Granitic Pegmatites in the Southeastern Adirondacks: Their Use as Indicators of Temperature, Pressure, and Fluid Conditions during a Late Stage of the Grenville Orogeny

A Thesis presented to the faculty
of the State University of New York
at Albany
in partial fulfillment of the requirements
for the degree of
Master of Science
College of Science and Mathematics
Department of Geological Science

John P. Mihalich
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ABSTRACT.

Granitic pegmatites associated with the Grenville orogeny intrude a mid-Proterozoic, upper amphibolite to near granulite facies terrane in the southeastern Adirondack Mountains of New York State. The inferred anatectic, fluid-saturated, and minimum melt characteristics of the pegmatites suggest that they can be used as analogs of the "granite" or "haplogranite" system, and as indicators of crustal temperature, pressure, and fluid conditions during a late stage of the Grenville orogeny. Bulk chemical compositions of seven simple, undeformed pegmatites in the southeastern Adirondacks indicate that they are generally depleted in silica and sodium, and enriched in potassium, relative to experimental data on phase equilibria of the granite-pure H_2O system, when $P_{H_2O}=7$ kb. Bulk chemical compositions of coexisting feldspars from four of the seven pegmatites indicate equilibrium conditions of crystallization. Zirconium and P_2O_5 contents of five of the pegmatites indicate minimum melt temperatures of 644-720o Centigrade, while zirconium concentrations of two of the pegmatites indicate temperatures of roughly 850o Centigrade. Appreciable amounts of fluorine and chlorine in biotites separated from three of the seven pegmatites suggest that additional volatile components present in the fluid phase during initial melting may have generated a silica- and sodium-depleted, and potassium-enriched minimum melt. Pegmatites in the southeastern Adirondacks which are enriched in sodium and silica, and depleted in potassium, intrude metasedimentary gneisses and amphibolitic gneisses associated with carbonates.

These pegmatites may have been produced under higher CO2 activity fluids, relative to the granite-pure H2O system. Sharp pegmatite-host gneiss contacts and disrupted foliations in the wall rock are indicative of an injection mechanism of emplacement. Similarities in chemistry and mineralogy between pegmatites and host gneisses suggest hydrothermal interaction between the intruding magma and the host gneiss. The mineralogy and chemistry of the host gneisses indicates an approximately 5 meter wide, sodium-enriched restite within the host gneisses on both sides of the pegmatites, and a less than one meter thick wide zone affected by an infiltrating, volatile-rich vapor phase, produced as the melt crystallized. The local nature of the pegmatites and association with high strain zones suggest that they represent the final "pulse" of the Grenville orogeny in the southeastern Adirondacks, which brought an upper level felsic slice to a lower, hotter region along ductile high strain zones, and initiated localized partial melting. Alternatively, the evidence for hydrothermal interaction between the pegmatites and their host gneiss is suggestive of low pressure intrusion, possibly related to uplift following the Grenville orogeny. The association of the some of the pegmatites with extensive migmatization and mylonites, and the lack of any chilled margins, however, indicate significant, albeit not necessarily peak, metamorphic pressure.

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CHAPTER 1. INTRODUCTION.

1.1. PURPOSE.

An important aspect of geology is the understanding of the physical conditions at which minerals and rocks form. The most important controlling factors, in addition to chemical composition, are pressure, temperature, fluid activity, and in some cases, stress due to grain deformation.

Mineral assemblages in many rocks have been studied and reproduced experimentally by varying the chemical composition and the physical conditions of pressure, temperature, and fluid activity in the laboratory. The grade of metamorphism has been defined according to the experimental data and their agreement with observations of the natural systems. According to Turner and Verhoogen (1960), temperature and load pressure generally increase simultaneously during regional metamorphism. Recently, however, petrologists have shown, based on the relationship between mineral zoning or overgrowth and tectonic implications, that the pressure-temperature path is not defined by a linear relationship (Spear et al, 1984). The rate of the change in pressure relative to the change in temperature is strongly dependent on the prevailing geothermal gradient, which in turn is subject to tectonic manipulation (Norris and Henley, 1976).

Chlorite, muscovite, and biotite are examples of hydrous minerals that will decompose to release the structurally bound water of hydration with a sufficient progressive increase in temperature. If this water is released into a system which is already at advanced metamorphic grade, it will act to lower the melting temperature of the rock and induce partial melting if the solidus of water-saturated melts is exceeded.

An interesting area of deep crustal rocks which has been brought to the earth's surface due to erosion of overlying material, and which represents relatively high metamorphic grade, is the Adirondack Mountains of New York State, a portion of the mid-Proterozoic Grenville terrane (Wiener et al, 1984). The rocks of the Adirondacks, which include originally shallow marine sediments, were subsequently deformed and buried, beginning approximately 1100 million years ago (Mclelland, 1986; Bohlen et al, 1985; Baer, 1981; Silver, 1969). This period of deformation, metamorphism, and igneous activity is termed the Grenville orogeny, and is believed to be the result of a collision between two separate continental masses, one aspect of the theory of Plate Tectonics (Dewey and Burke, 1973). Many geologists consider a modern analogue of the Grenville orogeny to be the Himalayan Mountain System, which has been caused by the collision between the northward moving Indian Plate and the Eurasian Plate (Baer, 1976; McLelland and Isachsen, 1980). Rocks presently being transformed and probably melted at roughly 25-50 kilometers below the surface of the Himalayas are probably similar in character and mineralogy to those of the Adirondacks, which are now exposed due to erosion and uplift of the former mountain system (Dewey and Burke, 1973). This report presents the results of an investigation as to whether data from experimental crystal-melt phase equilibria can be used to estimate the pressure and temperature conditions of crystallization of rocks believed to represent anatectic melts, provided that the geochemical characteristics of the "melt" and its host are understood.

The method used here combines the work of many experimental petrologists, which is summarized by Winkler (1979). The quartz-orthoclase-albite-anorthite ratios of water-saturated granitic pegmatites, determined from bulk chemical analysis, should plot along the cotectic line of the "granite" system, if these bodies are representative of former minimum temperature melts. The position of the cotectic line has been experimentally determined at various pressures, and is graphically displayed as a curve, which is projected onto the quartz-orthoclase-albite plane of the quartz-orthoclase-albite-anorthite tetrahedron. This tetrahedron represents the granite system. In order to evaluate the method, certain requirements must be satisfied. As an indicator of metamorphic conditions, the method requires that 1) the rocks represent anatectic melts--minimum temperature melt fractions derived from the metamorphic rocks in which they're emplaced--which then crystallized under equilibrium conditions, 2) the melt was saturated with water at the time of crystallization, and a fluid phase was present, 3) the fluid phase contained water as the volatile component, or the volatile components and proportions can be estimated, 4) the crystallized product is granitic in composition, i.e. it must contain the assemblage quartz + plagioclase + alkali feldspar, and 5) the rocks have not been subjected to significant deformation or metamorphism since the time of crystallization.

Migmatites are common in the southeastern Adirondacks, and small, but locally numerous, undeformed granitic pegmatite bodies intrude medium to high grade gneisses. It thus seems that temperatures, pressures, and locally fluid content were high enough during the closing stages of the Grenville orogeny to cause limited partial melting.

The composition of the pegmatites in the southeastern Adirondacks, their limited extent, and their distinction as the only igneous rocks formed at this time in the Grenville terrane attest to minimum melting conditions and limited fluids. If temperatures and fluid availability were higher, extensive anatectic igneous activity would be evident in the southeastern Adirondacks, and the composition of the melts formed would be more varied.

According to Winkler (1979), the ideas of Luth and Tuttle (1969), Mehnert (1968), Buddington (1963), Read (1957), and Sederholm (1908), testify to the close association of metamorphic rocks, migmatites, and granites in high grade metamorphic terranes. He states that "the spatial association of these rock types is due to processes occurring at similar pressure and temperature conditions in the deeper part of the crust", and that "experiments attempting to elucidate the origin of migmatites, and granites show that the process of anatexis is of major petrological importance." If, based on field observations in the southeastern Adirondacks, the propositions of Winkler and his predecessors are accepted, then the requirements of the method are partially satisfied; that is, the rocks used for this investigation represent anatectic melts which crystallized in the presence of a fluid phase. Previous estimates of temperatures and pressures prevailing in the southeastern Adirondacks during Grenville metamorphism, that is approximately 650 to 750o Centigrade and between 6.5 and 7.5 kilobars (Bohlen et al, 1985; Whitney and McLelland, 1973; Putman and Sullivan, 1979; Bohlen, 1982), indicate that anatexis was possible.

The apparent lack of zoning within the pegmatite bodies sampled, their coarse equigranular texture, and the homogeneity of mineral phases in thin section, are indicators of an equilibrium crystallization path, another requirement of the method. Finally, the pegmatites are undeformed, unmetamorphosed, and apparently unaltered, so their original granitic bulk composition is assumed to be preserved.

1.2. GEOLOGIC SETTING.

The region defined by the Adirondack Mountains, New York, is an upper amphibolite to near granulite facies terrane in the approximately 1.1 billion year old Grenville Province (Wiener et al, 1984), exposed on the North American craton (see Figure 1.2.1). The Adirondacks are characterized by a lower crustal anorthosite-charnockite massif core, which is roughly surrounded by structurally higher level quartzofeldspathic gneisses and metasediments, which were metamorphosed during the Grenville orogeny.

There is a disagreement as to whether the anorthosite-charnockite series is intrusive (Buddington, 1939; Isachsen et al, 1975), or represents basement rock, which was remobilized

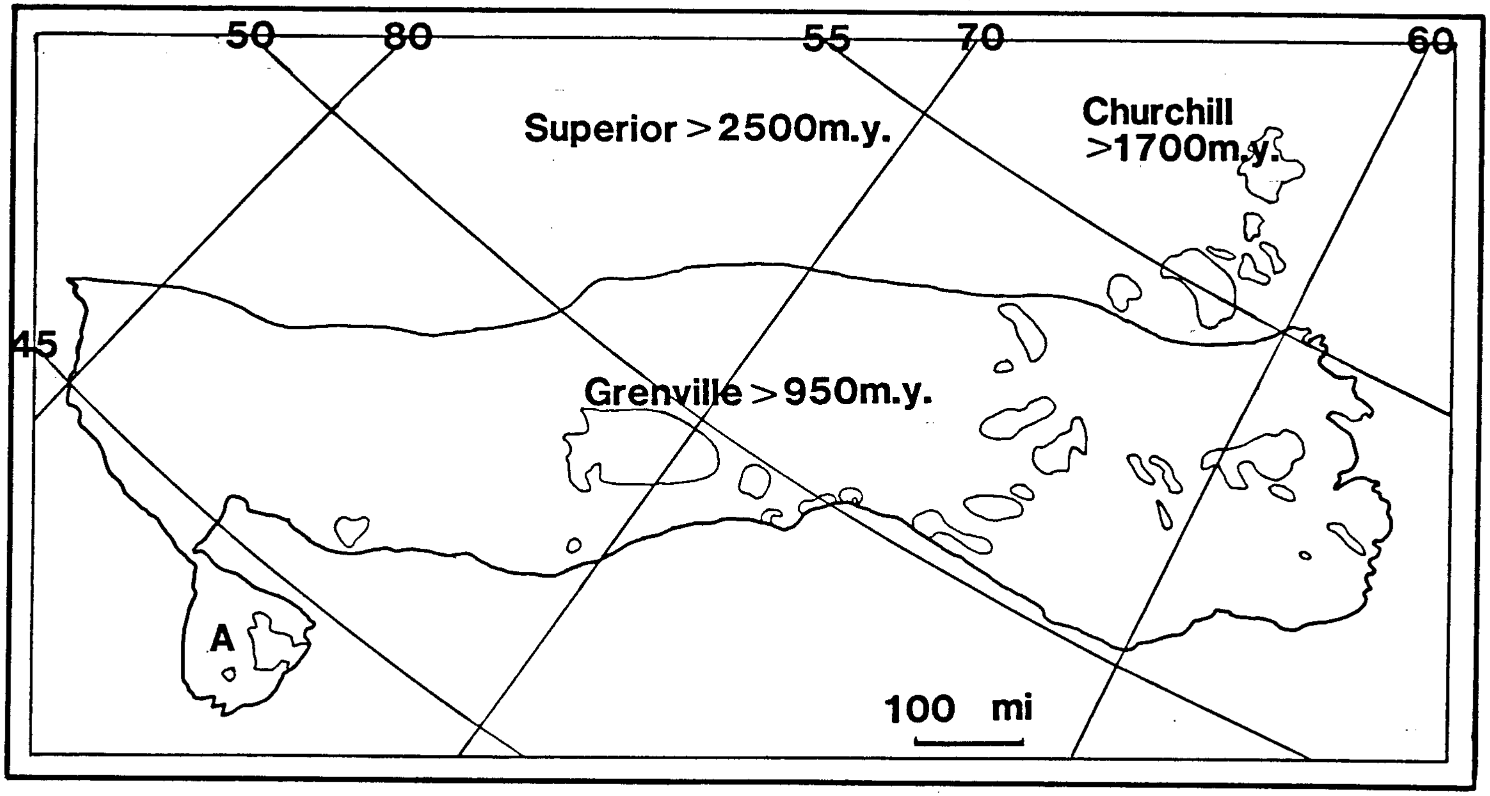


Figure 1.2.1. The Grenville Province in North America. Major anorthosite bodies are outlined. A:Adirondack Mts., New York (after Wynne-Edwards, 1964)

during the Grenville orogeny (Walton and de Waard, 1963). Younger granites and olivine-gabbros are believed to have intruded the anorthosite-charnockite series and the metasedimentary rocks.

1.3. PREVIOUS WORK.

The history of geology in the southeastern Adirondacks can be summarized in terms of three main phases of geologic mapping. In the early part of the century, most of the work was carried out by Kemp and Newland (1899), Cushing and Ruedemann (1914), and Alling (1927). In the early 1960's, Berry (1961) and Hills (1965) of Yale University, extended the mapping of their mentor, Walton (1961). More recently, the geology of the area has been compiled, reviewed, and reinterpreted by Wiener et al (1983), and Fisher (1985), based on mapping by Mclelland, Isachsen, and Whitney. This recent work attempts to correlate lithologic units of a metamorphic stratigraphy throughout the Adirondacks. However, highly strained rocks in the Adirondacks suggest that a simple litholgic stratigraphy is not preserved, and that, at best, lithologic/structural packages are separated by structural, not lithologic, boundaries (Baer, 1977).

Five of the seven pegmatites sampled for this investigation intrude gneisses between Putnam Station, N.Y. and Clemons, N.Y. in Washington County (see Figure 1.3.1). Kemp and Newland (1899) studied the geology of this county in the area of the eastern contact of Precambrian Adirondack rocks with overlying Paleozoic sediments. They describe three separate metamorphic series in this area. The first, termed the "Whitehall type" gneiss, is

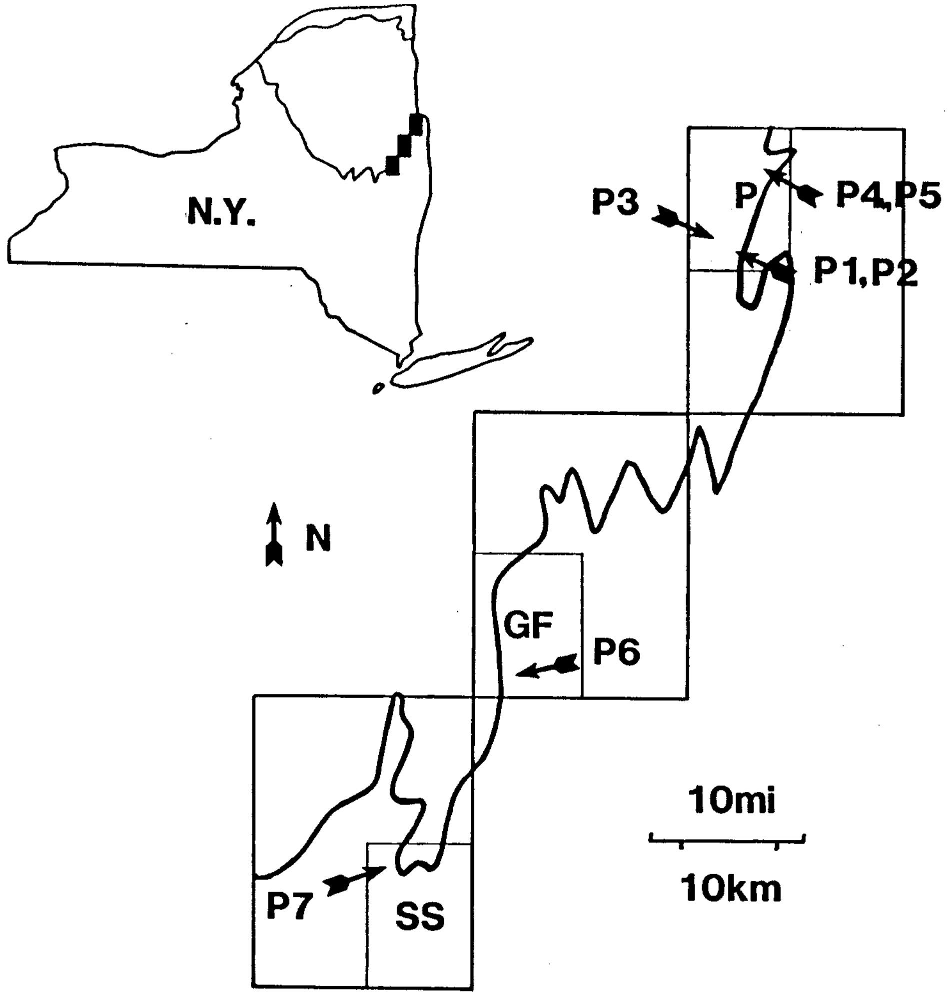


Figure 1.3.1. Location of pegmatites sampled, P1–P7. Putnam, Glens Falls, Saratoga Springs 7.5 minute quads. outlined.

considered metaigneous, containing microperthite, augite, hypersthene, hornblende, and quartz in the Ticonderoga, N.Y. area, with the addition of biotite to the south. Intense compression and shearing accompanies the addition of biotite. Garnet and "squeezed" pegmatites are common south of Putnam, N.Y. The second metamorphic series described consists of graphitic limestones and black schists which occur in the southern portion of Putnam township. Finally, the third metamorphic series consists of anorthositic and gabbroic gneisses recognized locally along the eastern shore of Lake George.

Berry (1961) studied the PreCambrian geology of the Putnam-Whitehall area, based on the idea that the "Whitehall type" gneiss is part of a comagmatic anorthosite-charnockite suite. He divides the "Whitehall type" gneiss of Kemp and Newland (1899) into a basement charnockitic gneiss and a sedimentary sequence, which was deposited on the basement and metamorphosed during the Grenville orogeny. Berry classifies the metasediments into the Lower and Upper Marble, separated by the Biotite Paragneiss, and overlain by the Alpha group, a sequence of mainly leucogranitic gneisses and garnet-sillimanite gneisses, and minor quartzites, marbles, and graphitic schists. The Beta group occurs in the southern portion of the area and is a thinner, correlative unit of the Alpha group. He recognizes an intrusive alaskite body in the central northern portion of the area, foliated along its contact with the overlying metasediments. The "Whitehall type" gneiss overlies the metasediments in most of the area, by virtue of a thrust surface which, according to Berry, allowed a

"Whitehall type" gneiss sheet of basement to override the metasediments in a northward direction. Three of the pegmatites sampled for this investigation (P1, P2, and P3) intrude the "Whitehall type" gneiss, and two of the pegmatites sampled (P4 and P5) intrude Berry's Alpha group.

Wiener et al (1983) and Fisher (1985) also consider the "Whitehall type" gneiss an exposure of the Grenville basement complex. In the southeastern Adirondacks, the "Whitehall type" gneiss is included in the Pharaoh Mountain Gneiss formation of the Piseco group, after Walton and de Waard (1963). The Piseco group is overlain by the Lake George group (Wiener et al, 1983).

Sample P6 intrudes a metapelite which outcrops southwest of Glens Falls, N.Y. along the Hudson River. According to Fisher (1985), this unit belongs to the Hague Gneiss member, of the Springhill Pond formation, of the Lake George group. The Hague Gneiss, as described by Alling (1918), is an aluminous gneiss which contains sillimanite, garnet, biotite, feldspar, and quartz, and is interlayered with quartzite, amphibolite, and granitic gneiss. It contains a thin marble unit in the Whitehall, N.Y. area. In terms of the structural stratigraphy outlined by Wiener et al (1984), the garnet-sillimanite gneiss unit, of the upper portion of Berry's Alpha group in the Putnam-Whitehall area, is correlative with the Hague Gneiss.

The final pegmatite sampled intrudes a metapelite north-northwest of Saratoga Springs, N.Y. Cushing and Ruedemann (1914) mapped the geology of Saratoga Springs and vicinity, where they were impressed by the predominance of garnetiferous schists over limestones and quartzites of the Grenville Series. They describe a ubiquitous association of schist and white metaigneous gniesses, which cut the schists intrusively and develop pegmatites, "forming injection gneisses and looking extremely like interbanded sediments".

Wiener et al (1983) place this unit within the Green Lake formation of the Lake George group. The Green Lake formation consists mainly of sillimanite-garnet-quartz-feldspar leucogneisses, and is considered by Wiener et al (1983) an upper unit of the Lake George group.

Although a lithologic correlation is not the main concern of this study, reconnaissance field work during sampling was necessary in order to understand the mode of origin and emplacement of the pegmatites, as well as their bulk chemistries, and the effect (if any) of emplacement on the host gneisses. A proposed correlation for the three general sub-areas from which the pegmatites were sampled is summarized in Figure 1.3.2, with references to the previous work in these areas.

This summary indicates that the Biotite Paragneiss of Berry (1961) in the Putnam-Whitehall area, can be correlated with the Older Paragneiss (Engel and Engel, 1953; Walton and de Waard, 1963), and the Lower Marble can be correlated with the Paradox Lake formation (Walton and de Waard, 1963), two ideas suggested by Berry (1961). The Alpha-Beta Group of Berry, according to his description, is similar to the Lake George Group (see Fisher, 1985; Weiner et al, 1984). The Green Lake formation may correlate with the Springhill Pond formation.

Figure 1.3.2. Correlation between rock units intruded by pegmatites P1 through P7.

Alpha Group	Lake George Group	Lake George Group
Leucogranitic gneiss, garnet sillimanite gneiss, with marble, graphitic schist, and quartzite. **P4,P5**	Springhill Pond Formation Bear Pond Schist Catamount Sillimanite Schist Bryant Lake Marble Swede Pond Quartzite Hague Gneiss Sillimanite, biotite, garnet, feldspar, graphite gneiss, interlayered with quartzite, amphibolite, marble, and granitic gneiss. **P6**	Green Lake Formation Sillimanite, garnet, quartz, feldspar gneiss, interlayered with quartzite, amphibolite, and pyroxene granulite. **P7**
Upper Marble		
Biotite Paragneiss	Treadway Mountain Formation Biotite, quartz, plagioclase gneiss.	
Lower Marble Quartz, feldspar, diopside, garnet, scapolite, graphite, sphene, interbedded with amphibolite, and graphite-bearing schist.	Paradox Lake Formation Dresden Granulite Amphibolite Johnsburg Marble Graphite, diopside, scapolite, feldspar, serpentine, talc. Sacandaga Quartzite	
"Whitehall-type" Gneiss **P1,P2,P3**	Pharoah Mountain Gneiss	
Berry, 1961.	see Fisher, 1985.	see Wiener et al, 1984.

CHAPTER 2. THE "GRANITE" SYSTEM.

The "granite" system is discussed by Winkler (1979). In summary, the idealized granite system consists of the components SiO2, NaAlSi3O8, CaAl2Si2O8, KAlSi3O8, and H2O, and forms the minerals quartz, plagioclase, and alkali feldspar. Tuttle and Bowen (1958) studied the equilibria among crystals, melt, and gas in the system SiO2-NaAlSi3O8-KAlSi3O8-H2O at various pressures. Shaw (1963), Morse (1970), Seck (1971), and Luth et al (1964) further defined details of this system, which is termed the "approximate" granite system by Winkler (1979), or the "haplogranite" system by Luth et al (1964).

Petrologists realized the importance of the anorthite component, CaAlSi3O8, in the granite system, and the experimental work of Stewart (1957,1967), Yoder (1968), Winkler and Lindemann (1972), and Winkler and Ghose (1974) defined phase relationships in a five component system which represents granitic or granitoid rocks. The contributions of Winkler and von Platen (1961), von Platen and Holler (1966), von Platen (1965), James and Hamilton (1969), and Piwinski and Wyllie (1970) to equilibrium phase relations and melting in natural rocks should also be noted.

The granite system can be represented on a tetrahedron; for example, Figure 2.1 is Winkler's diagram of the granite system at 5 kb of PH2O. The components SiO2, NaAlSi3O8, CaAl2Si2O8, and KAlSi3O8, abbreviated Qz, Ab, An, and Or, are plotted as pure end member components on the four corners of the tetrahedron. The fifth component, H2O, is assumed to be present in excess. Each face of the tetrahedron, therefore, depicts a four component

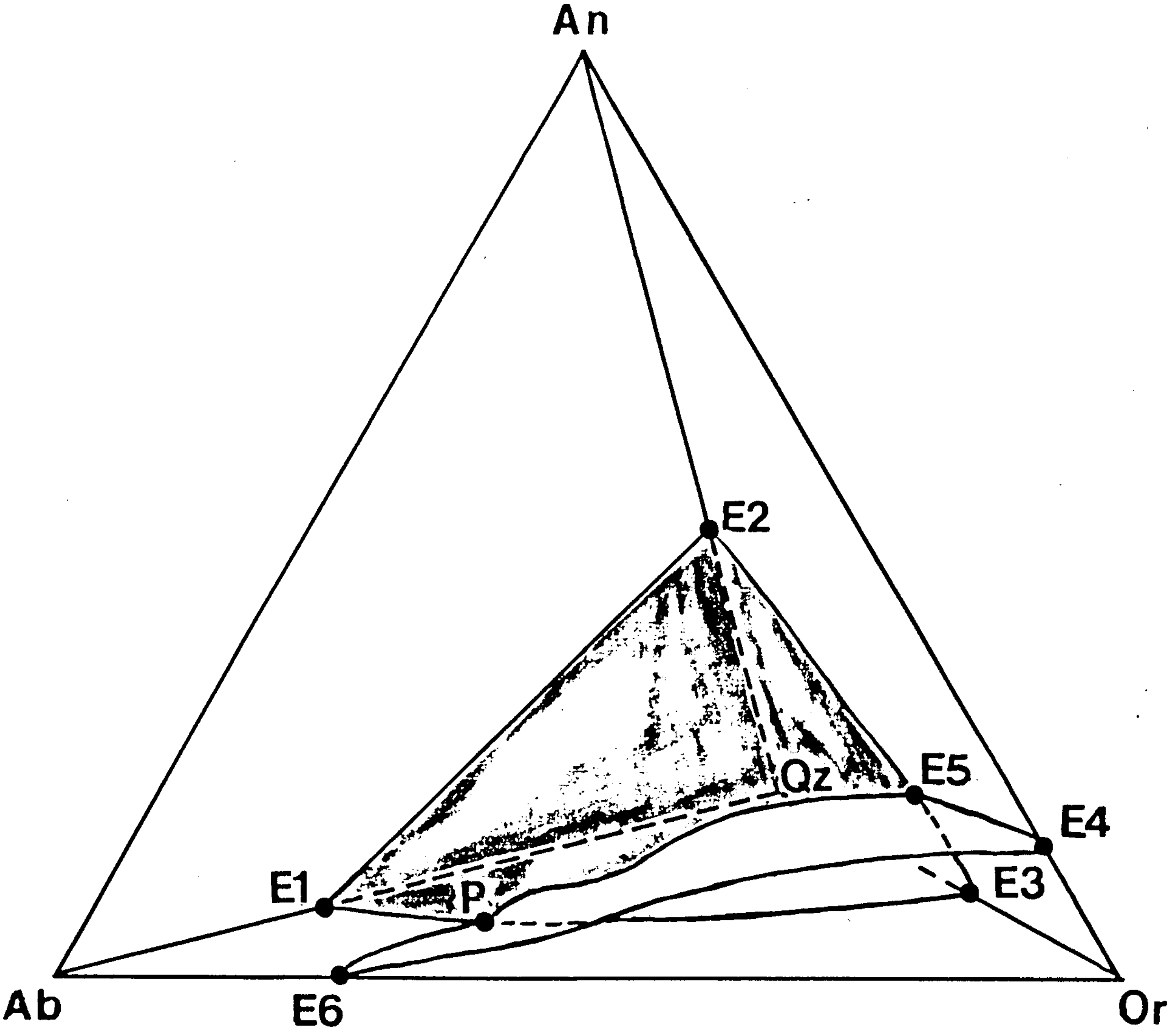


Figure 2.1. The granite system at 5 kb PH2O, after Winkler, 1979. E1 through E6 are eutectic points of the subsystems. P-E5 is the cotectic line. The shaded surfaces separate fields of stability for the phases quartz, plagioclase, and alkali feldspar.

subsystem projected from H2O.

A eutectic point represents the composition of the minimum temperature melt of a mixture of components, at a given PH2O. Points E1 through E6 of Figure 2.1 are eutectic points in the hydrous binary systems that form the edges of the tetrahedron; these points in turn are connected to one another by cotectic lines on the tetrahedron faces representing ternary hydrous systems. Cotectic lines define the composition of melt coexisting with two solid end members crystallizing in the presence of a vapor phase at fixed pressure. Point P is a eutectic above 3 kb. The intersections of the cotectic surfaces divide the space of the tetrahedron into a quartz space, an alkali feldspar space, and a plagioclase space.

For example, suppose a point within the alkali feldspar space represents the composition of a melt undergoing crystallization. For this composition, an orthoclase-rich alkali feldspar (K-feldspar for simplicity) crystallizes initially. The composition of the melt is directed away from the Or corner (approximately) with falling temperature as the Or component separates, until it hits a cotectic surface. If it hits the surface separating the quartz and K-feldspar spaces, K-feldspar plus Qz separate from the melt with falling temperature. The melt composition is then directed toward the cotectic line P-E5, where plagioclase appears and all three phases then crystallize from the melt as temperature falls. The melt then proceeds toward point P until crystallization is complete. This process acts in reverse if temperature is raised and granitic material with the same solid phases is melted.

During fractional crystallization, solid phases are effectively removed from the system as they crystallize, so that the composition of individual portions of the solid, when crystallization is complete, is different from the bulk composition of the original liquid, and the crystal aggregate is not compositionally homogeneous.

Winkler (1979) lists three possible reasons why there may be a departure of early stage melts from the relevant cotectic line of the experimental studies: 1) The early stage of anatexis has been surpassed with the loss of a solid phase, so that the melt does not coexist anymore with the three solid phases, i.e. plagioclase, alkali feldspar, and quartz; 2) The original gneiss did not contain the three components quartz, plagioclase, and alkali feldspar; and 3) The presence of other volatiles plus H2O may affect the minimum melt temperature and composition. For example, if 5% HCl is present in the fluid phase, the cotectic composition is shifted to about 5% less Qz, and with the addition of fluorine, the position of the boundary between quartz and alkali feldspar also moves away from the Qz apex (see Figure 2.4.1). A fourth reason can be added to Winkler's list. One or more of the solid phases of the granite system may be represented by another phase in the source rock, for example, the An component may be incorporated in hornblende, or the Or component in biotite or muscovite.

2.1. EFFECT OF ANORTHITE COMPONENT.

The position of the melt along the cotectic line P-E5 is dependendent on temperature. The position also reflects a particular An content of the melt and plagioclase crystals in equilibrium with it. An increase in the An/Ab ratio of the bulk composition increases the minimum melting temperature, and shifts the composition of the minimum melt along line P-E5 toward E5 (Winkler, 1979), and greater Or contents.

The temperature range for P-E5 at 7 kb is roughly 50o Centigrade. The plagioclase of gneisses is commonly An10 to An40, and therefore, an early stage anatectic melt will not reach a position close to E5. It is restricted to the lower temperature portion of P-E5, represented by a dark line on Figure 2.1.1. The maximum temperature at which such a melt can be represented along line P-E5 is roughly 660o Centigrade at 7 kb PH2O. The temperature range is roughly 25o Centigrade. This corresponds to a maximum anorthite component of the melt of no more than roughly 13% (see Figure 2.1.2).

2.2. EFFECT OF PRESSURE.

With an increase in PH2O, the An content of the eutectic in the Qz-An-Or-H2O subsystem increases and the temperature of the eutectic decreases (Winkler, 1979). The position of the cotectic line at various water pressures has been determined experimentally, and is shown in Figure 2.1.1, where the cotectic lines are projected onto the Qz-Ab-Or plane of the tetrahedron. The position of the cotectic line moves away from the quartz apex, pivoting on the relatively stationary point E5, whose Qz-Or ratio increases only slightly with an increase in PH2O. The composition of the minimum melting point (P in Figure 2.1.1)

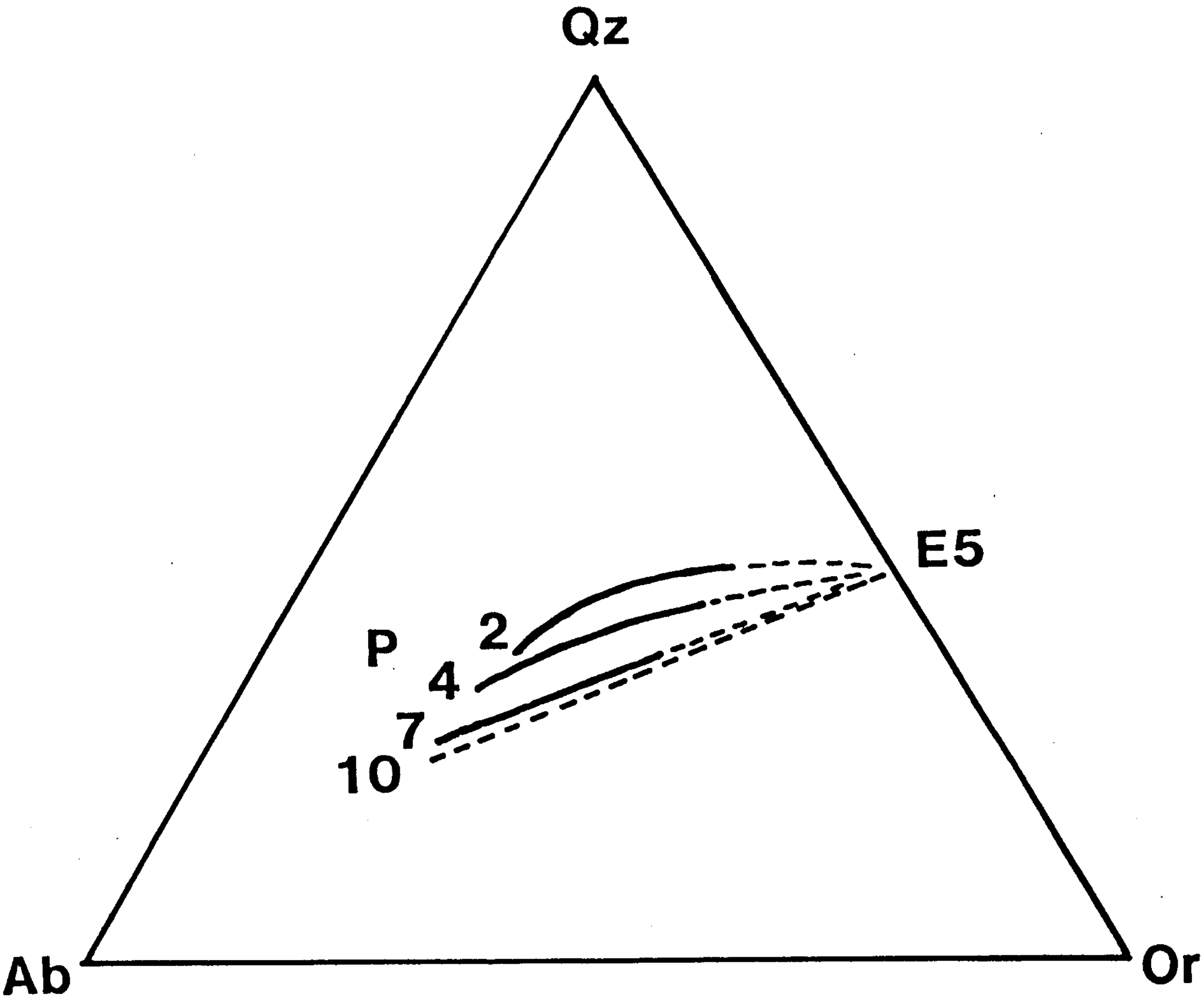


Figure 2.1.1. The cotectic line, P-E5, experimentally determined at various PH2O. Dark lines represent probable compositions of minimum melts, based on the anorthite content of the source of the melt, after Winkler, 1979.

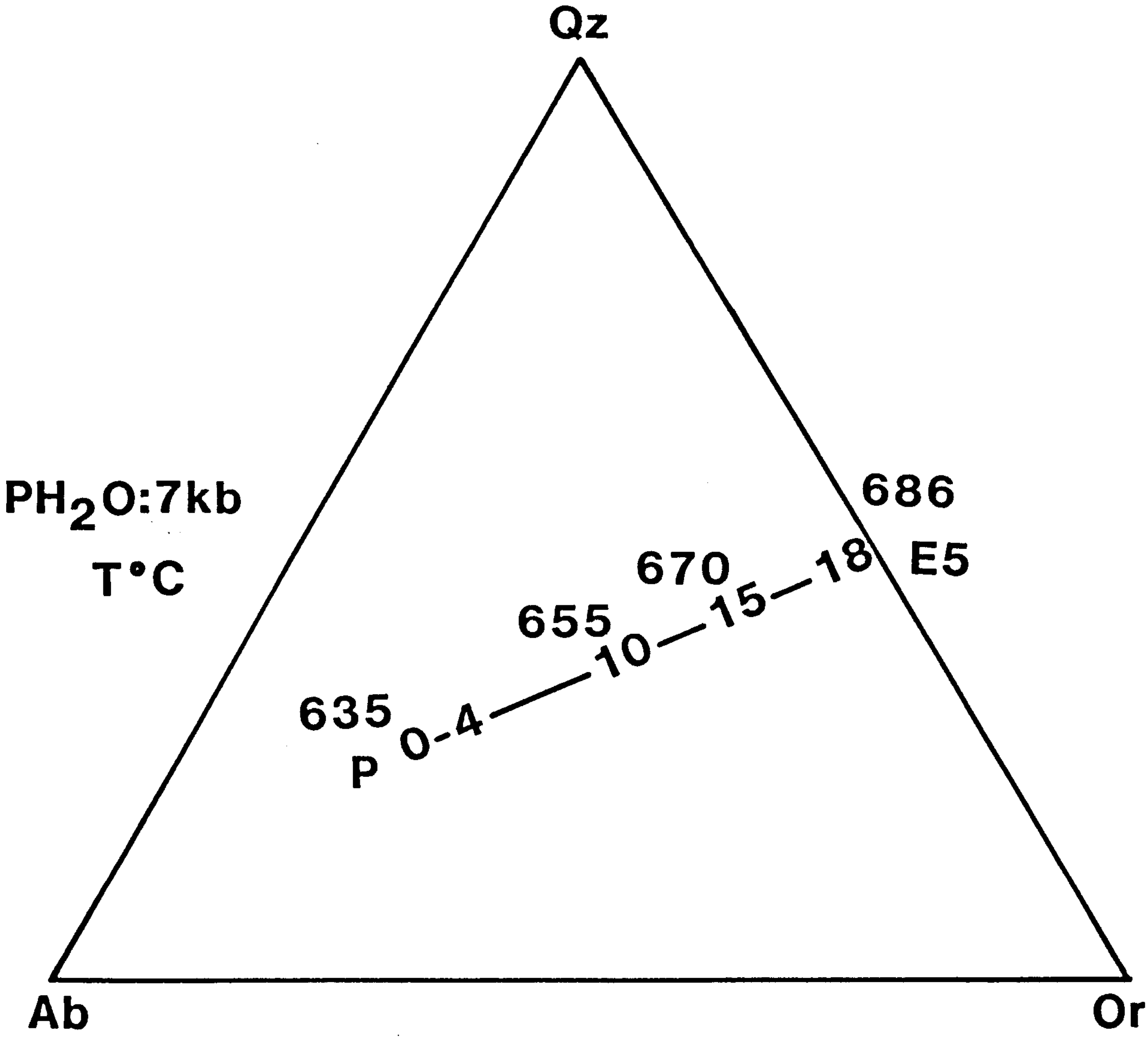


Figure 2.1.2. The temperature/composition range of minimum melts at 7 kb PH2O, after Winkler, 1979.

increases in Ab component, and the temperature of the minimum melting point (P) decreases, with an increase in PH2O.

According to Winkler (1979), "any melt formed at the beginning of anatexis of gneisses which can be represented in the system Qz-Ab-An-Or-H2O is situated on the cotectic line P-E5." This theory is the basis for the present investigation.

Johannes (1984) studied the beginning of melting in the granite system and the approximate granite system from 2 to 15 kb of PH2O. He found that the melting temperature in the An-free system decreased from 690o Centigrade to 630o Centigrade with an increase in PH2O. The solidus temperatures increase by roughly 3o Centigrade if the plagioclase starting material has a composition of An20, and by roughly 7o Centigrade if the plagioclase is An40. The total range of 50o Centigrade along P-E5 agrees with earlier studies.

Between pressures of 6 and 16 kb PH2O at 600o Centigrade, Johannes (1984) found that plagioclase is replaced by zoisite + kyanite + quartz, and that granite is replaced by zoisite + muscovite + quartz, a subsolidus hydrated assemblage. This, however, assumes excess water is present.

2.3. EFFECT OF WATER.

It is believed that an excess of water is not necessary to induce partial melting of gneisses in medium to high grade metamorphic terrains (Winkler, 1979; Brown and Fyfe, 1970; Johannes, 1984). According to Winkler (1979), only the total amount of melt formed by anatexis depends on the amount of water and other volatiles in the system. Migmatites, in most cases, represent areas where the maximum amount of melt was not formed, and in order for the maximum amount of melt to form, sufficient water must be present. At 3 kb, for example, the amount of H2O present in a saturated melt is about 8 weight percent (see Figure 2.3.1). If less than 8% water is present at 3 kb confining pressure, the total amount of granite system minerals cannot be melted, even if they account for all of the host rock, although a partial melt may form (Winkler, 1976). Johannes (1984) estimates that 1% water or less is sufficient to transform portions of a gneiss into leucosome. He states that anatectic magmas are usually not saturated with water, since temperature may exceed the H2O saturated solidus for the given pressure.

2.4. EFFECT OF ADDITIONAL VOLATILE COMPONENTS.

The addition of components other than H2O to the aqueous phase has a variable effect on phase equilibria of the granite system. Wyllie and Tuttle (1964) investigated the effect on the granite solidus of adding up to 5 weight percent of a second volatile component to the aqueous phase. At a pressure of 2.75 kb, the addition of HF, as well as Li2O or P2O5 lowered the temperature of the beginning of melting, while the addition of HCl or NH3 raised the temperature of the beginning of melting. The effect of SO3 as the second volatile component was negligible.

Swanson (1979) investigated the effect of the addition of CO2 to the aqueous phase on the granite system at pressures of 2 and 8 kb. With a vapor mixture of 96 weight percent H2O and 4

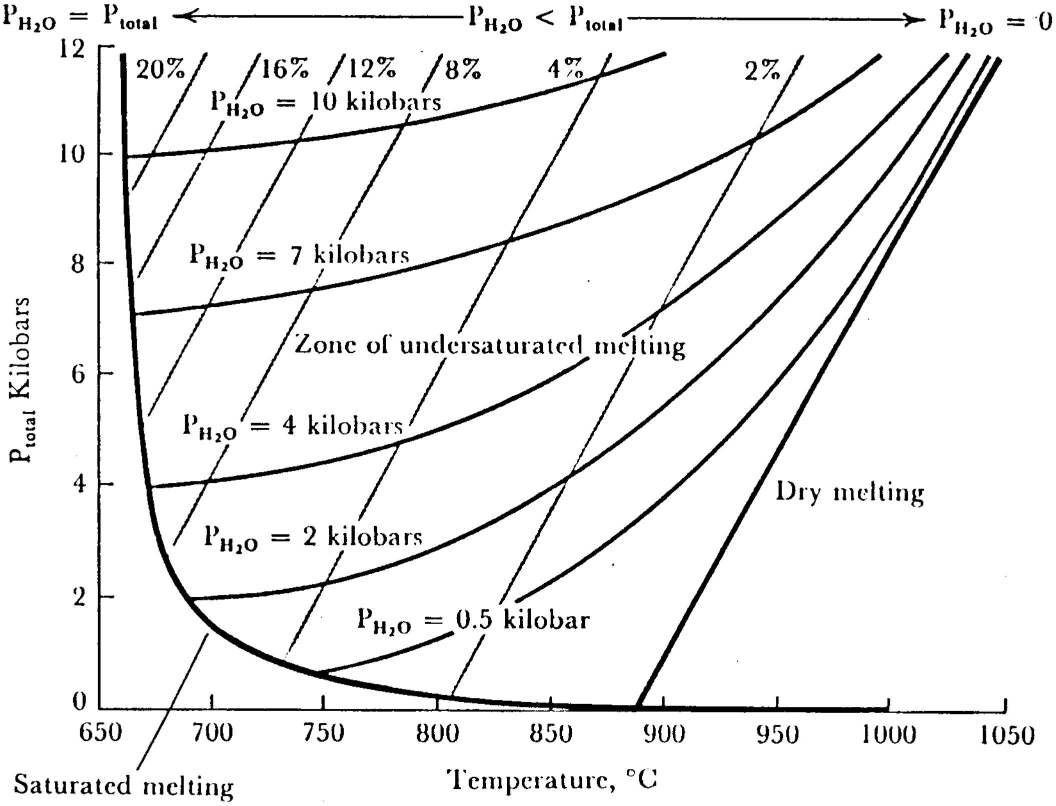


Figure 2.3.1. P-T conditions for water-saturated vs. water-unsaturated granitic melting, after Hyndman, 1972.

weight percent CO2, no appreciable effect was observed on the solidus or liquidus temperatures of the granite system at 2 kb. At 8 kb, however, solidus and liquidus phase assemblage fields in the granite system were shifted to higher temperatures by 20 and 50o Centigrade, respectively.

The addition of up to 10 weight percent boron to the aqueous phase of the granite system at 1 kb was studied by Chorlton and Martin (1978), who found that the solidus and the liquidus were lowered by roughly 125o Centigrade.

Although Wyllie and Tuttle (1964) found that the addition of HCl to the aqueous phase slightly raised the temperature of the beginning of melting in the granite system, Winkler found that the addition of roughly 5 weight percent HCl to the aqueous phase at 2 kb lowers the melting temperature of granite by as much as 10o Centigrade. The addition of HCl to the granite system also changes the composition of the minimum melt during anatexis towards a lower quartz content, and somewhat richer orthoclase content.

Finally, the effect of fluorine on phase relationships in the approximate granite system, Qz-Ab-Or, at 1 kb PH2O was investigated by Manning (1981). With increasing fluorine content, the position of the quartz-albite-orthoclase field boundary moves away from the quartz apex, and the geometry of the liquidus surface changes (see Figure 2.4.1). The minimum melting temperature and the minimum melt composition change progressively with increasing fluorine content, from 730o Centigrade and Qz37-Ab34-Or29 for the fluorine-free system of Tuttle and Bowen

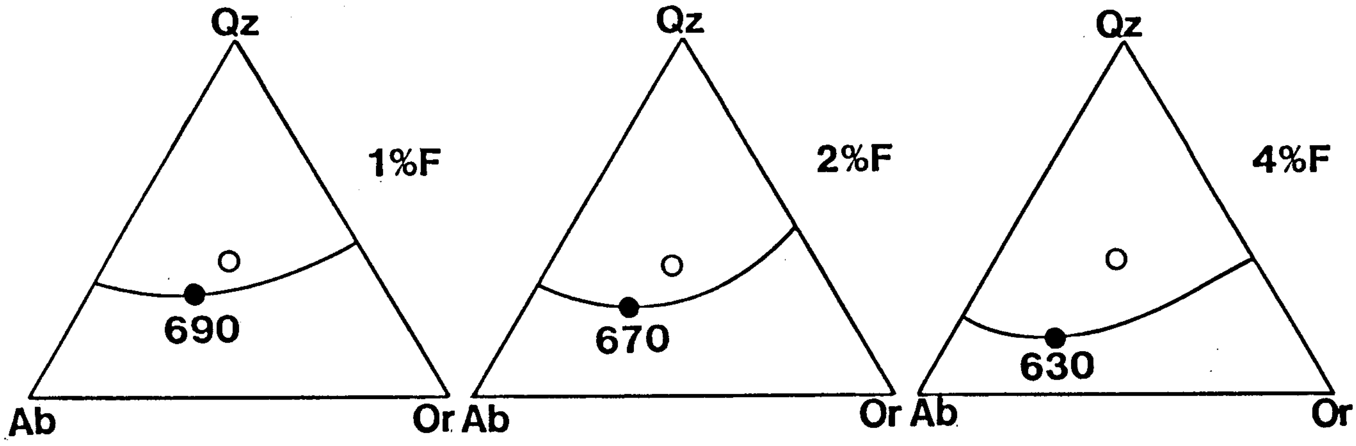


Figure 2.4.1. The granite system at 1 kb PH2O with the addition of fluorine. Open circles refer to the composition of the minimum melt of the fluorine-free system. Closed circles refer to the composition of the minimum melt in systems with fluorine, after Manning, 1981. Temperature in degrees Centigrade.

(1958), to 630o Centigrade and Qz15-Ab58-Or27 for the system with 4 weight percent fluorine added to the aqueous phase.

In the presence of an aqueous phase containing several of the additional volatile components, the effect of the constituents that tend to lower the melting temperature will be predominant (Wyllie and Tuttle, 1964). This is indicated by the fact that small concentrations of HF or Li2O lower the melting temperature appreciably, while higher concentrations of HCl, NH3, or CO2 are needed to raise the melting point appreciably.

According to Chorlton and Martin (1978), an increase in solidus and liquidus temperature due to the addition of secondary components in the volatile phase, such as CO2 or HCl, is caused by a reduction in the activity of water in the melt and the coexisting vapor phase. Apparently, the addition of 4% CO2 at 8 kb has a greater effect on the liquidus temperature, which it raises by 50o Centigrade, than on the solidus temperature, which it raises by 20o Centigrade (Swanson, 1979). The solubility of CO2 in the melt is lower than that of H2O, which results in an H2O-rich silicate liquid and a CO2-rich vapor phase at hypersolidus temperatures. HCl also has a low solubility in silicate melts and a strong affinity for associated aqueous phases (Fuge, 1977). This idea may explain the discrepancy in the behavior of HCl in the granite system as reported by Wyllie and Tuttle (1964) versus Winkler (1979, from the data of von Platen, 1965). HCl may lower the solidus temperature probably because of its affinity for sodium in albite (Wyllie and Tuttle, 1964), but increase the liquidus temperature of the granite system because of its lower solubility in silicate melts, relative to the associated vapor phase.

A decrease in the solidus or liquidus, however, by the addition of another volatile component to the aqueous phase is a result of the partitioning of that volatile into the melt phase, which modifies the structure of the melt (Chorlton and Martin, 1982).

In summary, chlorine and fluorine are likely extra volatiles which shift the minimum melt compositions toward less Qz content, but more Or and Ab content, respectively. Granitic rocks representing minimum melts which formed in the presence of a chlorine-rich aqueous phase should be somewhat depleted in Qz content, and somewhat enriched in Or content relative to the pure H2O-granite system. Those which formed in the presence of a fluorine-rich aqueous phase, however, should be somewhat depleted in Qz content, and somewhat enriched in Ab content, relative to the fluorine-free system.

CHAPTER 3. PEGMATITES.

A pegmatite is an assemblage of minerals that is remarkably coarser and texturally more irregular than is ordinarily observed in rocks with similar mineral assemblages (for example, granitic pegmatite vs. granite). Most pegmatites are granitic in composition, but gabbroic and anorthositic pegmatites are also reported (Silver, 1969; Beard and Day, 1986). A variety of rare minerals may be common and abundant in pegmatites.

The coarse-grained texture, typically granitic character, and the association of rare minerals are related to the idea that a separate vapor phase was present during the crystallization of a silicate melt. During the crystallization of granite, water and other volatiles are largely excluded from the solid phases and concentrated in the residual magma, along with other elements which cannot, because of ionic radius or charge, substitute for common elements in crystal lattices growing at the temperatures of magmatic crystallization (Turner and Verhoogen, 1960). This solution eventually crystallizes as granitic pegmatite, in which may be concentrated rare minerals. It is generally believed that a high water content lowers the viscosity and aids the rate of diffusion and crystal growth rate within a silicate liquid, while volatiles inhibit the nucleation rate of grains, hence the coarse-grained texture of pegmatites (Hyndman, 1972; Swanson, 1977). Growth of large crystals in the vapor space of a melt-H2O system has been demonstrated experimentally (Jahns and Burnham, 1969).

3.1. PHYSICAL CHEMISTRY.

The high water content of pegmatite magmas suggests that, under certain conditions, a separate vapor phase may be in equilibrium with solid and liquid phases of the magma. Thus, an introduction to the physical chemistry of systems with volatile components should aid in the understanding of the formation of pegmatites. The reader should consult Turner and Verhoogen (1960), Luth and Boettcher (1986), and Barker (1976) for a more detailed explanation of systems which contain volatile components. For the purposes of this investigation, it is sufficient to be familiar with the phases which may be present in the granite system during the formation of pegmatite and the external factors which control the existence of these phases.

Although the granite system consists of five components, the physical chemistry is better understood in terms of a two-component system, such as albite and water. The phase relationships are shown on the hypothetical phase diagram of Niggli (1943, from Turner and Verhoogen, 1960). (See figure 3.1.1.) For a bulk composition of X1, the system consists of a gas phase of composition X3 and a liquid or silicate melt phase of composition X2 at temperature T1. As the temperature is lowered, the liquid is enriched in water, until the condensation of vapor is complete at T2. The system now consists of a single liquid with composition X1. The liquid remains stable until temperature T3, when albite begins to crystallize. The solution cools along curve CA, and becomes richer in water. At temperature T4, a gas phase of composition X4 appears, and the

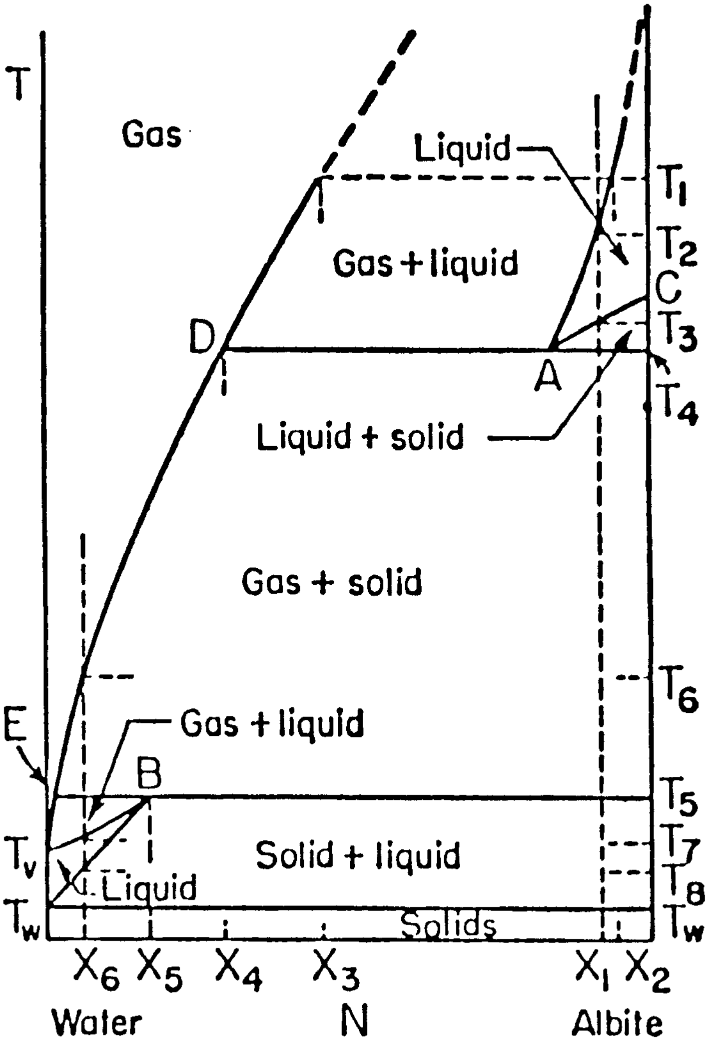


Figure 3.1.1. Phase equilibria of the binary system, albite-water, from Turner and Verhoogen, 1960.

system now consists of three phases, gas D, Liquid A, and solid albite. At point A, the system is univariant, with respect to pressure and temperature (i.e. if pressure is specified, then the system is invariant). The value of the vapor pressure can be determined from the temperature alone if the magma is univariant.

Notice that, as the temperature falls from T3 to T4, the magma is crystallizing and the dissolved volatiles are becoming more and more concentrated in the remaining liquid. Thus, vapor pressure in systems with volatile components rise with a decrease in temperature. When the vapor pressure exceeds the external pressure, the liquid becomes volatile-saturated, and will "boil". The phenomenon of boiling due to an increase in volatile concentration and vapor pressure during cooling is called the "second boiling" point (Turner and Verhoogen, 1960; Hyndman, 1972). This point is represented as point A on Figure 3.1.1, the invariant point (assuming P is fixed), discussed above. As heat is removed from the system during crystallization, more water goes into the vapor phase, and the amount of melt decreases, until the system consists of only two phases, a solid and a water-rich vapor.

The cotectic curve of the granite system, referred to in Chapter 2 of this thesis, is comparable to this invariant point. The curve defines the bulk compositions of liquid, crystallizing quartz, alkali feldspar, and plagioclase as solids, in equilibrium with a water-rich vapor phase.

3.2. MINERALOGY.

On the basis of mineralogy, pegmatites can be generally classified as simple or complex (Turner and Verhoogen, 1960). Simple pegmatites consist of quartz, alkali feldspar, and plagioclase, and minor micas. Rare minerals are absent or present as accessories. Complex pegmatites contain rare minerals in abundance and variety, such as lepidolite, spodumene, tourmaline, topaz, cassiterite, beryl, tantalite, columbite, uraninite, thorite, amblygonite, and allanite.

Simple pegmatites are more common in regionally developed migmatite complexes, while complex pegmatites are associated with granitic intrusions (Turner and Verhoogen, 1960). Pegmatites in the southeastern Adirondacks can be classified as simple pegmatites.

3.3. STRUCTURE.

Some pegmatites exhibit a zoned structure, which is defined by variations in mineralogy and texture. The zoned structure is thought to be the result of a crystallizing magma in a closed system (Turner and Verhoogen, 1960). As a portion of the melt crystallizes against the wall rock, the unmelted fraction is enriched in water and concentrated inward. Previously formed crystals may exchange with the vapor phase also. The core then represents the final stage of crystallization, in which an aqueous gas phase is confined. Zoning is best shown in podiform pegmatites and low angle dikes or sills. An example of this type is the Robbins Silica Prospect, in San Bernadino County, California (Jahns, 1953).

The results of this study suggest that, in an open system, the gas phase may be expelled along the margins of the intrusive body, which by itself appears to be unzoned.

The mineral zonation is contingent on 1) the phases present during crystallization, and 2) the composition of these phases. For example, in the presence of two immiscible fluid phases, silicate melt and vapor, chlorine is preferentially concentrated in the vapor phase, where it increases the solubility of sodium relative to potassium (probably in the form of NaCl), and thus causes an alkali fractionation, if exchange through the vapor phase occurs (Dujon and Lagache, 1986; Lagache and Weisbrod, 1977). This may result in plagioclase-rich margins and alkali feldspar and quartz-rich cores in pegmatites of the southeastern Adirondacks.

There does not seem to be a relationship between the size of the pegmatite and whether or not it is zoned. Zoning is apparently not determined by factors of regional geologic significance. For example, zoned and unzoned pegmatites are distributed throughout the Precambrian metasediments of the Mica Mine area, in the Van Horn Mountains, Texas (Flawn, 1950).

The forms of pegmatites include dikes, veins, lenses, stringers, and irregular masses of up to several miles in length and hundreds of feet in thickness (Turner and Verghoogen, 1960; Flawn, 1950). The form may be structurally-controlled.

3.4. PETROGENESIS.

Due to the fact that pegmatites vary in mineralogy, size, form, and rock association, several theories have been suggested to explain their origin and emplacement. According to Flawn (1950), there are four main theories to choose from in postulating a mode of origin and emplacement for pegmatite bodies.

3.4.1. THE GRANITIZATION THEORY.

The granitization theory contains two ideas on pegmatite formation, the first of which is the concept of pegmatite genesis by metasomatism. Metasomatism is an homogenizing process according to Flawn (1950), whereby country rock is transformed into pegmatite by a chemical change involving ion transfer or diffusion (Higazy, 1949; Perrin and Roubault, 1949; King, 1948). Pegmatites which are zoned or show evidence of injection, such as dilation or distortion of foliation, cannot be explained by metasomatism. Further, metasomatism does not explain the coarse pegmatite textures.

The second concept involves a low temperature liquid enriched in incompatible elements, which forms in advance of a wave of deep-seated granitization, and is emplaced by injection (Backlund, 1946).

One problem with the granitization theory is that it does not account for basic elements eliminated from the granitized rock. Neither does this theory account for the source of the granitizing fluid, or the fact that some pegmatites are not associated with granitoids (the "wave" of granitization). Coarse crystalline bodies grown from low temperature aqueous solutions or brines are not "granites" (Flawn, 1950).

3.4.2. THE ANATEXIS THEORY.

Pegmatites which are developed in migmatite complexes may have formed due to the partial melting of crustal rocks, and subsequent crystallization with a vapor phase present Flawn, 1950). Simple pegmatites, which intimately penetrate their host and are unrelated to larger granitic intrusions, probably originated locally via anatexis. Pegmatites which contain large concentrations of rare minerals were probably generated externally, unless there is evidence for an abundance of rare elements in the host rock.

Zoned pegmatites are probably not formed during anatexis, since most discussions of the anatexis theory cite an equilibrium fusion process that does not allow for large scale differentiation (Flawn, 1950). The anatexis theory differs from the magmatic theory in that the latter allows for differentiation via crystallization.

3.4.3. THE OPEN SYSTEM THEORY.

The open system theory suggests that pegmatites form from the successive crystallization of solutions moving through the system (De Ameida and Johnston, 1944; Quirke and Kremers, 1943). Zoned pegmatites are then the result of a progressive filling or precipitation of crystals from solutions changing in composition. A criticism of the open system theory is that some zoned pegmatites contain internal pod-like units completely surrounded by other units, a feature indicative of closed system crystallization (Flawn, 1950).

3.4.4. THE MAGMATIC THEORY.

Magmatic theorists contend that pegmatites form in a closed or restricted system, following from the injection of a granitic rest-magma (Flawn, 1950). Through crystallization in successive stages, the rest magma produces a sequence of mineralogical and textural units, provided the physical conditions during crystallization are favorable. This theory thus provides an explanation for zoned and complex pegmatites, as well as for those pegmatites which show signs of injection. However, this theory requires the association of a major granitic body which would have provided a rest-magma during its crystallization.

In summary, pegmatites do not all form as a result of the same process, or at least processes operating in a like manner. The mode of origin and emplacement postulated for any pegmatite studied must be supported by the mineralogy, form, structure, and host terrane with which it is associated.

The dike form of the pegmatites in the southeastern Adirondacks indicate that they have been injected, and their formation is thus supported by the magmatic theory of genesis. The association of these pegmatites with migmatite complexes in the southeastern Adirondacks and the lack of a larger granitic source magma suggests that their magmas formed due to locally fluid-saturated anatexis.

CHAPTER 4. PROCEDURE AND RESULTS.

4.1. FIELD WORK.

Geological reconnaissance in the Putnam, Glens Falls, and Saratoga Springs 7.5 minute quadrangles of the southeastern Adirondacks revealed the presence of undeformed granitic pegmatites intruding middle Proterozoic metamorphic rocks, as well as locally extensive migmatization, particularly within the metasedimentary units of the central and northern portions of the Putnam quadrangle and in the southwestern portion of the Glens Falls quadrangle (see Figure 1.3.1). The pegmatization and the migmatization are believed to be contemporaneous because of their local intimate association. The migmatites in the area of P6, for example, are locally pegmatitic.

The size and form of the pegmatite bodies vary considerably, from small, irregular apophyses to meter-thick dikes. Larger irregular bodies or pods with several meters of outcrop also occur. The mineralogy is relatively consistent, with varying amounts of quartz, alkali feldspar, and plagioclase, minor biotite and muscovite, and occasional garnet and hornblende. Mutually intergrown, euhedral to subhedral crystals of quartz, plagioclase, and alkali feldspar in the pegmatite are up to 5 centimeters long.

In outcrop, the pegmatites exhibit structural features characteristic of plutonic igneous rocks. They discordantly and concordantly intrude well foliated quartzofeldspathic gneisses, which contain variable amounts of hornblende, biotite, and garnet (see Figure 4.1.1). Aplitic textures are rarely observed, and

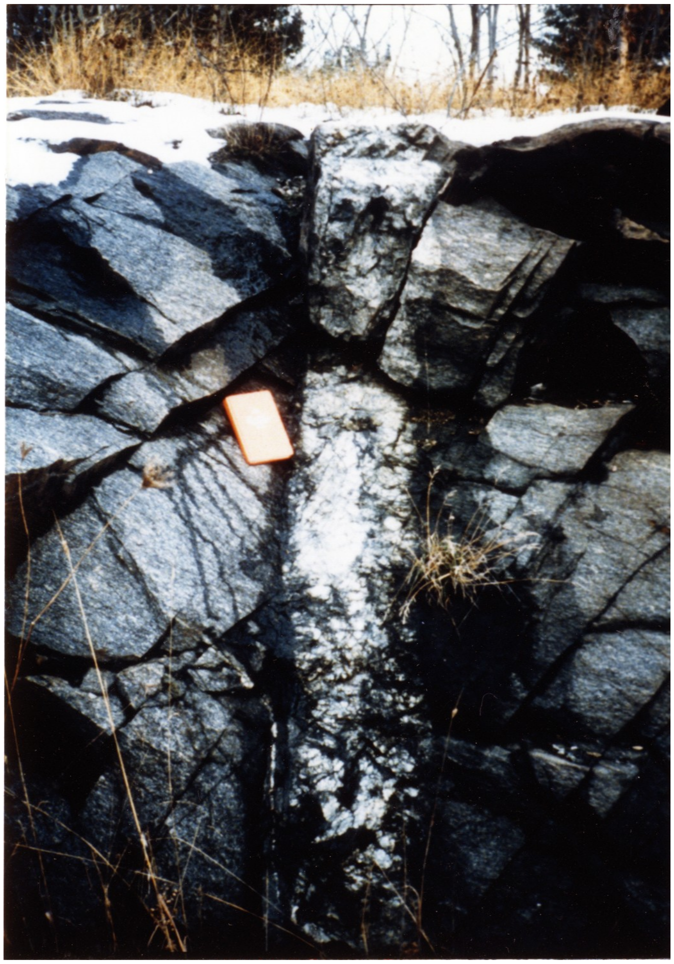


Figure 4.1.1. Pegmatite P2, showing dike form.

are present in thin stringers (roughly 2 centimeters wide) which extend from large pegmatite bodies into the country rock, and at the contact between some pegmatite dikes and their hosts, extending for a few centimeters along the foliation plane into the country rock. Some of the pegmatite intrusions contain small, irregular fine-grained pods which exhibit graphic intergrowths of quartz and plagioclase. The larger irregular pegmatite bodies contain xenoliths of the country rock, up to a meter in size. Near the town of Putnam Station, for example, along the contact between a leucogranitic gneiss and the structurally overlying metasedimentary units, xenoliths of graphitic marble, calc-silicate marble, and biotite paragneiss are observed within a pegmatitic granite (see Figure 4.1.2). This particular pegmatite appears to originate as a melted fraction of the leucogranitic gneiss. This is an example of a pegmatite which was not sampled due to its obvious heterogeneity and therefore difficulty in obtaining a representative sample.

Field relationships between the pegmatites and the gneisses into which they intruded were examined. In strike the pegmatites sampled are oriented roughly N-S to NE-SW, although pegmatites which strike roughly E-W are also observed (but which, by coincidence, were not appropriate for sampling). Commonly, a narrow (< 5 centimeters) concentration of mafic minerals occurs along both sides of the pegmatite-gneiss contacts, which may represent a reaction zone or an immobile or residual volume of material from the gneiss which did not enter the pegmatite magma phase during melting (see Figure 4.1.3). The host gneisses

Figure 4.1.2. Graphitic, granitic pegmatite containing xenoliths of biotite paragneiss (top central portion of photograph), intruding calc-silicate-rich metasediments.

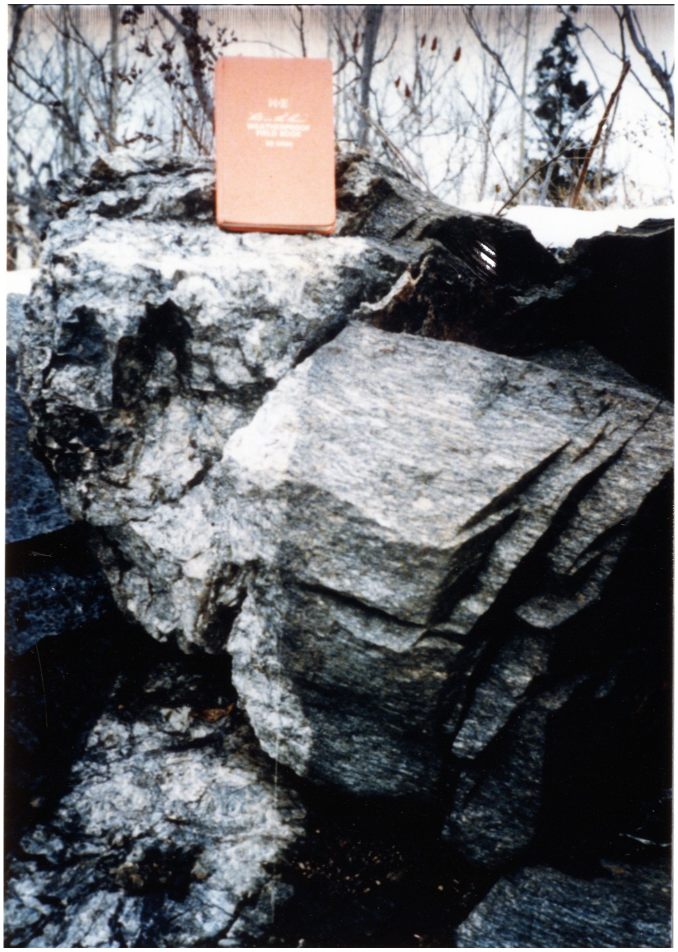


Figure 4.1.3. Mafic-mineral concentration at the contact between pegmatite P2 and host gneiss.

consist predominantly of quartz and feldspar, with commonly less than 35% mafic material. They are fine- to medium-grained (roughly 1 to 3 mm), and fairly equigranular.

At some of the contacts between host gneiss and pegmatite in the area studied, two megascopic foliations are produced by the preferred orientation of biotite and hornblende. The primary foliation parallels the regional foliation, while the secondary foliation seems to parallel the orientation of the pegmatite body.

Gneissic banding is better developed away from the pegmatite contacts. Foliation strikes roughly E-W in the Putnam-Whitehall area, roughly NE-SW in the Glens Falls area, and roughly N-S in the Saratoga Springs area where the pegmatites were sampled. Locally, the gneisses exhibit augen structure; this is generally absent adjacent to the pegmatites.

It is apparent from field observations that the pegmatites represent a late-stage magmatic event. Undeformed pegmatite bodies are found within the same outcrop as compositionally similar leucogranitic masses, which are sheared along the regional foliation (see Figure 4.1.4). Local portions of one foliated leucogranitic mass, which outcrops on the west side of Route 22 near the town of Putnam Station, have retained their initial pegmatitic, igneous character, and thus seem to be relicts which have escaped the effect of the regional deformation event. This is evidence for only one generation of pegmatite which continued until after the deformation event (see Figure 4.1.5). It is possible that the pegmatites and migmatites, as

Figure 4.1.4. (a) Undeformed pegmatite (oriented in a north-west direction of the photograph), extending from (or feeding) deformed pegmatites (oriented in a north-south direction of the photograph), which parallel the regional foliation; (b) strained pegmatite.

Figure 4.1.5. A partially deformed pegmatite body. Central and upper left-hand region of the photograph show undeformed portion, which extends into the deformed portion in the right-hand side of the photograph. This is a close-up of lower portion of photograph of Figure 4.1.4a.

isolated pockets of silicate melt, acted as zones of high fluid pressure, localizing strain and providing lubrication for the initiation of thrusting during a late stage of the Grenville tectonic event. Norris and Henley (1976) describe a similar process in reference to the development of crosscutting veins in the Haast Schists of Otago, New Zealand, each vein set associated with a unique episode of deformation. Here, a fluid phase may have entered and hydraulically wedged open fractures, transporting dissolved components prior to deposition as a vein assemblage. In the Adirondacks, the analogous "fluid" might be an anatectic melt with or without a coexisting vapor phase.

4.2. LOCATION OF SAMPLES AND SAMPLING TECHNIQUE.

Sampling was limited to those pegmatite bodies which do not exhibit deformation or megascopic features indicating obvious heterogeneity, such as mineral zoning in outcrop or inclusions of country rock.

Seven pegmatite dikes (P1 through P5) and sills (P6 and P7) were sampled (see Figure 1.3.1). They intrude in a similar orientation, striking N-S to N59oE, and dip between 38 degrees to the east and 74 degrees to the west, the sills being concordant to the foliation. P1, P2, and P3 intrude a fine- to medium-grained quartzofeldspathic gneiss, which locally contains biotite, hornblende, and garnet. P1 and P2 are located at the intersection of Route 22 and Clemons-Hulett's Landing Road, approximately 8 miles (13 km) north of Whitehall, New York, in the Putnam quadrangle.

P1 outcrops on the south side of the Clemons-Hulett's Landing Road, approximately 50 meters west of Route 22. Discordant to the foliation, which strikes roughly east-west and dips to the south, P1 appears as a vertical, tabular dike, which strikes N8oE. It is approximately 20 centimeters wide and maintains a fairly constant width.

P2 is observed on the north side of the Clemons-Hulett's Landing Road, approximately 100 meters east of Route 22. It is also tabular in shape and discordant to the regional foliation. It is roughly 30 centimeters wide, and vertical, striking N10oE. Its width remains constant throughout the outcrop.

Approximately 8 and a half miles (14 km) north of Whitehall is a paved road which heads due west. A stream intersects this road 500 meters west of Route 22. P3 is located roughly 500 meters upstream from this point, on the south side of the stream. On the lower part of the outcrop, P3 is nearly vertical, but it bends to the east toward the top of the outcrop. The bend occurs roughly at the contact between a fine-grained mafic unit and an overlying quartzofeldspathic augen gneiss. P3 maintains a fairly consistent thickness of 35 centimeters, and is discordant to the foliation.

P4 and P5 are located approximately 12 miles (20 km) north of Whitehall along Route 22 in the Putnam quadrangle. These two dikes are not readily observed from the highway. However, a massive pegmatite body, several meters in width, is observed from the highway. P4 is about 10 meters to the south of this massive body, and P5 is roughly 100 meters to the south of P4. Both intrude a well-foliated metasedimentary gneiss, with foliation striking roughly northeast, and dipping to the southeast.

P4 is roughly 9 centimeters thick and vertical, striking N59oE. P5 is approximately 22 centimeters thick, and strikes N50oE and dips 74 degrees to the northwest.

P6 outcrops on the south bank of the Hudson River, across from the Sherman Island Power Plant, roughly 3 miles (5 km) southwest of Glens Falls, in the Glens Falls quadrangle. The pegmatite can be reached via paved and dirt road from Speir Falls Road at exit 17N of the Adirondack Northway. P6 is approximately 38 centimeters wide, and is oriented roughly parallel to the regional foliation, which strikes N-S and dips 38 degrees to the east. Migmatitzation is extensive in this area.

P7 intrudes a metapelite on the north side of Granite Lake, a short distance to the east from Braim Road, which intersects Daniels Road roughly one and a half miles (3 km) northwest of Saratoga Springs, in the Saratoga Springs quadrangle. P7 is approximately 1 meter thick and parallels the regional foliation, which strikes N54oE, and dips to the southeast.

The pegmatites were etched with hydrofluoric acid and stained with sodium cobaltinitrate at various places along and across the outcrop to assure that they were grossly homogeneous. Outcrop heterogeneity on this scale induces a basic problem in obtaining a truly representative bulk sample.

Large representative hand samples to be used for bulk chemical analyses were collected along and across each body. Thin sections were cut from these samples for petrographic analysis and electron microprobe mineral analysis, in order to evaluate the extent of variation in feldspar composition throughout each pegmatite.

Modal analyses of samples from the pegmatites, slabbed and stained with sodium-cobaltinitrate for easier identification of quartz, potassium-rich feldspar, and plagioclase feldspar phases, indicated that these phases were present in proportions appropriate to granites. However, potassium-rich feldspar predominated over quartz or plagioclase in most cases, suggesting a potential problem using the presently available experimental data on the water-saturated "granite" system (that is, with H2O the only volatile) to estimate the pressure and temperature of formation of the pegmatite.

The host gneisses were sampled on a meter scale at several intervals away from their contact with the pegmatites in order to evaluate any effect of the introduction and/or crystallization of the pegmatite magma, viz: P1, at 5, 10, and 20 meters from the contact; P2, at 0 and 5 meters from the contact; P3, at 0, 5, and 20 meters from the contact; P4, at 0, 10, and 20 meters from the contact on the south side, and at 0 and 1 meter from the contact on the north side; P5, at 0, 1, and 3 meters from the contact; P7, at 0, 1, and 5 meters from the contact.

This sampling of the host gneisses reflects the accessibility of outcrop, the attitude of the foliation, and the presence of factors which may have affected the mineralogy, such as additional intrusive bodies, migmatization, or fault zones. For example, a gneiss which was intruded by two pegmatite dikes 40 meters apart, was sampled no further than 20 meters from the pegmatite under study, to limit any possible effect due to the presence of the adjacent pegmatite or its emplacement. In fact, the host rock of P6 is itself a heterogeneous migmatite, riddled with pegmatitic bodies, and was thus omitted from the systematic study of the effects of the pegmatites on the host gneiss.

The gneisses were sampled along strike of the foliation to insure that a lithologic/structural contact was not crossed, and thus that the mineralogy of gneisses of different units and possibly different origin was not being compared. In cases where the plane of the intrusive pegmatite body was dipping, the gneiss was sampled on the hanging wall side, since it would more likely show the effect of any exchange or metasomatic infiltration of material released from the crystallizing pegmatite magma.

4.3. PETROGRAPHIC ANALYSIS.

4.3.1. PEGMATITES.

Thin section examinations show that quartz, alkali feldspar, and plagioclase are present in all of the pegmatites sampled for this investigation. The pegmatites exhibit coarse-grained allotriomorphic granular texture, in which individual crystals are generally anhedral. The alkali feldspar is microcline microperthite. Although perthitic, alkali feldspar appears homogeneous in terms of the concentration of exsolution lamellae and the orthoclase content of the host phase. Plagioclase composition ranges from An12 to An28 (oligoclase), as determined by extinction angles on albite twins. Compositional zoning of plagioclase was not observed in any samples, although plagioclase is antiperthitic in P1 and P4 (see Figure 4.3.1). The maximum size of quartz, alkali feldspar, and plagioclase crystals in thin section ranges from 3 millimeters to greater than 15 millimeters. There does not seem to be any preferred orientation of quartz or feldspar phases. Muscovite is present in P4 and P5. Lepidolite is observed in P1 and P6 in grains up to .2 millimeter long.

Mafic phases comprise a minor (less than 10%) portion of the pegmatites. Biotite is commonly present as .5 to 1 millimeter long grains. Hornblende is observed in P4. Opaque phases are minor but common, reaching 3 millimeters long. Magnetite-ilmenite exsolution is observed in P5. Rutile is observed as euhedral prismatic crystals up to 2 millimeters long and as tiny needles in biotite and inclusions in large feldspar grains (see Figure 4.3.2).

Accessory phases include zircon, apatite, and allanite, the grains of which are euhedral to subhedral in form, and rarely up to .2 millimeters long. In P1, however, allanite reaches 3 millimeters and exhibits a characteristic metamict structure, due to radiation damage in its crystalline structure (see Figure 4.3.3). The zircon grains in the pegmatites, in general, show euhedral overgrowth features, as opposed to the rounded nature

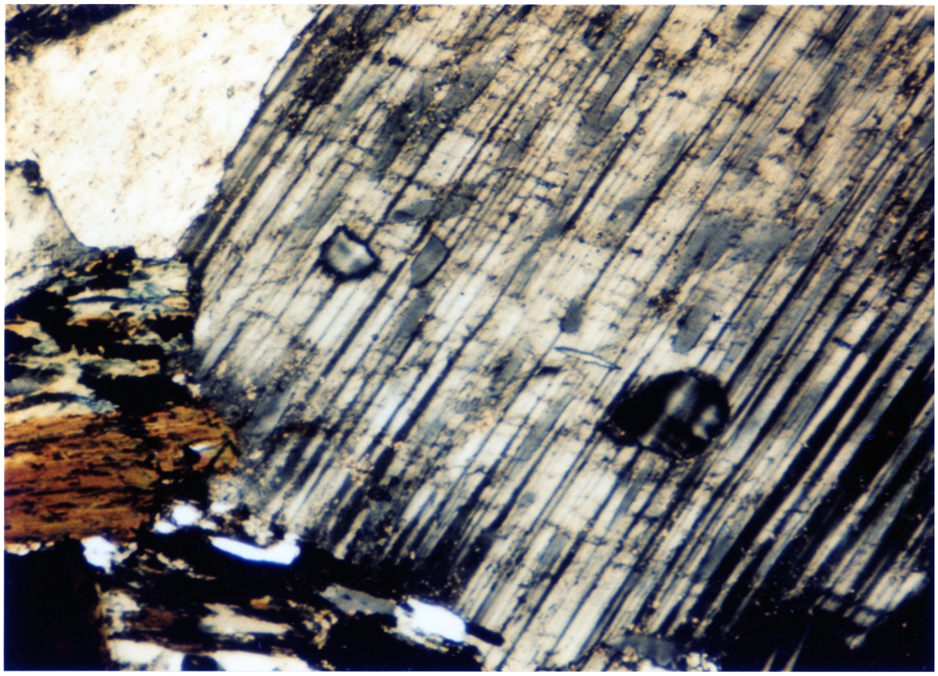


Figure 4.3.1. Antiperthitic plagioclase from pegmatite P5. Hornblende is at lower left corner of plagioclase grain. East-west length is 2 mm.

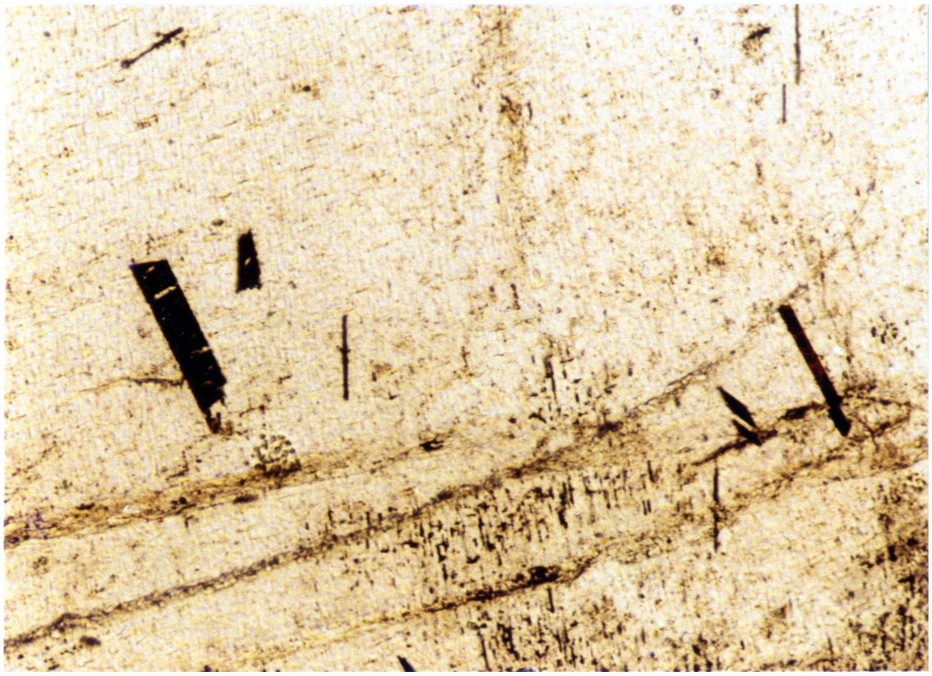


Figure 4.3.2. Rutile in alkali feldspar from pegmatite P4.
East-west length is 0.34 mm.

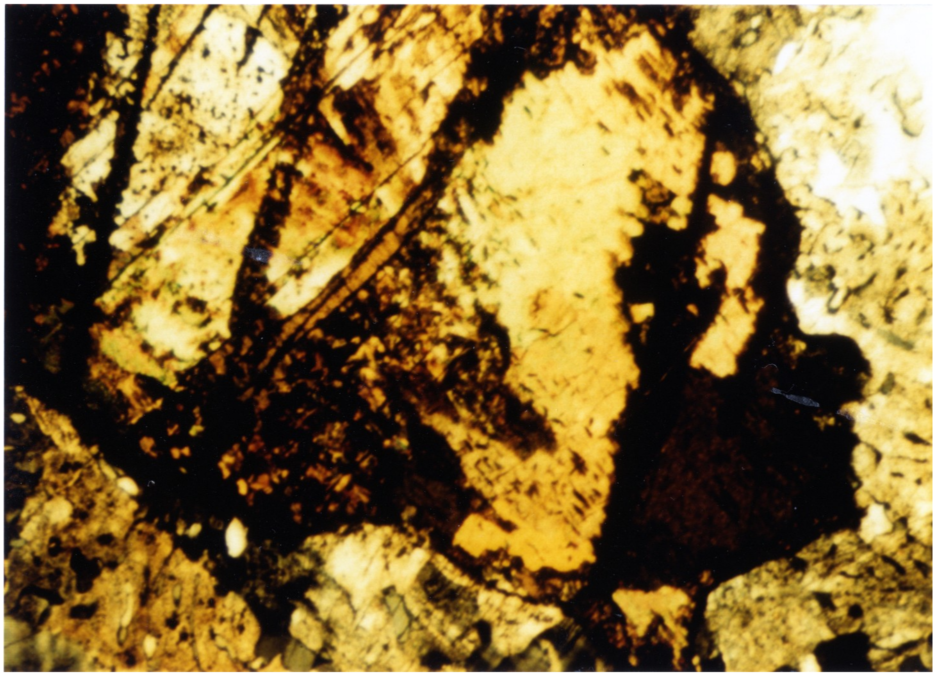


Figure 4.3.3. Allanite crystal from pegmatite P1.
East-west length is 3.4 mm.

which they exhibit in the host gneisses. The amount of zircon in each pegmatite seems comparable to the amount in the relative host gneiss.

Mafic and accessory minerals occur as inclusions within the large quartz and feldspar grains, indicating that they may have crystallized relatively early in the pegmatites' parageneses. Muscovite, however, is observed along quartz-feldspar and feldspar-feldspar grain boundaries as radiating masses, and may have formed synchronously with or later than the felsic minerals as a post-consolidation or deuteric phase.

4.3.2. HOST GNEISSES.

A petrographic analysis of the gneisses which host the pegmatites serves three main purposes. Generally, it is important to understand the effect of the formation or presence of the pegmatite melt on the mineralogy, chemistry, texture, and structure of the wall rock. Specifically, it is believed that in the presence of a volatile phase, where vapor pressure equals or is near lithostatic pressure, partial melting is induced if temperatures exceed the vapor-saturated solidus in the host rock. Chemical constituents enter the melt phase or are retained in residual solids, depending on their partition coefficients. A variation in the mineralogy of the gneiss toward a locus of melting should indicate relative affinities of chemical constituents for the phases present during the production and intrusion of the melt, or alternatively indicate chemical exchange occurring via the introduction of an intruded melt and a release of volatiles.

Finally, the mineralogy of the host rocks should be studied for indications that the pegmatites have formed approximately *in situ*, with minimal intrusion displacements, and are not the result of magmatic injection from removed or non-related sources.

The host rocks of each pegmatite were sampled and thin-sectioned at various intervals away from the contact as explained previously, and studied under a petrographic microscope. A modal analysis based on roughly 300 counts for each thin section at a 1 millimeter spacing, was used to estimate the approximate volume percentages of the minerals present.

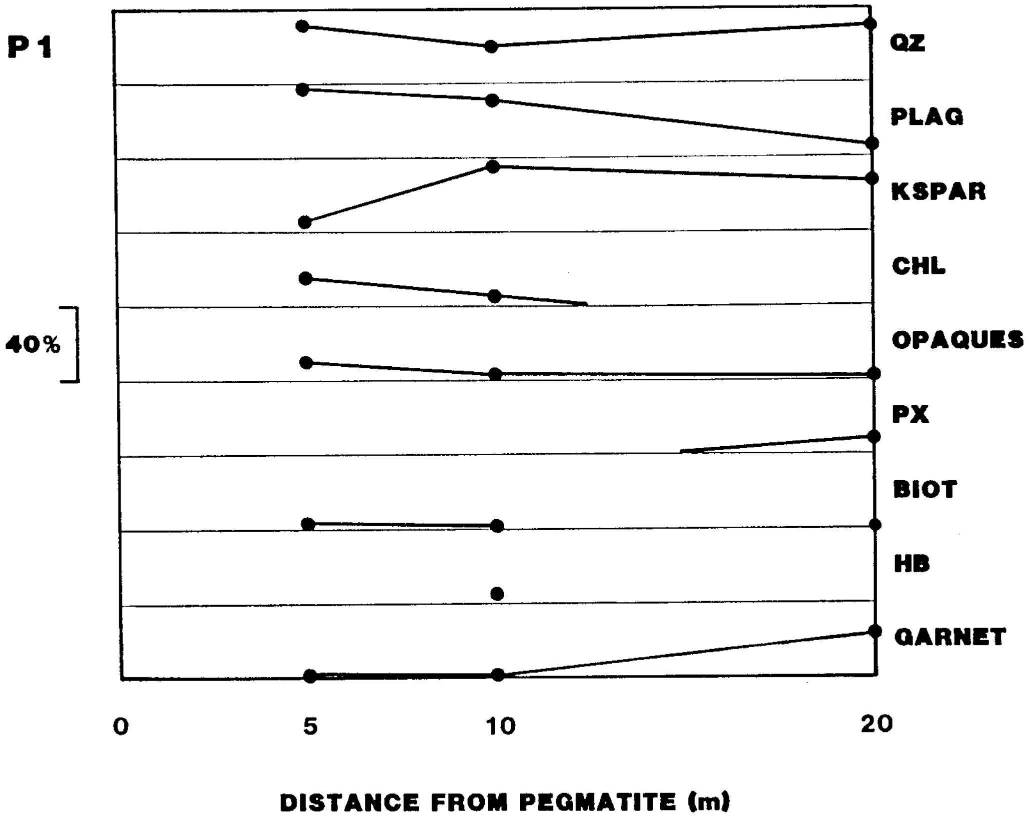
The general features of the petrography and the modal analyses are summarized here. Some specific examples and exceptions are noted as I deemed them important to the understanding of the effect of the production and emplacement of the melt.

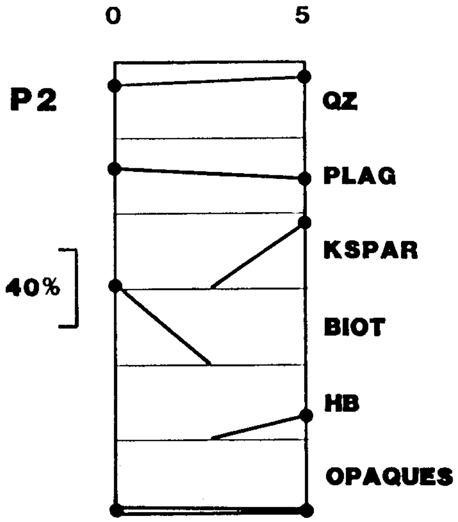
As a result of the petrographic study and the modal analysis, in addition to field observations, several apparent trends in the mineralogy, chemistry, texture, and structure were observed and are discussed in the following sections.

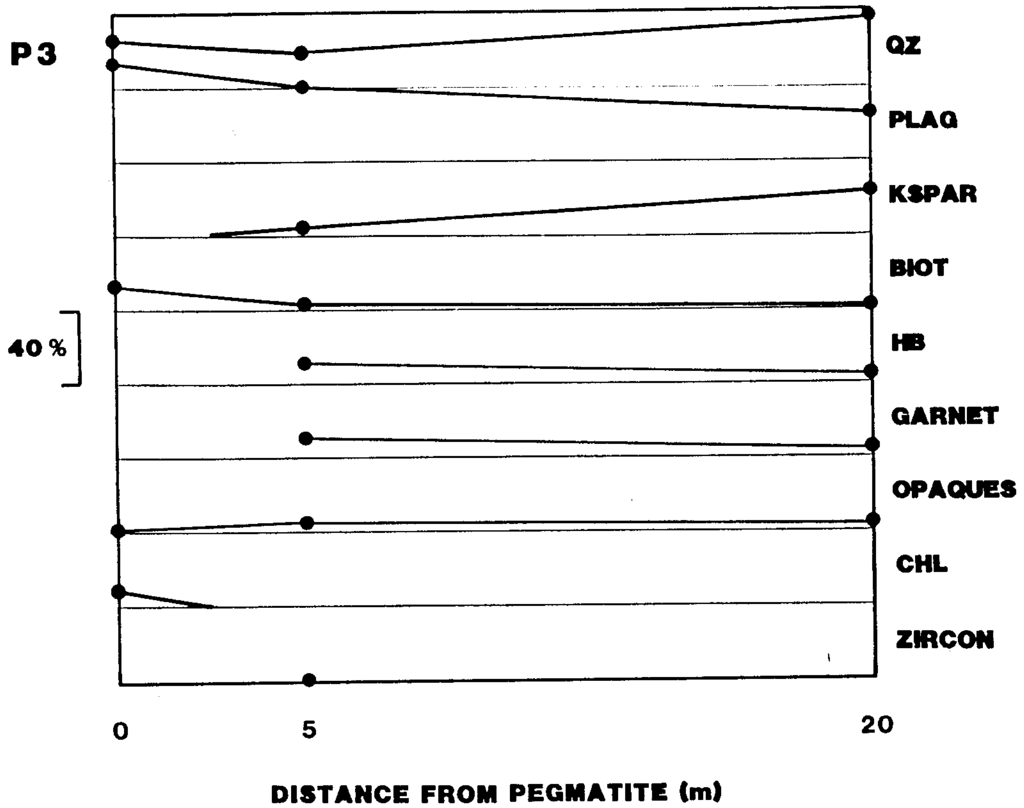
4.3.2.1. VARIATIONS IN MINERALOGY.

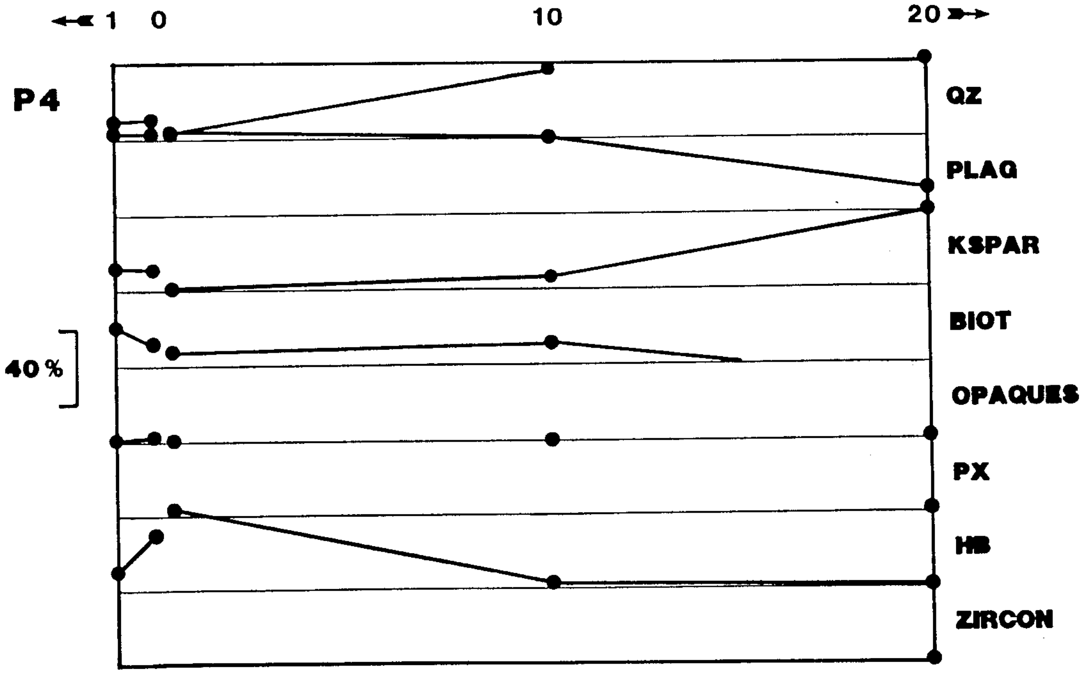
Mineralogically, four consistent trends are apparent within the host gneisses. Figure 4.3.2.1.1 graphically depicts the variation in mineralogy of the host rock, with distance away from the intrusive contact. In 5 of the 6 pegmatite-host rock systems studied, plagioclase increases toward the contact, by an average of roughly 25%. This increase begins roughly between 5 and 20 meters from the pegmatite. Alkali feldspar, in contrast,

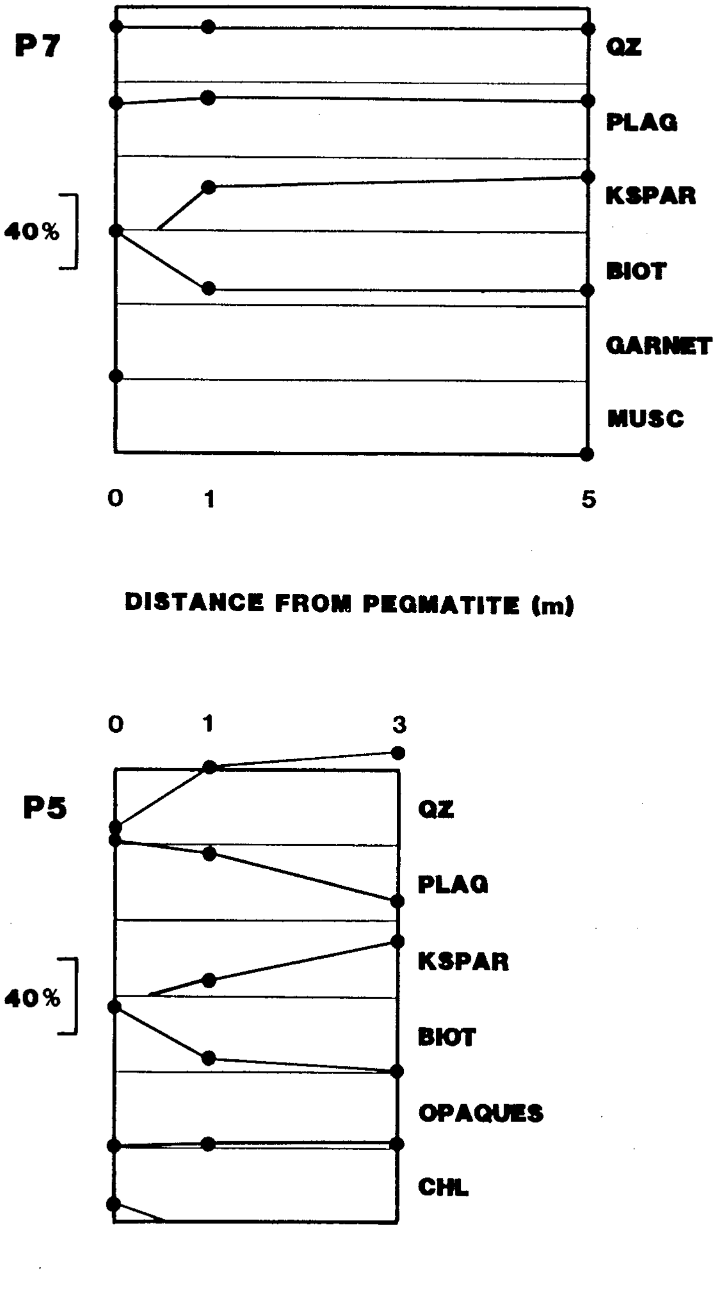
Figure 4.3.2.1.1. Variation in mineralogy of host gneisses for P1-P5 and P7.


P1
QZ
PLAG
KSPAR
CHL
OPAQUES
PX
BIOT
HB
GARNET
40%
0
5
10
20
DISTANCE FROM PEGMATITE (m)


0
5
P2
QZ
PLAG
KSPAR
BIOT
HB
OPAQUES
40%


P3
40 %
QZ
PLAG
KSPAR
BIOT
HB
GARNET
OPAQUES
CHL
ZIRCON
0
5
20
DISTANCE FROM PEGMATITE (m)


1
0
10
20
P4
40 %
QZ
PLAG
KSPAR
BIOT
OPAQUES
PX
HB
ZIRCON


P7
QZ
PLAG
KSPAR
40%
BIOT
GARNET
MUSC
0
1
5
DISTANCE FROM PEGMATITE (m)
0
1
3
P5
QZ
PLAG
KSPAR
40%
BIOT
OPAQUES
CHL

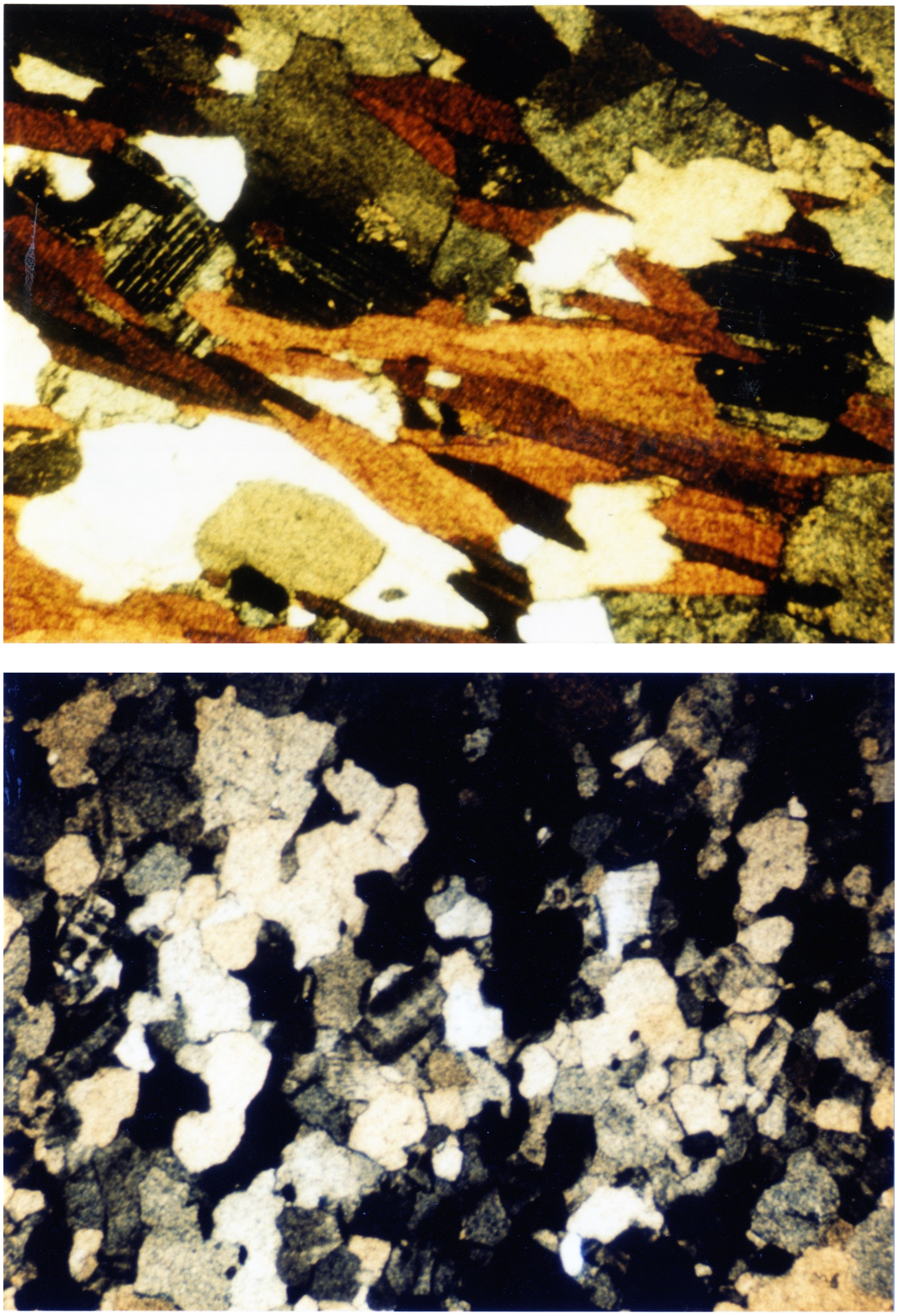


Figure 4.3.2.1.2. Variation in mineralogy of host gneiss away from the pegmatites. a) Biotite-plagioclase gneiss at the contact with pegmatite P7, and b) Quartz-microcline gneiss, roughly one meter from pegmatite P7. East-west length is 3.4 mm.

decreases in amount toward the contact by roughly 30% in all of the gneisses examined. Hydrous minerals such as biotite, hornblende, and chlorite increase toward the contact in all of the gneisses, usually within 5 meters of the pegmatite-gneiss contact, by up to roughly 40% (see Figure 4.3.2.1.2). This concentration is obvious in the field, where it is observed as a roughly 5 centimeter thick mafic-rich contact zone, as opposed to the variation in feldspars, which is a more subtle change seen only from the modal analyses of thin sections. Relatively "dry" minerals prevail away from the contact. Quartz, for example, decreases toward the contact in all cases by up to 40%. Garnet decreases by roughly 25% toward the contact of P1. Roughly 2.5% garnet is present at the contact of P7, but in no other contact samples is garnet observed. At the contact of P7 it increases from 0 to 12%. Pyroxene is observed in host gneisses of P1 and P4, but only at distances of greater than 10 meters from the intrusive contact. The following reaction describes the variation in mineralogy of the host gneiss of P1 (see Figure 4.3.2.1.1):

pyroxene + hornblende + garnet + $H2O$ + $O2$ = biotite + chlorite + opaques.

4.3.2.2. CHEMICAL VARIATIONS.

Although the above variations in mineralogy are obvious, they do not necessarily imply major changes in the bulk chemistry of the host gneiss related to pegmatite formation, such as an introduction of sodium from the crystallizing magma, which could account for the concentration of plagioclase in the host gneiss toward the pegmatite body. Actually, reduction of iron oxides and consumption of potassium-rich feldspar, could account for the high plagioclase to potassium-rich alkali feldspar ratios near the pegmatite, according to this reaction:

K-rich feldspar + Fe_3O_4 + H_2O = biotite + O_2.

Therefore, variations of the bulk chemistries of the host gneisses may aid in understanding the effect of the emplacement of the pegmatite on the host gneiss, or conversely, on the pegmatite itself, to distinguish these possibilities.

The bulk chemistries of the host gneisses were estimated from modal analyses and the use of the following 2 equations:

$X(i) = V(i)d(i)/V(T)d(T)$ and

$C(et) = X(i)C(ei) + X(j)C(ej) + \ldots + X(k)C(ek)$,

where i,j, and k are mineral phases present in thin section and T denotes total mineral phases present; X(i) is the weight fraction of phase i; V(i) is the volume percent of phase i determined from the modal analysis; V(T)d(T) is the summation of Vd of all phases present; d(i) is the density of phase i,; C(ei) is the percent of component e in phase i; and C(et) is the total weight percent of component e. Thus, there is a direct relationship between mineralogy and chemistry (compare Figures 4.3.2.1.1 with 4.3.2.2.1). The volume percentage of each mineral phase in the host rock was determined via modal analyses of thin sections, taken at several intervals away from the contact, as explained previously. The density values and the percentages of major oxides in each mineral phase is based on the chemical analyses of representative minerals from Deer et al (1966). Minerals from

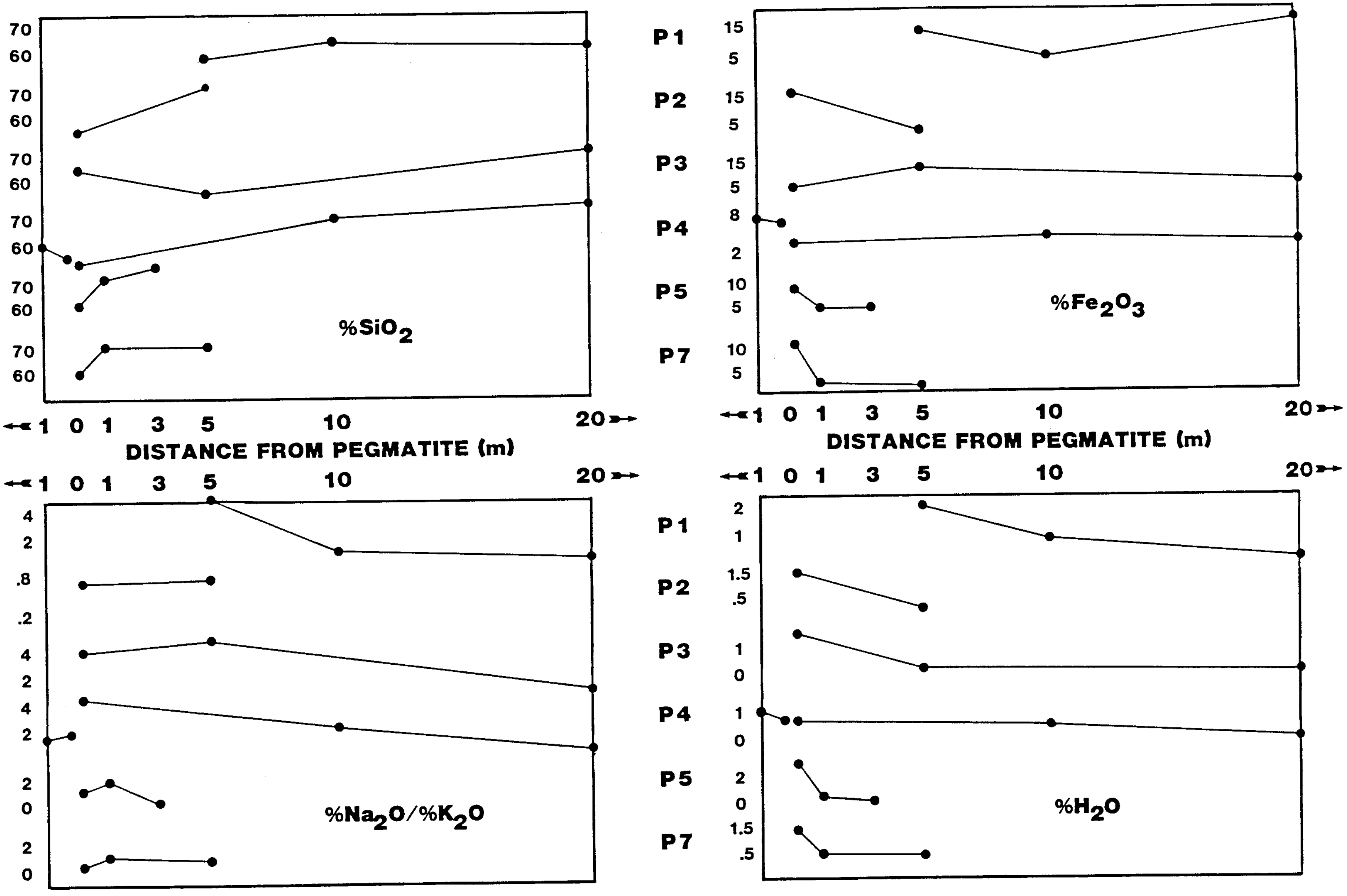


Figure 4.3.2.2.1.Chemical variations in host gneisses. Points represent thin sections from which estimates were made.

the host gneisses themselves were not analyzed chemically. The estimate is considered justified because the mineral phases of the host gneisses do not seem to show any compositional variation (petrographically) with distance from the pegmatites. The results of the calculations are listed in the Appendix.

A compilation of the information obtained from all of the host gneisses suggests that their bulk chemistries show common features that appear related to the emplacement of the pegmatites. There are four discernible effects (see Figure 4.3.2.2.1); namely:

First, the gneiss samples from the pegmatite-gneiss contacts show that the host is lower in SiO2 content in comparison to the gneiss away from the contact. Gneisses 5 and 7 indicate that this SiO2 depletion disappears at a distance of less than one meter from the contact, where silica increases by up to roughly 18%, and then remains relatively constant away from the pegmatite.

Gneiss 4, however, remains low in SiO2 up to one meter away from the pegmatite, and gneiss 3 indicates that it remains low in SiO2 up to 5 meters from the pegmatite. The pegmatite which intrudes gneiss 4 (P4) is also notably low in silica (less than 1% modal quartz). There is a large pegmatite body roughly 10 meters north of P4, which may be the reason for the depletion of SiO2 in the host rock. Also, this host gneiss is, in places, an amphibolite, which may have originally been poor in silica. Gneiss 3 also contains a mafic unit, roughly 15 meters thick, in the area of P3. Beyond 5 meters, the gneisses do not seem to have been affected by the pegmatites in terms of silica content.

Second, iron is concentrated in the gneisses, as are magnesium and titanium, at their contacts with the pegmatites. This concentration disappears, similar to the silica depletion, usually at a distance of less than one meter from the contact. The iron content in gneisses 4 and 1, however, remains constant. Iron content varies, but is relatively low in the gneisses away from the contact.

A common feature in most of the pegmatites is a low Na2O to K2O ratio. Since this is one of the main reasons for investigating the host rock, it is significant, and represents the third observed effect of emplacement. Although the variation in mineralogy away from the pegmatite suggests that Na2O/K2O may decrease immediately away from the contact, the variation in chemistry shows that this ratio is consistent up to roughly 5 meters away from the contact, where it then begins to decrease, and finally becomes characteristic of medium- to high-grade metapelitic gneisses by roughly 20 meters away from the contact (e.g. see Walther and Orville, 1982).

Finally, the water content in the gneisses is higher at the pegmatite-gneiss contacts than away from the pegmatites. Gneisses 5 and 7 show that there is a marked decrease within one meter away from the contact, while Gneiss 1 suggests that water content may remain relatively high up to 5 meters away from the pegmatite body.

4.3.2.3. TEXTURAL VARIATIONS.

Textural and structural trends are not as obvious, but some variations in the gneisses with distance from the contact are inferred to be the result of the formation and emplacement of the pegmatite bodies.

The gneisses are generally equigranular, with grain sizes of roughly .5 to .75 millimeter, but which may be up to 3 millimeters, and they typically exhibit a granular or granoblastic to decussate texture. The gneisses seem to exhibit a texture which is characteristically metamorphic, but which becomes equigranular and more characteristic of igneous rocks at the contact between the pegmatite and the host gneiss.

In felsic layers, quartz is usually coarser than the feldspar, while in mafic layers, felsic and mafic minerals are equigranular. Gneiss 1 becomes less equigranular away from the contact, with grain size ranging from 1.2 to 3 millimeters at a distance of 20 meters from the contact. Gneiss 4 also becomes less equigranular away from the contact with grain size ranging from .1 to 2 millimeters at a distance of 20 meters from the pegmatite. Gneiss 5 exhibits a general reduction in grain size and a decrease in equigranularity away from the contact to 3 meters away, where grain size ranges from .25 to 1 millimeter. The granoblastic texture of gneiss 5 at the contact becomes more decussate from the contact to 3 meters away. Gneiss 7 also becomes less equigranular from the contact to 5 meters away from the contact, where grain size ranges from .2 to 3 millimeters. Biotite flakes in gneiss 7 are roughly 2 millimeters long at the intrusive contact, but sharply decrease to .75 millimeter in length at one meter away from the contact.

4.3.2.3. STRUCTURAL VARIATIONS.

Foliation is better developed in the gneisses away from the pegmatite contacts. The biotite of most gneisses exhibits preferred orientation in 2 directions at the contact (see Figure 4.3.2.3.1). One orientation is roughly parallel with the regional foliation; the other is roughly parallel with the attitude of the intrusive body. Gneiss 4 exhibits a marked increase in metamorphic differentiation of mafic and felsic material with distance from the intrusive body. The thickness of gneissic bands increase to 30 millimeters at a distance of 10 meters from the contact. The biotite of gneiss 5 is randomly oriented at its contact with the pegmatite, while a marked foliation is observed in the gneiss away from the contact with the pegmatite.

With the exception of gneiss 3, augen of any type are not observed in the gneisses at their contact with the pegmatites. In gneisses 1 and 7, quartz and feldspar augen appear at 20 and 5 meters away from the intrusive contact, respectively.

In summary, the petrography indicates that the formation and/or emplacement of the pegmatite melts is accompanied by several consistent changes in the mineralogy, chemistry, texture, and structure of the host gneiss. The metamorphic character of the gneisses seems better preserved or stronger away from the pegmatite bodies, while at the pegmatite-gneiss contact, the gneiss seems to have acquired an igneous textural character, as a result of the former presence of the melt.

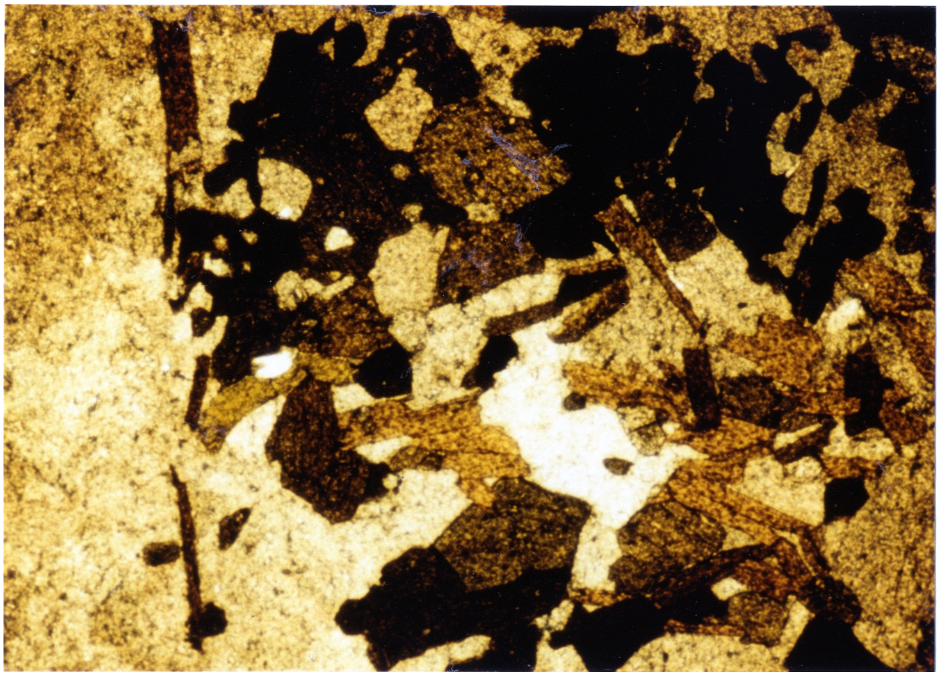


Figure 4.3.2.3.1. Biotite and hornblende in host gneiss of P4, showing 3 directions of orientation. Pegmatite is at extreme left of the photograph, oriented north-south of the photograph.
East-west length is 3.4 mm.

4.4. BULK CHEMICAL ANALYSIS.

The results (listed in Table 5.1) were obtained via atomic absorption on the direct current plasma machine at Lamont-Doherty Geological Observatory.

Approximately 2 to 5 kilograms of representative sample material from each pegmatite were crushed and split to approximately 100 grams, and reduced to a powder passing 200 mesh using a tungsten-carbide shatterbox. From each powdered sample, just in excess of 200 milligrams were removed and heated at 950o Centigrade for 30 minutes. This preignition step dries and oxidizes the sample by driving off H_2O and other volatiles.

Two hundred milligrams of powder from each heated sample were mixed in graphite crucibles with 800 milligrams of lithium metaborate, and fused in an oven at 1060o Centigrade for 15 minutes. The lithium metaborate induces fusion and aids in producing a homogeneous glass bead for subsequent analysis.

Each fused bead was dissolved in 100 milliliters of dilute nitric acid (4% HNO_3) spiked with a 10 ppm germanium internal standard. The germanium served as a measure of drift during analysis, when the samples were compared with a blank of dilute nitric acid and germanium. These solutions were analyzed for the trace elements P, Zr, Y, Rb.

One milliliter of each of these solutions was further diluted in 24 milliliters of dilute lithium carbonate solution, also spiked with 10 ppm of germanium as a standard. These dilutions were analyzed for the major elements SiO_2, TiO_2, Al_2O_3, Fe_2O_3, MnO, MgO, CaO, Na_2O, and K_2O, and for Ba and Sr in each sample.

4.5. MINERAL ANALYSIS.

4.5.1. FELDSPARS.

Comparison of the compositions of coexisting feldspars with experimental phase equilibria data can be used as a measure of equilibrium conditions during crystallization (Yoder et al, 1957) and as an indicator of the temperature of crystallization (Stormer, 1975). However, determinations of core and rim compositions of single grains are needed to indicate the range of compositional zoning in minerals, so that fractionation effects, xenocryst presence, or other indications of equilibrium restriction during crystallization can be assessed.

Experimentally determined tie lines connect coexisting alkali feldspar and plagioclase phases which crystallized in equilibrium with liquid and vapor (Yoder et al, 1957). Theoretically, tie lines which connect coexisting, naturally occurring feldspars should, if equilibrium was maintained during crystallization and no subsequent alkali exchange has occurred, roughly parallel the experimentally determined tie lines if the systems are analogous (Yoder et al, 1975).

Coexisting alkali feldspar-plagioclase phases from four of the seven pegmatites were analyzed by the electron microprobe at Rensselaer Polytechnic Institute to estimate bulk chemical composition and to check for compositional zoning. The probe operated with an accelerating potential of 15 keV, an emission current of 14.5 nA, and an electron beam spot diameter of 15 microns. The feldspars were analyzed for the following six components: SiO_2, Al_2O_3, K_2O, Na_2O, CaO, and total Fe_2O_3.

Feldspars analyzed from P2, P4, P5, and P6 were selected for their cleanliness and accessibility under the microprobe. Roughly 20 points per .5 millimeter square area on each of the eight grains were analyzed. A point count of exsolved plagioclase blebs as shown on a photograph of an alkali feldspar from P6, magnified 400 times, was used in conjunction with the microprobe data of the perthite to better estimate its bulk composition. The bulk feldspar compositions obtained in these analyses (=integrated weight percent) are listed in Table 4.5.1.

The results of the analyses show that the feldspars are relatively homogeneous, and lack systematic trends in chemical composition throughout the grains, as might be the case if growth zoning was present.

4.5.2. BIOTITE SEPARATES.

It is apparent that the composition of the aqueous phase coexisting with a crystallizing granite has an effect on the composition and solidus temperature of the minimum melt. Thus, in order to estimate the composition of the aqueous phase in contact with the pegmatites during crystallization, approximately 2 grams of biotite were separated from each of three pegmatites, P1, P5, and P7, and analyzed for fluorine and chlorine. The wet chemical analyses were performed by X-Ray Assay Laboratories of Don Mills, Ontario. The results are listed in Table 4.5.2. The pegmatites apparently crystallized in the presence of a vapor phase which contained fluorine and chlorine.

	P2		P5		P4		P6	
	Plag.	Kspar.	Plag.	Kspar.	Plag.	Kspar.	Plag.	Kspar.
ALBITE	94.3	6.0	75.0	5.5	70.2	12.8	78.9	18.1
ORTHOCLASE	3.9	93.9	9.3	93.5	2.4	85.6	5.0	79.8
ANORTHITE	1.8	0.1	15.7	1.0	27.4	1.6	16.1	2.1

FIGURE 4.5.1. Integrated bulk chemical compositions of coexisting feldspars. In terms of Ab, Or, and An.

	P1	P5	P7
Fluorine	2400 ppm	4200 ppm	2800 ppm
Chlorine	11500 ppm	4500 ppm	750 ppm

Table 4.5.2. Results of analyses of biotites separated from pegmatites P1, P5, and P7.

CHAPTER 5. DISCUSSION OF RESULTS.

5.1. BULK CHEMICAL COMPOSITIONS.

The results of the bulk chemical analyses of the seven pegmatites are fairly consistent, with the exception of P4, which has a lower silica content and a higher alumina content in comparison to the other samples. In general, the pegmatites studied here are depleted in sodium and silica, and enriched in potassium, relative to the experimental data on the granite system (see Winkler, 1979), as well as to the inferred minimum melt composition (White and Chappell, 1977) (see Table 5.1.1).

The mineralogy and bulk compositions of most of the pegmatites correspond to S-type granites (Chappell and White, 1974). Hornblende is absent, but biotite, muscovite, and garnet may be present. All are strongly peraluminous, with Al/(Na + K + (Ca/2)) greater than 1.1, and low Na/K ratios. None of the pegmatites have normative diopside, and they contain normative corundum. P4 and P5, both of which intrude a gneiss which is locally amphibolitic, are exceptions. Although peraluminous, the Na_2O to K_2O ratios of P4 and P5 are 0.55 and 1.06, respectively, roughly 2 times the Na_2O to K_2O ratios of the other 5 pegmatites. It should be noted that the CaO contents of P4 and P5 are higher (by roughly 2 times) than the other pegmatites studied. Approximately one-third of the mafic component of P5 is hornblende, in contrast to the other pegmatites. The characteristics of the S-type granites, relative to the I-type granites, are due to the removal of

Table 5.1. Bulk chemical analyses of pegmatites P1 through P7. "M" is bulk chemical composition of the "minimum melt" of White and Chappell, 1977. Elements Sr, Ba, Y, Zr, and Rb are listed in parts per million.

	P1	P2	P3	P4	P5	P6	P7	M
SiO_2	71.43	69.78	73.50	62.03	70.70	70.72	70.04	75.07
TiO_2	0.18	0.13	0.05	0.05	0.06	0.04	0.36	0.28
Al_2O_3	14.43	15.20	13.94	20.43	16.05	15.49	14.13	12.86
Fe_2O_3	1.65	1.68	0.31	0.88	1.38	1.63	2.68	1.45
MnO	0.02	0.02	0.01	0.02	0.02	0.06	0.03	0.03
MgO	0.42	0.62	0.05	0.62	0.46	0.21	0.99	0.36
CaO	1.01	0.59	0.14	3.04	2.15	1.03	0.08	1.11
Na_2O	2.31	2.03	2.23	4.19	4.36	3.35	1.58	3.08
K_2O	7.83	9.24	8.73	7.61	4.13	6.59	8.60	4.81
P_2O_5	0.02	0.08	0.03	0.01	0.05	0.03	0.04	0.06
Sr	186.26	145.47	190.05	580.98	287.81	281.83	192.43	
Ba	686.09	620.99	976.54	1122.29	334.95	752.03	1500.00	
Y	6.82	42.69	2.42	6.22	8.93	41.36	12.23	
Zr	38.46	48.23	7.24	20.00	17.04	237.47	244.43	
Rb	335.21	415.18	378.50	167.71	116.02	203.75	278.52	
Total	99.30	99.37	98.94	98.88	99.41	99.15	98.53	99.11

sodium into sea water and calcium into carbonates, and the subsequent relative enrichment of the sediments in aluminum, during sedimentary fractionation.

The normative amounts of quartz, orthoclase, albite, and anorthite (listed in Table 5.1.2), recalculated to 100%, are shown on a ternary diagram which represents the granite system at 7 kb of PH2O (see Figure 5.1.1.2). This diagram projects the bulk compositions of the pegmatites, P1 through P7, onto the Qz-Ab-Or plane from the An apex of the tetrahedron, using the method of parallel projection which considers the effect of the fourth component (see Hoffman, 1974). Also plotted on this diagram are bulk compositions of 34 granitic pegmatites reported in Washington's (1917) tables.

The compositions of all but one of the 7 pegmatites sampled plot in the alkali feldspar field of the liquidus on the ternary diagram, relative to the cotectic line at 7 kb PH2O. Estimates of crustal pressure in this area of the Adirondacks are as low as 6.5 kilobars (Bohlen et al, 1985; Bohlen, 1982; Putman and Sullivan, 1979; Whitney and McLelland, 1973). Although there is no evidence that these pegmatites formed at peak metamorphic conditions, their position on the ternary diagram deviates even more so from the experimentally determined cotectic curves for pressures lower than 7 kilobars.

5.2. DEVIANCE FROM THE COTECTIC CURVE.

The generally high K2O to Na2O ratio and lower silica content of the pegmatite samples relative to experimental data

Table 5.2. CIPW norms of pegmatites P1 through P7.

	P1	P2	P3	P4	P5	P6	P7
Quartz	24.00	19.80	30.60	0.60	24.00	23.40	23.40
Orthoclase	44.48	55.60	50,04	44.48	22.24	38.92	50.04
Albite	20.96	15.72	15.72	36.68	36.68	26.20	15.72
Anorthite	5.56	2.78	-	13.90	11.12	5.56	-
Corundum	-	1.02	2.04	-	1.02	1.02	2.04
Enstatite	1.00	1.00	-	1.00	1.00	1.00	1.00
Ferrosilite	2.64	2.63	-	1.32	2.64	2.62	5.28

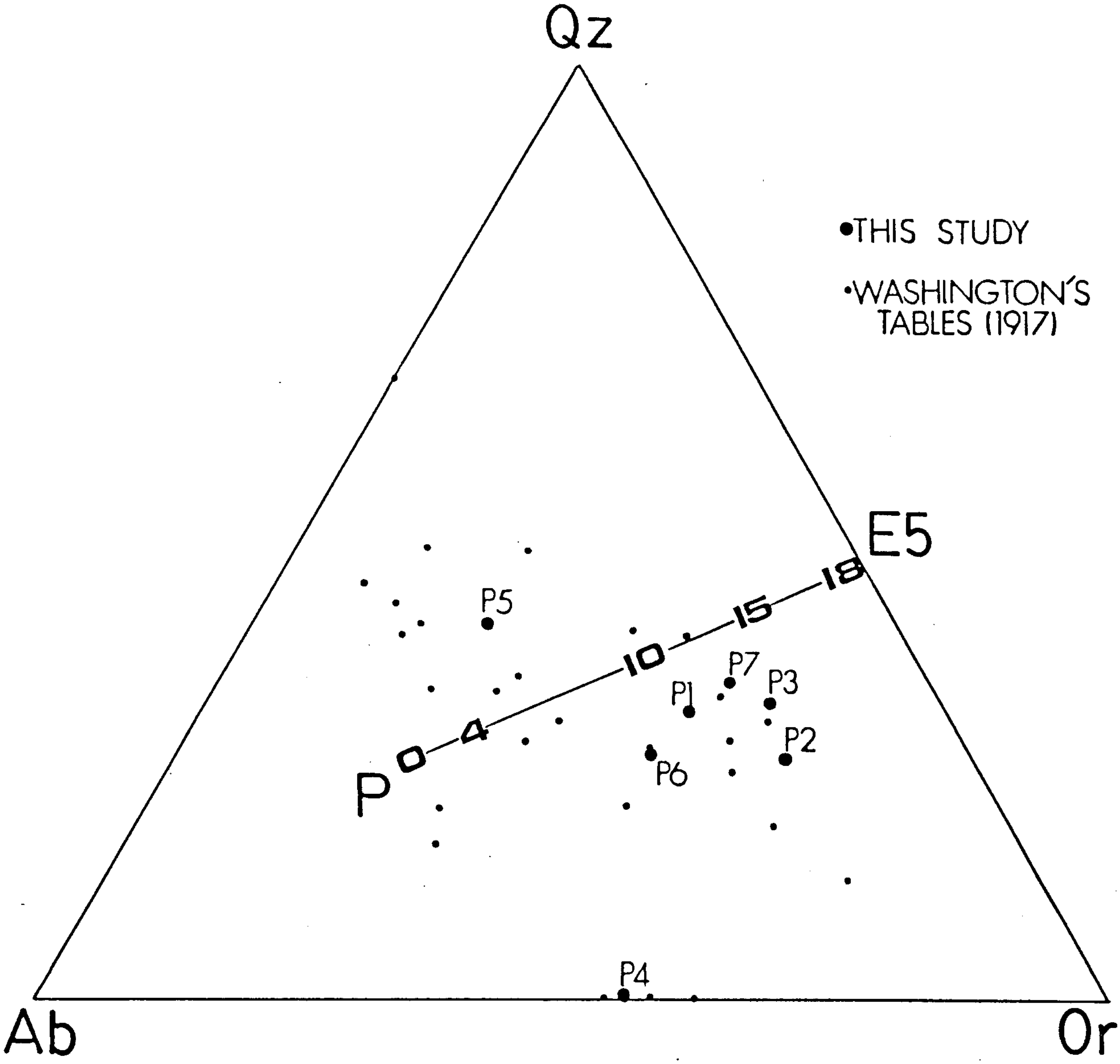


Figure 5.1.1.2. The bulk chemical compositions of pegmatites P1 through P7, plotted onto the Qz-Ab-Or plane of the granite system tetrahedron. Also plotted are pegmatites, granitic in composition, reported in Washington's tables (1917).

indicates that circumstances during initial melting, emplacement, or crystallization must have affected the phase equilibria of the pegmatite system, and caused the bulk chemical compositions to deviate from their position projected from the experimentally determined cotectic line for 7 kb PH2O (see Figure 5.1.1.2). Note that the bulk compositions of the granitic pegmatites from Washington's tables also do not form any coherent cotectic line trend.

In review, there are 5 reasons why points which represent water-saturated minimum melts may deviate from the curve (Winkler, 1979), assuming that the pressure estimate is correct: 1) The source did not originally contain all three components quartz, plagioclase, and alkali feldspar as solid mineral phases, or the early stage of anatexis has progressed to the point of the loss of one of the solid phase, so that the melt no longer coexists with the 3 solid phases quartz, plagioclase, and alkali feldspar; 2) One or more of the solid phase components in the granite system is represented by another phase in the source rocks, such that it is prohibited from entering the melt at minimum melting temperatures (e.g. the Or molecule in biotite or muscovite; 3) The melts did not crystallize under equilibrium conditions, i.e. a phase was removed or fractionated from the system as it crystallized; 4) Volatiles other than H2O were present in the fluid phase; 5) The original pegmatite composition was altered due to deformation, metamorphism, or metasomatism.

5.2.1. AVAILABLE CHEMICAL COMPONENTS AND TEMPERATURE.

The high ratio of alkali feldspar to plagioclase suggests that, possibly, an adequate amount of potassium but not sodium was present in the source rock during melting, and as the sodium-bearing phases were consumed, excess potassium was incorporated in place of sodium and calcium as melting proceeded. However, since plagioclase is present as a major component in the host rock gneisses, an adequate amount of sodium, as well as calcium, seems to have been available.

If temperatures were raised above minimum melting, minerals such as biotite or hornblende in the source would enter the melt phase and contribute to the Or or An content, respectively, in the melt, and possibly cause it to depart from the minimum melt curve, if potassium-rich feldspar or plagioclase, respectively, were absent in the host rocks (Winkler, 1979).

There is an absence of extensive, more diverse igneous activity, which would be expected if temperatures were higher. A qualitative microprobe analysis of coexisting feldspars in several of the pegmatites suggests that temperatures of crystallization remained low. The Ab component of the coexisting feldspar pairs, plotted on the diagrams of Stormer (1975) and Bohlen and Essene (1977), can be used as a geothermometer for the pegmatites (see Figure 5.2.1.1). Temperature estimates of 450 to 600o Centigrade are determined, but an error estimate of plus or minus 50o

Centigrade is appropriate, due to the difficulty of reintegrating perthites (Bohlen and Essene, 1977). For example, if some of the albite component has completely exsolved from the alkali feldspar, and was not accounted for during reintegration, then these temperature estimates are somewhat low. Furthermore, these diagrams do not account for the anorthite component, which could increase the temperature estimates by roughly 50 to 100o Centigrade (Green and Usdansky, 1986). These results are consistent with the 650-700o Centigrade estimates made by others (Putman and Sullivan, 1979; Bohlen, 1982).

The trace element data from the chemical analyses of the pegmatites (see Table 5.1.1) can also be used as a geothermometer. Recent experiments on the kinetics of apatite and zircon dissolution in felsic melts show the dependence of the solubilities of these phases on absolute temperature, as well as melt composition (Harrison and Watson, 1984; Harrison and Watson, 1983). Based on the data of these experiments, the zirconium concentrations of pegmatites P1 through P5 indicate temperatures of 650-700o Centrigrade, while those of pegmatites P6 and P7 indicate temperatures of approximately 850o Centigrade. The P2O5 contents of the pegmatites indicate temperatures of 644-720o Centigrade. Although temperatures of 650-750o Centigrade are characteristic of minimum melts (Winkler, 1979), the temperature estimates of 850o Centigrade show that minimum melting conditions were, at least in the cases of pegmatites P6 and P7, exceeded in the southeastern

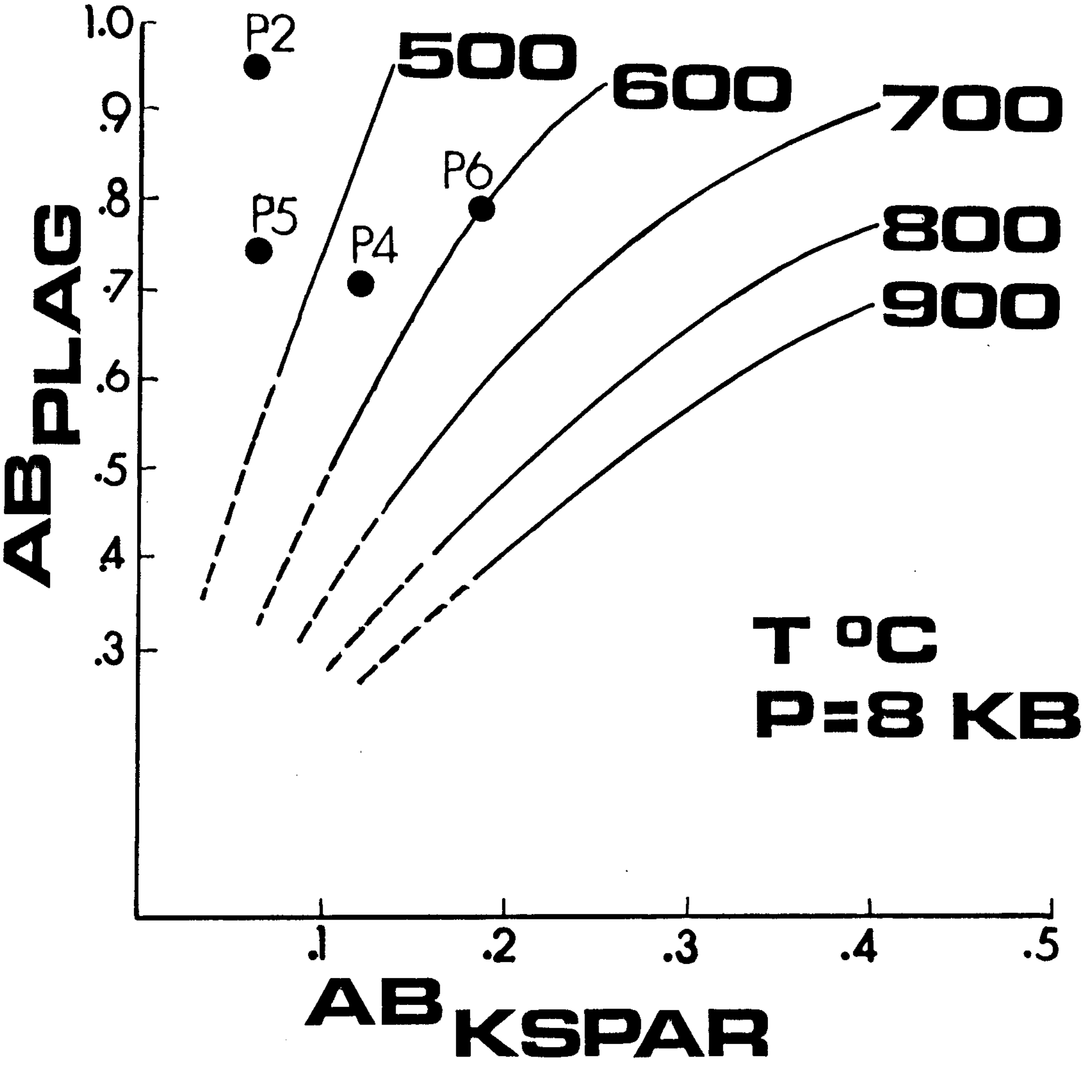


Figure 5.2.1.1. **Distribution of albite in coexisting feldspars as a function of P and T, after Stormer, 1975, and Bohlen, 1977. Data from southeastern Adirondacks are plotted.**

Adirondacks. This high temperature is a probable cause for the deviation from the cotectic curve on the ternary diagram.

5.2.2. REMOVAL OF MAFIC MATERIAL.

The amount of mafic material in the pegmatites, estimated from a point-count of slabs from each pegmatite sampled, ranges from 0 to roughly 18%, with an average of 6%. Most, of the mafic material is biotite (90% or more), except for P5, in which biotite comprises about one third of the total mafic content. The remainder of the mafic minerals of the pegmatites consists of ferromagnesian oxides, hornblende, and garnet.

Since the relative percentage of alkalis, calcium, and silica in the pegmatites is a determining factor for one of the intentions of this investigation, the alkali, calcium, and silica content in biotite, hornblende, and garnet, as well as the mafic components is accounted for on the Q-Ab-Or-An projections. The chemical compositions of the mafic minerals are based on representative mineral analyses of Deer et al (1962).

Mafics generally separate relatively early from granitic melts, or at least have an early crystallization paragenesis (Naney and Swanson, 1980); thus, by subtracting them from the bulk chemistry, the minimum melt is better represented in respect to experimental (mafic-free) systems. The effect of this subtraction is minor, with the exception of P1 and P7, which contain 10% and 18% mafic material, respectively, and the deviation from the cotectic curve is still observable.

5.2.3. EQUILIBRIUM DURING CRYSTALLIZATION.

Microstructural features which suggest equilibrium are 1) common grain contacts between minerals, especially smooth, clean contacts; 2) the lack of evidence of one mineral replacing another; 3) the presence of stable shapes of grains and inclusions; and 4) the presence of a relatively small number of minerals, so that the assemblage obeys the Phase Rule, and 5) the lack of mineral zoning (Vernon, 1976).

The mineral paragenesis of the pegmatites and the host rocks, compared at various distances from the pegmatite bodies, seem to support the above criteria. Only the relative abundances of mineral phases change. For example, the amount of biotite plus opaque minerals in the host rocks seems to increase toward the pegmatites. The ratio of biotite to opaques, however, also increases toward the pegmatites, indicating reduction of magnetite to form biotite, for instance, in an area saturated with water and K-feldspar (Vernon, 1976). This reaction might seem to defy criterion 2) from above as a case of crystal fractionation. However, this criterion is unreliable alone, because a partial replacement could indicate that one of the reactants (H2O, for example) became exhausted at the contact during crystallization of the pegmatite (Vernon, 1976).

According to Vernon (1976), the best criteria for chemical equilibrium are that a) all phases of an assemblage are in contact with each other; b) there exists a similar composition for edges of different grains of the same phase;

c) there is a similar distribution of components between a pair of phases in different parts of the rock volume; d) consistent temperatures are obtained from a study of oxygen isotope ratios between different minerals in different parts of the rock volume; and e) there are stable shapes of grains and inclusions of the minerals concerned.

Since petrography was the only source of analysis of the host gneisses, some of these criteria cannot be applied. However, optical properties show that the plagioclase (and all other minerals) are unzoned, so that b) and c) are probably fulfilled. Furthermore, the composition of the plagioclase (determined via the Michel-Levy technique) remains constant in the pegmatites and the host gneisses, at all distances from the pegmatites, which supports criterion c) from above.

Figure 5.2.2.1. is a plot of experimentally determined tie lines connecting coexisting feldspars in equilibrium with melt and vapor at a pressure of 5 kb (Yoder et al, 1957). This diagram can be used as a reference for feldspar phase equilibrium under conditions estimated for the pegmatites from the southeastern Adirondacks, where somewhat higher pressures have been estimated (Bohlen and Essene, 1977), if it is understood that, with higher pressures and lower temperatures, the tie lines make a larger angle to the Or-An side of the feldspar system triangle (Yoder et al, 1957). As shown by Figure 5.2.2.1., the pegmatite feldspars analyzed by microprobe are connected by lines which roughly parallel the

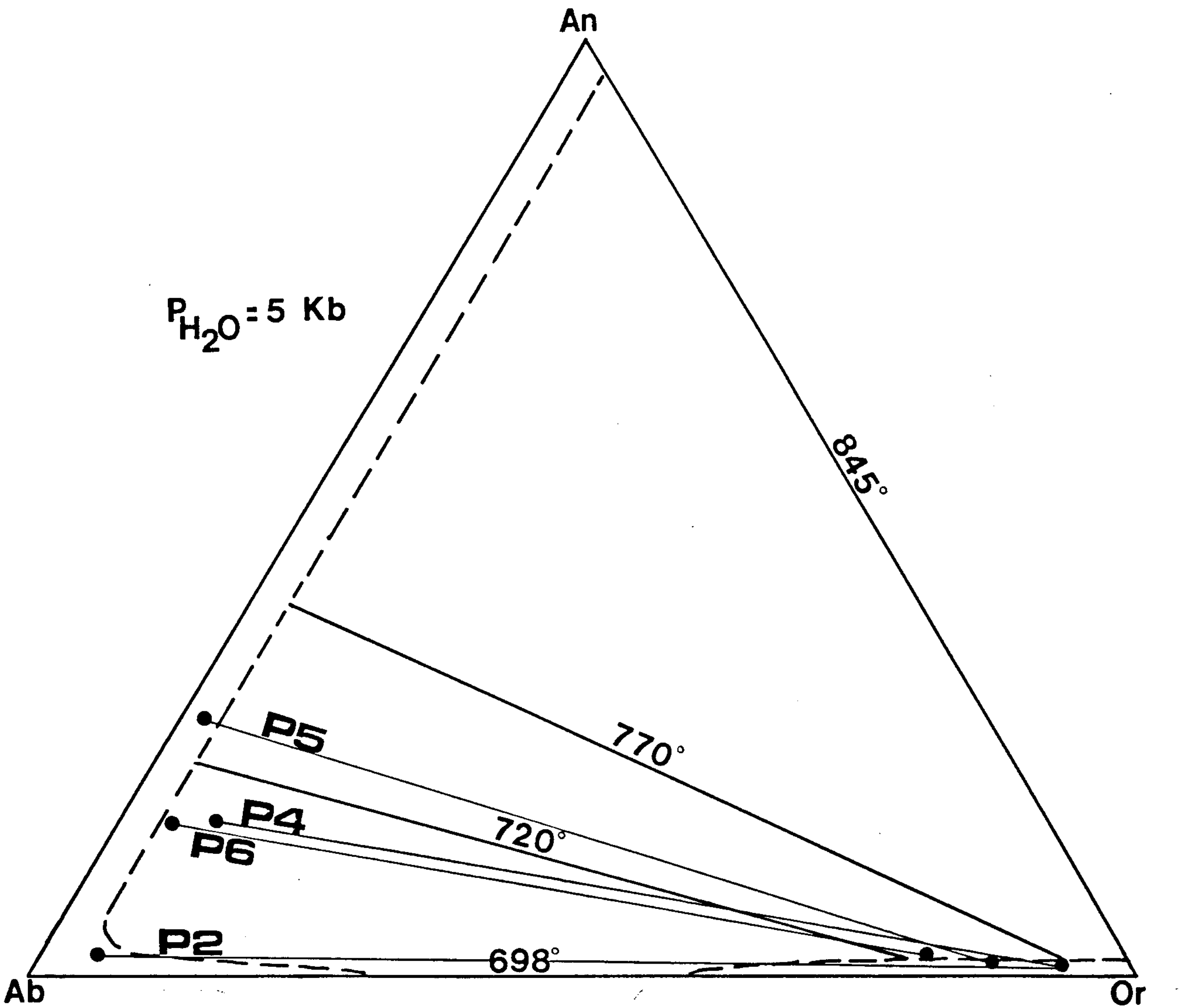


Figure 5.2.2.1. Experimentally determined tie lines connecting coexisting feldspars in equilibium with liquid and vapor, afterYoder et al, 1958,T°C. Data from pegmatites plotted.

experimentally determined tie lines. Thus, the pegmatite feldspars appear consistent with a bulk equilibrium crystallization of these bodies from a melt.

Finally, there is no textural evidence to suggest crystal-liquid separation or interrupted crystallization (e.g. phenocrysts are absent; the pegmatites are equigranular).

5.2.4. COMPOSITION OF THE VAPOR PHASE.

Experimental evidence suggests that the composition of the vapor phase has an affect on the composition and temperature of the coexisting minimum melt of the granite system (refer to Chapter 2.4 of this thesis). The mineral paragenesis of the pegmatites suggests that fluorine and chlorine may be important second volatile components. Thus, biotite separates from P1, P5, and P7 were analyzed for fluorine and chlorine content. The results are listed in Table 4.5.2., and shown graphically in Figure 5.2.3.1.

If fluorine is present in the system, the position of the quartz-alkali feldspar boundary moves away from the quartz apex. The composition of the minimum melt becomes enriched in Ab content, and the temperature of the minimum melt decreases progressively, with an increase in fluorine (see Figure 2.4.1). The addition of chlorine causes the minimum melt to become enriched in Or content and depleted in Qz content (Winkler, 1979). The temperature of the minimum melt decreases slightly with the addition of chlorine to the fluid phase.

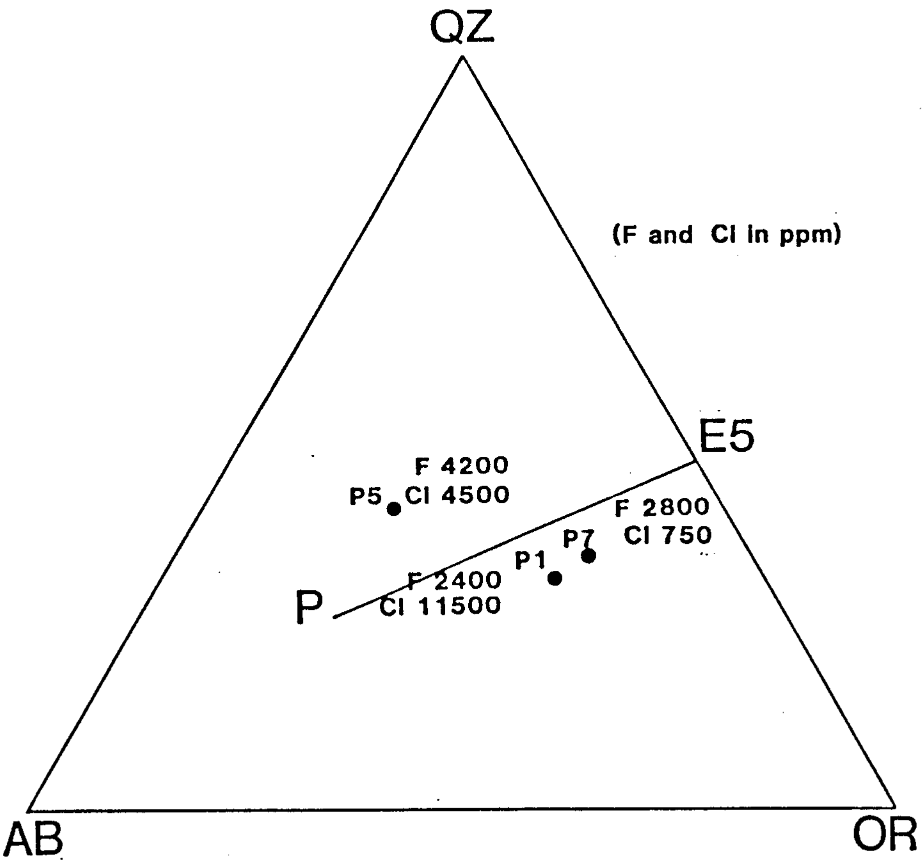


Figure 5.2.3.1. The data from biotite separates from pegmatites P1, P5, and P7, plotted with bulk chemical compositions on the Qz-Ab-Or plane of the granite system tetrahedron.

According to the experimental data of Wyllie and Tuttle (1964) and Burnham (1967), the effects of CO2 in the vapor phase are opposite to those of the fluorine and chlorine. The presence of CO2 will 1) raise the solidus and liquidus temperatures, especially at high pressures, and 2) depress the aqueous solubilites of the alkalies, as well as aluminum and silica, but to a lesser extent. This decreases the chance for alkali metasomatism.

Notice, however, that the fluorine content of biotite from P5 (4200 ppm) is higher than that of P1 (2400 ppm) and P7 (2800 ppm) (see Figure 5.2.3.1). This, at first, seems to be in contrast to the effect of the addition of fluorine, which is to deplete the melt in silica content. However, the tendency for fluorine to be partitioned into crystalline or liquid phases, relative to chlorine and CO2, which have higher affinites than fluorine for the vapor phase (Burnham, 1967; Fuge, 1977), indicates that the effects of chlorine and CO2 must be stronger in this case than that of fluorine, relative to vapor phase metasomatism. Furthermore, the fluorine content in biotite is not necessarily proportional to that of the melt. There is considerably less biotite in P5 than in P1 or P7, and thus, fluorine probably was not as effective in changing the minimum melt composition of P5 as it was in changing the compositions of P1 and P7. The relatively high sodium content of P5 is consistent with the idea that minimum melts are somewhat enriched in sodium with increasing amounts of fluorine in the aqueous system (Manning, 1981).

Therefore, a likely cause for the departure of the bulk compositions from the cotectic line, in addition to temperatures above minimum melting in the cases of P6 and P7, is the presence of other volatiles, such as fluorine, chlorine, or CO2 during initial melting.

Recent experimental data suggests that the composition of any coexisting vapor phase may also have affected the chemical composition of the pegmatites at a later stage of formation, such as during emplacement or crystallization (Lagache and Weisbrod, 1977; Dujon and Lagache, 1986). At magmatic temperatures and pressures, it is shown that, if a chlorine-rich aqueous phase present in the granite system is a homogeneous fluid, there is no alkali fractionation. However, in the event of a heterogeneous fluid, consisting of a melt phase and a vapor phase, sodium will enter the vapor phase, while potassium remains in the melt phase. The pegmatites may represent the liquid melt phase, while a vapor phase would be generated from the system as the melt crystallized and caused "second boiling", driving off a sodium, volatile-rich vapor, which would then infiltrate the host rock. This may account for the mineralogic/chemical variation in the Na2O to K2O ratios of the host gneisses observed within roughly 5 meters from the pegmatite-gneiss contacts, if this is condidered to be the result of infiltration during crystallization. However, petrographically, the alkali feldspar in the host gneisses do not show signs of replacement by plagioclase. The concentration of hydrous mafic phases in the host gneisses at the contacts suggest that infiltration may well have been effective within <1 meter.

Burnham (1967) attributes alkali fractionation to a temperature gradient, whereby potassium would be transferred to a low temperature region, and sodium in the opposite direction. The studies of Orville (1963) support this conclusion. Based on these ideas, the fractionation would have occurred during the emplacement stage, with sodium migrating toward the "hot" intruding pegmatite melt, and exchanging with potassium. However, there is no direct evidence of a temperature gradient during emplacement of the pegmatites, such as dike chilled margins (Hyndman, 1972).

This variation in Na2O to K2O ratios beginning roughly 5 meters from the contacts, is therefore considered a result of hydrothermal interaction between the host gneiss and the intruding magma. It is possible that, due to the presence of chlorine in the vapor phase associated with the intruding pegmatite magma, potassium entered the melt phase, while sodium partitioned into the vapor phase, a process determined from the experiments of Dujon and Lagache (1986) and Lagache and Weisbrod (1977). Beyond 5 meters, the host gneisses appear unaffected by the intruding magma.

5.2.5. ALTERATION OF THE MINIMUM MELT COMPOSITION.

Based on field evidence and petrography, the pegmatites analyzed for this investigation are relatively undeformed and unmetamorphosed. It is thus assumed that the minimum melt composition is preserved, except as it may have been changed by second boiling and vapor fractionation effects.

According to the granitization theory (Mehnert, 1968), metasomatism is caused by the introduction of an externally derived potassium-rich liquid. In the southeastern Adirondacks, an external source of potassium does not seem to exist, with the exception of the original anatectic melt itself. Due to the composition of the vapor fraction during melting, the melt was enriched in potassium, and depleted in sodium and silica, relative to the experimentally determined minimum melt composition of the pure H2O-granite system. There is no need to invoke a mechanism of metasomatism to account for their chemical composition.

CHAPTER 6. ORIGIN AND EMPLACEMENT OF PEGMATITES OF THE SOUTHEASTERN ADIRONDACKS.

Granitic pegmatites have been described in relation to larger granitic bodies, where they are commonly massive and zoned (Orville, 1960; Hyndman, 1972), or as constituents of regionally developed, medium to high-grade metamorphic terranes, where they are usually tabular and discordant (Turner and Verhoogen, 1960). Their complexity and variability prevent models designed for a particular pegmatite or zone of pegmatites from being used to define the origin and history of all pegmatites (King, 1948). A model is presented here to explain the geologic features associated with pegmatites in the southeastern Adirondacks in relation to their origin and history.

This investigation has defined several characteristics of the pegmatites in the southeastern Adirondacks which must be considered when designing a model. Their typical structure, relation to host rock, mineralogy, and geochemistry are summarized here, followed by a model of pegmatite genesis.

Although features typically associated with the injection of hotter magma into colder country rock, such as chilled margins and contact metamorphic aureoles (Hyndman, 1972), are not observed in relation to these pegmatites, their tabular form, the presence of xenoliths, but lack of phenocrysts, and sharp discordant contacts indicate they have been injected.

It is not surprising that evidence for a discernible difference in temperature between pegmatite and host rock is absent in the southeastern Adirondacks. Four features support the idea that the pegmatites were generated at or near their site of emplacement, and thus, a regional variation in temperature would not be expected. The pegmatites show a complete gradation between deformed, locally deformed, and completely undeformed, which indicates that their generation was coincident with the late stage of a regional tectonic event, assuming that there was only one generation of pegmatite. A similar pegmatite intruding Grenville age rocks in the western Adirondacks was dated as 860 Ma by the potassium-argon isotopic dating method (Long and Kulp, 1962). The pegmatites in the southeastern Adirondacks occur in a region of medium to high grade metamorphism, where migmatization is locally intense. Physical conditions for anatectic saturated, or near saturated, melts surely prevail. Finally, the compositional effects of their interaction with the host gneisses during emplacement are observed within 5 meters from the pegmatite-gneiss contacts.

However, the interaction between the intuding pegmatites and their host gneisses within 5 meters of the pegmatite-gneiss contacts indicates hydrothermal activity, which is characteristic of low pressure intrusions (De Almeida and Johnston, 1944; Quirke and Kremers, 1943).

The sharp contacts mask a progressive but well-defined variation in mineralogy away from the contact on both sides of the intrusions. Sharp contacts are marked by mafic absent pegmatite, juxtaposed adjacent to thin contact zones, not more than 5 centimeters thick, with high concentrations of mafic material relative to the remainder of the wall rock. Thin sections show that these mafic-enriched zones are composed almost entirely of biotite and/or hornblende and plagioclase. In the wall rock, the amount of mafic material decreases sharply away from these zones, and then steadily until it comprises usually less than 10% of the rock 5 to 20 meters away from the intrusive contact. Modal plagioclase exhibits a consistent, steady decrease over the same distance.

Biotite in these contact zones ranges from randomly oriented to oriented in two directions: a major foliation parallel to the regional foliation, and a minor foliation which is roughly parallel to the orientation of the intrusion.

Bulk chemical analyses indicate the following geochemical characteristics must be explained by any model for the formation and emplacement of the pegmatites in the southeastern Adirondacks: 1) a generally high K2O to Na2O ratio compared to that of Winkler (1979) as well as the wall rock gneisses, 2) a minor depletion in SiO2, and 3) temperatures of crystallization ranging from 644-850o Centigrade.

Analyses of fluorine and chlorine in separates of biotite from the pegmatites, indicate that the volatiles in the pegmatite system may have contained a considerable amount of components other than H2O and, thus, that the melting temperature and melt composition may have been affected.

Processes called upon to explain the origin of pegmatites include anatexis, granitization, magmatism, and crystallization of solutions moving through an open system (King, 1948; Flawn, 1950; Jahns, 1953; Orville, 1960; Jahns and Burnham, 1969), which were discussed in Chapter 2 of this thesis.

The first concept of granitization contends that the volume occupied by the pegmatite was formerly occupied by country rock, which has been metasomatized to the point of homogeneity. The injection features, sharp contacts and reoriented foliations, of the pegmatites in the southeastern Adirondacks cannot be reconciled with this concept, which is therefore concluded not to be important to their formation. The second concept of granitization involves a low temperature, volatile-rich liquid formed in the advance of a wave of granitization. The extent of metamorphism and the granitioid character of some rock formations in the Adirondacks may be consistent with the inital stages of the granitization theory, but this idea does not agree with a late emplacement of the pegmatite, or their very limited scale.

The open system theory is associated with low pressure, hydrothermal systems (De Ameida and Johnston, 1944; Quirke and Kremers, 1943). There appears to have been an interaction between the intruding pegmatites and their host gneisses within roughly 5 meters from the contacts, which suggests hydrothermal activity and possibly low pressure. However, the association of the pegmatites with extensive migmatization and mylonite zones in the southeastern Adirondacks (Berry, 1961) and the lack of chilled margins (Hyndman, 1972, p. 141) indicate that they formed under the influence of a significant amount of pressure (albeit not necessarily under peak metamorphic pressure).

The major process controlling the formation of pegmatites in the southeastern Adirondacks is anatexis with limited volatiles, which is believed to explain simple quartz-feldspar pegmatites present in some migmatite terranes (Turner and Verhoogen, 1960). The magmatic process is also called upon to explain the injection features of the pegmatites in the southeastern Adirondacks.

The model includes the following steps: the initiation of melting, the accumulation of granitic liquid, injection, interaction between the magma and the host gneiss (i.e. contamination of the magma or further accumulation of magma), and the loss of a volatile phase subsequent to crystallization.

Experiments on quartzofeldspathic gneisses show that, in the presence of water, with a rise in temperature, melting begins along grain boundaries (Mehnert et al, 1973; Busch et al, 1974). The composition of the melt is granitic. The presence and release of an aqueous phase during regional metamorphism is a concept well-recognized from field observations and experimental studies (Fyfe and Turner, 1958; Norris and Henley, 1972; Walther and Orville, 1980), and thus it is probably present along grain boundaries and in other pore spaces, such as stress-induced fractures. Since this water is lost with progressive metamorphism, only hydroxyl water may remain (Walther and Orville, 1980).

The volatile phase or component apparently contained components other than H2O, such as HF, HCl, and P2O5, which tend to lower the melting temperature to aid in the formation of initial melts, and produce a minimum melt somewhat depleted in SiO2 and enriched in sodium or potassium, relative to the experimental data on the granite system (Manning, 1981; Wyllie and Tuttle, 1964).

In order to account for a source of volatile material in a region of limited fluids (see Valley and O'Neil, 1984; Newton et al, 1980), it is suggested that a less dehydrated felsic slice from a higher crustal level is lowered along high strain zones. At this deeper level, the felsic slice is subjected to higher temperatures, and partial melting is induced.

Recent experimental evidence (Watson, 1982; Jurewicz and Watson, 1984) suggests that, depending on free energy relationships of solid-solid and solid-liquid interfaces, silicate melt introduced into a crystalline medium will either 1) infiltrate the material along tubules connecting grain edges if the melt is mafic, or 2) collect as pools at grain edges and fail to segregate from the solid unless directed stress is induced on the system if the melt is felsic.

It is postulated that during anatexis granitic melt initially accumulated in unconnected pools at grain edges throughout appropriate solid host material (gneisses) of the pegmatites. The melt is enriched in volatiles and alkalies, since they tend to migrate toward regions of low stress (Hyndman, 1972).

The melts intruded due to a north-south compressional event, or in response to uplift following the Grenville orogeny.

The sodium-enriched region within 5 meters from the contact indicates hydrothermal interaction between the intruding pegmatite and the host gneiss.

Recent experiments on feldspars in contact with a chlorine-rich aqueous solution show that, in the presence of two immiscible fluids, one gas and one liquid, the distribution coefficient of sodium to potassium in the gas is greater than that coefficient in the liquid (Lagache and Weisbrod, 1977; Dujon and Lagache, 1986). Apparently, if a melt separates or generates a separate vapor phase, vapor/liquid partition relations will favor sodium into the vapor phase, while potassium remains in the liquid phase. These workers call upon this process as a possible mechanism for alkali metasomatism.

This mechanism of alkali fractionation accompanying crystallization may have caused the lowering of the Na2O to K2O ratio in the host gneisses roughly 5 meters beyond the contact, but there is the lack of evidence for relict potassium feldspar being replaced by sodium feldspar in the gneisses. Rather, the sodium-enriched region of the host gneiss may represent a restite, a volume of material unable to enter the intruding melt phase due to the affinity of sodium for chlorine, present in the vapor phase associated with the intruding magma. There is likewise a tendency for potassium to enter the melt phase (Dujon and Lagache, 1986; Lagache and Weisbrod, 1977).

This model suggests that the narrow (<1 meter) mafic mineral-enriched, silica-depleted wall rock, is the result of the infiltration of a volatile-rich vapor phase, which separated from the melt as it crystallized. This volatile-rich vapor phase reduced iron oxide phases and consumed a potassium-rich alkali feldspar phase (and possibly deposited a mafic component from the melt) to form the plagioclase, biotite, hornblende-rich zone within the host.

The model therefore consists of 1) silica- and sodium-depleted initial melting due to the presence of chlorine or fluorine in a local fluid, 2) stress-induced injection of anatectic fluid-saturated magma, 3) interaction between the intruding magma and the host gneiss (i.e. contamination of the magma by potassium from the host gneiss, or further accumulation) from up to roughly 5 meters away from the intrusion due to the presence of chlorine or fluorine, and 4) infiltration of the host gneiss by a vapor phase released from the crystallizing magma within one meter of the pegmatite-host gneiss contact.

This model assumes that the pegmatites are Grenville age, based on 1) the association of the pegmatites with Grenville migmatization and mylonitization (Wiener et al, 1984; Grant et al, 1984), 2) a K-Ar age of 860 Ma from a pegmatite in the western Adirondacks (Long and Kulp, 1962), and 3) the absence of chilled margins of the pegmatites, which is indicative of medium to high pressures (Hyndman, 1972).

An alternative interpretation of the origin of the pegmatites is that they intruded under lower pressures, possibly during uplift following the Grenville orogeny. This alternative is viable if 1) the generation of pegmatites in the southeastern Adirondacks is later than that in the western Adirondacks, or 2) the age reported above represents crystallization, rather than significant cooling. Further support for lower pressure intrusion is the evidence for hydrothermal interaction between the intruding magma and the host gneiss (De Almeida and Johnston, 1944; Quirke and Kremers, 1943). East-west lineations in the southeastern Adirondacks indicate that transport was predominantly east-west. This is in contrast to the north-south transport required by the first interpretation.

CHAPTER 7. GEOLOGIC HISTORY.

Several models have been developed to explain the tectonic history of the Adirondacks as a southern extension of the Grenville Province (Wynne-Edwards, 1964; Baer, 1972; Dewey and Burke, 1973; Baer, 1976; Mclelland and Isachsen 1980; Mclelland, 1986). The presence of an upper amphibolite to granulite grade of metamorphism in uplifted and eroded crust, in addition to the association of anorthosite and charnockite, have made it difficult to explain the tectonics of the Grenville Province in terms of a modern analogue, where direct knowledge of the lower crust is lacking.

It is generally believed that a continent-continent collision, similar to that now effecting central Asia and India, is the best explanation for the pressure and temperature conditions of metamorphism represented in the Grenville Province (Dewey and Burke, 1973). Recently, however, it was suggested that the collision of two continents is unlikely without first, an arc-continent collision (Burke and Kidd, 1980). Central Asia and the Indian subcontinent, for example, are separated by calc-alkaline granitic plutonic belts, which decrease in age to the south, and could represent remnant island arcs, which have accreted onto the Eurasian plate (Xu et al, 1985). Furthermore, the position of modern island arcs predicts that an arc-continent collision would usually precede a continent-continent collision.

Thus far, few models describing the tectonic evolution of the Grenville Province include an arc terrain (see Moore and Thompson, 1980; Crawford and Hoersch, 1984).

Radiometric ages obtained from the Adirondacks, using various methods of geochronology, range from roughly 860 to 1350 Ma (McLelland, 1986; Baer, 1981; Silver, 1969; Gaudette et al, 1969; Long and Kulp, 1962. It has been postulated that this represents more than one orogenic event during Grenville time (Baer, 1972), but because of the variety of methods used, few well-defined, separate, age-dependent events have been determined (see Grant, 1984).

The pegmatites of this study apparently represent a late magmatic stage, a final "pulse", of the Grenville orogeny. The pegmatite bodies range from deformed, to locally deformed (see Figure 4.1.5), to undeformed. The similarity and intimate association of the deformed and undeformed pegmatites, and the idea that the dry conditions of Adirondack metamorphism would not accommodate more than one episode of pegmatite formation, suggest only one period of magmatism. This is supported by the occurrence of pegmatites with similar structural characteristics and interpretation in other medium to high grade metamorphic terranes (Sandiford and Wilson, 1984; 1985).

Actually, the predominantly dry conditions represented by metamorphic rocks in the Adirondacks (Valley and O'Neill, 1984) imply that a structural mechanism must be called upon to explain the formation of the pegmatites in order to account for a source of fluid. A possible scenario is as follows:

As a final pulse of the Grenville orogeny, a less dehydrated slice (or slices) of felsic crust (possibly containing muscovite, for example) from somewhat shallower crustal levels than the rest of the already-tectonized and dehydrated Adirondacks, was thrust along a ductile high strain zone into a slightly deeper and hotter level. This shallower crustal slice provides a source of volatiles described by the model, and the high strain zones allow for the localized movement of fluid. Subsequent rise of the isotherms in the displaced slice then resulted in partial melting, dehydration, and locally, pegmatite formation.

Locally mylonitic high strain zones have been mapped in the southeastern Adirondacks, as well as in other parts of the Adirondacks, particularly where the pegmatites from this investigation were sampled (see Berry, 1961), and a relationship between high strain zones and pegmatites may exist.

This appears to be a natural analogue of the experimental work of Jurewicz and Watson (1984) on the distribution and infiltration of felsic melts. They suggest that felsic melts will not accumulate and infiltrate host rock, but rather, accumulate as "pools" at grain edges, unless they're subjected to an external stress, which would allow them to infiltrate. The local migmatitic character of the southeastern Adirondacks may represent relatively "non-accumulated" melt, while the pegmatites may represent relatively localized melt which has been able to accumulate and infiltrate (i.e. intrude) host rock. The ductile high strain zone then possibly marks the mechanism by which to lower a felsic slice to higher temperature levels, induce partial melting, and remove volatile material from the less-dehydrated felsic slice, so that pegmatite anatectic melts could form. Frictional heating along the thrust plane may have also aided anatexis. The highly volatile character of pegmatites may have initiated or aided their injection, once fluid pressure exceeded rock pressure, a mechanism suggested by Norris and Henley (1976), Knapp and Knight (1977), and Walther and Orville (1982).

The similar orientation of the pegmatites (N-S to NE-SW) indicates the presence of a tensional structural element. If, as Berry (1961) contends, a late stage thrust was in a northerly direction in the southeastern Adirondacks, then the orientation of these pegmatite dikes may represent an approximate east-west least-compressive stress direction. However, roughly east-west trending lineations are also observed in the southeastern Adirondacks, as are roughly east-west oriented pegmatite intrusions, similar to those which trend roughly north-south. These suggest an approximate north-south least-compressive stress direction. Therefore, two separate regional stress events may be recorded in the southeastern Adirondacks. This study does not attempt to relate, specifically, structure and magma generation in the southeastern Adirondacks, but suggests that the two may be related as they are in other areas (see Xu et al, 1985; Sandiford and Wilson, 1986, 1985).

CHAPTER 8. SUMMARY AND CONCLUSION.

It is believed that the generally high potassium to sodium ratios and relative silica depletions of the pegmatites in the southeastern Adirondacks are due to the composition of the fluid phase, dissolved and extracted within the parent granitic melt during the initial melting stage of pegmatite formation. This is supported by 1) the experimental data of Wyllie and Tuttle (1964), Chorlton and Martin (1978), Swanson (1979), and Winkler (1979), on the effect of addition of second volatiles to an aqueous phase in contact with granitic melts, and 2) the fact that analyses of biotite mineral separates from several of the pegmatites indicate that fluorine and chlorine were present during crystallization of the pegmatites. Although there are trace amounts of lepidolite in the pegmatites, its low abundance suggests that the presence of other volatiles or incompatible elements, such as lithium or boron, was not a significant factor during pegmatite formation, although biotite can accommodate an appreciable amount of lithium.

The pegmatite melts may have accumulated and intruded their host rocks in a condition where vapor pressure was near or equal to rock pressure, and it fractured the rock, but intrusion is at least partly the result of the stress of tectonic thrusting, due to the association of the pegmatites with high strain zones (Berry, 1961). Crystallization caused the melt to separate into a volatile-rich vapor phase and a liquid phase. The volatile phase then escaped from the system by, for example, infiltrating the country rock or moving along the walls of the dike within one meter of the contacts. The infiltration of a vapor phase resulted in the reduction of iron oxides and the consumption of potassium-rich alkali feldspar to form a biotite-rich zone in the host gneiss at the contact.

Although plagioclase predominates over potassium-rich alkali feldspar at the pegmatite-gneiss contacts, and decreases away from the contact, the estimated bulk chemistries of the host gneisses indicate that the potassium to sodium ratio is approximately constant up to within roughly 5 meters from the pegmatite. There is then a decrease in sodium and a subsequent increase in potassium in the host gneiss, away from the pegmatite, beyond roughly 5 meters from the pegmatite-gneiss contact. This is the result of hydrothermal interaction between the inruding magma and the host gneiss from up to roughly 5 meters from the pegmatite-gneiss contact, and the tendency for potassium to enter the melt phase due to the composition of the fluid present during initial melting and intrusion.

The potassium enrichment, silica-sodium depletion may be enhanced by alkali fractionation, which may be controlled by the composition of the vapor phase during crystallization (Dujon and Lagache, 1986; Lagache and Weisbrod, 1977). This would suggest that the decrease in the sodium to potassium ratio observed in the host gneisses beyond roughly 5 meters from the contact, may be the result of infiltration. The lack of evidence for the metasomatic replacement of alkali feldspar by plagioclase, however, indicates that this was not a major cause for the observed chemical variation within the host gneisses up to 5 meters from the contacts.

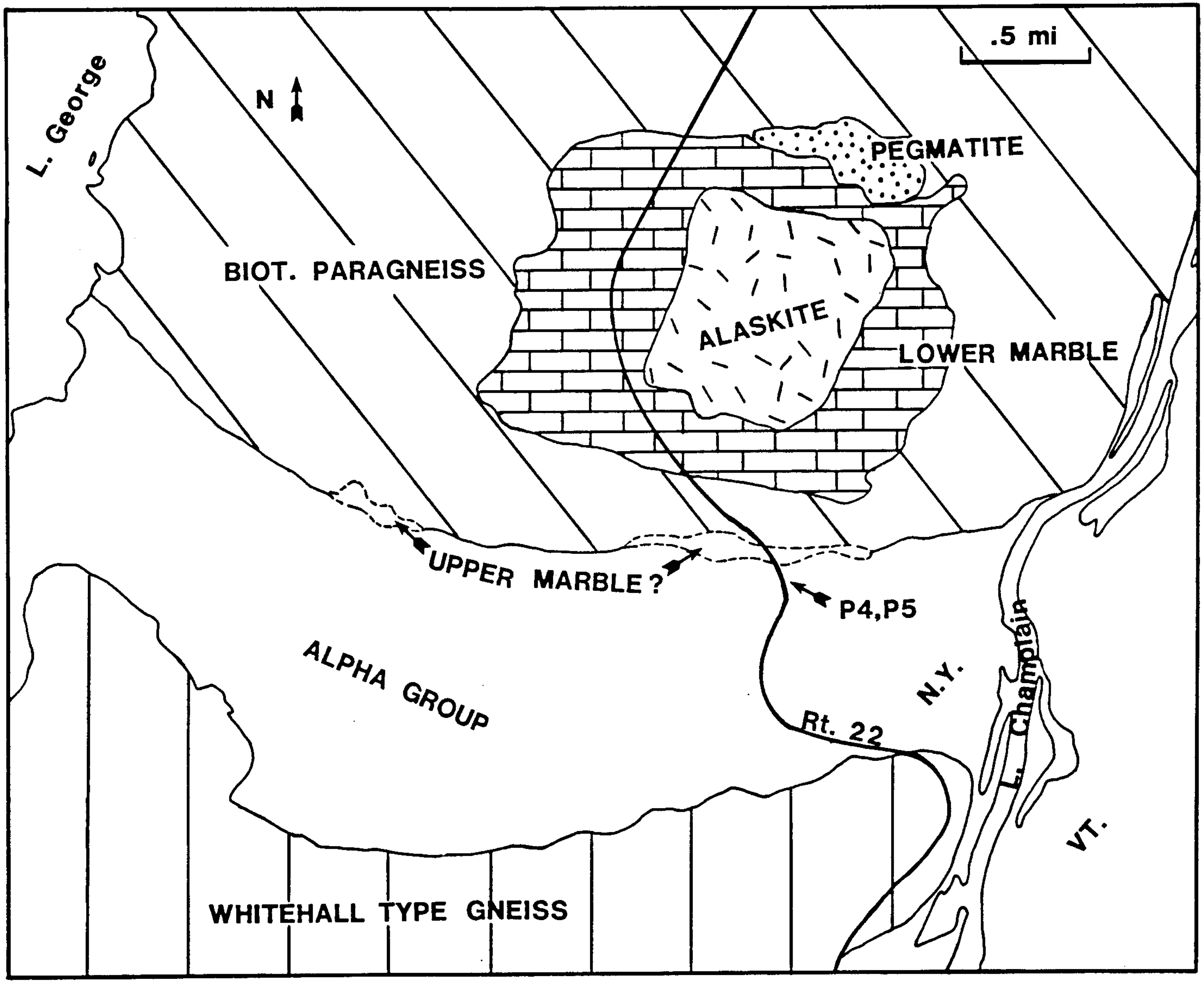
An exception to the potassic, silica-poor character of most of the pegmatites studied is P5, whose bulk composition plots in the Qz field of the Qz-Ab-Or ternary diagram at 7 kb of PH2O. Another exception is P4, which has less than 1% modal quartz. Both contain at least 2 times as much calcium as the other 5 pegmatites.

These two pegmatites intrude a locally amphibolitic gneiss of Berry's (1961) Alpha group, a sequence of leucogranitic gneisses and garnet-sillimanite gneisses, and minor quartzites, marbles, and graphitic schists. The position of pegmatite P5 within the metsedimentary package suggests that the possible presence of CO2 in the volatile phase may have affected its minimum melt composition, because the activity of H2O would have been lowered (Valley and O'Neil, 1984; Newton et al, 1980; Swanson, 1979; Wyllie, 1977). Based on the geologic maps of Berry (1961) and Fisher (1970), P4 and P5 intrude the Alpha group roughly 200 meters above its contact with the underlying Biotite Paragneiss. The Alpha group and the Biotite Paragneiss are locally separated by the Upper Marble in other areas of the southeastern Adirondacks (Berry, 1961). The Upper Marble unit may contain bands of amphibolite and quartzite (Berry, 1961). Field reconnaisance in this area indicates a low-lying, swampy region at the contact between the Alpha group and the underlying Biotite Paragneiss (see Figure 8.1). This low-lying region may represent the location of the since eroded Upper Marble unit, and suggests that a CO2-rich fluid may have been available during the generation of pegmatites P4 and P5. This may also account for their relatively high calcium contents.

It is conluded, therefore, that the addition of CO2 as a second volatile causes minimum melt compositions to plot on the Qz side of the quartz-alkali feldspar boundary relative to the pure H2O-granite system, while the addition of chlorine and fluorine causes minimum melt compositions to plot in the potassium-rich and sodium-rich feldspar fields, respectively, relative to the quartz-alkali feldspar boundary of the pure H2O-granite system.

The postulated anatectic origin of the pegmatites is supported by 1) the intimate association of pegmatites and migmatites in the southeastern Adirondacks, 2) the lack of associated, larger granitic sources, 3) the lack of chilled margins, 4) the general correspondence to possible minimum melts if fluorine and chlorine are taken into account, 5) the general correspondence of feldspar tie lines and temperature to experimental equilibria of crystallization from melt, and 6) the interaction between pegmatite and host gneisses roughly not more than 5 meters from the pegmatite-gneiss contacts.

Figure 8.1. Generalized geologic map of the Putnam Station, NY area, in the northern section of the Putnam 7.5 quadrangle. Note location of pegmatites P4 and P5, above the contact between Berry's (1961) Biotite Paragneiss and Alpha Group; dashed, swampy region (Berry's Upper Marble?); location of Alaskite after Fisher et al, 1971, geologic map of NY State.


.5 mi
L. George
N
PEGMATITE
BIOT. PARAGNEISS
ALASKITE
LOWER MARBLE
UPPER MARBLE ?
P4,P5
ALPHA GROUP
N.Y.
L. Champlain
Rt. 22
VT.
WHITEHALL TYPE GNEISS

Sharp contacts between pegmatite dikes and host gneisses, contact marginal zones, and distorted regional foliations at the contacts are evidence that some of the pegmatites were injected. Their association with high strain zones suggests that they represent accumulations of granitic melt which may have been allowed to intrude their hosts due to the induction of tectonic stress. These tectonically late melts may be related to the thrusting of an upper level, less dehydrated crustal slice, along a ductile high strain zone in the southeastern Adirondacks, as a final pulse of the Grenville orogeny in this area. This pulse would have induced partial melting and caused a limited late-stage magmatic event, represented by local deformed to undeformed granitic pegmatites and some migmatites.

Alternatively, evidence for the hydrothermal interaction between the intruding magma and the host gneiss may indicate low pressure intrusion, possibly caused by uplift following the Grenville orogeny. Although there is no evidence that the pegmatites formed under peak metamorphic conditions, the association of these pegmatites with migmatization and mylonites, and the lack of chilled margins, indicates that the pegmatites are at least mesozonal (Hyndman, 1972). Furthermore, only the position of P5 on the ternary diagram of Winkler (1979) suggests pressure lower than 7 kilobars, and this is probably due to a lower activity of H2O and a higher activity of CO2. This does not eliminate the possiblity of a low pressure intrusion; rather, the deviation of the compositions of the pegmatites from the cotectic curve of the postulated 7 kilobar pressure must be explained by means other than an incorrect pressure estimate.

It is concluded finally that, as indicators of pressure and fluid conditions during orogeny, these pegmatites provide qualitative, but not quantitative, information. Trace element data, however, indicate temperatures of 644-720o Centigrade for pegmatites P1 through P5, and temperatures of 850o Centigrade for pegmatites P6 and P7 (Harrison and Watson, 1984, 1983). This indicates temperatures well above minimum melting conditions, one of the requirements of the method used during this investigation. Due to the lack of experimental data on the granite system, comparable to the temperature and pressure estimates for this area of the Adirondacks, it is necessary to study pegmatites which apparently did not crystallize in the presence of volatile components other than H2O to obtain quantitative estimates of the temperature and pressure of crystallization.

8.1. FUTURE WORK.

This thesis accounts for the chemistry and mineralogy of the pegmatites, as well as for the chemistry and mineralogy of their host gneisses, specifically in reference to the effect of pegmatite formation.

A study of the precise relationship between these pegmatites and high strain zones in the southeastern Adirondacks would act as a test for 1) the suggested mechanism of emplacement, and 2) the postulated late-stage tectonic event of the Grenville orogeny in the southeastern Adirondacks.

"And will this whole damned world fall down Before we've learned to share what we've found" (China Crisis, 1985).

APPENDIX. Estimated bulk chemical composition of the host gneisses at various distances from the pegmatites. The numbers in the second column indicate distance, in meters, from the pegmatite from which thin sections were made.

		P1	P2	P3	P4	P5	P7
SiO2							
	1				61.70		
	0				56.50		
	0		54.46	64.66	53.19	62.45	60.40
	1			72.43	61.10		
	3					76.62	
	5	57.89	72.24	55.72			71.48
	10	63.58			71.42		
	20	61.01		72.48	76.23		
TiO2							
	1				0.80		
	0				1.03		
	0		2.66	0.70	0.40	0.91	1.00
	1					0.96	0.27
	3					1.27	
	5	2.71	0.55	1.51			0.25
	10	1.51			0.50		
	20	0.85		0.76	0.73		
Al2O3							
	1				17.96		
	0				15.47		
	0		13.48	15.64	15.91	18.53	14.93
	1					11.81	15.00
	3					9.40	
	5	11.96	11.72	15.24			14.84
	10	15.26			13.49		
	20	12.73		12.62	10.62		
Fe2O3							
	1				7.35		
	0				6.91		
	0		17.52	5.14	3.85	9.22	11.52
	1					5.64	2.85
	3					5.68	
	5	12.77	2.75	12.72			2.58
	10	4.69			4.18		
	20	16.28		6.5	3.64		
MgO							
	1				4.10		
	0				6.97		
	0		3.14	4.14	3.42	7.52	3.11
	1					1.61	0.84

	3					0.25	
	5	6.30	0.96	3.72			0.76
	10	2.35			1.63		
	20	2.79		0.18	0.46		
CaO							
	1				2.84		
	0				4.84		
	0		0.78	1.67	7.12	1.35	1.00
	1					1.18	1.29
	3					0.62	
	5	1.12	2.41	2.99			1.12
	10	1.85			1.74		
	20	2.52		1.51	0.99		
Na2O							
	1				5.10		
	0				4.69		
	0		2.55	5.13	5.35	4.26	3.03
	1					3.80	4.52
	3					2.54	
	5	3.39	3.12	4.47			4.16
	10	4.23			4.46		
	20	1.45		3.35	2.56		
K20							
	1				3.52		
	0				2.43		
	0		3.57	1.18	1.17	3.08	4.82
	1					1.78	3.62
	3					3.37	
	5	0.65	3.93	0.87			4.36
	10	4.04			1.95		
	20	2.61		2.82	4.95		
H2O							
	1				1.09		
	0				0.76		
	0		1.56	1.66	0.72	3.11	1.50
	1					0.71	0.51
	3					0.12	
	5	2.13	0.21	0.30			0.58
	10	0.87			0.59		
	20	0.13		0.21	0.11		

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