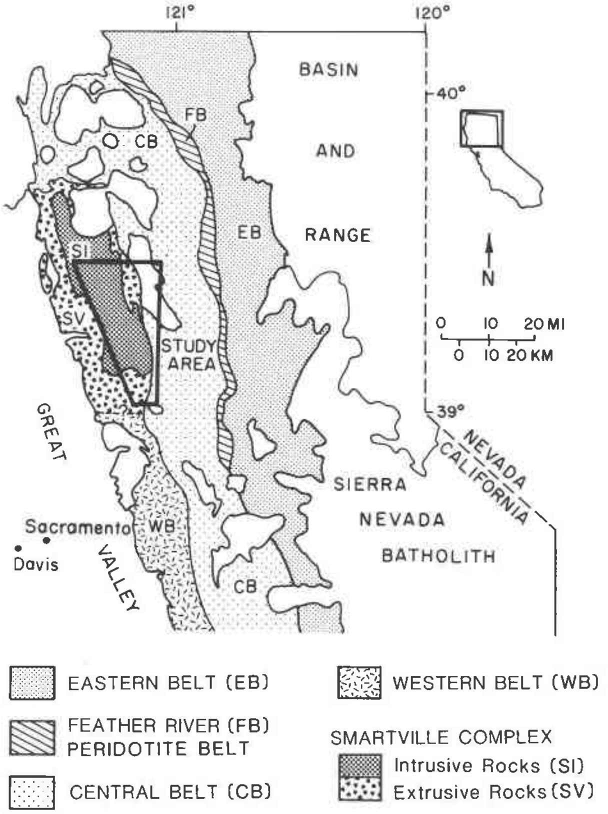

Fig. 1. Location of the Smartville complex in the northern Sierra Nevada. The study area is outlined.

Smartville complex. We will propose that these pegmatites are not subsolidus features, but mafic analogues of granitic pegmatites (Jahns and Burnham, 1969), perhaps implying hypersolidus evolution of a fluid phase in a mafic magma.

GEOLOGIC SETTING AND OCCURRENCE OF THE PEGMATITES

The Smartville complex is a rifted, Late Jurassic, ensimatic arc located in the northern foothills of the Sierra Nevada (Fig. 1). It consists of a carapace of mafic to intermediate volcanic rocks intruded by a plutonic and hypabyssal core (Moores, 1972; Beard and Day, 1982, 1983; Beard, 1985; Menzies et al., 1980; Xenophontos and Bond, 1978; Day et al., 1985; Xenophontos, 1984). The intrusive core of the Smartville includes a series of older metamorphosed massive diabases, fine-grained diorites, and gabbros. These were intruded by a sheeted dike complex and a series of coeval tonalite and gabbro-diorite plutons during intra-arc rifting (Beard and Day, 1983; Beard, 1985) (Fig. 2).

The gabbro-diorite plutons range from olivine gabbro intrusions, 200–1000 m in diameter, to elongate plutons, 25 km long, that are zoned from olivine gabbro in their cores to quartz diorite at their margins. Gabbroic pegmatites occur primarily in the olivine gabbros. Pegmatites of any composition are rare or absent in other rock types within the zoned plutons.

Host gabbro

Petrography. The olivine gabbros are fresh rocks with grain sizes ranging from 1 to 5 mm. Igneous foliation, defined by the preferred orientation of tabular minerals, especially plagioclase, is commonly well developed. Lensoidal, discontinuous anorthosite layers are present in some places. The gabbro consists of calcic bytownite or anorthite, clinopyroxene, and olivine (Fo_{65-80}). Orthopyroxene is present in minor amounts in most samples. Most orthopyroxene occurs as thin reaction rims around olivine. Some orthopyroxene in the reaction rims forms symplectic intergrowths with magnetite, suggesting some olivine reaction occurred at subsolidus temperatures (Haggerty, 1976; Ambler and Ashley, 1977). Chromian magnetite or magnetite is present in small amounts in most samples. Pargasitic amphibole is a common late-crystallizing interstitial phase. Some amphiboles are strongly zoned, and compositions in single grains may range from pargasite to actinolite. Biotite and apatite are minor accessory minerals in some samples.

Snoke et al. (1982) recognized a suite of calc-alkaline ultramafic to dioritic intrusive complexes in the Sierra Nevada and Klamath Mountains. On the basis of criteria laid out in their study and discussed in Beard (1985), the gabbroic plutons of the Smartville complex, including the olivine gabbros that host the gabbro pegmatites, are part of that suite. Arguments concerning the genesis, tectonic significance, and emplacement history of these intrusives are given in Snoke et al. (1982) and Beard (1985) and are beyond the scope of this paper. However, the conclusions of these studies relating to the olivine gabbros are relevant to this investigation. First, the olivine gabbros in the Smartville complex are mineralogically and geochemically identical to olivine gabbro cumulates found as xenoliths in volcanic rocks and in intrusive complexes in modern arcs (Beard, 1985; Beard and Day, 1984; also see Chivas, 1977; Stern, 1979; Arculus and Wills, 1980). It appears, therefore, that the olivine gabbros of the Smartville zoned plutons and of the Sierra-Klamath intrusive complexes, in general, are cumulate rocks that do not represent liquid compositions. Secondly, the Smartville olivine gabbros, along with the other cumulate rocks in the Sierra-Klamath intrusive complexes, lack cumulate structures and textures and commonly have brecciated contacts or otherwise deformed wall rocks (James, 1971; Snoke et al., 1981; Beard, 1985). Ultramafic dikes, including dunites, are not uncommon. The cumulate rocks of the Sierra-Klamath intrusive complexes, including the Smartville gabbros, are interpreted as deep-seated cumulate piles that were mobilized and re-emplaced as crystal mushes at a higher level in the crust.

Occurrence and structure of the pegmatites

Gabbro pegmatites occur in the olivine gabbros as podiform or, rarely, tabular bodies less than 1 m in maxi-

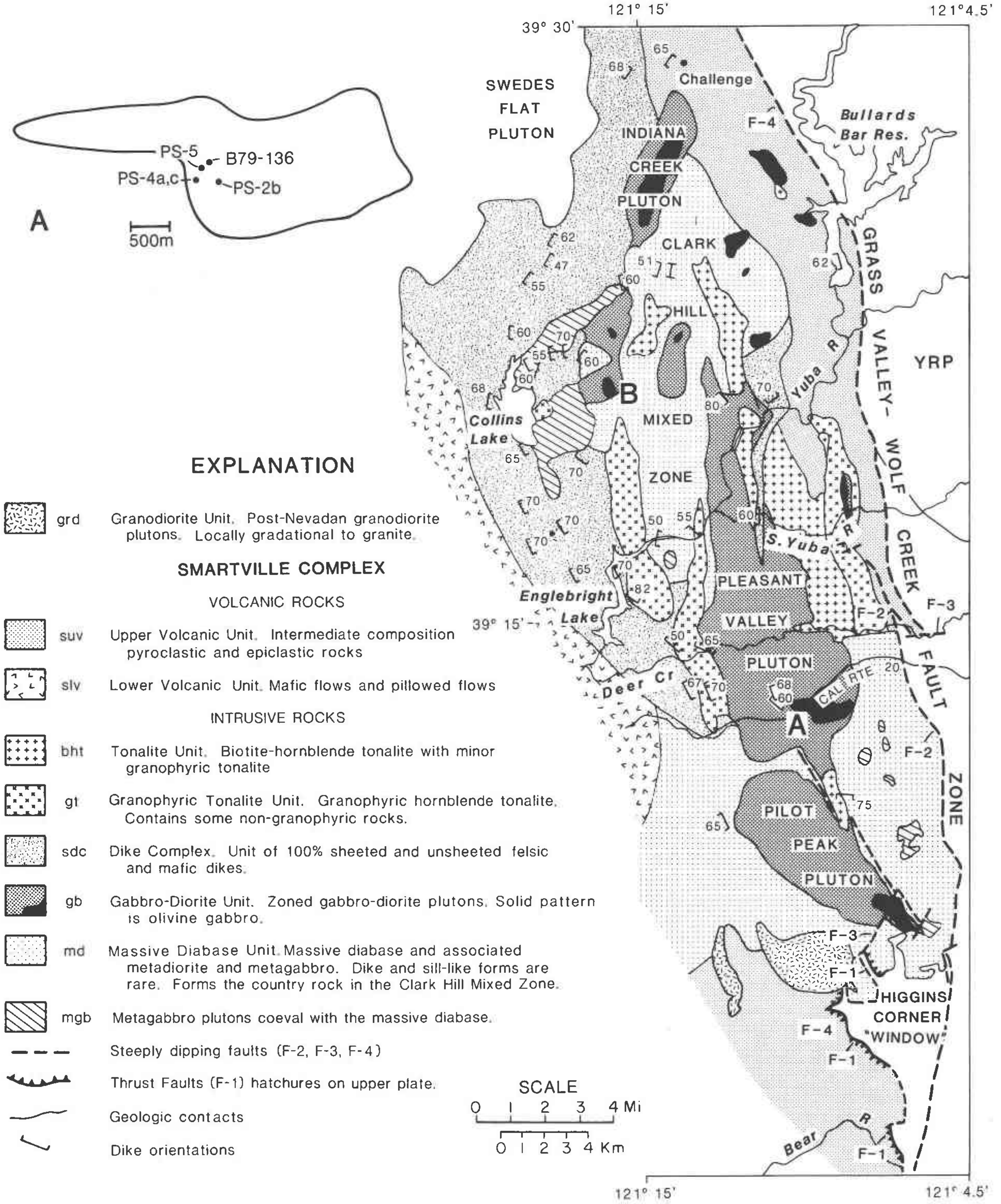

Fig. 2. Geology of the Smartville intrusive complex. Pegmatites occur in olivine gabbro intrusions and in the olivine gabbro cores of the larger gabbro-diorite plutons. Inset A (enlargement of loc. A) shows sample locations in the olivine gabbro core of the Pleasant Valley pluton. Location B is where sample PS-1 was collected.

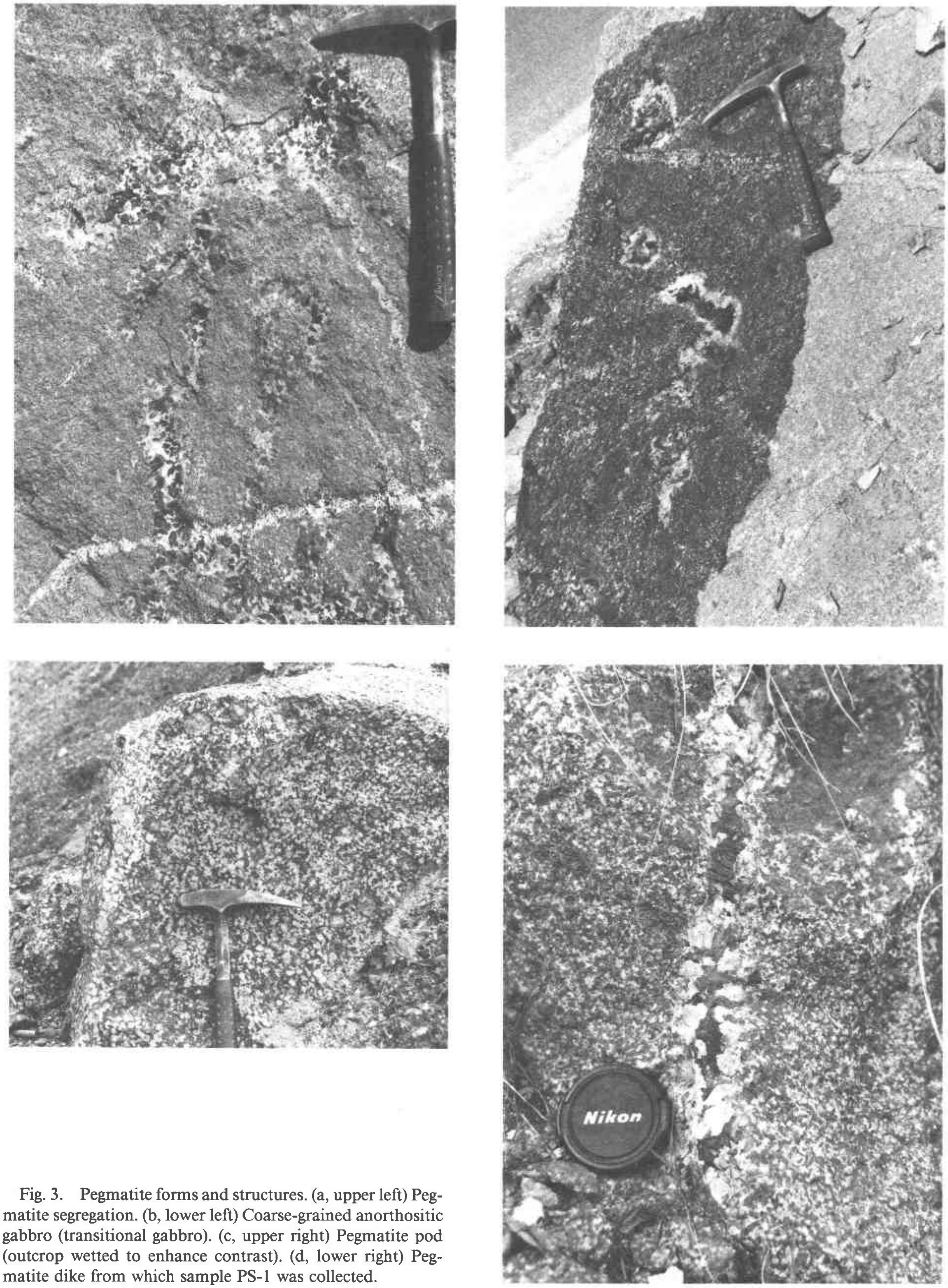

Fig. 3. Pegmatite forms and structures. (a, upper left) Pegmatite segregation. (b, lower left) Coarse-grained anorthositic gabbro (transitional gabbro). (c, upper right) Pegmatite pod (outcrop wetted to enhance contrast). (d, lower right) Pegmatite dike from which sample PS-1 was collected.

mum dimension. In some, areas up to 5% of the gabbro is pegmatitic. Six host-pegmatite pairs were collected for this study. One pair (PS-1) was collected from the olivine gabbro core of a small pluton just west of the northern end of the Pleasant Valley pluton (Fig. 2). Five other pairs (PS-2b, PS-4a, PS-4c, PS-5, B79-136) were collected from the olivine gabbro core of the Pleasant Valley pluton (Fig. 2).

The most common pegmatites in the olivine gabbros are irregular stringers and segregations (Fig. 3a). These range from clusters of a few crystals, 1 to 2 cm in size, to meter-scale segregations containing crystals up to 10 cm long. The contacts between the pegmatitic segregations and the host gabbro are gradational, marked by increasing grain size and, less commonly, by a decreasing color index toward the pegmatite. Anorthositic gabbro, consisting of 1- to 2-cm euhedral plagioclase crystals enclosed in clinopyroxene, occurs in the transition zone between some of the larger pegmatites and their host gabbro (Fig. 3b). There is no indication of either brecciation or chilling at the contact between pegmatites and their hosts.

Pegmatite pods are also common (Fig. 3c). They range in size from 5 to at least 30 cm. Their contacts with the host gabbros are sharp and are marked by an increase in plagioclase followed by an increase in grain size toward the pegmatite. Some of the pods are elongate and may be rods in three dimensions. Others have irregular shapes and differ from the segregations, described previously, mainly by having sharp contacts with their host. All of the pegmatites collected from the Pleasant Valley pluton are pods, although PS-2b, PS-4a, and PS-4c are somewhat irregular.

Pegmatite sills and dikes are rare. All of those observed had widths of less than 10 cm and many were only a single grain (1 to 2 cm) wide. Sample PS-1 was collected from the dike shown in Figure 3d.

Most pegmatite pods and dikes are zoned, with anorthositic rims and pyroxenitic cores. In granitic pegmatites, zoning has been shown to reflect the order of crystallization, with outer zones forming earlier and at higher temperatures than the inner zones (Taylor et al., 1979). The order of crystallization implied by the zoning of the gabbro pegmatites is the same as that observed in the transitional gabbros shown in Figure 3b.

A few of the pegmatites contain enclaves of finer-grained gabbroic material. In the irregular segregations, this material has the texture and mineralogy of the gabbro host and probably reflects the irregular contact between the pegmatite and its host in three dimensions. In the pegmatite pods, however, the included material is a finer-grained (0.2 to 2 mm), more nearly idiomorphic granular gabbro consisting of plagioclase and clinopyroxene. In some cases, clinopyroxene crystals up to 2 cm long are enclosed in the fine-grained gabbro (Fig. 4).

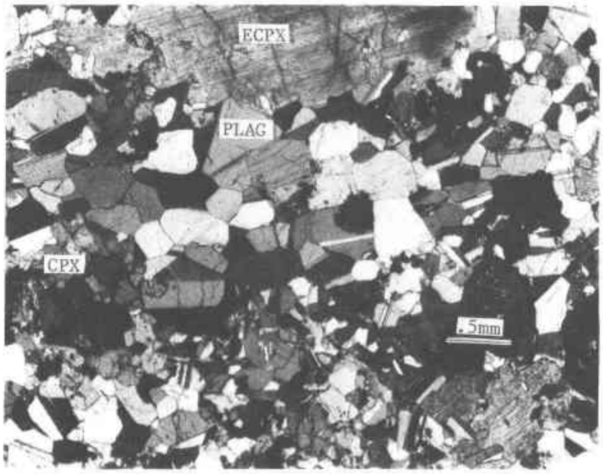

Fig. 4. Photomicrograph of aplite from the core of a small gabbro pegmatite. Note idiomorphic granular texture and gabbroic mineralogy. Plag = plagioclase; cpx = clinopyroxene; ecpx = early-formed, large clinopyroxene crystal floating in aplitic matrix.

Granitic aplites having textures similar to the fine-grained gabbro enclaves are common in the cores of granitic pegmatites. Granitic aplites form late in the crystallization history of pegmatites and have been interpreted as pressure quenches of the pegmatite melt brought about by the catastrophic loss of a fluid phase from the system (Jahns and Tuttle, 1963; Jahns and Burnham, 1969; Taylor et al., 1979). We suggest that the "aplitic" gabbro enclaves in the cores of Smartville gabbro pegmatites may have formed in a similar fashion.

Petrography of the pegmatites

The igneous mineralogy of the gabbro pegmatites is the same as that of their host gabbros. Clinopyroxene and calcic plagioclase are the dominant phases. Orthopyroxene is less abundant than clinopyroxene and makes up less than 5% of most pegmatites. Olivine is present in small amounts. Pargasitic amphibole occurs as interstitial crystals, as a partial replacement of clinopyroxene, and, rarely, as monomineralic veins that cut the pegmatites. Oxide minerals, biotite, and apatite are accessory phases.

Petrographic differences between the pegmatites and their host gabbros include the modal abundances and habits of orthopyroxene and olivine. Olivine is a minor phase in most pegmatites, occurring only as relics within large orthopyroxene crystals. Orthopyroxene is much more abundant than olivine in the pegmatites while the reverse is true in the host gabbros. Orthopyroxene occurs primarily as reaction rims on olivine in the host gabbros whereas discrete grains of orthopyroxene, as well as reaction rims, occur in the pegmatites.

Unlike the olivine gabbro host rocks, the pegmatites—and particularly the cores of the pegmatites—are altered. Augite and pargasitic hornblende are partially replaced by actinolite. Orthopyroxene is partially replaced by actinolite and/or chlorite. Plagioclase is commonly fresh but is locally sausseritized or replaced by carbonate. Olivine is partially replaced by tremolite, talc, chlorite, or serpentine.

Table 1. Plagioclase analyses

	PS-1		PS-2b		PS-4a		PS-4c		PS-5		B79-136	
	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host
SiO_2	45.32 (0.93)	45.84 (0.77)	45.48 (1.00)	45.77 (1.34)	45.67 (1.40)	45.53 (0.65)	46.37 (0.97)	45.42 (0.88)	44.81 (0.97)	44.62 (0.65)	45.54 (1.22)	45.68 (0.97)
Al_2O_3	35.16 (0.84)	34.83 (0.62)	34.29 (0.77)	34.10 (1.02)	34.65 (0.55)	34.59 (0.73)	33.87 (0.51)	34.61 (0.70)	35.24 (0.72)	35.12 (0.62)	34.84 (1.03)	34.94 (1.04)
FeO**	0.35 (0.41)	0.48 (0.14)	0.28 (0.30)	0.35 (0.13)	0.43 (0.26)	0.34 (0.28)	0.42 (0.10)	0.35 (0.12)	0.33 (0.09)	0.36 (0.08)	0.31 (0.29)	0.27 (0.08)
CaO	19.21 (0.82)	18.64 (0.48)	18.28 (0.35)	18.12 (0.55)	18.23 (0.59)	18.27 (0.53)	17.59 (0.62)	18.29 (0.47)	19.20 (0.92)	19.44 (0.38)	18.17 (0.81)	18.21 (0.48)
Na_2O	0.61 (0.50)	0.91 (0.20)	1.27 (0.27)	1.32 (0.38)	1.09 (0.37)	0.89 (0.22)	1.33 (0.39)	0.93 (0.36)	0.78 (0.48)	0.58 (0.17)	1.05 (0.47)	0.97 (0.34)
Total	100.60	100.70	99.52	99.66	100.07	99.79	99.66	99.68	100.42	100.14	99.92	100.07
Cations per eight oxygens												
Si	2.08	2.10	2.11	2.12	2.11	2.11	2.14	2.10	2.07	2.05	2.10	2.10
Al	1.90	1.88	1.87	1.86	1.88	1.89	1.84	1.89	1.91	1.91	1.89	1.90
Fe	0.01	0.02	0.01	0.01	0.02	0.01	0.02	0.01	0.01	0.01	0.01	0.01
Ca	0.95	0.92	0.91	0.90	0.90	0.91	0.87	0.91	0.95	0.96	0.90	0.90
Na	0.05	0.08	0.11	0.12	0.10	0.08	0.12	0.08	0.07	0.05	0.09	0.09
Total	5.00	5.00	5.01	5.01	5.00	5.00	5.00	5.00	5.01	5.01	5.00	4.99
An	94.6 (4.4)	91.6 (2.2)	89.5 (2.3)	88.3 (3.3)	90.3 (3.3)	91.9 (2.0)	88.0 (3.5)	91.6 (3.2)	93.2 (4.2)	94.9 (1.5)	90.5 (4.1)	91.2 (3.0)
Max. An**	97.0	93.1	90.6	90.8	92.4	93.2	90.0	94.3	95.6	96.6	93.5	93.3

Note: The number of analyses for all samples was 9, except for PS-1 (pegmatite, 8 analyses; host, 11 analyses). Analyses in weight percent; the numbers in parentheses are two times the calculated within-sample standard deviation. FeO* = total Fe as FeO. Small discrepancies in cation totals and calculated mineral components in this and subsequent tables reflect rounding errors.

** Max. An is the most calcic spot analysis from each sample.

Table 2. Olivine analyses

	PS-1		PS-2b		PS-4a		PS-4c		PS-5		B79-136	
	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host
SiO_2	38.98 (0.43)	38.19 (0.48)	37.72 (0.42)	38.13 (0.45)	38.13 (0.42)	37.98 (0.23)	38.06 (0.58)	38.14 (0.34)	38.82 (0.08)	38.69 (0.06)	38.22 (0.09)	38.24 (0.06)
FeO*	20.47 (0.56)	24.38 (0.46)	26.42 (0.48)	26.21 (0.38)	24.20 (0.35)	24.23 (0.32)	23.75 (0.43)	23.81 (0.57)	21.67 (0.53)	21.06 (0.20)	34.64 (0.39)	24.62 (1.05)
MnO	0.58 (0.03)	0.52 (0.16)	0.66 (0.09)	0.58 (0.02)	0.58 (0.11)	0.53 (0.09)	0.55 (0.06)	0.63 (0.18)	0.57 (0.08)	0.52 (0.04)	0.57 (0.12)	0.62 (0.13)
MgO	40.37 (0.24)	36.84 (0.30)	35.05 (0.58)	36.75 (0.26)	37.30 (0.65)	36.96 (0.40)	37.72 (0.44)	37.42 (0.04)	39.52 (0.47)	39.75 (0.31)	36.80 (0.43)	37.01 (1.03)
Total	100.41	99.93	99.85	100.67	100.21	99.70	99.89	100.00	100.64	100.03	100.24	100.48
Cations per four oxygens												
Si	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00	1.00
Fe	0.44	0.54	0.59	0.58	0.53	0.53	0.52	0.52	0.47	0.46	0.54	0.54
Mn	0.01	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Mg	1.55	1.44	1.39	1.40	1.46	1.45	1.47	1.46	1.52	1.53	1.44	1.44
Total	3.00	3.00	3.00	3.00	3.00	3.00	3.00	3.00	3.00	3.00	3.00	3.00
Fo**	77.9 (1.3)	72.9 (0.3)	70.3 (0.1)	70.8 (0.3)	73.3 (0.5)	73.1 (0.2)	73.80 (0.5)	73.7 (0.1)	76.5 (0.5)	77.1 (0.1)	72.7 (0.3)	72.8 (1.3)

Note: The number of analyses for all samples was 3. Analyses in weight percent; the numbers in parentheses are two times the calculated within-sample standard deviation. FeO* = total Fe as FeO.

** Fo was calculated from the atomic ratio Mg/(Mg + Fe).

Planar arrays of secondary fluid inclusions are present in some samples of plagioclase. A reconnaissance study of eleven polished thick sections suggests that fluid inclusions are restricted to plagioclase and are abundant only where there is independent evidence for late-stage alteration. Only a few of the fluid inclusions are as large as 2 μm in diameter and most were much smaller. Bubbles are mostly less than 10% by volume of the inclusion, consistent with low temperatures of entrapment, and daughter minerals are rare.

MINERAL CHEMISTRY

Olivine, plagioclase, augite, orthopyroxene, and amphibole were analyzed using a Kevex energy-dispersive spectrometer on an ARL-EMX-SM electron microprobe located in the geology department at the University of California, Davis. Operating conditions were 15-keV filament voltage and 300-nA beam current. Beam diameter was 3–10 μm for the mafic phases and 15–30 μm for plagioclase. Unless otherwise noted, tabulated analyses are averages of three to twelve several spots from one to three grains. MAGIC IV (Colby, 1968) was used to reduce the data. All minerals except amphibole were analyzed for all six host and pegmatite pairs.

Plagioclase. Plagioclase analyses are summarized in Table 1. In all cases the plagioclase is very calcic, ranging in composition from An_{88} to An_{95} in both the gabbros and pegmatites. Plagioclase contains between 0.3 and 0.5 wt% FeO. K_2O was below the detection limits (0.2 wt%) of the energy-dispersive system. Although core-to-rim zoning was not detected optically, anorthite content varied by 1.5 to 4.4 mol% in most samples, suggesting some inhomogeneity. Plagioclase compositions in the pegmatite and host are the same for all of the analyzed pairs within the limits of sample variation.

Olivine. Olivine compositions are summarized in Table 2. Olivine is unzoned and ranges between Fo_{70} and Fo_{78} in the pegmatites and between Fo_{71} and Fo_{77} in the host gabbro. CaO and NiO contents of the olivine were below detection limits of the EDS (less than 0.2 and 0.3 wt%, respectively). MnO content ranges from 0.5 to 0.7 wt%. In five of the samples studied, the difference in olivine compositions between the pegmatites and the host gabbro is less than 0.6 mol% Fo. Pegmatite PS-1, however, contains olivine considerably more magnesian that that in its host gabbro.

Clinopyroxene. Clinopyroxene analyses are given in Table 3, and quadrilateral compositions are plotted on Figure 5. Clinopyroxene quadrilateral compositions are $En_{43-45}Fs_{8-12}Wo_{43-47}$ in the pegmatites and $En_{43-45}Fs_{10-14}Wo_{43-47}$ in the host gabbros. With the exception of pegmatite PS-1 ($En_{45}Fs_8Wo_{47}$), the ranges for pegmatite and gabbro are virtually identical. The Al_2O_3 content of the clinopyroxene ranges from 2.2 to 3.0 wt%. Cr_2O_3, Na_2O, and MnO contents are less than 0.3 wt% and approach detection limits (0.2 wt%) in many cases. TiO_2 content ranges from 0.3 to 0.5 wt%. The total quadrilateral components within a single sample typically vary by 1 to 4 mol%. For most pegmatite and host pairs, within-sample variability (two sigma) exceeds any compositional difference between host and pegmatite. The clinopyroxene in pegmatite PS-1, however, is considerably more calcic and less ferroan than clinopyroxene in the host.

Orthopyroxene. Quadrilateral orthopyroxene compositions are $En_{73-78}Fs_{21-25}Wo_{1-3}$ in the pegmatites and $En_{74-78}Fs_{20-24}Wo_{2-3}$ in the host gabbro (Table 4, Fig. 5). Total quadrilateral components in most orthopyroxenes vary by 1 to 3 mol%. The within-sample variation (two sigma) of quadrilateral components exceeds any compositional difference between pegmatite and host.

Amphibole. Selected amphibole analyses are given in Table 5. Amphiboles from PS-2b, PS-4a, and PS-4c pairs and from B79-136 pegmatite were analyzed. The amphibole in all of these samples is characterized by strong compositional zoning. Optically, the zoning is manifested by a color change from brown or olive-green cores through a zone of green amphibole to green or blue-green actinolitic rims. Figure 6a is a plot of A-site alkalis vs. Si. Ti, like the total alkalis, varies inversely with the SiO_2 content of the amphiboles (Fig. 6b). These data show that the composition of the amphibole ranges from paragasitic hornblende to actinolite (Leake, 1978). The concentration of samples in the high- and low-Si regions of the diagrams largely reflects sample bias because cores and rims were preferentially analyzed. A microprobe traverse across one grain in sample B79-136 suggests that the complete range of compositions is present.

DISCUSSION

Equilibrium between host and pegmatite

Each of the gabbro pegmatite pods collected from the Pleasant Valley pluton contains minerals that are identical in composition, within the limits of observed variation, to the minerals in the host rock. The mineralogical similarities within each pegmatite and host pair are especially striking given the considerable variation of mineral compositions within the host gabbro pluton. For example, host gabbro samples PS-5 and PS-2b were collected within 200 m of one another yet olivine and plagioclase in PS-5 (Fo_{77} and An_{95}) are considerably more mafic than in PS-2b (Fo_{70} and An_{88}). The similarity of host and pegmatite mineral chemistry, therefore, suggests that equilibrium between host and pegmatite was attained on a local scale and that the pegmatites formed in place. This interpretation is consistent with the lack of intrusive contacts or intrusive form of the pegmatite pods.

In contrast, the mineral phases in the pegmatite dike, PS-1, are considerably more mafic than those in the host gabbro (olivine; Fo_{78} vs. Fo_{73}; plagioclase, An_{95} vs. An_{92}; clinopyroxene, $En_{45}Fs_8Wo_{47}$ vs. $En_{44}Fs_{13}Wo_{43}$; orthopyroxene, En_{78} vs. En_{75}). Sample PS-1 is the only pegmatite studied that has an intrusive form and may have crystallized from material that had equilibrated elsewhere in the cumulate pile.

Table 3. Clinopyroxene analyses

	PS-1		PS-2b		PS-4a		PS-4c		PS-5		B79-136	
	Pegmatite *N* = 7	Host *N* = 7	Pegmatite *N* = 6	Host *N* = 7	Pegmatite *N* = 6	Host *N* = 6	Pegmatite *N* = 7	Host *N* = 6	Pegmatite *N* = 6	Host *N* = 7	Pegmatite *N* = 6	Host *N* = 6
SiO_2	51.50 (1.73)	51.44 (0.30)	51.86 (1.40)	51.20 (0.67)	51.85 (0.70)	51.69 (1.38)	51.67 (0.75)	51.23 (0.59)	52.09 (0.68)	51.58 (0.55)	51.34 (1.35)	51.46 (0.83)
TiO_2	0.29 (0.30)	0.46 (0.09)	0.30 (0.34)	0.47 (0.08)	0.42 (0.18)	0.39 (0.24)	0.44 (0.17)	0.40 (0.13)	0.41 (0.15)	0.42 (0.18)	0.45 (0.17)	0.37 (0.13)
Al_2O_3	3.02 (2.18)	2.53 (0.34)	2.16 (0.99)	2.66 (0.58)	2.36 (0.48)	2.48 (0.98)	2.42 (0.80)	2.64 (0.49)	2.17 (0.22)	2.54 (0.17)	2.59 (1.07)	2.39 (0.61)
Cr_2O_3	0.12 (0.29)	0.16 (0.31)	0.14 (0.32)	0.15 (0.30)	0.16 (0.27)	0.14 (0.33)	0.23 (0.24)	0.28 (0.33)	0.19 (0.42)	0.29 (0.36)	0.13 (0.33)	0.04 (0.18)
FeO*	4.89 (1.85)	7.87 (0.58)	8.43 (2.63)	8.16 (1.52)	8.34 (1.23)	7.58 (1.30)	7.75 (1.25)	7.71 (1.48)	7.28 (0.85)	6.46 (1.87)	7.90 (1.51)	7.42 (1.87)
CaO	23.35 (2.57)	21.37 (0.86)	21.51 (2.86)	21.61 (1.73)	21.36 (1.38)	22.16 (2.01)	22.05 (2.04)	21.56 (1.83)	21.88 (1.29)	22.28 (2.40)	21.63 (2.38)	22.11 (2.66)
MnO	0.12 (0.26)	0.12 (0.26)	0.16 (0.26)	0.20 (0.28)	0.26 (0.32)	0.21 (0.27)	0.18 (0.27)	0.22 (0.25)	0.14 (0.28)	0.16 (0.26)	0.20 (0.20)	0.29 (0.08)
MgO	16.13 (1.40)	15.62 (0.49)	15.35 (1.32)	15.19 (0.59)	15.65 (1.13)	15.50 (0.92)	15.39 (0.72)	15.58 (0.76)	16.01 (0.91)	15.99 (1.19)	15.42 (1.28)	15.48 (0.83)
Na_2O	0.10 (0.22)	0.18 (0.15)	0.10 (0.10)	0.12 (0.20)	0.18 (0.12)	0.12 (0.14)	0.15 (0.16)	0.23 (0.16)	0.15 (0.17)	0.25 (0.24)	0.12 (0.27)	0.13 (0.14)
Total	99.53	99.73	100.01	99.74	100.59	100.26	100.28	99.84	100.34	99.96	99.78	99.67
Cations per six oxygens												
Si	1.89	1.90	1.92	1.90	1.91	1.91	1.90	1.89	1.91	1.90	1.90	1.91
Al^{IV}	0.11	0.10	0.08	0.10	0.09	0.09	0.10	0.11	0.09	0.10	0.10	0.01
Al^{VI}	0.02	0.01	0.01	0.02	0.01	0.02	0.01	0.01	0.01	0.01	0.01	0.01
Fe^{3+}	0.07	0.07	0.05	0.06	0.07	0.07	0.07	0.09	0.06	0.09	0.06	0.07
Ti	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Cr	0.00	0.01	0.00	0.01	0.01	0.00	0.01	0.01	0.01	0.01	0.00	0.00
Fe^{2+}	0.08	0.18	0.21	0.19	0.19	0.17	0.17	0.15	0.16	0.11	0.18	0.16
Ca	0.92	0.85	0.85	0.86	0.84	0.88	0.87	0.85	0.86	0.88	0.86	0.88
Mg	0.88	0.86	0.85	0.86	0.86	0.85	0.85	0.86	0.88	0.88	0.85	0.85
Mn	0.00	0.00	0.01	0.01	0.01	0.01	0.01	0.01	0.00	0.01	0.01	0.01
Na	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.02	0.01	0.01
En	45.2 (4.0)	44.1 (1.4)	43.2 (2.9)	43.0 (1.3)	43.9 (2.4)	43.4 (2.4)	42.7 (2.1)	44.0 (1.8)	44.7 (2.5)	44.9 (2.6)	43.6 (3.4)	43.6 (2.4)
Fs	7.7 (3.1)	12.5 (0.8)	13.3 (3.8)	13.0 (2.5)	13.1 (1.9)	12.0 (2.1)	12.2 (1.9)	12.2 (2.3)	11.4 (1.1)	10.2 (2.8)	12.5 (2.4)	11.7 (2.9)
Wo	47.0 (3.5)	43.4 (1.7)	43.5 (6.4)	44.0 (3.4)	43.1 (3.3)	44.7 (3.8)	44.5 (3.8)	43.8 (4.0)	43.9 (2.4)	45.0 (5.4)	43.9 (4.8)	44.7 (5.2)

Note: N = the number of analyses. Analyses in weight percent; the numbers in parentheses are two times the calculated within-sample standard deviation. FeO* = total Fe as FeO. Cations were normalized to four in order to calculate apparent Fe^{3+}. En, Fs, and Wo are calculated from the atomic ratios of Mg, Fe, and Ca.

P-T conditions of pegmatite formation

The pressures under which the pegmatites formed can be estimated from the metamorphic history of the Smartville complex and from the regional association and crystallization history of the gabbro-diorite plutons. Temperatures can be estimated from pyroxene and, qualitatively, from amphibole chemistry.

The pegmatites probably formed at relatively low pressures, almost certainly less than 5 kbar and possibly near 1 kbar. Beiersdorfer (1983) and Xenophontos (1984) suggested that the regional prehnite-pumpellyite facies metamorphism of the Smartville volcanic pile occurred at pressures between 1 and 3 kbar, although the low-pressure limit for such metamorphism is poorly constrained (Schiffman and Liou, 1980). The association of the gabbro-diorite plutons with a hypabyssal dike complex and with several plutons of granophyric tonalite also suggests emplacement at shallow levels (Beard and Day, 1983; Beard, 1985). The Al_2O_3 content of amphiboles from monzodiorites in the gabbro-diorite plutons suggests pressures of 3 ± 2 kbar (Beard, 1985; Hammarstrom and Zen, 1985).

Experimental work on a wide range of basaltic compositions has shown that amphibole is a hypersolidus phase under water-saturated conditions at pressures greater than 800–1000 bars (Yoder and Tilley, 1962; Eggler, 1972; Gilbert et al., 1982). The relationship between the upper-temperature stability limit of amphibole and the basalt solidus is a function of basalt composition and oxygen fugacity, but under all conditions studied, if the melt is water saturated, amphibole crystallizes if P_{H_2O} exceeds 2 kbar. The textures of the pargasitic amphiboles in the gabbro pegmatites (i.e., replacement of clinopyroxene, interstitial crystals) suggest that they formed at conditions just above and, perhaps, below the solidus (Wones and Gilbert, 1982). The late crystallization of amphibole in the gabbros and pegmatites is consistent with low-pressure crystallization.

The temperatures at which the pegmatites formed can be estimated from the compositions of coexisting clinopyroxene and orthopyroxene and from amphibole chemistry. Table 6 gives pyroxene temperatures calculated using the pyroxene geothermometer of Lindsley (1983). Temperatures were calculated using the 1-atm thermometer. Correction factors are approximately +4°/kbar for clinopyroxene and +3.5°/kbar for orthopyroxene. One-atmosphere clinopyroxene temperatures from the pegmatites range from 970 to 1016°C with a mean of 992°C, whereas those from the host gabbros range from 970 to 1020°C with a mean of 1001°C. Given the errors associated with the thermometers (inhomogeneity of pyroxene, Fe^{3+} calculations, analytical error, uncertainty in the thermometer), these values can be considered identical.

Orthopyroxene temperatures are considerably more variable and generally lower than those given by clinopyroxene. Pegmatite temperatures (exclusive of PS-1) range from 830 to 1030°C with a mean of 948°C. Host gabbro temperatures range from 800 to 1005°C with a mean of 906°C. Orthopyroxene in pegmatite sample PS-1 yields a temperature of 700°C, much lower than any other sample.

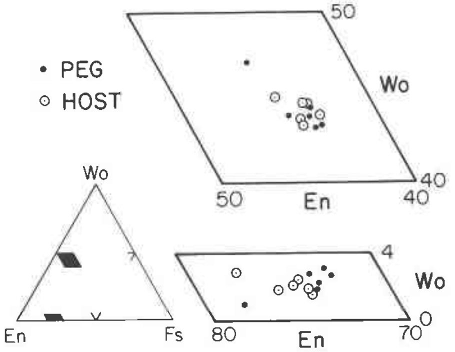

Fig. 5. Orthopyroxene and clinopyroxene compositions. En, Fs, and Wo components were calculated from atomic ratios of Mg, Fe, and Ca. En, Fs, and Wo were not corrected for the presence of additional components.

Five samples, including two pegmatites and three host gabbros, yield orthopyroxene temperatures less than 900°C. In all five cases, the analyzed orthopyroxene is from orthopyroxene-magnetite reaction rims around olivine, and the low temperatures clearly reflect the subsolidus origin of the reaction rims. For such samples, the two-pyroxene thermometer is not valid, and the clinopyroxene temperature may represent a minimum estimate.

The seven of twelve samples whose orthopyroxene and clinopyroxene temperatures are in agreement, within the 100° limits of combined error suggested by Lindsley (1983), give temperatures ranging from 975 to 1022°C with a mean of 990°C. The temperatures given by the two-pyroxene thermometer are 10° higher than the 1-atm basalt solidus (Wright and Okamura, 1977) and approximately 100° above the 1-kbar water-saturated solidus.

Oxygen-isotope data from pegmatite sample B79-136 are consistent with high temperatures of equilibration. Plagioclase, clinopyroxene, and whole-rock samples show $^{18}O/^{16}O$ enrichments (relative to SMOW) of 6.79, 5.93, and 6.49‰, respectively (A. Williams, written comm.). These data are typical magmatic values, providing no evidence for either high-temperature or low-temperature alteration. The fractionation between plagioclase (An_{90}) and clinopyroxene suggests an apparent temperature of 562°C (Matthews et al., 1983, p. 640), consistent with closure of the minerals with respect to oxygen exchange at lower temperatures than crystallization of the pyroxenes.

Amphibole compositions in the gabbros and pegmatites place only qualitative constraints on late magmatic and postmagmatic temperatures. Experimental work on a variety of basalt compositions has shown that amphiboles with Ti, Al, and A-site alkali contents similar to the pargasitic amphiboles analyzed here form at temperatures of between 700 and 1000°C at 5 kbar water pressure (Helz, 1973, 1976; Gilbert et al., 1982).

Table 4. Orthopyroxene analyses

	PS-1		PS-2b		PS-4a		PS-4c		PS-5		B79-136	
	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host	Pegmatite	Host
SiO_2	53.80 (0.74)	53.35 (1.87)	53.39 (0.94)	53.72 (0.93)	53.48 (0.52)	53.69 (1.06)	53.81 (1.07)	53.67 (0.42)	53.43 (0.69)	53.67 (1.51)	53.53 (0.31)	54.43 (0.08)
TiO_2	0.00 —	0.30 (0.05)	0.33 (0.12)	0.27 (0.05)	0.30 (0.26)	0.35 (0.14)	0.36 (0.04)	0.23 (0.40)	0.37 (0.08)	0.21 (0.10)	0.29 (0.10)	0.30 (0.06)
Al_2O_3	3.03 (0.54)	1.87 (0.54)	1.50 (0.11)	1.32 (0.21)	1.49 (0.18)	1.50 (0.13)	1.31 (0.28)	1.57 (0.13)	1.80 (0.27)	1.61 (0.22)	1.40 (0.17)	1.58 (0.33)
FeO^*	14.00 (0.16)	15.36 (0.54)	16.09 (0.75)	15.72 (0.72)	15.86 (0.93)	15.29 (0.38)	15.43 (0.31)	15.95 (0.49)	16.28 (0.34)	13.05 (1.29)	15.94 (0.29)	15.06 (0.17)
CaO	0.49 (0.39)	1.10 (0.08)	0.94 (0.24)	0.98 (0.25)	1.54 (1.40)	1.27 (0.64)	1.42 (0.29)	0.81 (0.46)	1.36 (0.09)	1.27 (1.40)	1.13 (1.16)	0.94 (0.06)
MnO	0.41 (0.06)	0.37 (0.08)	0.45 (0.06)	0.43 (0.21)	0.41 (0.14)	0.43 (0.06)	0.38 (0.05)	0.33 (0.23)	0.49 (0.08)	0.36 (0.11)	0.50 (0.06)	0.44 (0.03)
MgO	29.22 (0.52)	27.74 (0.40)	27.29 (0.62)	27.34 (0.62)	26.99 (0.74)	27.90 (0.31)	27.26 (0.33)	27.40 (0.71)	26.90 (0.69)	28.65 (0.99)	27.27 (0.67)	28.57 (0.20)
Total	100.95	100.10	99.98	99.79	100.06	100.45	99.97	99.97	100.60	98.83	100.04	101.31
	Cations per six oxygens											
Si	1.90	1.91	1.93	1.94	1.93	1.92	1.94	1.93	1.92	1.93	1.93	1.93
Al^{IV}	0.10	0.08	0.06	0.06	0.06	0.06	0.06	0.07	0.08	0.07	0.06	0.07
Fe^{+3IV}	—	0.01	0.01	0.01	0.01	0.02	0.01	—	0.01	—	0.01	0.01
Al^{VI}	0.02	—	—	—	—	—	—	0.00	—	0.00	—	—
Fe^{+3VI}	0.08	0.07	0.6	0.05	0.06	0.06	0.04	0.05	0.06	0.05	0.06	0.06
Ti	—	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Fe^{+2}	0.33	0.39	0.42	0.42	0.41	0.38	0.42	0.43	0.42	0.39	0.41	0.38
Ca	0.02	0.04	0.04	0.04	0.06	0.05	0.06	0.03	0.05	0.05	0.04	0.04
Mg	1.54	1.48	1.47	1.47	1.45	1.49	1.46	1.47	1.44	1.54	1.47	1.51
Mn	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.02	0.01	0.02	0.01
En	78.0 (1.0)	74.7 (0.5)	73.8 (1.0)	74.2 (0.2)	72.9 (1.4)	74.6 (1.1)	73.8 (0.2)	74.2 (1.2)	72.7 (0.6)	77.7 (1.2)	73.6 (1.9)	75.8 (0.6)
Fs	21.0 (0.3)	23.2 (0.3)	24.4 (1.3)	23.9 (0.4)	24.0 (1.2)	22.9 (0.2)	23.4 (0.3)	24.2 (0.5)	24.7 (0.7)	19.8 (1.6)	24.2 (0.5)	22.4 (0.3)
Wo	0.9 (0.8)	2.1 (0.1)	1.8 (0.1)	1.9 (0.6)	3.0 (2.7)	2.4 (1.2)	2.8 (0.5)	1.6 (0.9)	2.7 (0.1)	2.5 (2.8)	2.2 (2.2)	1.8 (0.3)

Note: The number of analyses for all samples was 3. Analyses in weight percent; the numbers in parentheses are two times the calculated within-sample standard deviation. FeO^* = total Fe as FeO. Cations were normalized to four in order to calculate apparent Fe^{3+}. En, Fs, and Wo are calculated from the atomic ratios of Mg, Fe, and Ca.

Table 5. Selected amphibole spot analyses

	PS-2b		B79-136			PS-4a				PS-4c		
	Host core	Peg. core	Pegmatite			Pegmatite		Host		Peg. Core	Host	
			Core	Inter.	Rim	Core	Rim	Core	Rim		Core	Rim
SiO_2	42.16	42.83	44.28	46.36	52.37	44.69	51.95	43.88	48.92	42.88	43.19	49.68
TiO_2	2.05	2.25	2.65	1.03	0.42	2.68	0.31	2.63	0.75	2.65	2.88	0.61
Al_2O_3	11.58	11.31	9.45	7.93	3.69	9.18	3.03	10.37	6.56	10.45	10.41	4.74
FeO^*	11.58	12.25	12.91	12.52	11.92	12.50	11.84	12.85	11.77	11.86	13.02	12.49
CaO	11.87	12.38	12.20	12.59	12.82	12.05	12.30	12.09	12.10	12.29	12.04	11.61
MnO	0.15	0.22	n.d.	n.d.	n.d.	0.24	n.d.	0.26	n.d.	0.23	0.18	n.d.
MgO	14.84	14.19	14.49	14.94	16.96	14.10	17.23	13.29	16.51	14.25	13.68	17.10
K_2O	1.24	1.31	1.01	0.78	0.32	1.05	0.19	1.36	0.56	1.26	1.27	0.57
Na_2O	1.97	1.66	1.34	1.31	0.12	1.42	0.24	1.57	0.93	1.43	1.17	0.78
Cl	0.26	0.30	n.d.	n.d.	n.d.	0.26	n.d.	0.34	n.d.	0.21	0.30	n.d.
Total	97.71	98.71	98.33	97.47	98.63	98.18	97.02	99.07	98.09	97.53	98.15	97.57
						Cations per 23 oxygens						
Si	6.25	6.30	6.50	6.82	7.47	6.58	7.53	6.44	7.70	6.37	6.39	7.23
Al^{IV}	1.75	1.70	1.50	1.18	0.53	1.42	0.48	1.57	0.93	1.63	1.61	0.77
Al^{VI}	0.27	0.26	0.14	0.20	0.09	0.17	0.04	0.23	0.18	0.19	0.21	0.04
Ti	0.23	0.25	0.29	0.11	0.05	0.30	0.04	0.29	0.08	0.30	0.32	0.07
Fe	1.44	1.51	1.59	1.54	1.42	1.54	1.43	1.59	1.42	1.47	1.61	1.52
Mn	0.02	0.03	—	—	—	0.03	—	0.03	—	0.03	0.02	—
Mg	3.28	3.11	3.17	3.28	3.61	3.08	3.78	3.01	3.56	3.15	3.02	3.71
Ca	1.88	1.95	1.92	1.99	1.96	1.90	1.91	1.90	1.87	1.96	1.91	1.81
K	0.23	0.25	0.19	0.15	0.06	0.20	0.03	0.26	0.10	0.24	0.24	0.11
Na	0.57	0.47	0.38	0.37	0.03	0.41	0.07	0.43	0.26	0.41	0.34	0.22
Total	15.91	15.83	15.67	15.63	15.22	15.63	15.25	15.72	15.48	15.75	15.67	15.46

Note: Analyses in weight percent. Peg. = pegmatite; inter. = intermediate between core and rim; n.d. = not determined; FeO^* = total Fe as FeO.

Presence of a silicate melt

Because some gabbro pegmatites appear to form at subsolidus temperatures (Bow et al., 1982), it is important to ascertain whether the Smartville pegmatites formed in the presence of a melt. Pyroxene thermometry indicates that they formed above the basalt solidus, but this does not rule out the possibility that the pegmatites are high-temperature, subsolidus replacements of refractory cumulates. The zoning in the pegmatites suggests, but does not demonstrate, crystallization from a melt. The differences in chemical composition implied by the higher orthopyroxene : olivine ratio in the pegmatites compared to the host gabbros rules out isochemical replacement, but the differences could have developed by subsolidus reaction with a silica-rich aqueous fluid.

The strongest evidence for the presence of a mafic silicate melt during pegmatite formation are the gabbroic "aplites" found in the cores of some pegmatite pods. These "aplites," as suggested earlier, may be analogous to aplites in granitic pegmatites formed by catastrophic loss of fluid from the magma. If the aplite analogy is valid, then (1) the mineralogy of the gabbro aplites should resemble the mineralogy of the pegmatites more closely than the host rock, (2) the aplites should show evidence of having formed during or after the late stages of pegmatite formation, and (3) the aplites should show some evidence of the quench mechanism by which they formed.

The gabbro aplites, as noted earlier, consists of pyroxene and plagioclase and lack olivine. Thus, they resemble the olivine-poor pegmatites more closely than the olivine-rich host gabbros. Large clinopyroxene crystals included in some aplites (Fig. 4) suggest that the aplites formed slightly later than the closely related pegmatites. Some of the large clinopyroxene grains included in the aplites are broken, suggesting some high-energy mechanism for aplite formation that would be consistent with catastrophic loss of a fluid phase. The absence of high-temperature pargasitic amphibole in the gabbro aplites is consistent with crystallization under the fluid-poor conditions that would follow catastrophic loss of a fluid phase.

Fluid phase

In granitic systems, the presence of a fluid phase is commonly considered necessary for pegmatite formation. Conversely, the presence of pegmatite is considered evidence for the presence of a free fluid phase (Jahns and Burnham, 1969; Jahns, 1982). However, direct evidence for the presence of a fluid phase at magmatic conditions is commonly elusive. Some low-pressure pegmatites contain cavities, and gem-bearing pegmatites in southern California show evidence of late crystallization directly from an aqueous-fluid phase (Jahns and Burnham, 1969). Other indications of an aqueous-fluid phase are the simultaneous crystallization and corrosion of a mineral phase at different places within a pegmatite and fluid inclusions in pegmatite minerals. We have found no direct evidence for the presence of a fluid phase associated with the silicate melt, but some aspects of the mineralogy and the crystallization of the pegmatites suggest that the activity of H_2O was elevated.

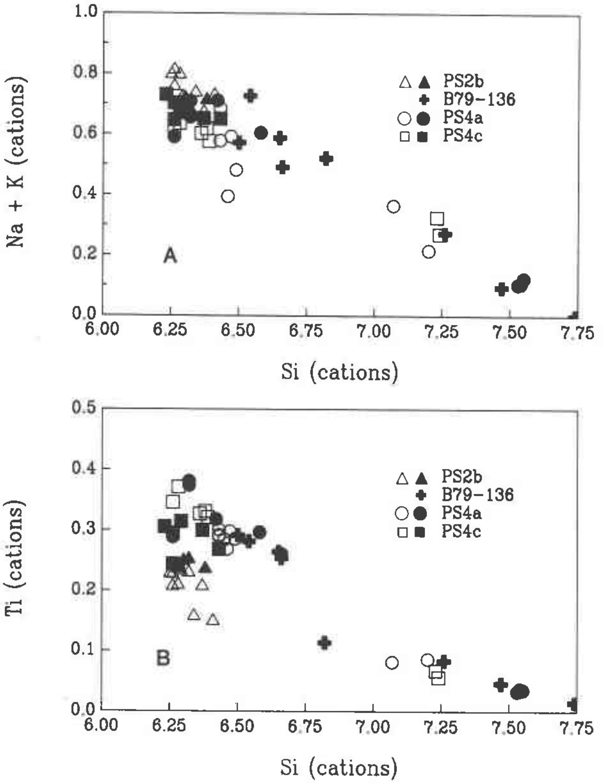

Fig. 6. Amphibole chemistry calculated on a 23-oxygen basis. (a) A-site alkali cations versus Si. (b) Ti cations versus Si. Open symbols are the host gabbro; closed symbols are pegmatite.

The presence of magmatic amphibole in the gabbro pegmatites indicates that the silicate melt was not anhydrous. One might argue that the restriction of amphibole crystallization to conditions near the solidus indicates undersaturation of H_2O. On the other hand, the pressures at which the pegmatites formed, as indicated by the regional metamorphic grade and the close temporal and spatial association with hypabyssal rocks, may have been quite low, perhaps as low as 1 kbar. The crystallization of magmatic amphibole at low lithostatic pressure requires high water activity in the melt. Decreasing the activity of water raises the minimum pressure at which amphibole can be stable in a basaltic melt (Gilbert et al., 1982). We conclude that the presence of magmatic amphibole in the gabbroic pegmatites and their host rocks implies high water activities and is consistent with water saturation during the last stages of crystallization.

Arculus and Wills (1980) noted that the compositions of plagioclase and olivine in cumulate gabbro xenoliths in andesites from the Lesser Antilles are substantially more calcic and magnesian, respectively, than plagioclase and olivine in layered mafic complexes. They attributed this difference to modification of the plagioclase liquidus and solidus induced by high water pressures (Johannes, 1978). Plagioclase and olivine compositions in the Smartville pegmatites and the host gabbros are similar to those reported by Arculus and Wills (1980) (Beard and Day, 1984; Tables 1 and 2).

The presence of subsolidus hydrous alteration assemblages in the gabbro pegmatites is strong evidence for the presence of a fluid phase. As noted earlier, this alteration is commonly restricted to the cores of the pegmatites, but, in some samples, involves the entire pegmatite. Evidence of that alteration is commonly lacking in the host gabbro. Fluid inclusions in plagioclase appear to be associated preferentially with the alteration in the pegmatites. Hence, either a fluid phase was present locally during pegmatite crystallization and cooling, or fluids migrated preferentially through the podiform pegmatites as they cooled.

Finally, if aplites in granitic pegmatites form by a mechanism involving fluid loss and if the analogy with the gabbroic aplites proposed earlier is complete, the presence of a fluid phase in the gabbro pegmatites is implied.

IMPLICATIONS FOR PEGMATITE FORMATION

Granitic pegmatites

Jahns and Burnham (1969) and Jahns (1982) proposed a model for the formation of granitic pegmatites that has won widespread acceptance (Černý, 1975; Uebel, 1977). The central premise of the model is that granitic pegmatites form from a silicate melt in the presence of an aqueous-fluid phase. Evolution of the fluid may result from crystallization in a magma that was originally undersaturated, although some pegmatites, such as those formed during partial melting of metamorphic rocks, may be saturated with respect to H_2O from the outset. Most of the pegmatite presumably forms from the constituents of the silicate melt. Although the model provides for end-stage crystallization directly from the fluid phase, the principal role of the aqueous fluid is to enhance the transport of material and to increase the growth rates of the crystals. In many pegmatites, there is evidence for gravitational separation of the fluid phase from the silicate melt. The pegmatite may form along the hanging wall of a dike where fluid is concentrated while aplites are crystallizing in a fluid-poor region along the footwall. Other aplites form at low temperatures in the cores of the pegmatites (Jahns and Tuttle, 1963; Taylor et al., 1979) when the fluid phase is lost catastrophically during the last stages of pegmatite formation.

Mafic pegmatites in other gabbroic intrusions

Cameron and Desborough (1964) suggested that pegmatitic dunite pipes in the Bushveld Complex formed as the result of subsolidus metasomatic replacement of the gabbroic host. Bow et al. (1982) interpreted pegmatitic pods associated with the Howland reef of the Stillwater Complex as subsolidus replacements, based on textural evidence. Gabbro pegmatite dikes in the Duke Island Complex (Irvine, 1974) and the Skaergaard Complex (McBirney and Noyes, 1979) were interpreted as filter-pressed, water-saturated intercumulus liquids. Viljoen and Scoon (1985) suggested that extremely Fe-rich pegmatites

Table 6. Pyroxene temperatures (°C) calculated at 1 atm

		Clinopyroxene				Orthopyroxene				2 Pyroxene
		En	Fs	Wo	*T*	En	Fs	Wo	*T*	*T*
PS-1	Pegmatite	53.4	4.8	41.8	1000	81.6	17.9	1.0	700	n.a.
	Host	50.7	10.4	38.9	1030	77.6	20.1	2.3	950	990
PS-2b	Pegmatite	48.2	11.9	39.9	970	76.4	21.7	1.9	830	n.a.
	Host	49.1	11.1	39.8	990	76.1	21.9	2.0	830	n.a.
PS-4a	Pegmatite	50.0	10.8	39.2	1015	75.3	21.5	3.2	1030	1023
	Host	49.8	9.7	40.6	975	77.5	19.9	2.6	980	978
PS-4c	Pegmatite	49.3	10.1	40.6	970	75.6	21.5	2.9	990	980
	Host	51.4	9.2	39.4	1020	76.2	17.7	1.7	800	n.a.
PS-5	Pegmatite	50.4	9.2	40.4	995	75.2	22.0	2.8	960	978
	Host	52.3	6.8	40.8	1005	79.7	17.7	2.6	1005	1005
B79-136	Pegmatite	49.7	10.6	39.7	1000	76.1	21.5	2.4	930	960
	Host	50.3	9.2	40.5	990	78.4	19.6	1.9	870	n.a.
	Average				997				906	988
	Average pegmatite (except PS-1)				990				948	
	Average host				1002				906	

Note: Compositions are recalculated and temperatures calculated after Lindsley (1983). n.a. = not applicable because orthopyroxene is a subsolidus phase.

in the Bushveld Complex represent highly fractionated, fluid-saturated intercumulus melts concentrated in areas disturbed by the late-magmatic disruption of the Bushveld cumulate pile. In all of these instances, the presence of pegmatite is used to infer the presence of an aqueous fluid phase. As noted by Viljoen and Scoon (1985), the major difference between granitic and gabbroic pegmatites is that granitic pegmatites apparently form only from silicate melts while at least some gabbroic pegmatites may form by subsolidus processes.

Formation of gabbro pegmatites in the Smartville complex

Any model for the formation of the gabbro pegmatites in the Pleasant Valley pluton must account for the following observations: (1) The mineral chemistry of pegmatite and host rock pairs suggests that the pegmatites were in equilibrium with their hosts and formed in place. (2) The presence of gabbro aplites, the high apparent temperatures recorded by pyroxene thermometers, and the zoning of the pegmatites suggest that the pegmatites crystallized from a melt. (3) The scarcity of olivine and the abundance of orthopyroxene in the pegmatites relative to the gabbros suggests that the pegmatites contain more SiO_2 than the host gabbros. (4) Several lines of evidence suggest that the pegmatites formed in the presence of elevated fluid pressures.

Replacement models for the origin of the pegmatites would fail to account for the apparent equilibrium between pegmatite and host and would be inconsistent with evidence suggesting that the pegmatites crystallized from a melt. Filter pressing may account for the segregation of the melt from which the pegmatite dikes such as PS-1 formed, but the other pegmatites formed from a melt in local equilibrium with their gabbro hosts and must have formed essentially in place. Although these pegmatites may be slightly enriched in SiO_2 relative to their hosts, they do not show the extreme fractionation of the Fe-rich pegmatites in the Bushveld Complex (Viljoen and Scoon, 1985).

We propose that the gabbro pegmatites in the Pleasant Valley pluton formed in place from a fluid-enriched intercumulus melt by a mechanism analogous to that proposed by Jahns and Burnhan (1969) for granitic pegmatites. The Jahns and Burnham model accounts for several important characteristics of the Smartville pegmatites including (1) the presence of a silicate melt, (2) the presence of aplitic gabbros in the cores of some pods, and (3) the inferred presence of an aqueous-fluid phase. Their model, however, does not address some of the problems peculiar to the formation of the Smartville gabbro pegmatites. First, the Smartville pegmatites are scarcely fractionated relative to their host rocks. It is not possible, therefore, to appeal to fractional crystallization as a mechanism for concentrating the volatile species. Nonfractional (e.g., eutectic) crystallization, however, might have led to the evolution of a fluid phase. Second, judging from their mineral chemistry, the composition of the Smartville pegmatites more closely resembles their cumulate host rocks than that of any of the more-fractionated rocks in the Smartville zoned plutons. Finally, the Jahns-Burnham model does not suggest a mechanism for the segregation of a fluid-saturated melt from its host.

The loss of confining pressure accompanying the reemplacement of the olivine gabbro cumulates may have played an important role in the evolution of a fluid phase. Fluid saturation may have occurred in the following manner: (1) Prior to their remobilization, the gabbro cumulates began to cool and to equilibrate with an intercumulus melt in a magma chamber at unknown depth. (2) Continued crystallization would cause an increase in the volatile content of the remaining liquid. (3) As the cumulates were remobilized and intruded to a higher level of the crust, the drop in pressure would decrease the solubility of the volatile species in the remaining melt and a fluid phase may evolve.

The reasons why the Smartville pegmatites—as well as certain other pegmatites formed from intercumulus melts, e.g., Skaergaard (McBirney and Noyes, 1979) and Duke Island (Irvine, 1974)—are so mafic are unclear. One might speculate that that the mafic compositions are a reflection of complex interactions involving resorption and back reaction between the intercumulus melt and the cumulus crystals during cooling.

The mechanism by which intercumulus melt segregates prior to pegmatite crystallization is not known. Perhaps, as suggested by Viljoen and Scoon (1985), disruption of a cumulate pile facilitates the migration and accumulation of any remaining intercumulus melt. Profound disruption of the Smartville olivine gabbro cumulates certainly must have occurred during their remobilization and emplacement to higher levels of the crust. Even in the relatively well-studied granitic pegmatite systems, little is known about the mechanisms of melt separation. We have observed that pegmatitic granites are common in schlieren, among other places, suggesting that disruption or flow in a magma can facilitate the segregation of a fluid-saturated melt.

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