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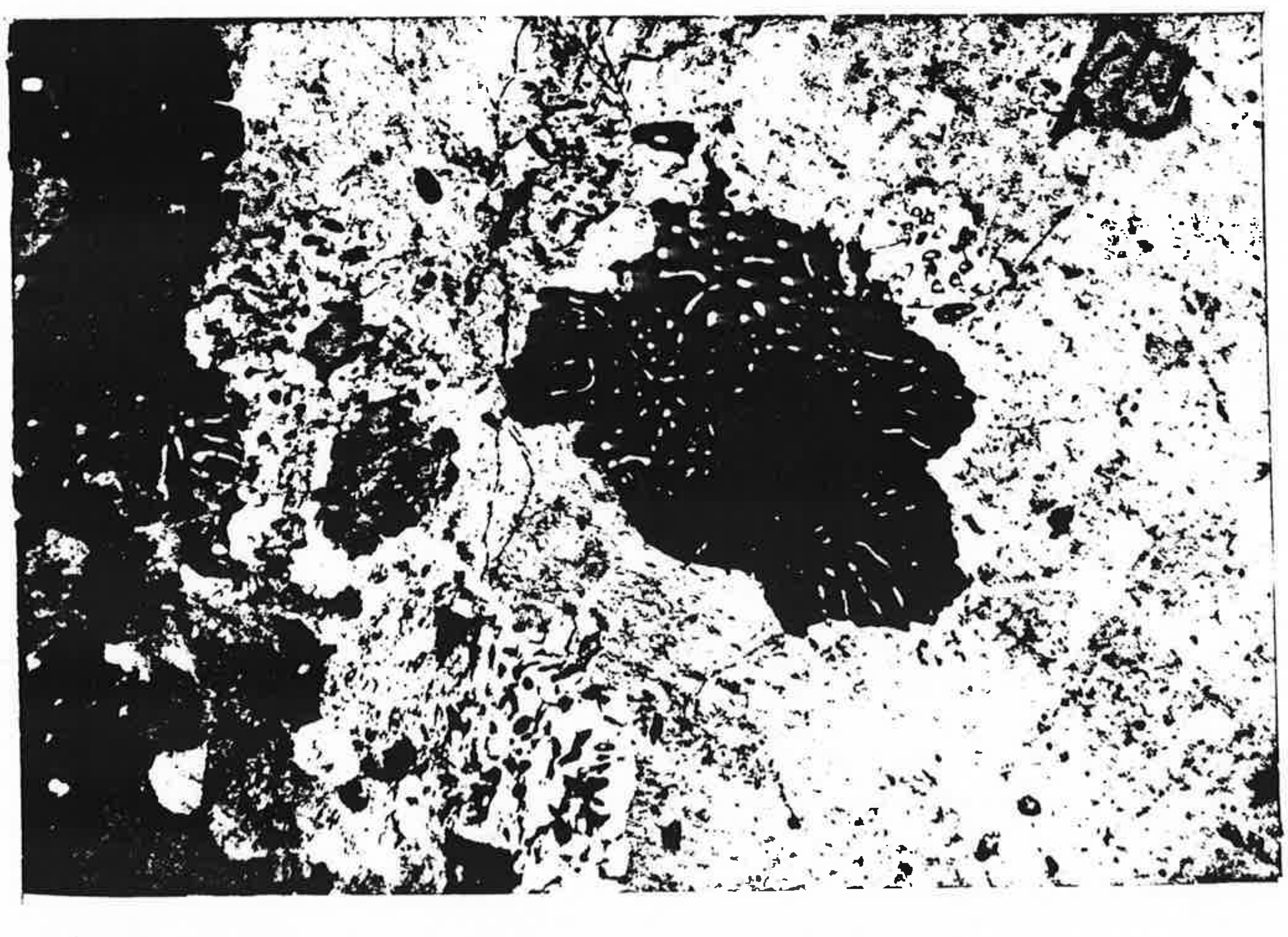
A FIELD GUIDE TO MYRMEKITE

BY LORENCE COLLINS



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Preface

The contents of this "field guide" are extracted in part from a much larger volume of material which I hope to publish as a book. In order to save time and money I have utilized the same figure numbers as in the original manuscript, and therefore, many figures are missing. For example, the first illustration in the "guide" is Figure 5. I have also omitted some figures that are described in the text which are either photomicrographs or glossy prints of outcrops because these photographs do not copy well. However, seeing these pictures are not essential to understanding the text.

You may find it helpful, if you have time, to look at several of the references listed in the back of the "guide", pages 77-79, before going on the field trip. In alphabetical order these include the following: Grout (1941), Henshaw (1942), Jahns and Wright (1951), Myers (1978), Phillips (1974), and Schermerhorn (1960). These articles describe or give background information for the kinds or rocks or topics we will discuss in the field.

The ideas that I will discuss in the "guide" and in the field are "brand new", and so I expect to be "ONE" against the "Crowd", but hopefully, I can challenge you to "see" rocks in a new way.

Lorence G. Collins

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INTRODUCTION

Myrmekite, the vermicular intergrowth of quartz in plagioclase, was first reported by Michel-Levy (1874), who referred to it as "quartz vermicule" (like worm tubes). Later Sederholm (1897) gave it the name "myrmekite", meaning "wart-like". Historically, myrmekite has been ignored in discussions of the origin of granitic rocks even though lengthy and complex arguments have been presented (e.g., Read, 1948; Raguin, 1957; Walton, 1955; Wegmann, 1963; Mehnert, 1968; Pitcher and Berger, 1972). This is because myrmekite generally constitutes less than one percent of any granitic rock, and therefore, investigators have not felt that a mineral intergrowth composing such a small volume could possibly have any major effect on the chemical composition of the whole rock or on its evolutionary history. If myrmekite is mentioned at all, it is regarded as a possible, late-stage deuteric alteration of a granitic rock that has solidified or nearly completely solidified from a magma. This concept has been so widely accepted that it has not been challenged, and therefore, efforts to explain the origin of myrmekite have been concentrated on the mechanism of myrmekite-formation during the deuteric process rather than relating myrmekite to the whole process of granite formation. To suggest that the origin of granite and myrmekite might be inter-related has not even been considered.

Explaining the origins of myrmekite-bearing granitic rocks or of myrmekite has been perplexing, but the failure to produce explanations that are satisfactory to everyone has not been the fault of experimental petrologists but of field geologists who have not described or explained their own observations well enough (Marmo, 1971) or who have assured the experimentalists that myrmekite-bearing rocks were created by magmatic processes when that assurance has been an assumption rather than a proven fact. Consequently, most experimental work on granitic rocks has begun with a hydrous melt in which the water has not acted as a flux (in the sense of a fluid passing through a solid) but as a solvent of component oxides or natural mineral fragments in a closed system (Marmo, 1971). Moreover, the solvent has been at high temperatures and pressures, under supercritical conditions. The results of this experimental work have been spectacularly successful in explaining relationships in non-myrmekite-bearing granitic rocks where temperatures above the melting point have obviously existed, but the problem is that these data do not apply to many plutonic, myrmekite-bearing, granitic rocks that have formed from processes that occur in an open system at temperatures and pressures below the melting point (300° to 525° C)(Barth, 1956).

Various theories for the origin of granitic batholiths have been proposed, including: (1) magmatic differentiation and physical emplacement of residual granitic liquids, (2) anatexis of a geosynclinal pile of metasedimentary rocks of granitic bulk chemical composition, and (3) granitization of a non-granitic host rock by aqueous fluids that bring in Si, K, and Na and change the rock to granite by metasomatic processes. Summaries of these theories and their unexplained problems have been presented in many different sources and need not be further amplified here (e.g., Grout, 1941; Bateman and Eaton, 1967; Kistler and others, 1971, Hyndman, 1972; Best, 1982). In each theory conservation of mass (the "room problem") has been considered, and on the basis of that consideration, the evidence suggests that granitic rocks may be formed by any one of the above methods, depending upon the geologic environment and field relationships. However, problems with each of these methods in many geologic terranes suggest that an alternative explanation may be possible for the origin of some granitic batholiths. Rather than magmatic differentiation, anatexis, or granitization on a batholithic scale, I suggest that some calc-alkalic granitic rocks have evolved from former magmatic rocks of gabbro or diorite composition by a new kind of granitization process which I propose to be called "hydrothermal differentiation". In this process hydrothermal fluids modify the compositions of solid but sheared gabbro, diorite, or other mafic rocks, so that Ca, Al, Mg, and Fe are subtracted. This results in a volume reduction and a concomitant enrichment in K, Na, and Si in the recrystallized residue. By changing batholithic mafic rocks to granitic rocks in situ, there is no "room problem", since shouldering aside of the mafic rocks is not necessary to emplace the residual granitic rocks.

In previous considerations of the origin of granite by "granitization" an emphasis has been placed on the addition of Si, K, and Na to the mafic rocks. It has been noted that this addition results in a loss of Ca, Al, Mg, and Fe, but generally this loss is regarded to be only an apparent loss and more the result of dilution by the K, Si, and Na rather than an actual subtraction of the Ca, Al, Mg, and Fe (Marmo, 1971). Arguments for this dilution and in favor of this particular kind of granitization have included the following:

(1) the lack of evidence in some rocks for the replacement of ferromagnesian silicates, whereas calcic plagioclase commonly shows evidence for replacement by K-feldspar, quartz, and albite.

(2) the coexistence in other altered rocks of secondary epidote, calcite, chlorite, sericite, sphene, and other mineral products into which Ca, Al, Mg, and Fe from altered primary ferromagnesian silicates could reside.

(3) the use of balanced chemical equations to show that the products of the alteration reactions can account for all components of the original rock without any loss of Ca, Al, Mg, and Fe.

(4) the absence of any evidence of a "basic front", indicating where released Ca, Al, Mg, and Fe could have gone.

(5) the experimental data that show that ferromagnesian silicates and calcic plagioclase melt at higher temperatures than quartz, K-feldspar, muscovite, and sodic plagioclase. These data supposedly suggest that K, Si, and Na should be more mobile and Ca, Al, Mg, and Fe less mobile, resulting in a "basic behind", and

(6) the granitic rocks have high temperature fabrics that suggest a magmatic origin.

In response to the above six arguments, the following counter arguments are given:

(1) The absence of the replacement of ferromagnesian silicates in granitized rocks is more apparent than real. In rocks being granitized the ferromagnesian silicates are gradually replaced by quartz, but this replacement is not recognized because it does not "look like" replacement. Where quartz replaces ferromagnesian silicates, it occurs as interstitial grains adjacent to the ferromagnesian silicates or as "inclusions" inside the disappearing ferromagnesian silicates, forming

a poor- to well-developed sieve texture (or symplectite). In support of the replacement of ferromagnesian silicates is the evidence in some terranes of a ghost stratigraphy that shows that the emplacement of the granite is by passive processes rather than by physical injection or by shouldering aside of the former basic rocks (Pitcher and Berger, 1972).

(2) The presence of secondary epidote, chlorite, sericite, calcite, and sphene, and (3) the writing of balanced chemical equations involving these secondary minerals show that the Ca, Al, Mg, and Fe released from primary minerals to form these secondary minerals can be fully accounted for in the mass reactions. Therefore, these elements may not have been lost from the system in some places where myrmekite and other deuteric mineral products have been formed. However, these altered rocks have not been granitized to any great extent, and the nearly constant chemical composition confirms this. Nevertheless, if these partially altered rocks are traced into strongly granitized rocks, both the secondary alteration minerals and the primary ferromagnesian silicates and zoned plagioclase generally disappear or decrease in abundance as K-feldspar, albite, myrmekite, and quartz increase in abundance. This suggests that where granitization occurs, Ca, Al, Mg, and Fe are subtracted from the system.

(4) The absence of a "basic front" is primarily a problem of not recognizing what a basic front looks like. Furthermore, the separation between granitized rocks and the basic front may be so great that in most places, if the basic front can be seen, its connection with granitized rocks cannot be proved. I hope to show later by indirect

evidence that appinitic intrusives, lamprophyre dikes, and mafic calc-alkaline volcanic rocks are the basic fronts containing the elements lost from nearby or deep-seated granitized rocks. Usually, because of their contrasting chemical compositions, appinitic intrusives, lamprophyre dikes, and mafic calc-alkalic volcanic rocks are considered to be totally unrelated to granites. This supposition is because petrologists are accustomed to think in terms of magmatic histories for these rocks that involve magmatic differentiation rather than to a possible hydrothermal process of subtraction at temperatures below the melting point of the source rock.

(5) Although the experimental data that are derived from melts at high temperatures in closed systems may not be applicable to some granitic rocks, there are several experimental studies in lower temperature ranges (180^{o} to 650^{o} C) which may help to explain the movements of elemental materials even though the experiments have been performed in closed systems. For example, in granitic rocks subjected to a temperature gradient from 180^{o} to 620^{o} C and at constant pressure of 2 kbar, Si, Fe, Al, K, and Na are found to migrate to the lower temperature end while Ca (from plagioclase) and Mg (from biotite) migrate to the high temperature end (Johann and Winkler, 1965). In the process albite becomes separated from K-feldspar and occurs in the temperature range of 420^{o} to 470^{o} C while K-feldspar and quartz form below 420^{o} C. Khitarov (1957) showed that Mg cannot be removed from biotite under hydrothermal conditions at approximately 0.5 kbar and 430^{o} to 460^{o} C, although silica and Na under these conditions show a greater mobility than K. However, at temperatures of 350^{o} to 700^{o} C

and at pressures of 2 kbar, a rock rich in Ca and containing two co-existing alkali feldspars shows an opposite tendency for K and Na. If the rock is Ca-rich, the coexisting vapour phase is enriched in K relative to Na so that a warmer, Ca-rich rock becomes depleted in K while a cooler, Ca-poor rock becomes enriched in K-feldspar (Orville, 1963). O'Neill (1948) found that whether Na replaces K-feldspar or K replaces Na-plagioclase was dependent upon the Na or K ion concentration and not on whether the solutions were basic, neutral, or acidic. But Currie (1968) found that whether Na or K moved was dependent more on the pressure. At water vapour pressures of 400 to 3,500 bars and at temperatures of 400^{o} to 650^{o} C, the higher pressure favors the release and movement of K. The higher pressure (up to 30 Kb) also causes less silica to be released (Wendlandt and Eggler, 1980).

Because these data are primarily derived from granitic rocks containing biotite as the only ferromagnesian silicate and because these data are derived from closed, hydrous systems, the data can act only as a first approximation of the way in which hydrous fluids would act on olivine, pyroxenes, hornblende, and biotite in gabbro, diorite, or other basic rocks that are subjected to shearing stresses in an open system. These hydrous fluids may also act differently where the mineral grains are interlocked tightly against each other rather than as a powder or disoriented fragments that are not directly bounded by other crystals. Moreover, the relative movements of Ca, Al, Mg, Fe, Na, K, and Si from rocks containing olivine, pyroxenes, hornblende, biotite, and calcic plagioclase may be somewhat different if HF, HCl, P_2O_5, CO_2, or B were added to the water instead of just water alone.

Gore (1968) has already shown that boron salts aid the movement of K in granitic rocks. In the absence of supporting experimental work on open systems, the above experimental work on closed systems suggests that movements of elements from sheared basic rocks during alteration by hydrothermal fluids may be as follows:

(a) Mg would be the first to move out, but Fe and Al would also be carried away because the destruction of olivine, pyroxenes, hornblende, and biotite would release these elements as well. Some Fe, Al, and lesser amounts of Mg would remain behind in residual biotite.

(b) Ca from destroyed hornblende, augite, and plagioclase would move out at a faster rate than Na, since Na is generally not abundant in hornblende or augite and since Na would tend to stay in remnant plagioclase to form the more stable sodic plagioclase at lower temperatures. The greater subtraction of Ca than Na from plagioclase would also release some Al, but in some rocks the Al tends to remain behind to form Al-rich minerals such as muscovite, garnet, tourmaline, or sillimanite.

(c) K from destroyed biotite would move out ahead of Na since plagioclase (a source of Na) is not affected in early stages by hydrous fluids as much as the ferromagnesian silicates. The amounts of Na relative to K in escaping fluids may be a function of the degree to which the K is moving into and replacing the plagioclase to form microcline. If microcline replaces plagioclase, then Na would be displaced to the escaping fluids.

(d) Si is the slowest to move and tends to remain behind in quartz.

The effect of these relative movements in early stages of hydrous fluids moving through sheared basic plutonic rocks at temperatures below the melting point is to form fluids rich in Ca and Mg, relatively rich in Fe, Al, K, and Na and poor in Si. These proportions are typical of appinitic rocks, lamprophyres, and feldspathoid-bearing mafic volcanic rocks that are commonly associated with granitic terranes (see discussion in later section).

(6) The high-temperature fabric in many granitic rocks is probably a relic texture and is not recognized by many geologists to be a low-temperature pseudomorph of the former high-temperature hypidiomorphic framework (Schermerhorn, 1960). Older minerals (hornblende, biotite, normally zoned plagioclase, monoclinic K-feldspar, quartz) in more basic rocks have been recrystallized to form younger, low-temperature minerals (albite resulting from decalcification of plagioclase, triclinic K-feldspar, myrmekite, chlorite, quartz) in granite. Because geologists have not recognized that these low-temperature minerals have been superposed during metamorphic processes on to a high-temperature fabric, errors in interpretation have been made by geologists who have assumed that the recrystallized granitic rocks are magmatic and who have tried to explain the changes in mineralogy and chemical composition by using data that apply to magmatic rocks.

Several different geothermometers indicate that myrmekite-bearing granitic rocks form at temperatures which are far below the melting point of granite (Barth, 1951, 1962; Stormer, 1975; Currie, 1971; Tracy and others, 1976). The presence of myrmekite is emphasized here because this plagioclase-quartz intergrowth in future experimental studies may be shown to form only in rocks below melting temperatures, and therefore, its presence may also prove to be useful as a geothermometer of a certain low-temperature range when other minerals normally used as geothermometers are absent. Geothermometers that are known to be useful include coexisting feldspar pairs and garnet-biotite pairs that occur in the myrmekite-bearing granitic rocks. Metamorphic minerals (i.e., cordierite, kyanite, staurolite) and sulfides that occur in wall rocks would also help to establish a limiting temperature range for the adjacent myrmekite-bearing granitic rocks. Temperature measurements of myrmekite-bearing granitic rocks include the following: 300° to 380° C for feldspar pairs in the Giants Range batholith of Minnesota (Griffin, 1967); 430° C for feldspar pairs in the Woodson Mountain granodiorite of the Southern California batholith and 400° to 450° C in other anatectic granites in Norway and Sweden (Barth, 1956); less than 500° C for sulfides in veins adjacent to granitic gneisses (Marmo and Hyvarinen, 1953); 475° to 550° C for garnets in a two-mica granite in Nevada (Ghent and others, 1979); 490° C for garnet-biotite pairs in the Lowe granodiorite (Weigand, personal communication, 1982); $425^{\circ} \pm 25^{\circ}$ C for feldspar pairs and $525^{\circ} \pm 25^{\circ}$ C for garnet-biotite pairs in granitic rocks near Augusta, Maine (D. S. Barker, 1964; Ferry, 1978, 1979). Because microcline cannot be formed experimentally at temperatures above 525° C (Goldsmith and Laves, 1954) and because coexisting albite and myrmekite adjacent to microcline cannot be post- or pre-microcline (Rogers, 1961), the myrmekite-bearing granitic rocks must have formed at temperatures below 525° C. All of these temperature measurements support Schermerhorn's (1960) contention that a low-temperature fabric has been superposed onto a high-temperature framework and give evidence that myrmekite-bearing granitic rocks may form by hydrothermal processes at temperatures below the melting point of granite. Furthermore, electron-microprobe measurements of the coexisting feldspars in the myrmekite-bearing granite near Temecula (discussed in later sections) give values of 4 to 8% albite in K-feldspar and 87 to 89% albite in plagioclase. These values indicate temperatures of formation in the range of 400 to 500^{0} C, using Stormer's (1975) charts, which confirm the above correlations.

Marmo (1971) has suggested that the K in granitized rocks comes from sedimentary rocks or from basic igneous rocks containing orthoclase and that the movement of this K from its source rock to cause the granitization in another rock results in a simultaneous granodioritization of the source rocks. I agree that K in granitized rocks may come from sedimentary rocks, but I suggest that the K comes via an intermediate step of incorporation in the magmatic rocks, gabbro or diorite. Most mantle-derived gabbro and diorite do not have enough K to make large granitic bodies. However, the assimilation of K from sedimentary rocks by an intruding gabbro or diorite pluton may locally enrich these rocks in K which becomes incorporated in biotite (10 to 30%). This biotite would then be the secondary source of K when these biotite-bearing basic rocks are sheared and recrystallized to form microcline in residual myrmekite-bearing granitic rocks. On the other hand, if the magmatic basic rocks are devoid of biotite (or orthoclase), the K must be transported to the sheared basic rocks if these rocks are to be granitized. The transport and addition of Na and/or Si along with K, however, may not be necessary because Na and Si would be residually concentrated in the basic rocks as Ca, Al, Mg, and Fe are subtracted.

THE RELATIONSHIP OF MYRMEKITE TO THE ORIGIN OF GRANITE

Although there is an apparent relationship between the presence of lamprophyres (and perhaps carbonatites) and myrmekite-bearing granitic rocks, it is not clear from this association how the formation of myrmekite is related. In most myrmekite-bearing rocks no transition stages are ever found that show how myrmekite is formed. Consequently, at least ten hypotheses have been presented to explain its origin (Smith, 1974; Phillips, 1974, 1980b; Hibbard, 1979, 1980b; Collins, 1983). I the following section I describe a 10-cm wide transition zone between tonalite and granite that occurs 2.5 km southwest of Temecula, Cali-

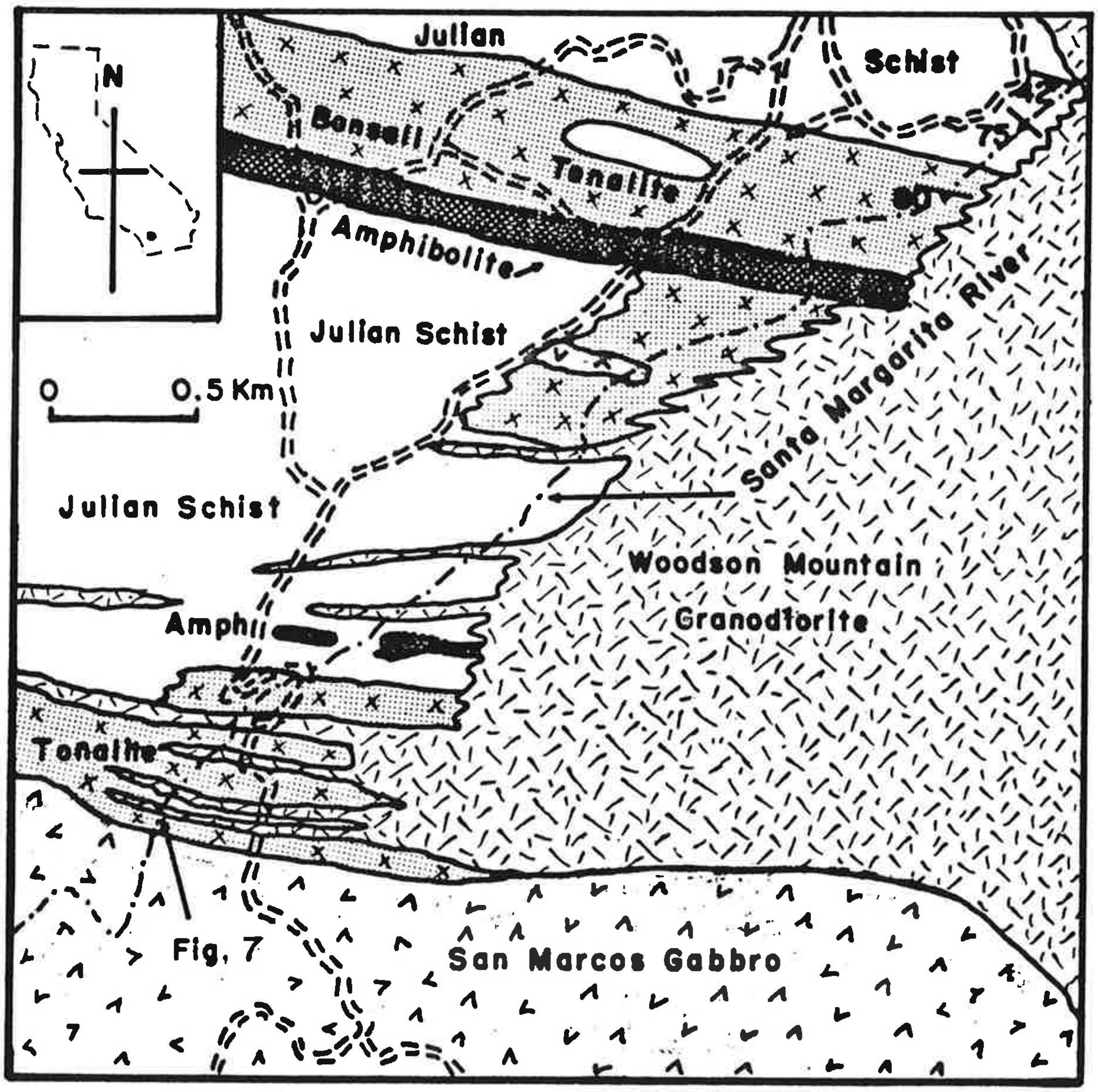


Figure 5

fornia. In this transition zone all stages in the formation of myrmekite can be found.

The area containing the 10 cm transition zone is shown on Figure 5. In this area excellent exposures of all rock types occur along the Santa Margarita River canyon, but soil cover and dense chaparral away from the canyon prevent locating contacts precisely. The oldest (Paleozoic ?) rocks are metasedimentary and composed of quartzose or feldspathic (oligoclase), well-laminated rocks in which mica is a prominent constituent. These rocks were intruded by diabasic sills and dikes, and then later the sediments were metamorphosed to form schists (the Julian schist) and the diabases were metamorphosed to form amphibolites (Jahns and Wright, 1951; E.S. Larsen, 1951); see Figure 6, stage 1. Intruded into these rocks are the San Marcos gabbro and the Bonsall tonalite (Figure 6, stage 2). The San Marcos gabbro is variable having both plagioclase and mafic rock layers. The so-called "Bonsall tonalite" is not all tonalite, however. Adjacent to the gabbro it is a hornblende-pyroxene diorite and adjacent to the schist it is a biotite-hornblende tonalite. The diorite is massive and relatively uniform, but where the diorite grades into the tonalite, the tonalite has a faint foliation which is parallel to the schistosity of the amphibolites and the Julian schist. At the tonalite-amphibolite contact, the tonalite is full of partially assimilated amphibolite xenoliths, but at the tonalite-schist contact, the tonalite rarely contains xenoliths, and these disappear within 2 m of the contact. The abrupt disappearance of schist xenoliths and the accompanying increase in mica suggest that the tonalite resulted from assimilation

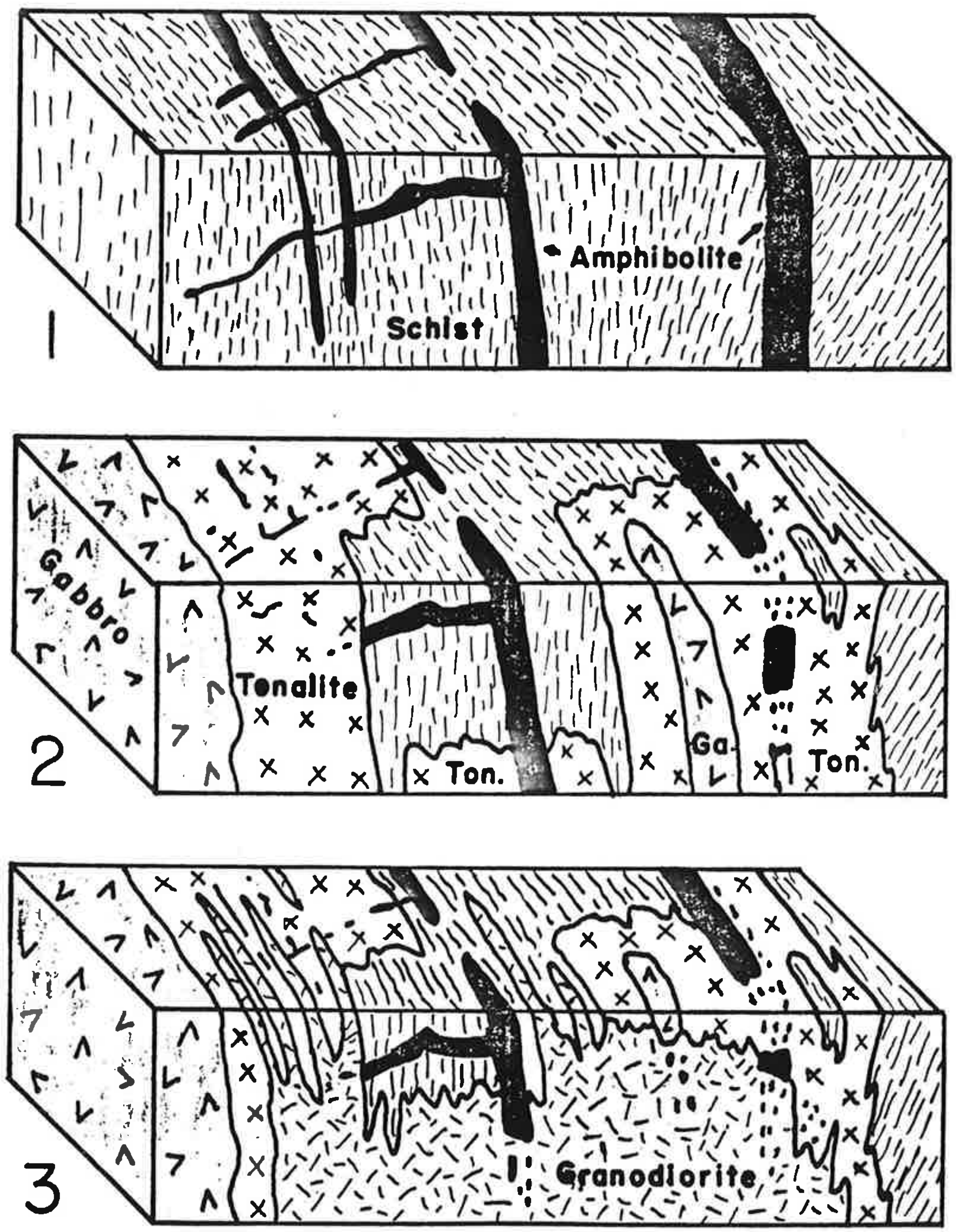
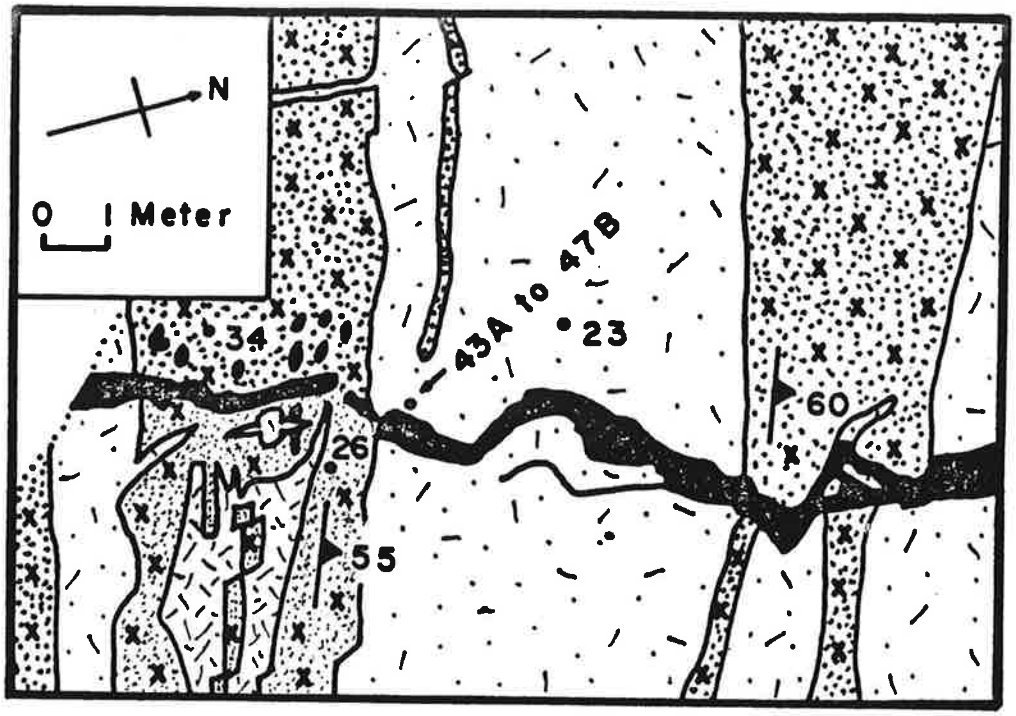

Amphibolite
Schist
1
Gabbro
Tonalite
Ton.
Ga.
Ton.
2
Granodiorite
3

Figure 6.

of the schist by the original diorite magma. If this is true, perhaps the small percentages of quartz (0 to 5%) that occur in the tonalite may be quartz that has been derived by incorporation of the schist. Some quartz, however, may result from hydrothermal alteration that accompanied the formation of the Woodson Mountain granodiorite (see next section).

Appearing to intrude the Julian schist, amphibolite, gabbro, diorite, and tonalite is the Woodson Mountain granodiorite (and granite) (Figure 6, stage 3). Dike-like bodies of the granodiorite extend into the older rocks. In the following sections I present evidence to indicate that the granodiorite and its dike-like bodies are replacement in origin, forming from sheared tonalite at temperatures below the melting point (430° C) (Barth, 1956) and inheriting the high temperature fabric of the tonalite. Part of this evidence is a ghost stratigraphy of the amphibolite dikes, Julian schist, and tonalite that occur in the Woodson Mountain granodiorite east of the Santa Margarita River canyon (Figure 5) although contacts with these rock types to the west are sharp as if they represented the border of an intrusive granodiorite magma.

Figure 7 shows a close-up of the field relationships between the Woodson Mountain granodiorite and granite and the Bonsall biotite-hornblende tonalite; see Figure 5 for location of Figure 7. In the close-up area the tonalite has intruded the Julian schist and amphibilite and incorporated the amphibolite as xenoliths. Remnants of the Julian schist bordering the amphibolite are found in nearby exposures but are not shown on Figure 7. Field evidence that the Woodson Mountain



EXPLANATION

- Woodson Mountain granite
- Woodson Mountain granodiorite
- Bonsall biotite tonalite
- Bonsall bio.-hornbl. tonalite
- amphibolite xenoliths

Figure 7.

granodiorite and granite resulted from the hydrothermal modification of the Bonsall tonalite instead of by physical injection of a magmatic granitic body into the tonalite includes the following:

(a) The older amphibolite "dike" extends (across Figure 7) from the tonalite through the granite and back into the tonalite without any evidence of breakage or rotation of fragments in the granite. Likewise, there is no dilation or separation between the tonalite wall rocks on either side that would indicate displacement by an intruding magma.

(b) West of the amphibolite "dike" in the central, thick granite mass is a ghost island remnant of the tonalite (1 m west of specimen 43A on Figure 7) which extends parallel to the nearby biotite-hornblende tonalite mass. This remnant is about 2 to 3 cm wide and 8 m long, but its width is greatly exaggerated on Figure 7 in order to show its physical features. This narrow remnant is uncontorted and preserves a faint foliation that is parallel to the foliation in the nearby massive biotite-hornblende tonalite. A fracture offsetting this remnant projects to an equivalent fracture in the adjacent biotite-hornblende tonalite. Although no trace of the fracture extends through the granite between the two tonalite bodies, the granite cannot be a magmatic rock because the alignment of the fracture plane in the two places shows that this thin remnant has not been physically displaced by dilation. The former fracture plane has been destroyed when the granite was formed by replacement of the tonalite that once bordered the fracture. The lack of contortion in the thin island remnant also supports a replacement origin for the granite.

(c) Although the biotite tonalite east of the amphibolite "dike" appears to be penetrated by granodiorite veins and dikes, the unmatched walls, the absence of any dilation, and the isolated granodiorite pod adjacent to the amphibolite "dike" (1 m southwest of specimen 26) that cuts both the tonalite and an amphibolite xenolith indicate that the granodiorite has formed by replacement of the tonalite.

(d) No xenocrysts of hornblende occur in the granodiorite and granite adjacent to the tonalite or amphibolite contacts, and no hornblende reaction rims occur along these contacts. The absence of these features eliminates the possibility of assimilation by a granitic melt or the extraction of granitic components from the tonalite by partial melting.

(e) The many "re-healed" fractures east of the amphibolite "dike" (lower left quarter of Figure 7) suggest that the tonalite was more strongly sheared in this area than to the west. The loss of hornblende from the tonalite in the eastern area also supports a greater cataclasis and recrystallization in this area. The orientation of the fracture patterns suggest that the shearing was partly controlled by a faint foliation in the tonalite that parallels the schistosity in the nearby Julian schist. As in (b) above, the "re-healed" fracture planes occur in the biotite tonalite but not in the recrystallized granodiorite and granite.

(f) There are no high-temperature contact metamorphic minerals (garnet, sillimanite, kyanite, etc.) in the Julian schist adjacent to the Woodson Mountain granodiorite and granite. The low-temperature measurements of 430° C on the feldspar pairs in the granodiorite (Barth, 1956) indicate that temperatures were not high enough to form a metamorphic aureole that commonly occurs around large magmatic bodies.

All of the above suggests that the sharp contacts and cross-cutting relationships shown on Figure 7 are misleading, since these kinds of features are normally used as evidence for magmatic injection (see Roddick, 1982). More likely, the sharpness of the contacts occurs because different parts of the tonalite have been subjected to different degrees of severity of cataclasis. The sharp contacts represent distinct boundaries between those rocks that have not been as strongly affected by the cataclasis, as along a fault line. Perhaps the cataclasis was episodic, so that each successive movement and crushing allowed different amounts of fluid to be introduced, and subsequently, these fluids created different degrees of replacement and recrystallization depending upon the degree of cataclasis---first converting biotite-hornblende tonalite to biotite tonalite, then to biotite granodiorite, and finally to granite containing only a small remnant of the original biotite (1 to 3%).

Because sharp and cross-cutting contacts are found nearly everywhere, it is difficult in most places to show that the Woodson Mountain granodiorite and granite are not magmatic rocks, particularly when all that is seen east of the Santa Margarita River for many hundreds of square meters is granodiorite. Even in thin section the high-temperature fabric and remnant xenocrysts of zoned plagioclase grains from the tonalite give the impression of a magmatic rock instead of a rock pseudomorph on which a low-temperature fabric has been partially "telescoped" (Schermerhorn, 1960). To show that the granodiorite and granite were originally tonalite, a place was found where remnant biotite-hornblende tonalite bordering the amphibolite "dike" was converted gradually to biotite tonalite, then granodiorite, and finally granite in a 10 cm transition zone (43A to 47B, Figure 7). The same kind of transition zone also occurs in a 1.5 cm wide zone bordering the thin island tonalite remnant, but this transition is too abrupt to be of much help in separating the stages of recrystallization and replacement.

Part of the difficulty of understanding the replacement process in the 10 cm transition zone is the microscopic size of the grains which are being replaced. Viewed under the lowest power, the field of view is a circle about 2.5 mm in diameter. Because of this limitation, observing the changes across the 10 cm zone is like looking at a complex cathedral glass window through a "tunnel peep hole" and trying to remember what the sum of all the peep hole images would look like if they were combined into a complete picture. In order to see all mineral relationships at the same time, I photographed each thin section at low power (3.5x) with more than 100 overlapping images. Prints from these images were enlarged 21x and pasted together in a mosaic 52 cm across. Then, grain-by-grain, I mapped the mosaic and plotted the locations of each mineral species: quartz, biotite, hornblende, plagioclase, microcline, and myrmekite. What emerged are patterns of mineral changes, abundances, and textures that clearly show all mineral relationships in the transition from tonalite to granite. The changes in modal volumes of the minerals are summarized on Figure 8. The mineral-grain maps revealed the following relationships:

1. The transition is not a smooth uniform conversion. Various parts of the tonalite are in more advanced stages of change than others. In the middle of the transition, island remnants of relatively unaltered plagioclase xenocrysts of the tonalite are surrounded by an aureole of

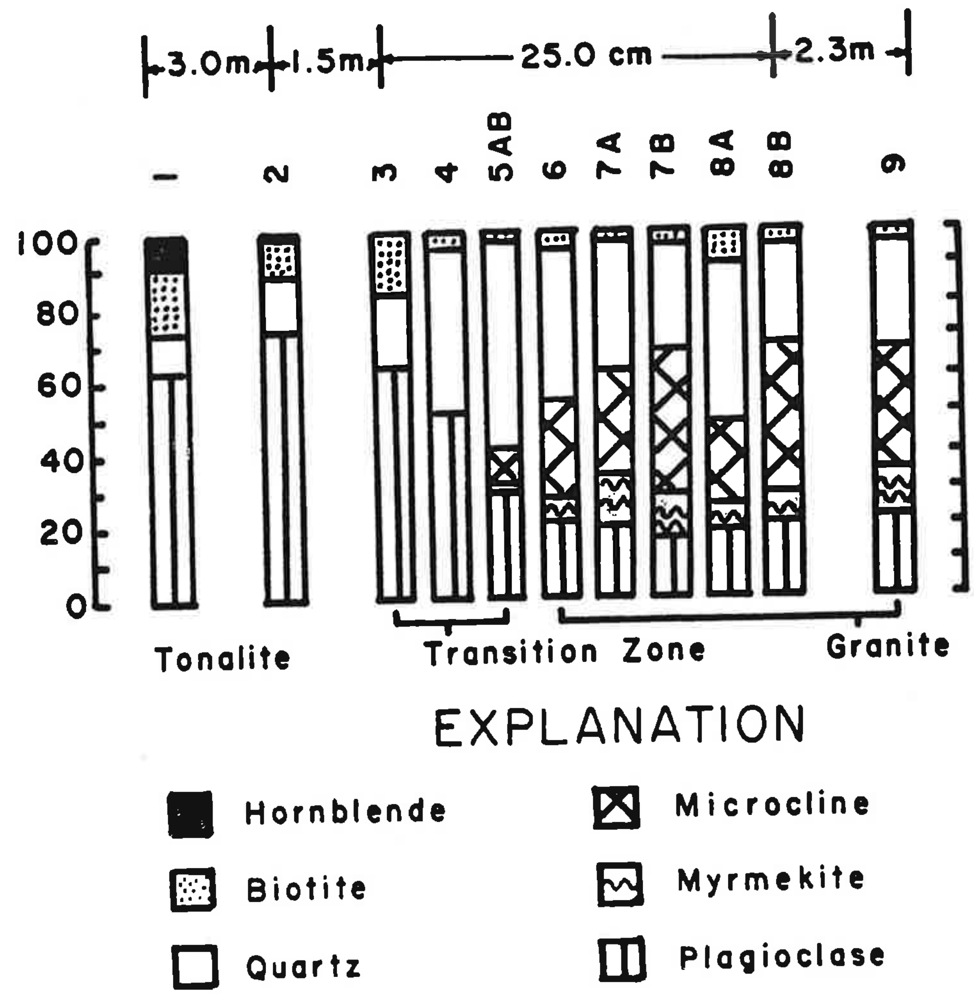


Figure 8. Graphic display of modal analyses of tonalite, transition zone, and granite in the Santa Margarita River canyon. (No. 1 = specimen 34; no. 2 = specimen 26; nos. 3 to 8b = specimens 43A to 47B; no. 9 = specimen 23; specimen locations are shown on Figure 7).

grains in process of being converted. The non-uniform conversion probably resulted because cataclastic crushing of the tonalite was non-uniform. Therefore, local differences in cataclasis probably affected the ability of fluids to reach or penetrate certain grain boundaries.

2. The primary hornblende, biotite, and plagioclase grains of the tonalite are not uniform in size but show a seriate texture (0.1 to 2.0 mm). Biotite (2 to 18%), hornblende (8 to 20%), and quartz (trace amounts) are commonly intergrown, but in some places, small, isolated biotite grains are enclosed by plagioclase. Apatite is rare. Magnetite and sphene are absent except where tonalite grades to diorite.

3. Zoisite inclusions are common in the cores of the large plagioclase grains in the tonalite but across the transition zone towards the granite the zoisite disappears (a loss of Ca and Al).

4. The zoning in the plagioclase grains (cores An_{39}, rims An_{22}) in the tonalite is not consistent from grain to grain. This is a function of where the grain is cut in thin section and of its grain size. Large plagioclase grains show more zoning (bands of different composition) than small grains. This suggests that the plagioclase composition is slightly different from grain to grain, and these variations would affect the replacement process.

5. Nearer the tonalite where plagioclase first shows indications of conversion, the large zoned plagioclase grains show incomplete alteration. Remnant zoning is preserved in some places, but a loss of zoning and of albite twinning occurs in other places to form areas of mottled extinction. However, the outline of the large grain is still preserved so that the changes that took place are internal, rather

than the result of an external introduction of a different plagioclase mineral from a granite magma.

6. In the transition adjacent to tonalite, quartz increases as hornblende and biotite decrease. In some places the quartz occurs as a poorly developed sieve-texture in the hornblende, which does not occur in the unaltered tonalite or diorite. Across the transition zone toward granite, hornblende disappears before biotite, and where hornblende has disappeared (to form biotite tonalite), the biotite concentrations also exhibit quartz sieve-textures. Isolated biotite grains enclosed by plagioclase, however, show no apparent alteration and continue through the transition zone into granite. The disappearance of most biotite grains coexisting with hornblende in the transition zone suggests that this biotite (decreasing from as much as 18% to as low as 1 to 3%) is the source of K for the microcline in the granite (see also Figure 8). The quartz that first appears as a poor sieve-texture in the biotite and hornblende increases from trace amounts in the tonalite to as much as 38% in the granite. Because the disappearance of hornblende and biotite correlates with the appearance of residual quartz, the irregular stringers, pencils, and clots of quartz in the granite are presumeably former sites of irregular patches of hornblende and biotite in the tonalite.

7. All plagioclase grains, either enclosing biotite or free of biotite, show complete gradual changes in physical appearance through the transition zone. First, they have normal zoning and albite twinning in the tonalite (Figure 9). Then they become broken with bent albite twin lamellae and strained extinction. Later they lose their zoning and

twinning and have a mottled extinction, then become "pitted", then clear and untwinned, and finally, show albite twinning in the granite. In the middle of the 10 cm transition zone all plagioclase grains, regardless of their size and degree of cataclasis, show the "pitted" appearance; Figure 10. The "pits" are tiny, spherical holes, 0.001 to 0.005 mm in diameter, and occur in the transition zone prior to where microcline becomes abundant in the granite. The textural appearance (grain size and shape) of the pitted plagioclase grains is no different from the adjacent unpitted plagioclase grains in the tonalite. On the opposite side in the granite the textural appearance of the recrystallized plagioclase grains (free of holes) is also the same except that some of the plagioclase grains have been replaced by microcline and myrmekite.

8. From the tonalite toward the granite the first visual appearance of microcline is in the interstices between plagioclase grains and in the cores of altered plagioclase grains showing mottled extinction. In the cores the microcline occurs as irregular patches which are randomly distributed. The lack of microcline in the tonalite and its gradual appearance inside plagioclase grains in the transition zone supports an internal replacement rather than an external engulfing of plagioclase xenocrysts by a penetrating granitic magma. The lack of microcline in the tonalite also rules out an origin by exsolution.

In the transition zone adjacent to the granite a few isolated, large (1.5 mm) microcline grains do occur in the midst of abundant altered plagioclase grains, but these large microcline grains are not more than 0.2 cm from the granite where the microcline modal abundance (30 to 35%) becomes greater than that of plagioclase (25 to 30%); see Figure 8.

9. Myrmekite is absent not only where microcline first occurs in the interstices and as irregular islands in the plagioclase, but also where the large, isolated microcline grains first appear near the granite. Only where microcline becomes abundant does myrmekite occur, and then myrmekite is found either as tiny "aggregate" grains between two large microcline grains or as larger, wart-like protrusions "growing" on a large plagioclase grain and enclosed by a microcline grain (Figure 11). In the adjacent area of altered tonalite are numerous aggregate tiny grains of "pitted" plagioclase of the same size and shape as the aggregate myrmekite grains enclosed by microcline. These aggregate, pitted plagioclase grains are broken fragments of former larger grains in the tonalite. In the granite these grains lose their pitted appearance. Some of them become tiny myrmekite grains enclosed in a single microcline grain or between two microcline grains; others recrystallize as small, clear, albite twinned plagioclase grains that are outside of any microcline grains and free of quartz vermicules. So it appears that the formation of myrmekite from the aggregate altered plagioclase grains is a function of whether these aggregate grains are surrounded by microcline. It is clear from the thin sections that the aggregate "pitted" plagioclase grains that become myrmekite are derived from fragments of former large plagioclase grains in the tonalite and are not the result of exsolution from microcline. This is clear not only from the textural evidence but also because in the

early part of the transition zone the volume of aggregate myrmekite grains far exceeds what could logically exsolve from the volume of adjacent microcline that surrounds the myrmekite.

The other type of myrmekite, the wart-like growths on plagioclase, does not occur unless the myrmekite is enclosed by microcline and unless other microcline grains are nearby. This is suggested because isolated microcline grains are myrmekite-free.

10. Remnants of unreplaced plagioclase occur in some microcline grains in the granite. These remnants are albite lamellae (An_{0-5}) which are scattered in random distribution and which have irregular shapes typical of replacement processes (Alling, 1932, 1938). Also, in some microcline grains remnant quartz ovoids, looking like ghost myrmekite, may be found.

In summary, the conversion of the tonalite to granite (Figure 8) shows that the modal volume of plagioclase has been reduced from 60% in the tonalite to less than 35% (including myrmekite) in the granite. Hornblende disappears early in the transition, but biotite gradually decreases in amount from as much as 18% in the tonalite to 1 to 3% in the granite. Quartz has increased from trace amounts to 25 to 38%, and microcline from 0 to as much as 35%. In this process K has been transferred from the tonalite to the granite, and large quantities of Ca, Al, Mg, Fe, and other elements have been carried out of the system because of the disappearance of hornblende, the destruction of most of the biotite, and the decalcification of the plagioclase; see Figure 12. (S. White, 1975, has also found that naturally deformed portions of oligoclase porphyroclasts have lost Ca relative to undeformed portions during recrystallization.) The loss of elements is confirmed by chemical analyses and electron-microprobe studies discussed in the next two sections. However, before

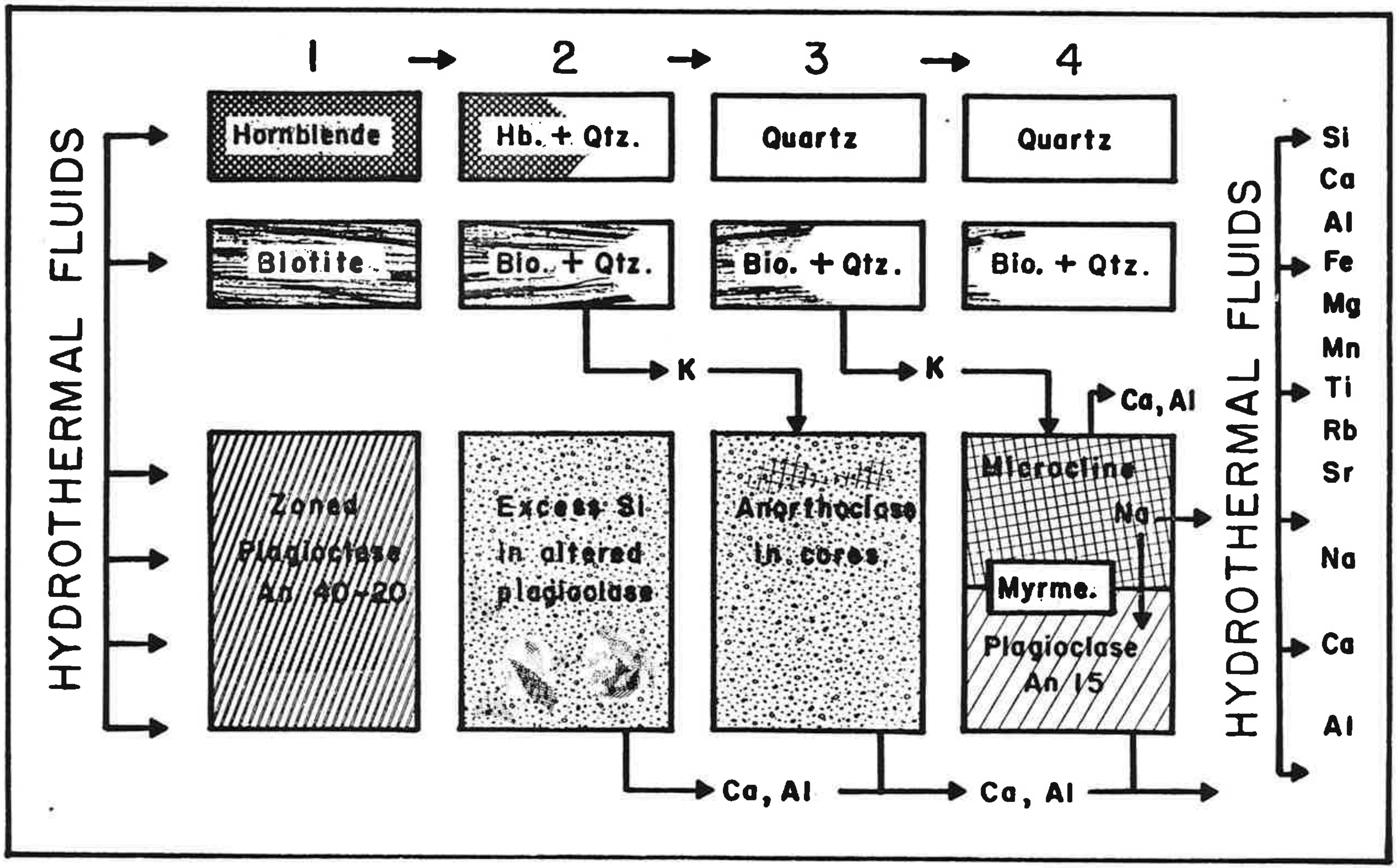


Figure 12. Schematic flow chart showing stages in the conversion of a biotite-hornblende tonalite (or diorite) in stage 1 to a myrmekite-bearing granite in stage 4.

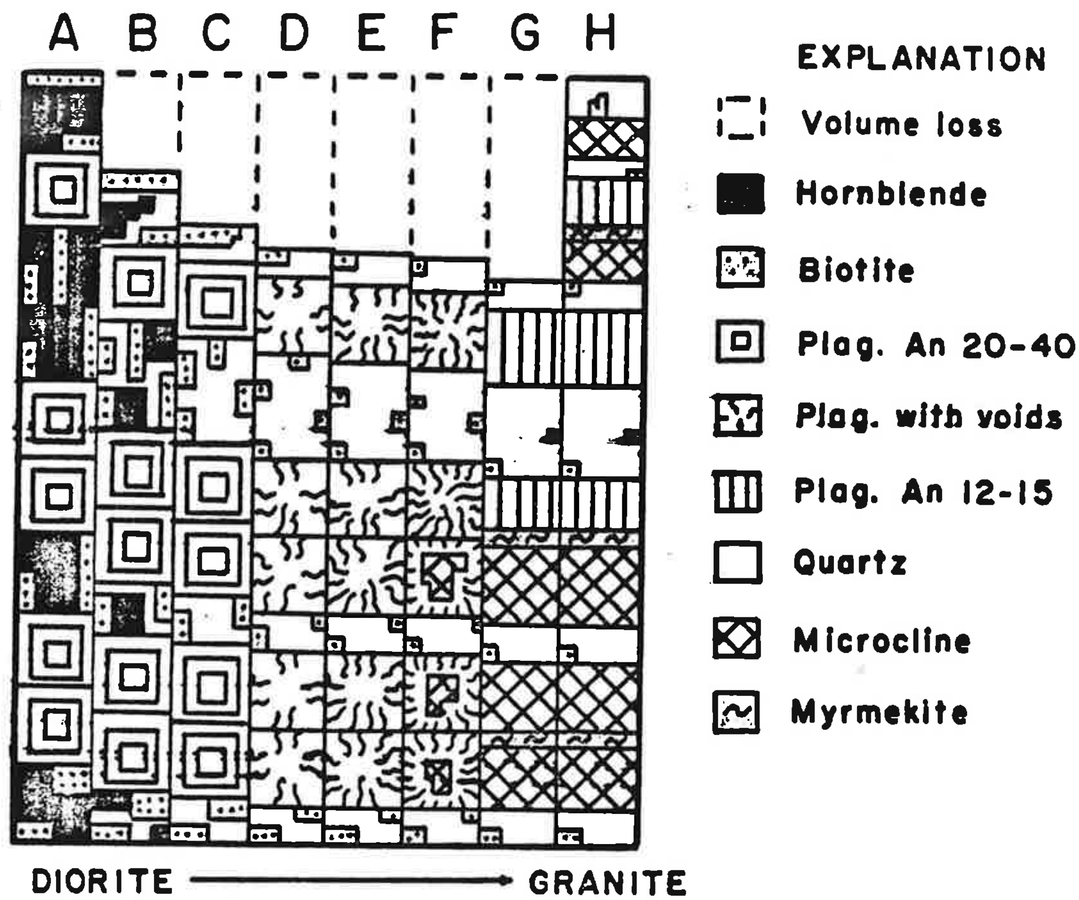
the chemical studies are considered, volume losses that accompany the modal changes shown on Figure 8 need to be estimated. As a first approximation Figure 13 A-H schematically shows relative volume losses that occur during the recrystallization of a theoretical diorite containing 35% hornblende, 15% biotite, and 50% plagioclase.

Figure 13-B is the first stage of recrystallization in which the bulk of the hornblende (25 of the 35%) and a small portion of the biotite (2 of the 15%) are recrystallized to form quartz. In that process a volume loss of 12.8% occurs as the result of subtraction of K, Ca, Al, Fe, Mg, and other elements from these mafic minerals. The rock is now a quartz diorite or biotite-hornblende tonalite.

In Figure 13-C all hornblende is destroyed to form quartz, and the original 15% biotite is reduced to 9%. This results in the quartz percentage increasing to 21.2% and the volume loss to 19.8%. Now the rock is a biotite tonalite.

Figure 13-DEF are transition rocks in which little outward change is apparent, but internally many changes take place inside the plagioclase crystals. In early stages K has entered the cores to form anorthoclase in a few places. In later stages, Figure 13-F, microcline makes its first appearance. Biotite is shown to be reduced to 3% and plagioclase to 45%, while quartz is increased to 23.7% and the volume loss to 23.3%.

Figure 13-G is the final stage. Here microcline has replaced about half of the original plagioclase grains, and myrmekite is abundant, either trapped between microcline grains or "growing on" the plagioclase. All plagioclase crystals that are unreplaced by micro-



	A	B	C	D	E	F	G	H
Volume loss	-	12.8	19.8	22.2	22.7	23.3	26.2	-
Hornblende	35.0	10.0	-	-	-	-	-	-
Biotite	15.0	13.0	9.0	5.0	4.0	3.0	2.0	2.7
Plag. An 20-40	50.0	50.0	50.0	-	-	-	-	-
Plag. with voids	-	-	-	50.0	50.0	45.0	-	-
Plag. An 12-15	-	-	-	-	-	-	16.0	21.7
Quartz	-	14.2	21.2	22.8	23.3	23.7	24.1	32.6
Microcline	-	-	-	-	-	5.0	27.6	37.4
Myrmekite	-	-	-	-	-	-	4.1	5.6
	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

Figure ~~15~~ 13.

cline have recrystallized as clear grains exhibiting albite twinning (see discussion in previous sections). In this final stage the original biotite is reduced to 2.0%, and quartz is increased to 24.1%. The final total volume loss is about 26.2%. The 26.2% volume loss has an important bearing on the development of a model for the origin of granitic diapirs discussed in a later section. Moreover, the volume loss is also significant in that when the volume of all remaining minerals is recalculated to 100%, shown in Figure 13-H, the result is a rock having the composition of a typical granite containing about 2.7% biotite, 32.6% quartz, 21.7% plagioclase, and 37.4% microcline. Therefore, if chemical analyses of this rock were made, these analyses would suggest that much Si, Na, and K were added to a tonalite during replacement, when in fact, the apparent introduction of Si, Na, and K is the result of a volume loss in the original rock because of the subtraction of Ca, Al, Mg, and Fe.

In later sections theoretical chemical analyses of stages A, B, C, D, and H, shown on Figure 13, are utilized for comparison with chemical analyses with other rock types.

A MODEL FOR THE ORIGIN OF MYRMEKITE

The conversion of biotite-hornblende tonalite to granite is diagrammatically summarized in Figure 16. The illustration simulates the mineralogical and textural changes that occur in the conversion. The cataclastic textures shown in the figure are simplified to show mottled extinction or breakage of grains, but bent twin lamellae, shear zones, and mortar texture are not illustrated although they occur in thin section. The decreasing heights of the diagrams symbolize relative volume losses that result because of subtraction of elements; see Figure ~~8~~ 13. If changes in the hornblende and biotite are disregarded

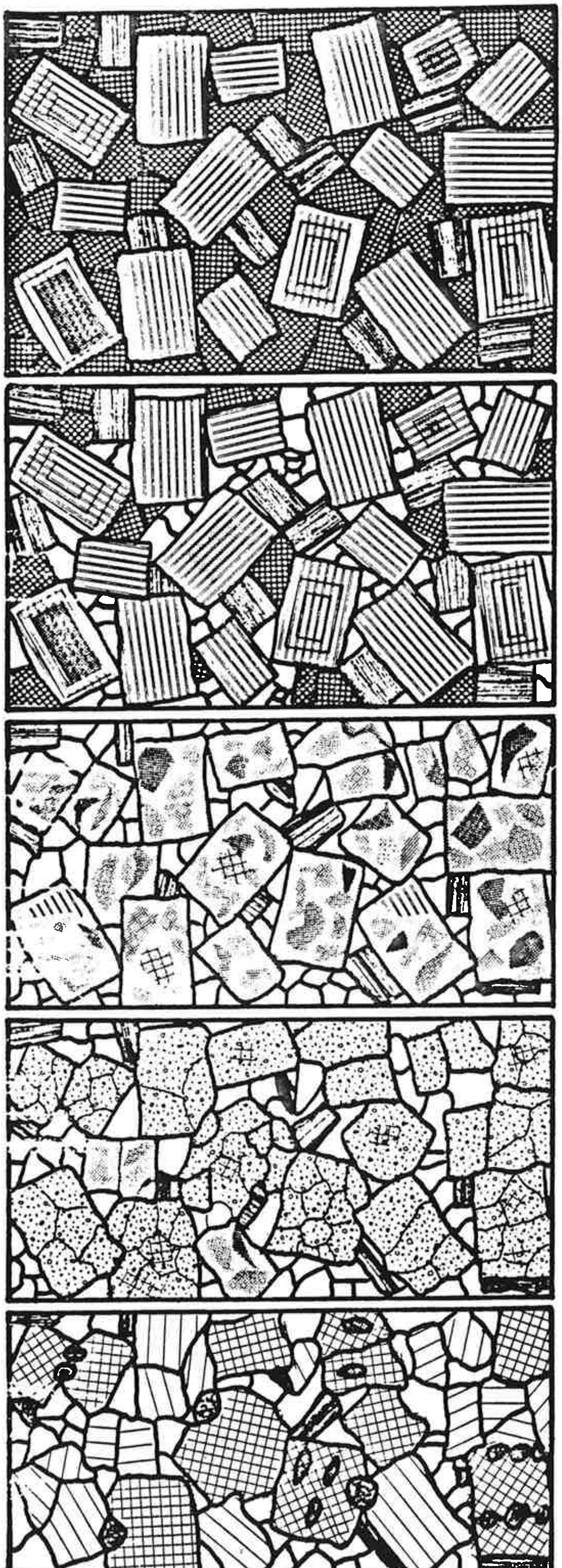


Figure 16.

1. DIORITE

Biotite

Hornblende

Plagioclase

zoned An_{40-20} & twinned

2. TONALITE

Partial conversion of biotite & hornbl. to quartz

Sieve textures may occur in mafic silicates

3. TRANSITION ZONE

Hornblende gone

Plag., mottled

Anorthoclase forms in plagioclase cores

Voids & excess silica form in all plagioclase grains

Anorthoclase & microcline form in plagioclase cores. Bio. decreased

4. GRANITE

Microcl. replaces plag. from inside out; myrmekite forms on rims

Remnant plag. forms perthite lamellae

Plag. converts to An_{15}

for the moment, the process of myrmekite formation is suggested to begin with cataclasis of the solid tonalite and with the introduction of hydrothermal fluids to remove Ca and Al along plagioclase grain boundaries. As Ca and Al are removed, the rim becomes slightly more sodic because of the relative increase of residual Na proportional to the loss of escaping Ca. The key factors in the whole process are that Na and Si are not added to replace Ca and Al and that sodic plagioclase has lower density (2.61 to 2.64 for An_{0-20}) than more calcic plagioclase (2.64 to 2.67 for An_{20-40}). Therefore, as the outer rim of a plagioclase grain in the transition zone loses Ca and Al, its structural framework tends to expand to form a less dense array. However, because the tonalite is solid (although cataclastically broken), most grains that are being altered are confined between other solid mineral grains, and therefore, expansion is narrowly limited or impossible. Because the volume of the original grain then is fixed to or nearly to its initial shape, the structural framework in the rim is fixed in volume. Moreover, the subtraction of Ca and Al from the rim does not cause the structural framework in the rim to shrink because Ca and Al from the core diffuses outward to the rim to replace the Ca and Al that have been lost to passing fluids. In this process the density and An content of the rim remain nearly constant, and therefore, because the volume also remains nearly constant, the grain is able to retain its rigidity and strength under confining pressures.

The above relationships and the process of converting a zoned plagioclase grain to the various stages of replacement and recrystallization are schematically summarized in Figure 17. In all stages

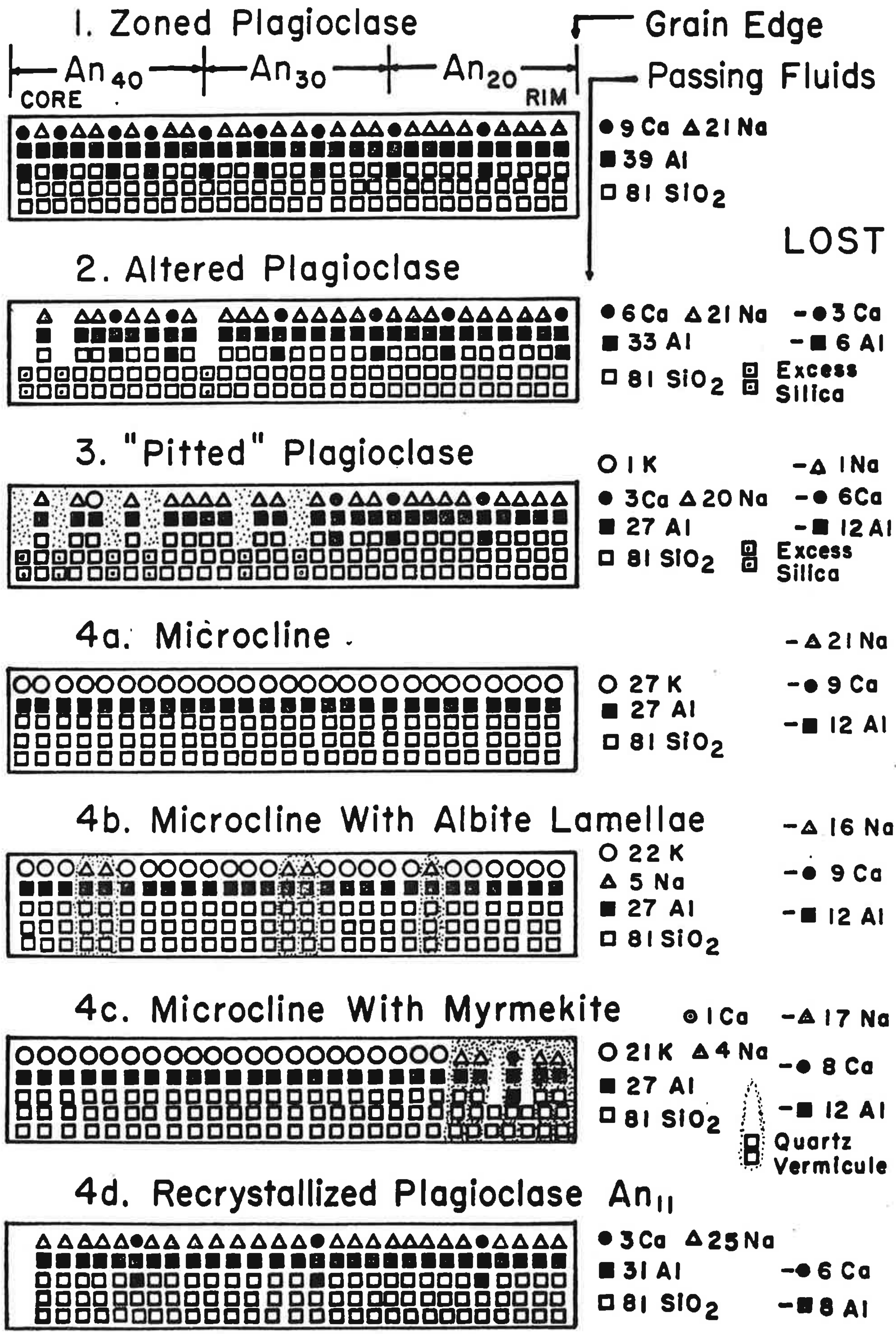

1. Zoned Plagioclase
Grain Edge
An40
An30
An20
CORE
RIM
Passing Fluids
9 Ca 21 Na
39 Al
81 SiO2
LOST
2. Altered Plagioclase
6 Ca 21 Na
33 Al
81 SiO2
-3 Ca
-6 Al
Excess Silica
3. "Pitted" Plagioclase
1 K
3Ca 20 Na
27 Al
81 SiO2
-1 Na
-6Ca
-12 Al
Excess Silica
4a. Microcline
27 K
27 Al
81 SiO2
-21 Na
-9 Ca
-12 Al
4b. Microcline With Albite Lamellae
22 K
5 Na
27 Al
81 SiO2
-16 Na
-9 Ca
-12 Al
4c. Microcline With Myrmekite
1 Ca
21 K 4 Na
27 Al
81 SiO2
-17 Na
-8 Ca
-12 Al
Quartz Vermicule
4d. Recrystallized Plagioclase An11
3 Ca 25 Na
31 Al
81 SiO2
-6 Ca
-8 Al

Figure 17.

silica is kept constant (81 SiO_2). Stage 1 at the top of the figure shows an original plagioclase grain with a core composition of An_{40} and the rim, An_{20}. Stage 2 shows the grain after partial extraction of Ca and Al. Because electron-microprobe data have shown that the more calcic core of a plagioclase grain becomes less Ca- and Al-rich in the transition zone, this suggests that Ca and Al in the core diffuse out to the rims to take the place of Ca and Al atoms that have been lost to the fluids moving along grain boundaries. During this stage uneven diffusion from place to place would destroy the albite twinning and zoning and cause mottled extinction (Figure 17, stage 2; and Figure 16, no. 3). As diffusion proceeds the An content of the core would be reduced gradually to equal the An_{20} composition of the rim (stage 2), but eventually diffusion outward of Ca and Al makes the interior more sodic (stage 3, core An_0, rim An_{20}). Therefore, the grain would show reverse zoning in later stages of alteration. As more and more Ca and Al are transferred from the core to the rim, the lattice structure of the interior would expand as the core becomes relatively more sodic, since sodic plagioclase is less dense than more calcic plagioclase. Here the expansion is possible because the disappearance of Ca and Al creates a volume loss, and the expansion is in the interior where space is being created and not along the exterior where adjacent solid grains and confining pressures make the expansion impossible. However, calculations show that the expansion is not sufficient to compensate for all the volume loss; therefore, tiny spherical holes are formed in the inner structural framework (stage 3, Figure 17; Figure 16, no. 4; and Figure 11. The loss of Ca and Al from the core also results in a small percentage of excess silica which probably initially is fixed in place as an intimate intergrowth with the now more sodic plagioclase (stage 3, Figure 17). This excess silica would not be detected optically as quartz in a separate phase because of its small quantity and being scattered throughout the plagioclase structure, but could be detected by an electron-microprobe since more SiO_2 would be recorded than would balance the Na_2O, CaO, and Al_2O_3 that remain. (A method of checking by x-ray diffraction whether this excess silica is quartz has not yet been devised.) Calculations show that more than enough excess silica is released in final stages to form the quartz vermicules in myrmekite so that no additional silica need be introduced from an outside source.

The diffusion outward of Ca and Al from the core to the rim seemingly should stop when both core and rim have the same An content because at that point a chemical gradient would no longer exist. However, the electron-microprobe studies show that Ca and Al continue to occur in less and less amounts in the cores (An_{2-8}) even though the rims are more calcic (An_{16-20}). Continued outward diffusion of Ca and Al from the core may occur because in the expanded lattice of the more sodic plagioclase structure in the core, Ca and excess Al are less stable, and therefore, these elements diffuse to the denser lattice structure of the more calcic plagioclase in the rim.

The fact that the plagioclase does not compensate for its loss of Ca and Al in the interior by an inward collapse of the structure is explained in part by the tendency of the interior framework to expand as it becomes more sodic and in part by an analogy with a geodesic dome. A building roof that is constructed as a geodesic dome has a network of polygonal forms, and this arrangement creates an arching, rigid structure that does not collapse even though no supporting columns underlie the dome. In like manner the rigid, polygonal array of Si-O and Al-O tetrahedral units in the rim, which stays at the same approximate An composition, prevents the collapse of the crystal into the core and provides the structural strength to preserve the shape of the original crystal.

Next K ions released from destroyed biotite grains reach those places in the tonalite where the plagioclase cores are filled with tiny spherical holes (Figure 17). (This process may be started in stage 3 where symbolically one K has been added to the core.) Four alternative replacement end-products are possible: microcline, microcline containing albite lamellae, myrmekite bordering microcline, or recrystallized plagioclase of low An content; see Figure 17, stages 4a, b, c, and d. The end-product that is formed is a function of the direction in which K enters the plagioclase core, the degree of alteration of the core, the quantity of K entering the core, the availability of escape routes for Ca, Al, and Na that are displaced by K, and the degree to which these elements are carried away before the escape routes are cut off.

Depending upon the above variables, some altered plagioclase grains may be entirely replaced from the inside out to form large "primary appearing" microcline grains (stage 4a, Figure 17). The "holes" in the altered cores (stage 3) provide ample room for the expanded lattice structure of microcline, unlike along the rims of an altered plagioclase grain where space for expansion is limited or not available. As microcline forms in a plagioclase core, the expanded structure in which the large K^{+1} ions are stable would tend to make the smaller Ca^{+2}, and Na^{+1}, and excess Al^{+3} ions unstable and force them to the margins of the plagioclase grain; if fluids are available to carry away all these elements, the whole grain may be completely replaced by microcline. (On Figure 17, stage 4a, less K (27) occurs than the sum of Na plus Ca (9 + 21) (stage 1) because microcline has a lower density (2.56) than plagioclase (2.61 to 2.63) and therefore, has an expanded lattice structure that nearly fills the same space for the same number of residual Si and Al atoms.)

If in stage 4a all 9 Ca and 21 Na atoms are displaced by K and if more than 12 Al atoms are also lost, then excess silica would remain (not shown on Figure 17). This could result in residual quartz blebs in the microcline structure as a fifth alternative end-product that is observed in the Woodson Mountain granite. These quartz blebs may appear to be unreplaced portions of quartz vermicules in myrmekite, but in actuality are formed separate from myrmekite.

The formation of residual albite lamellae (An_{0-5}) in microcline (perthite) would result where K enters the altered plagioclase grain from several directions and surrounds islands of plagioclase before all of the Na can escape. This should not be unexpected since the ionic charge of Na^{+1} is like K^{+1} and would be tolerated in the microcline structure.

Where cataclasis has broken large plagioclase grains into an aggregate mosaic of tiny, disoriented, altered grains, K can readily enter along the fractures and replace the plagioclase along many different avenues from different directions. If microcline surrounds these grains so that fluids do not have access to them, then residual Ca and Na cannot escape. As the mineral recrystallizes, excess silica relative to the concentration of Ca and Al atoms would form quartz vermicules and convert the tiny grains into myrmekite (stage 4c). Such aggregate myrmekite grains are most likely to occur between two centers of K-replacement because the outward advancing microcline formation (moving from inside a plagioclase core out toward the rim) would trap portions of former plagioclase rims between the two microcline grains and cut off escape routes for the Ca and Al. If fluids have access so that the broken plagioclase fragments can adjust their composition by losing Ca and Al and gaining Na, then the tiny plagioclase grains occur as quartz-free enclosures in the margins of the microcline rather than as myrmekite.

In stages 1, 2, and 3, Figure 17, it has been assumed that no Na has been carried away by fluids that are altering the tonalite. This is because the expanded plagioclase structure, which is formed where Ca and Al are removed, favors the retention of Na since sodic plagioclase is less dense than more calcic plagioclase. This assumption is consistent with a similar observation made by Ferry (1979, page 132). However, when K begins to replace the cores of altered plagioclase grains (stages 4a, 4b, and 4c), the Na that is initially retained is released for the first time. This Na from the plagioclase would behave chemically like K released from altered biotite and would move to other altered plagioclase grains that are filled with tiny holes (stage 3). Here, the Na could enter the structural framework and combine with the excess silica to convert these altered plagioclase grains to recrystallized albite-oligoclase. In Figure 17, stage 4d, the final composition is shown to be An_{11} if 3 Ca atoms remain, but it would be An_{14} if 4 Ca atoms remain, and these values approximate the observed An_{12-15} compositions measured by electron-microprobe studies. Altered plagioclase grains (stage 3) that have not lost as much Al relative to the amount of Ca that is lost may be more easily replaced by sodic plagioclase than by microcline since more Al (31 Al in stage 4d) is in the plagioclase per unit volume than in microcline (27 Al in stage 4a). However, as Na replaces cores of some grains and as K replaces cores of other grains, all grains lose their "holes" (stage 3) and recrystallize to the four optically clear products (stages 4a, b, c, and d) that are found in granite. Where this occurs, both microcline and plagioclase grains seem to be primary as if they crystallized from a magmatic rock. And some of these recrystallized plagioclase grains still enclose biotite xenocrysts from the tonalite, but the plagioclase grains enclosing this biotite now have a uniform An_{12-15} composition instead of the former zoning (cores An_{39}, rims An_{22}). Because in most places geologists have not observed stages 2 and 3 but have only seen rocks representing stage 1 and stages 4a, b, c, and d with a knife edge contact between them, geologists have been misled about the origin of myrmekite and its significance.

The sharp contacts between light-colored, myrmekite-bearing granitic rocks and older, darker, more mafic rocks occur because of contrasting degrees of shearing of the mafic rocks prior to replacement and recrystallization. The separation between crushed and uncrushed rock in fault zones is usually sharply delineated, and therefore, the recrystallization of the crushed mafic rocks to form myrmekitic granitic rocks leaves sharp contacts of light-colored rock against the uncrushed mafic rock. If the zone of crushed mafic rock cuts across the structure, the myrmekite-bearing granitic rock formed by recrystallization would also cut across the structure. For these reasons, sharp contacts, cross-cutting relationships, and contrasting mineralogies and chemical compositions need not indicate an intrusion of magma (see Roddick, 1982).

Still to be explained is the myrmekite partially enclosed by microcline grains while seeming to "grow on" large, adjacent plagioclase crystals. This happens where K has replaced altered plagioclase cores at one end of a crystal (stage 4a), and Na released from another site has replaced the altered plagioclase at the opposite end (stage 4d). As the K advances through the crystal, moving Ca, Na, and excess Al ahead of it, eventually empty sites into which the Ca, Na, and Al can move are plugged by Na moving into the crystal from the opposite direction. Where that occurs, the remaining excess silica that cannot be incorporated into the plagioclase structure would become quartz vermicules in myrmekite trapped between microcline and the recrystallized plagioclase grains (An_{12-15}). On Figure 17, stage 4c would be continuous with stage 4d but the recrystallized plagioclase would be placed so that the myrmekite would be between the microcline and the recrystallized plagioclase.

Widenfalk (1969) showed that the plagioclase portion of myrmekite has a higher An content than that in adjacent quartz-free plagioclase where the An content of the plagioclase in the myrmekite is greater than An_{15}. This is consistent with the model since Na entering from one side of an altered plagioclase crystal and K from the other is likely to trap more Ca in the recrystallized plagioclase of the myrmekite than in the quartz-free plagioclase. Moreover, the entering Na would also tend to reduce the An value in the quartz-free plagioclase below that in the adjacent myrmekite.

Not yet mentioned is a third type of myrmekite found in some granitic plutons. This is "rim myrmekite" that borders zoned plagioclase grains in granodiorite or diorite which contains only small percentages of interstitial microcline (Figure 18). Rim myrmekite occurs in the bulk of the Woodson Mountain granodiorite pluton east of the Santa Margarita River (Figure 5). The absence of granite here is probably because the original, unreplaced rock was diorite that was deficient in biotite instead of tonalite rich in biotite. The absence of large amounts of biotite in the diorite would eliminate a source of K needed to form the large modal volumes of microcline in granite. However, the Woodson Mountain granodiorite containing rim myrmekite is gradational to the granodiorite and granite containing "aggregate myrmekite" and "wart-like myrmekite" projecting into microcline but "growing on" plagioclase. Therefore, a different hypothesis is not needed to create rim myrmekite. The interstitial microcline would represent places where K has entered plagioclase rims that have been formerly altered and "pitted". As in Figure 17, the K would displace the residual Na, Ca, and Al ahead of the K until empty sites are full. The excess silica remaining in the altered but now recrystallized rims would form quartz vermicules against the relatively unaltered zoned plagioclase cores. The presence of preserved zoning in the plagioclase xenocrysts from the former diorite suggests that cataclasis was not as severe in the bulk of the Woodson Mountain granodiorite as in the border areas where biotite-rich tonalite was recrystallized as massive granite and dike-like granite bodies.

QUARTZ-PLAGIOCLASE CORRELATIONS IN MYRMEKITE

Although other investigators have not recognized the importance of volume-for-volume replacement relationships between microcline and plagioclase, discussed in the previous sections, they have attempted to explain the correlation between the volume of quartz vermicules and the An content of the plagioclase in myrmekite. In doing so, other investigators have used the following equations in which plagioclase (Na^{+1}, Ca^{+2}) is shown to replace microcline:

$$KAlSi_3O_8 + Na^{+1} = NaAlSi_3O_8 + K^{+1}$$

$$2KAlSi_3O_8 + Ca^{+2} = CaAl_2Si_2O_8 + 4SiO_2 + 2K^{+1}$$

microcline plagioclase quartz

myrmekite

However, as has been shown in the previous sections, microcline (K^{+1}) in many rocks clearly replaces plagioclase where myrmekite is found. But the above equations have caused other investigators to disregard the microcline replacement of plagioclase because the reverse of the equations shows that replacing calcic plagioclase by K^{+1} should consume quartz rather than produce it. In Figure 48 a single curve is drawn

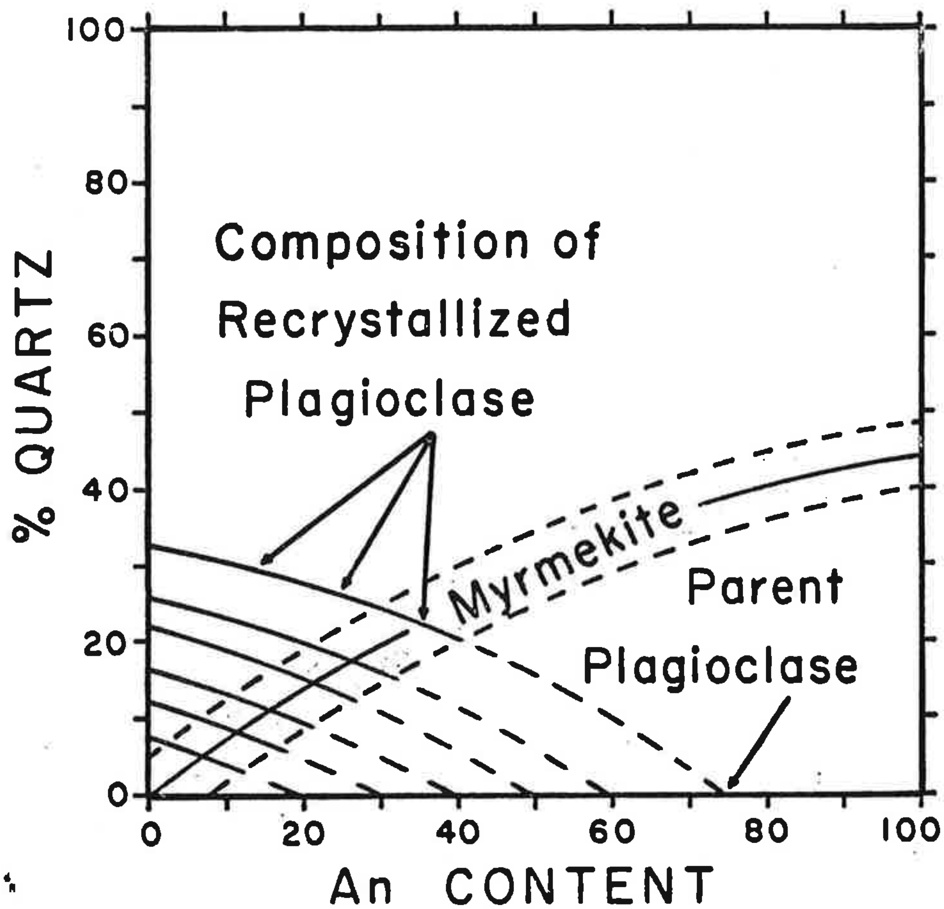


Figure 48. Volume percent quartz in vermicules of myrmekite plotted against An content of parent plagioclase of original rock and against An content of recrystallized plagioclase. Six curves in lower left trace decrease in An content of parent plagioclase as Ca and Al are removed to form recrystallized plagioclase of lower An content. The loss of Ca and Al is accompanied by an increase in excess silica which results in quartz vermicules forming in myrmekite; see Figure 17. Dashed lines from lower left to upper right show the field for quartz vermicules in recrystallized plagioclase for various myrmekite compositions. Compositions below this field do not exist while those above result in quartz-plagioclase textures that are discussed in a later section. Solid line in the myrmekite field traces compositions predicted for balanced chemical equations (see text).

(bounded by a field outlined by dashed lines, lower left to upper right) to illustrate the relationship between the volume of quartz produced and the An content of plagioclase on the basis of the above equations. If this theoretical curve is compared with measured values for myrmekite in different rock types, most data points plot above the curve, but some plot below it. The trend of plotted points, however, parallels the curve. That an exact correlation has not been found everywhere has been explained on the basis of errors in measurements and on the difficulties of observing myrmekite in the third dimension (Sederholm, 1916, page 137; Phillips, 1964; Phillips and Ransom, 1968; Ransom and Phillips, 1969; Shelley, 1969; Ashworth, 1972; D. S. Barker, 1970).

If the abundance of quartz in vermicules and the An content of the so-called "primary plagioclase" in the granite at Temecula are plotted on Figure 48, the measured values of An_{12-15} and 9 to 12% quartz plot just slightly above the curve. However, the plagioclase in the granite is not "primary" but is the secondary product of a hydrothermal replacement process in which zoned plagioclase grains, averaging An_{30}, have been recrystallized to An_{12-15}. What is apparent from this information is that the replacement process tends to form myrmekite which is associated with recrystallized plagioclase whose composition is about half the An content of the parent plagioclase.

To illustrate how the replacement process changes "parent plagioclase" to recrystallized plagioclase, six curves are drawn on Figure 48 which cross the single curve derived from the equations. The An value where each curve originates at the base of Figure 48 is the An content of the parent plagioclase. Each curve is constructed by determining the density of plagioclase at the parent An value, computing weights of elemental oxides in the parent, and then calculating what volumes of plagioclase at successively lower An contents can be made from the available oxides, keeping Na_2O constant and equal to the amount initially in the parent plagioclase. In each calculation CaO and Al_2O_3 are lost, and excess silica remains. The volume of quartz that can be formed from the excess silica is calculated and plotted against the An content of the associated recrystallized plagioclase. Smooth curves are then drawn between calculated points of measurement. The fact that measured values fall above or below the single curve probably results from differences in the amount of replacement by K_2O and differences in the degree to which CaO and Al_2O_3 are subtracted relative to the original Na_2O content of the parent plagioclase. Myrmekite then occurs in a field that extends from lower left to upper right rather than being restricted to the single curve that bisects the field.

No rocks are known to plot in the far lower right portion of Figure 48. This is because in most places the hydrothermal process that removes Ca and Al to produce residual silica goes beyond any minimum amounts of subtraction and because, once started, the alteration generally proceeds until K can be introduced. This occurs when the process has gone so far that the plagioclase grains recrystallize with An values about half that of the parent plagioclase.

Cargo Muchacho Mountains

In the Cargo Muchacho Mountains of southeastern California, two metasedimentary-appearing units have been mapped: (1) the Tumco formation containing a quartzose rock having the composition of a metagraywacke, and (2) the Vitrefax formation containing quartzite, sericite schist, kyanite-tourmaline schist, and amphibolite (Henshaw, 1942; Dillon, 1975). Examination of more than 350 thin sections, however, suggests that these formations are derived from the adjacent diorite and quartz monzonite plutons. Unsheared diorite and quartz monzonite can be traced through different zones of progressively greater degrees of cataclasis and recrystallization as follows. In early stages partially sheared diorite and quartz monzonite become myrmekite- and epidote-bearing. Then, in zones of increased shearing, primary K-feldspar crystals in the former quartz monzonite remain as augens in a foliated, cataclastic matrix of quartz, albite, epidote, myrmekite, and remnant mafic minerals. In zones of more intense alteration muscovite and quartz replace biotite, plagioclase, and K-feldspar to form muscovite-quartz phyllites. Finally, in areas of greatest alteration the phyllites are recrystallized as kyanite-bearing quartzites or pure quartzites. See Figures 52 and 53 for possible volumes of quartz and kyanite that can be formed from plagioclase.

In the quartzite, the kyanite forms radiating crystal sprays or occurs as druses in vugs. Dumortierite, black tourmaline, and magnetite are common accessories, but epidote is totally absent where kyanite forms. The kyanite-bearing zones represent places where nearly all Ca has been removed from the former diorite or quartz monzonite; however, a

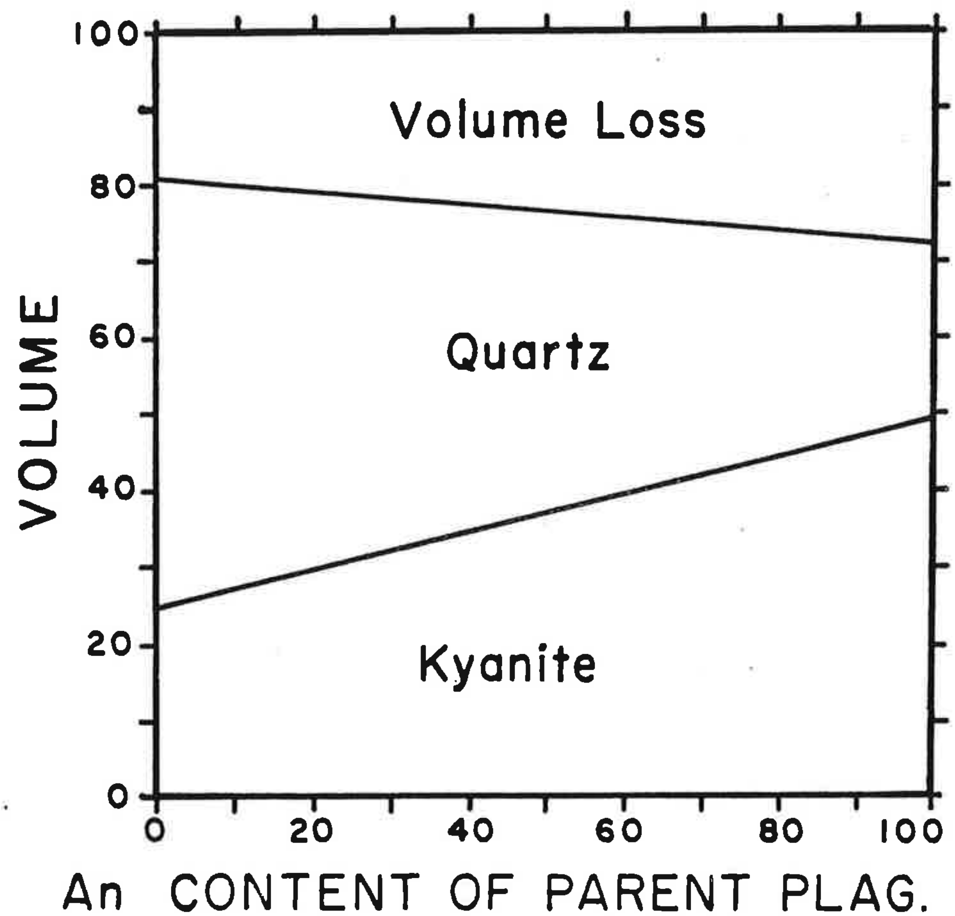


Figure 52. Plot of volumes in cc of quartz, kyanite, and volume loss that result from the conversion of 100 cc of parent plagioclase An_{0-100} and from the loss of Ca and Na during recrystallization.

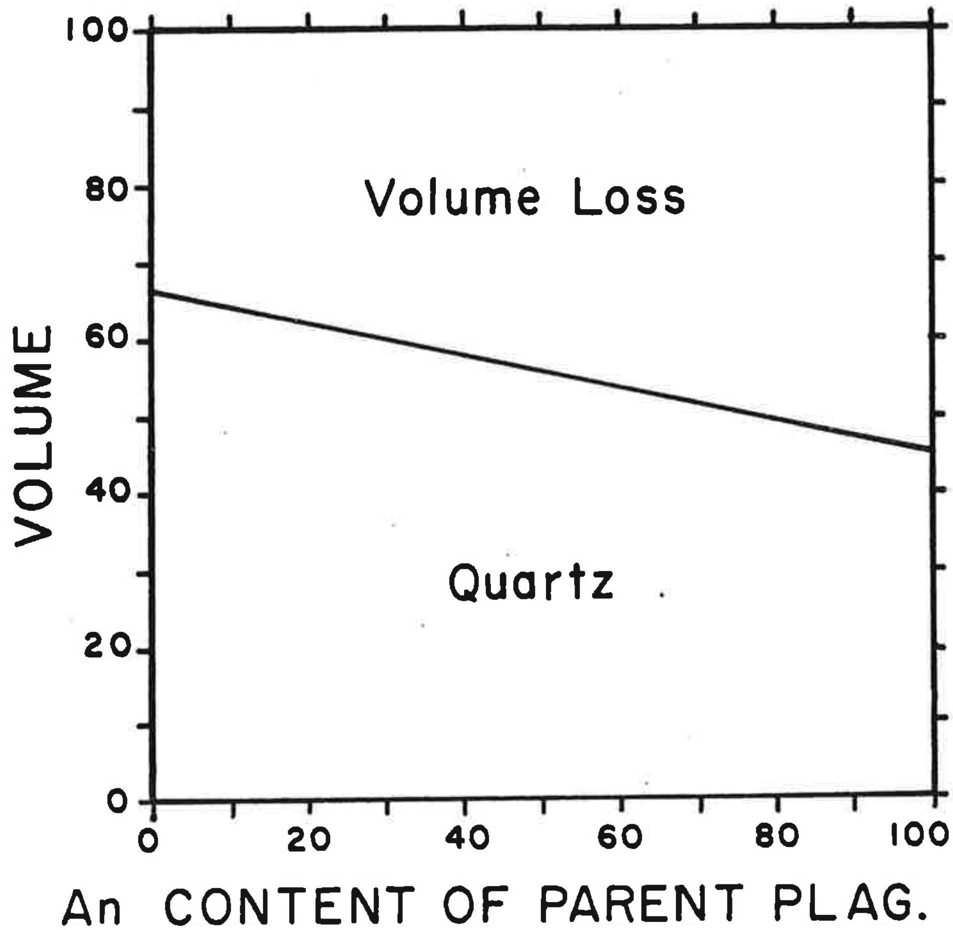


Figure 53. Plot of volumes in cc of quartz and volume loss that result from the conversion of 100 cc of parent plagioclase An_{0-100} as all Ca, Al, and Na in the parent plagioclase are lost during recrystallization.

small amount of Ca may remain in occasional concentrations of apatite. The abundance of kyanite in the quartzite unit (30 to 40%) is consistent for the amounts shown on Figure 52 for the recrystallization of parent plagioclase An_{30-40} in the former diorite and quartz monzonite.

The rocks in the Cargo Muchacho Mountains have undergone multiple episodes of faulting, shearing, crushing, and recrystallization. Figure 132 shows several major faults trending N74^{o}W that cut six different granitic rock types, but many more shear planes parallel to these faults are not shown. Moreover, these faults are representative of only the more recent fault activity because not illustrated are (1) fault systems trending N32^{o}W parallel to the fault west of Vitrefax Hill and parallel to the San Andreas fault and (2) three other sets of closely spaced but older, less distinct, fault systems that trend N15^{o}E, N45^{o}E, and nearly east-west. At the scale of the map, putting all of these additional fault systems in their proper place would make the map unreadable. In many places extreme cataclastic textures are still preserved, but in other places the recrystallization has eliminated most of the evidence for cataclasis. The granitic rocks 1 km east and northeast of Vitrefax Hill are examples of where extreme cataclastic crushing is preserved. Here the granitic rocks are so crushed that remnant K-feldspar crystals of the former quartz monzonite look like phenocrysts in a volcanic porphyry.

Almost all rocks in the Cargo Muchacho Mountains contain myrmekite, either rim myrmekite in less disturbed areas or well developed, wart-like myrmekite bordering K-feldspar grains in strongly recrystallized zones. Former plagioclase grains, if they once existed in former diorite or quartz monzonite, have been replaced by K-feldspar during the repeated episodes of shearing and hydrothermal differentiation.

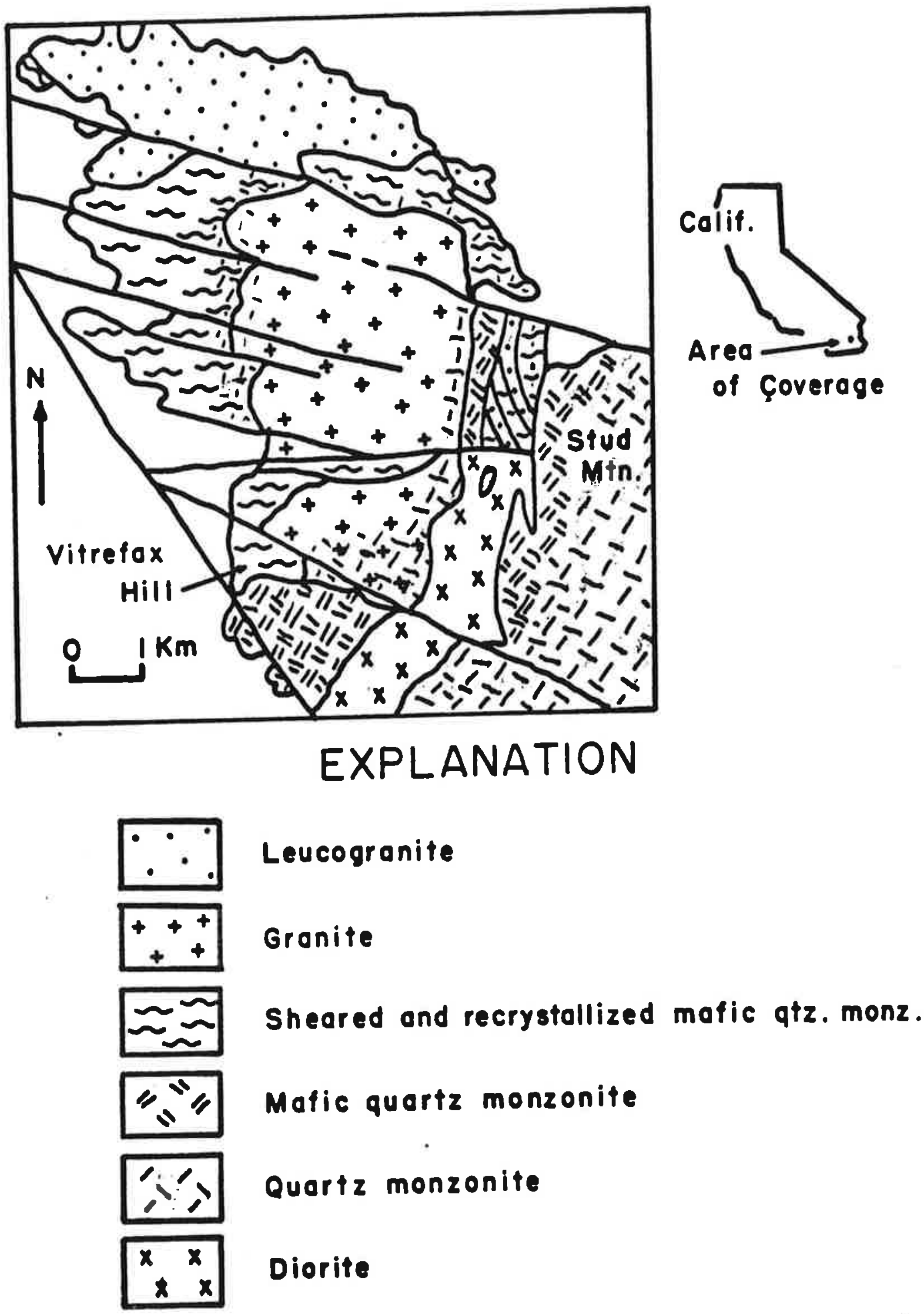
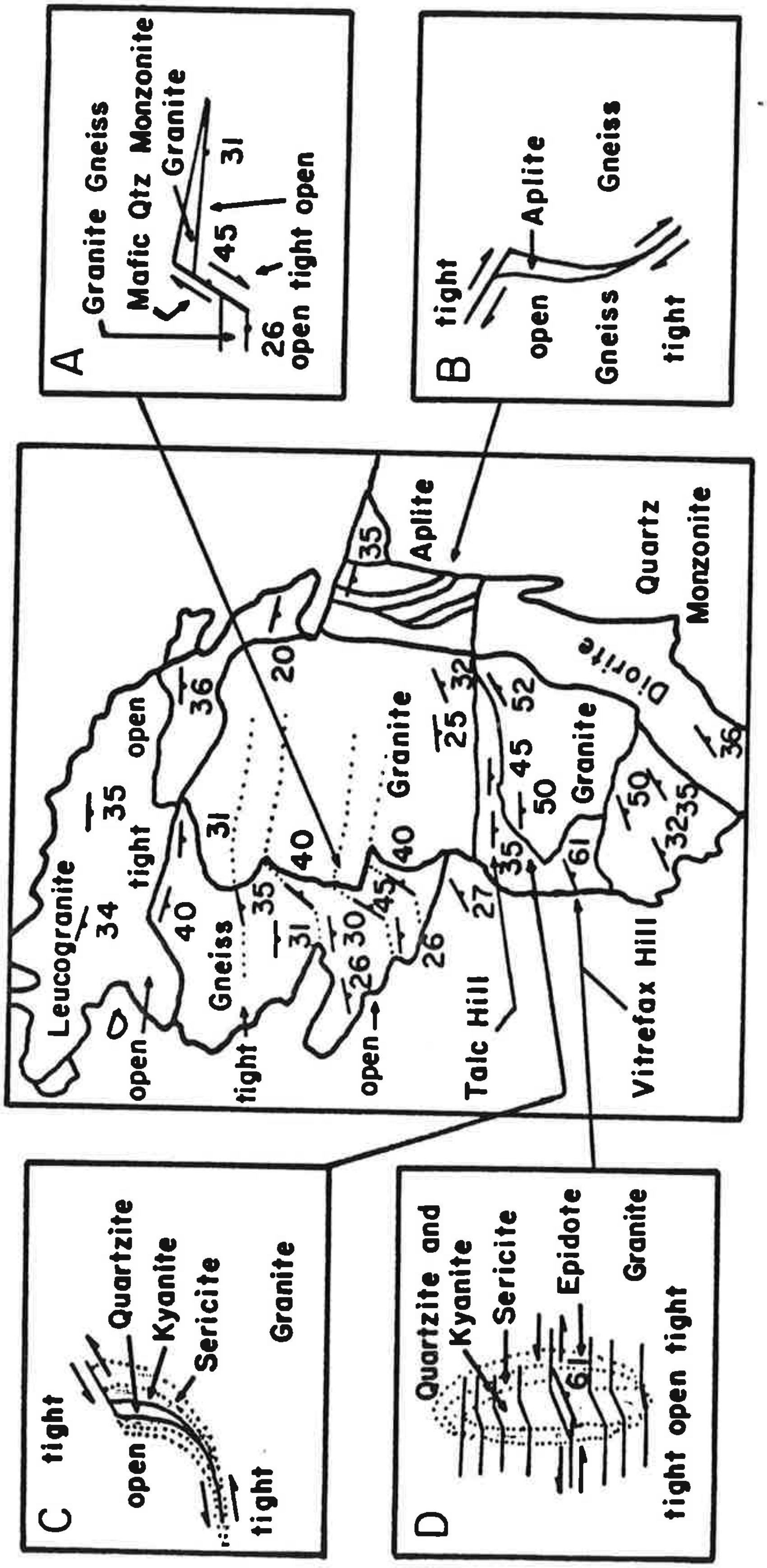


Figure 132. Generalized geologic map of a portion of the Cargo Muchacho Mountains. Henshaw (1945) mapped two metasedimentary rock units in this same area (the Vitrefax and Tumco formations), but these are interpreted as intensely sheared and recrystallized mafic border facies of a former quartz monzonite pluton.

The rocks labelled "sheared and recrystallized mafic quartz monzonite" on Figure 132 or "gneiss" on Figure 133 were mapped earlier by Henshaw (1942) as the metasedimentary Tumco and Vitrefax formations. Dillon (1975) later lumped them together as the same formation. After study of more than 350 thin sections in the area, it is clear that the Tumco and Vitrefax formations are not metasedimentary but modified plutonic igneous rocks that have been subjected to different degrees of cataclastic shearing, hydrothermal alteration, and recrystallization (see Myers, 1978). However, they are so well foliated in some places that they look "metasedimentary". Evidence for their non-sedimentary origin is discussed below.

The structural control for the hydrothermal differentiation is illustrated in four different localities (Figure 133ABCD). Figure 133A is an enlarged view of the structural relationships that occur between a granite pluton and the sheared and recrystallized quartz monzonite (the Tumco formation of Henshaw, 1942). The contact is sharp along its southwestern border and easily mapped because of the contrasting light color of the granite and its abrupt change to the dark, mafic, well-foliated quartz monzonite gneiss. However, minerallogically the two rocks are essentially the same except for their hornblende, biotite, and quartz contents. As hornblende and biotite decrease, quartz increases. Figure 134 shows how the modal compositions change from west to east, from the gneiss into the granite. In the western part, the foliation strikes nearly east-west and dips shallowly 21 to 31^{o} to the south. Eastward, as the foliation pivots to a N45^{o}E strike adjacent to the granite pluton, the dip increases to 40 to 45^{o}, the biotite and hornblende con-



A
Granite Gneiss
Mafic Qtz Monzonite
Granite
45
31
26
open tight open
B
tight
open
Aplite
Gneiss
Gneiss
tight
Leucogranite
34
open
tight
35
open
36
20
40
31
Gneiss
tight
35
31
30
26
40
45
40
Granite
25
32
52
45
50
Granite
61
50
32
35
36
27
26
open
Talc Hill
Vitrefax Hill
35
Aplite
Quartz Monzonite
Diorite
C
tight
open
tight
Quartzite
Kyanite
Sericite
Granite
D
Quartzite and Kyanite
Sericite
Epidote
Granite
61
tight open tight

Figure 133.

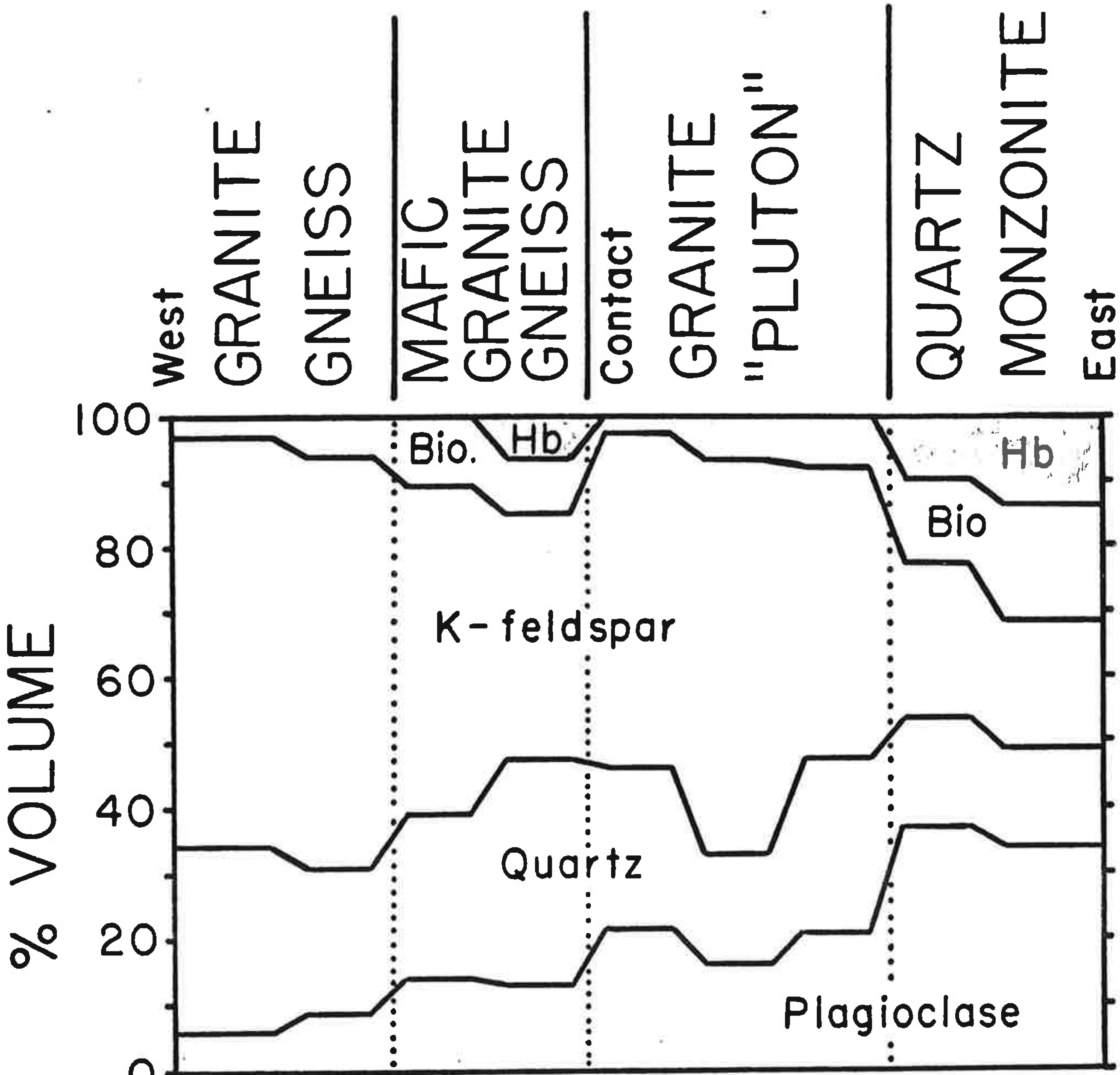


Figure 134. Schematic diagram showing modal changes in composition from west to east across: (1) the granite gneiss (Tumco Formation of Henshaw, 1942), (2) the mafic granite gneiss (mafic quartz monzonite at the contact), (3) the granite pluton, and (4) the quartz monzonite at the eastern border of the granite pluton. See area west and east of Figure 80 A.

tents increase, and the rock becomes progressively a dark, mafic quartz monzonite. Adjacent to the granite contact where the foliation of the gneiss has the maximum deviation in the "S-curve", the hornblende and biotite also have the greatest abundance. Then at the contact, but in the granite, the foliation makes an abrupt swing back to an east-west direction, hornblende nearly completely disappears, and biotite is markedly diminished in abundance. The granite here is massive and myrmekitic. However, progressively eastward a faint foliation and remnant cataclastic textures appear. At the eastern border of the pluton hornblende reappears, and both hornblende and biotite increase until the rock becomes a mafic quartz monzonite.

Northward along the western "contact" of the pluton, the "S-curve" flattens until the strike of the foliation in the "wall rock" gneiss equals the strike of the foliation in the granite. As the foliation in the bordering gneiss becomes more nearly parallel to that of the granite, the granite gradually loses its massive appearance and progressively contains more biotite and hornblende. Eventually, northward where the strike of the foliation of both rocks are the same, the granite grades into the mafic quartz monzonite. From the compositional changes shown at the western contact, it is apparent that the greatest deviation in the "S-curve" is the tightest part of the structure. The granite pluton formed in the most "open" part of the structure where fluids subtracted Ca, Al, Mg, and Fe to the greatest degree from a former mafic quartz monzonite. In some places the recrystallization (in the solid state) has been so great that evidence for former cataclasis has been destroyed, or else where myrmekite is absent, the chemical composition has shifted to granite and temperatures have been high enough that melting and recrystallization have destroyed the cataclastic textures and myrmekite.

In the north and northeastern part of the Cargo Muchacho Mountains the leucogranite shows a similar gradational relationship with the foliated mafic quartz monzonite. Here the foliation gradually shifts from a southwest strike to a northwest strike. In the northwestern part, the foliation strikes N70^{o}W; in the north central part, the strike is nearly east-west; and in the northeastern part, the foliation swings back to a N70^{o}W direction. The central part is "tight" and both extremities of the leucogranite are "open" so that along strike as the foliation shifts to N70^{o}W the hornblende and biotite in the quartz monzonite gradationally disappear ("feather-out"). As the hornblende and biotite disappear, residual magnetite increases in abundance. Interfingering bands of leucogranite in the quartz monzonite increase in abundance near the "open" areas and have sharp contacts against the quartz monzonite (across strike). Myrmekite is abundant in the leucogranite, and garnet occurs in trace amounts in coarser pegmatite facies. The hydrothermal differentiation seems to occur as the result of fluids penetrating along foliation planes in the "open" areas of the "S-curve".

On the eastern side of the granite pluton, Figure 133B, two leucogranite aplite zones (30 to 50 m wide and more than 1 km long) occur with sharp borders against dark, foliated or massive, mafic quartz monzonite. Southward where their trends swing from a north-south stike to a northwest-southeast strike, the easternmost aplite narrows, pinches out upward, and plunges under the quartz monzonite rocks in Stud Mountain

(Figure 32). The westernmost aplite can be traced gradationally southward into quartz monzonite in a "feathered" relationship. In the transition to the aplite, hornblende and biotite in the quartz monzonite disappear through a zone of cataclasis. In the aplite the recrystallization and annealing of the former crushed granitic residue eliminates the evidence of the former cataclasis. In the adjacent sheared mafic quartz monzonite (the Tumco formation of Henshaw, 1942) remnant K-feldspar grains appear to be augen metacrysts in a well-foliated gneiss, but both northwestward and southeastward the well-foliated augen gneiss can be traced gradationally into massive mafic quartz monzonite containing occasional angular xenoliths of diorite. Myrmekite is common in the well-foliated gneiss as are cataclastic textures. The cataclasis is so strong in some places that the diorite xenoliths are drawn out into flattened sheaths, and the former coarse-grained texture of the mafic quartz monzonite (crystals as much as 2 cm across) become a fine-grained texture (0.5 to 1 mm across) with occasional remnant augens of K-feldspar (1 cm across); see also Myers (1978).

The aplite zones are interpreted to be low pressure sites in which the structure was more open so that aqueous fluids could subtract the Ca, Al, Fe, and Mg to a greater extent than in the adjacent more mafic, foliated gneiss. In the gneiss remnant biotite and hornblende grains are crushed, drawn out into planar fragments, and partially converted to residual quartz; and plagioclase and K-feldspar are commonly filled with tiny epidote crystals.

In addition to the two large aplite dikes, hundreds of smaller aplite and pegmatite dikes cut through the massive and well-foliated mafic quartz monzonite that borders the granite pluton, and these dikes represent another episode of localized hydrothermal differentiation (see Figure 135). Because exposures are excellent, pegmatites can be traced for several hundreds of meters from regions in the north in mafic quartz monzonite where they are well developed and thoroughly interlaced to regions in the south where they are scarce and in the process of formation (in the diorite south of the two aplite dikes, Figure 133B). Where pegmatites are in early stages of formation, thin sections show that the host rock is cataclastically granulated. In the transition zone hornblende forms a quartz sieve texture and then disappears, leaving only residual quartz as interstitial grains between broken feldspar fragments. Biotite also diminishes in abundance forming additional quartz. Potassium, released from the recrystallized biotite, produces additional microcline in altered plagioclase grains as myrmekite forms. Dusky, primary, Carlsbad-twinned "phenocrysts" of orthoclase or microcline, containing thin plagioclase lamellae, are also sheared and become recrystallized as clear grains with distinct cross-hatch twinning.

In the field these proto-pegmatite zones have a grey color. Where they were more intensely altered, introduced hydrothermal fluids removed soluble elements, caused annealing and recrystallization, and converted the "grey zones" to coarser, cream-colored pegmatites and aplites. Although these granitic dikes have sharp contacts agains the dark gneissic quartz monzonite (or diorite), in no place are there any rotated blocks of the quartz monzonite or diorite wall rock in the pegmatites that would indicate forceful injection of granitic magma from an outside source (see Figure 136 and also Ramberg, 1952, page 258). Where the dikes cut through the foliation, large feldspar crystals (2 to 4 cm long) in the pegmatites can be seen in the field to project into the wall rock and be parallel and similar in size and shape to feldspar grains in the quartz monzonite a few centimeters away from the contact. However, the mafic material of the wall rock almost abruptly ceases along a sharp "contact" line (Figures 136 and 137). In such places remnant traces of sheared wall rock fragments containing the mafic minerals may occur as islands in the pegmatite, but these mafic islands still are oriented parallel to the structure of the mafic minerals in the adjacent wall rock; see Figure 138. Where the pegmatites are parallel to the foliation of the wall rock, thin wisps of the wall rock can be traced into the pegmatite where the dark minerals feather to a thin line and disappear (Figure 138).

The locations of the pegmatites and aplites are structurally controlled by the degree of hydrothermal differentiation of the host rock. The pegmatites and aplites are absent in the more massive and feldspathic parts of the pluton but are frequent where the pluton once contained abundant mafic constituents. The space for the pegmatites and aplites was created by the differential shrinkage of the more mafic parts of the quartz monzonite pluton in comparison to the lesser amounts of shrinkage in the more feldspathic parts. This shrinkage would result from the general and greater pervasive loss of Ca, Al, Fe, and Mg in the mafic-rich areas where myrmekite formed and where the mafic silicates were converted to residual quartz and magnetite.

Near the fault contact between the granite pluton and the "sheared" and recrystallized mafic quartz monzonite" southeast of Talc Hill is a vertical quartzite unit (10 to 30 m wide and several 100 meters long); Figure 133C. Bordering the quartzite on both sides is a kyanite-rich zone that may also contain abundant tourmaline (schorlite) and trace amounts of magnetite and apatite. Adjacent to the kyanite-bearing zone is a sericite schist which locally contains rare andalusite crystals as much as 2.5 cm long.

The presence of both kyanite and andalusite in close proximity in the sericite schist suggests that the temperature of formation is less than 500^{0} C; see Figure 139. This is the same approximate upper limit for the temperature of formation of myrmekite that has been determined from other geologic thermometers in other geologic terranes; see previous sections.

The andalusite-bearing sericite schist grades laterally eastward into less altered rocks in which remnant K-feldspar augens are still preserved of the same size and shape as the K-feldspar crystals in the adjacent granite across the fault contact. On the west side of the quartzite the sericite schist grades laterally into biotite granite gneiss or hornblende diorite that locally contains abundant tourmaline. The sericite schist zones are wider on the eastern side than the western side of the quartzite. The granitic gneiss to the west and the granite to the east contain myrmekite and epidote. To the south where the fault curves westward, the quartzite thins and disappears, and then the kyanite-bearing envelop also disappears. Still further southwest the outer sericite schist envelop diminishes in thickness, but a thin, poorly sericitized zone can be traced in the massive granite for several 100

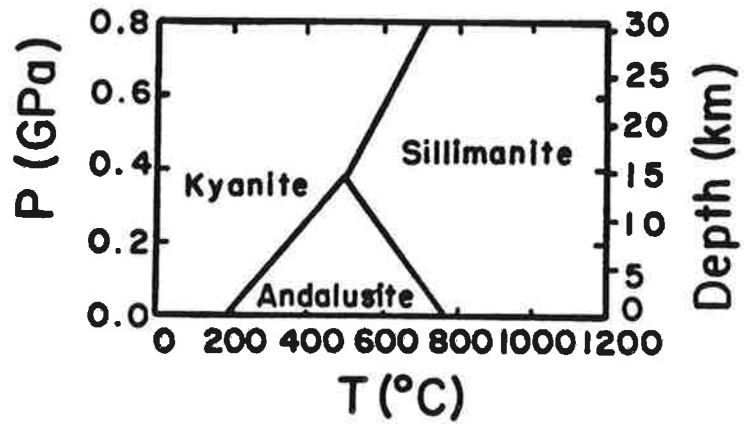


Figure 139. Al_2SiO_5 stability diagram (modified after Holdaway, 1971).

meters to the southwest where it disappears under stream alluvium. It is clear that the sericite schist is not a metamorphosed pelitic sediment but rather the result of hydrothermal alteration of the granite.

The quartzite has a sugary texture and outwardly looks like a metasediment. Dillon (1975) thought that he could recognize remnant cross-bedding in some places. However, thin sections show no remnant sand grains with secondary quartz overgrowths. Gradationally from the sericite schist to the quartzite, the quartz grains first occur as tiny grains in the interstices; then they become larger and partially enclose the sericite; then they become still larger (1 to 2 mm in diameter) and engulf remnant sericite. Finally they become clear quartz grains in pure quartzite or are interstitial to kyanite.

In some places kyanite crystals project into cavities in the quartzite which clearly indicates that the quartzite as well as the sericite schists are the result of hydrothermal alteration of the granite rather than the result of metamorphism of sandy and pelitic sediments. Kyanite formed by regional metamorphism does not occur in cavities since the high load pressure would not permit openings to occur.

On the north side of Talc Hill (named for the talc-like softness of the sericite schist), the sericite schist can be traced along strike eastward into biotite augen (K-feldspar) gneiss and then into more massive granite. In the granite the mineral grains that surround the K-feldspar augens are less cataclastically broken, and the K-feldspar grains do not stand out as augens. Thin sections show that muscovite (sericite) replaces the biotite in the transition between the sericite schist and the granite.

The quartzite unit is interpreted to be the most open part of the structure where hydrothermal fluids came through a fault zone and subtracted all elements except SiO_2. In the transition where K-feldspar augens also break down to form sericite or kyanite, micrographic intergrowths of quartz with angular or oval shapes appear in the K-feldspar grains.

In some places the transition between quartzite layers and granite is a sharp contact without any intervening kyanite or sericite schist zone. But thin sections across the granite-quartzite contact show that the feldspars in the granite exhibit micrographic textures and are in the process of breaking down to form quartz.

Isolated quartzite "dikes" (10 cm to 1 m wide) in the granite are not massive, milky "bull" quartz (although "bull" quartz veins locally cut the granite). Instead they have the same "sugary" appearance just like the larger quartzite masses that are surrounded by kyanite or sericite schist. These isolated quartzite "dikes" are interpreted to be narrow fault zones in which hydrous fluids have locally altered the granite so intensely that only quartz is left as a residue. In some places these isolated quartzite zones grade into areas where the quartz grains contain sericite inclusions like those found near the sericite schists. The local presence of tourmaline suggests that boron may be an essential consituent of the fluids for the alteration to be so extreme as to form quartzite from the granite.

A similar alteration zone occurs at Vitrefax Hill; the structural control to the alteration here is suggested in Figure 133D. As in Figure 133C, the alteration envelopes surrounding the central quartzite are asymmetric; the sericite schist zone is wider on the east side than on the west side. However, the kyanite-rich zones are not restricted to an outer envelop around the quartzite but are interlayered with the quartzite in the more "open" part of the structure. On the east side of Vitrefax Hill the sericite schist grades into a sericitized granite and then outward into an epidotized granite. The epidote content progressively increases to as much as 40 percent of the rock before decreasing again to 3 to 10 percent in less altered, myrmekite-bearing granite and quartz monzonite that surrounds Vitrefax Hill. Figure 140 shows the modal changes in composition from the quartzite core through the kyanite, sericite, and epidote zones into the granite.

Mafic quartz monzonite 100 m west of Vitrefax Hill locally contains crushed zones that are weakly sericitized and in early stages of hydrothermal alteration. Locallized veins of magnetite and muscovite indicate pathways along which fluids subtracted the mafic constituents (Ca, Al, Fe, and Mg). In other parts of the Cargo Muchacho Mountains similar magnetite veins contain epidote concentrations as well as traces of copper sulfides. In still other places the veins contain quartz, epidote, pyrite, gold, and scheelite.

On Vitrefax Hill locally staurolite and garnet as well as biotite and garnet were found in the sericite schist. This means that all compositional fields of the AKF diagram are preserved in the envelop surrounding the central quartzite unit; see Figure141. However, these are not fields representing primary sediments of different compositions but fields representing different stages of hydrothermal alteration of the biotite- and hornblende-bearing granite and mafic quartz monzonite as

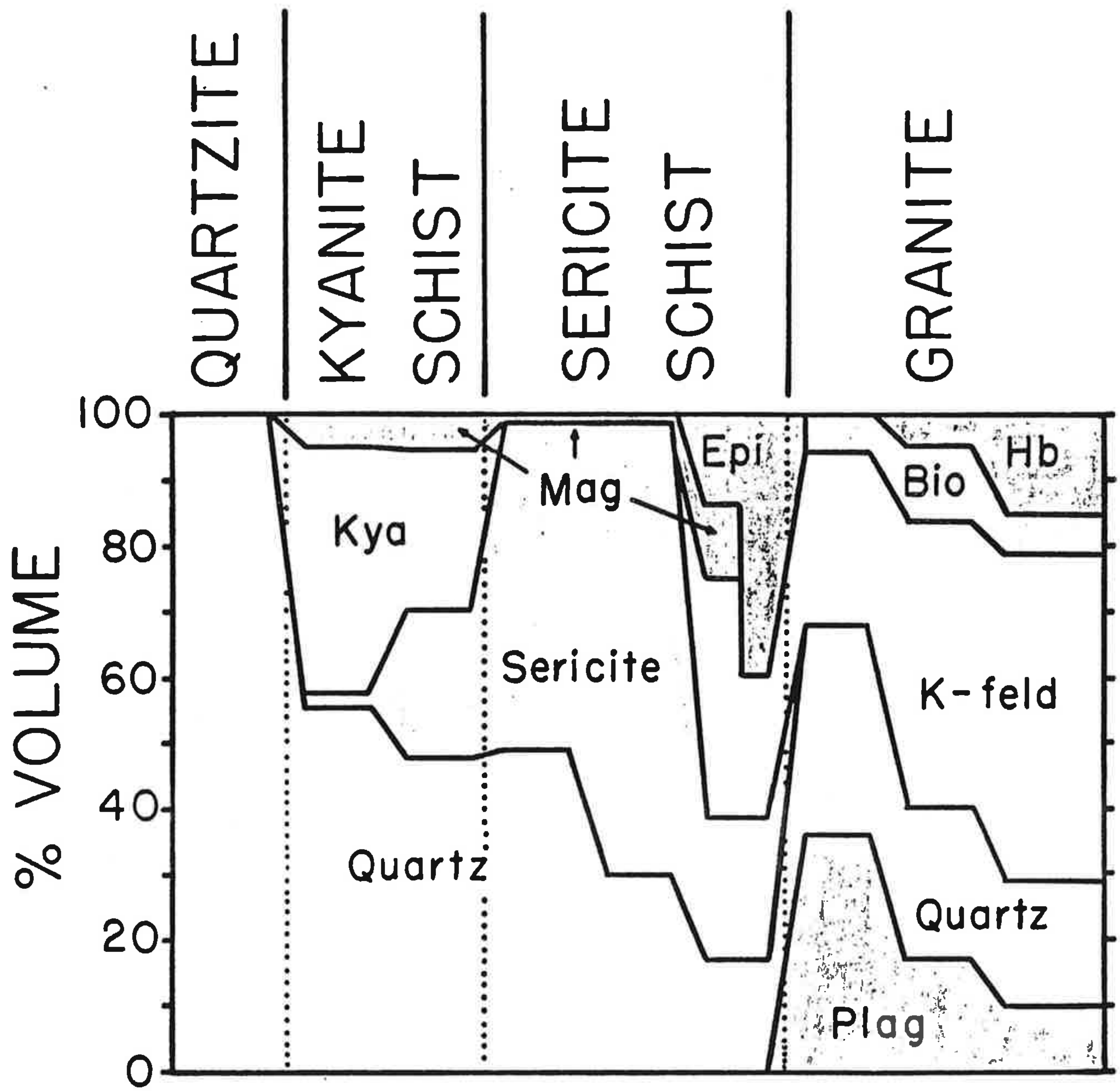


Figure 14Q. Schematic diagram showing modal changes in composition in zones occurring on and adjacent to Vitrefax Hill and extending from the quartzite and kyanite-rich core eastward into granite; see enlarged sketch, Figure 80 D.

the compositions of these parent rocks shifted from the K-F side to the A-corner of the AKF diagram; see arrows on Figure 141.

The kyanite deposits in the Cargo Muchacho Mountains represent an area where the conversion of granite and quartz monzonite to a quartz-rich peraluminous rock is locally intensified. The common association of tourmaline with these aluminum-rich rocks, and the simultaneous occurrence of nearby myrmekite-bearing rocks suggest that boron is a necessary ingredient in the hydrothermal fluids to form these strongly peraluminous rocks. This observation is supported by the fact that tourmaline is also found (1) in the quartz-rich pegmatitic zones adjacent to the hydrothermally altered, myrmekite-bearing gabbro and diorite in the Lake Isabella pluton, (2) in the myrmekite-bearing gabbro and pegmatites in the Pala area, (3) in the myrmekite-bearing, kyanite- or sillimanite-bearing Wissahickon schist in Pennsylvania (Wyckoff, 1952; Postel and Adelhelm, 1944), and (4) in the peraluminous metamorphic rocks surrounding the myrmekite-bearing Cooma granite in Australia (Pidgeon and Compston, 1965).

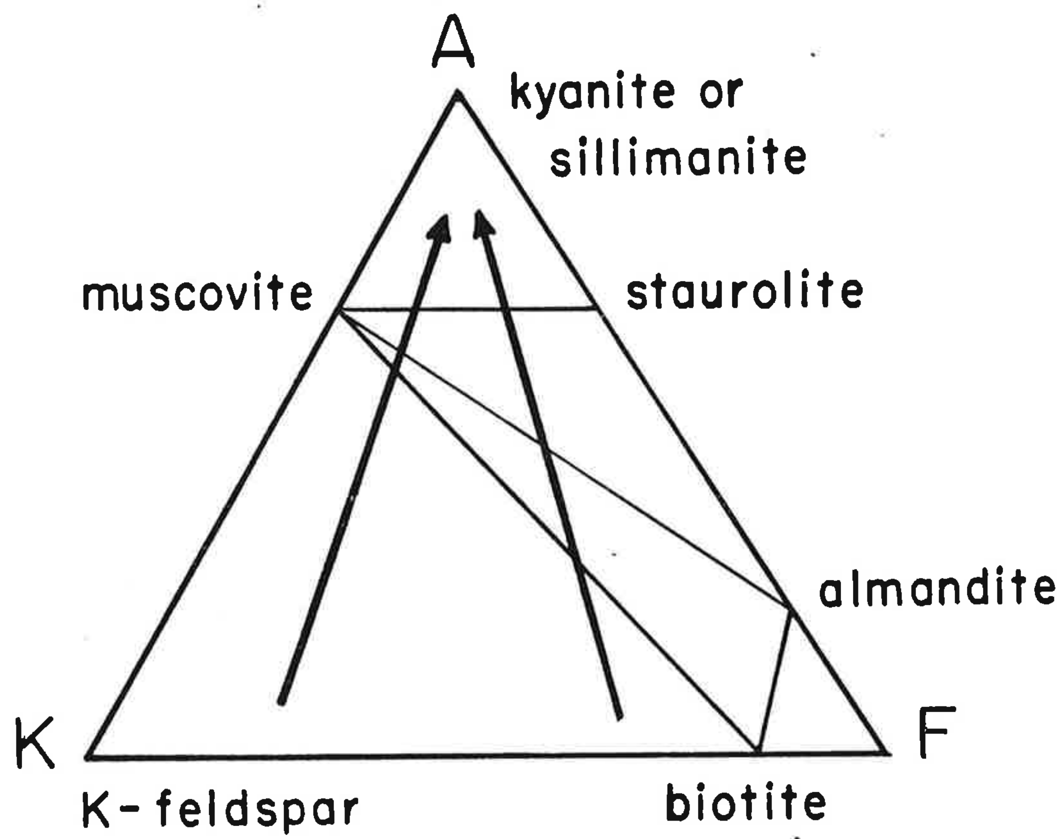


Figure 141. AKF diagram showing schematically how hydrothermal differentiation modified the composition of a biotite granite in the Cargo Muchacho Mountains to form kyanite-rich quartzites. Intermediate sericite, sericite-garnet, and sericite-garnet-staurolite rocks are locally preserved. The kyanite- or sillimanite-bearing Wissahickon schist in Pennsylvania may have been formed by the same kinds of compositional changes rather than by the metamorphism of pelitic sediments since the Wissahickon schist is tourmaline and myrmekite-bearing.

Pala Area

The geology of the Pala pegmatite area has been described by Jahns and Wright (1951) in the following manner (as extracted in excerpts):

Metamorphic Rocks

"The metamorphic rocks of the Pala district are thin, elongate remnants of a once extensive, sedimentary terrane. The remnants consist of quartzite, quartz conglomerate, meta-arkose, quartz-mica schist, quartz-mica-amphibole schist, and feldspathic amphibole schist, and are now transected and enclosed by younger igneous rocks. These metamorphic rocks are most abundant in a discontinuous, broadly curving belt" that "...separates gabbroic rocks on the south from dominantly granodioritic rocks on the north." "...Quartzite is dominant in the western part of this belt, schist in the eastern part." "...Bedding and schistosity in the quartzitic rocks are essentially parallel. On a larger scale, the trends of schistosity, of individual beds of quartzite, and of masses of schist and other metamorphic rocks also are parallel. Linear elements, mainly rows of biotite blades, amphibole needles, or stretched pebbles in conglomerate, are locally very well developed. Their plunges are generally very steep. The rocks are tightly folded on a small scale, as shown in many outcrops, but their broad structure appears to be simple."

Igneous Rocks

Gabbroic Rocks.

"Gabbro, norite, and associated intrusive rocks of basic composiition are very abundant in the vicinity of Pala, where they occur as composite plutons of circular to roughly elliptical plan. "...Individual rock types include medium- to coarse-grained gabbro and norite, olivine gabbro and norite, hornblende-rich gabbro and norite, hornblende-biotite gabbro and norite, and quartz-bearing gabbro and norite." "...The most common rock

PALA AREA


Granodiorite
Kqb
Qal
Kqd
Leuco-granodiorite
Kqm
PKs
Gabbro
Metasedimentary Rocks
Salmons Can.
Chief Mountain
Kqp
Kt
Mc Gee Can.
Mc Gee Ranch
Mc Gee Flat
Kcb
Granodiorite
Castro Can.
ELSINORE FAULT ZONE
Little Chief Mountain
Tonalite
Hiriart Mountain
Gabbro
Slice Mountain
PKq
Ashley Ranch

type ... is a dark-gray, medium-grained, equigranular, homogeneous olivine-hornblende-hypersthene gabbro. Its principal constituents, listed in order of decreasing abundance, are calcic plagioclase, hornblende, olivine, augite, and hypersthene. Accessory minerals include biotite, ilmenite, magnetite, pyrite, pyrrhotite, and spinel. "Flow layering is well developed in some of the hornblendic rocks. "...Of even more local occurrence are nodlular, orbicular, and auto-injection structures." "The gabbro, norite, and associated basic rocks contain inclusions of quartzite and schist, and appear as apophyses in larger masses of these rocks. Thus they are clearly younger than the metamorphic series, although they are the oldest of the plutonic rocks exposed in the Pala district."

Tonalite

At least two varieties of tonalite are present in the area". "...One ...is a medium-to coarse-grained, moderately light-gray rock that contains scattered ovoid inclusions of gabbro, quartzite, and schist. It forms dikes and thick, tongue-like masses on Hiriart Mountain..." "The other...is fine-to medium-grained, medium to dark gray, and in most places is distinctly schistose. It is characterized by abundant discoid to elongate inclusions of darker, finer-grained rock, which are strung out parallel to the planar structure of the host rock." "...The relatively dark, fine-grained tonalite consists typically of andesine or labradorite, biotite, hornblende, quartz, augite, sphene, and zircon. The coarser-grained, lighter-colored tonalite consists of more sodic plagioclase (in theandesine-oligoclase range), biotite, hornblende, orthoclase, and rare augite and hypersthene. Accessory minerals are sphene, apatite, zircon, and magnetite." "The dikes and other masses of darker gray, finer-grained tonalite are distinctly younger than the enclosing gabbroic rocks, and are in sharp contact with them in most places. The trends of these contacts are closely reflected by the well-developed schistosity and streaked pattern of the platy inclusions that are characteristic of the tonalite."

Granodiorite

"Coarse-grained granodiorite underlies much of the northern part of the district, where it appears as the eastward and southeastward tapering prong of a very large intrusive mass." "...The rock is rather uniformly coarse-grained and light gray, with abundant phenocrysts of potash feldspar as much as half an inch in diameter. Its constituent minerals, listed in order of decreasing abundance, are oligoclase, quartz, orthoclase and microcline, biotite, muscovite, and hornblende, with accessory apatite, magnetite, phene, and zircon. The mafic minerals rarely amount to more than 8 percent of the total, and in most places are much less abundant.

"Typically the large granodiorite pluton is massive, but has well-developed schistosity and foliation along its borders, generally in a belt not more than 1,200 feet wide. Although very irregular in detail, these planar features are essentially parallel, and are oriented in crude conformity with the adjacent wall rock contacts.

"Inclusions of schist, quartzite, tonalite, and gabbroic rocks are widespread, but are nearly everywhere less abundant than those in the tonalites. Where the granodiorite is schistose, the inclusions are oriented in essential conformity with this structure, but elsewhere they show no consistent orientation. The relative ages of the rocks are further demonstrated by dikes of granodiorite that locally cut the tonalites and gabbroic rocks, and by a thick intrusive prong of granodiorite on the southeast corner of Hiriart Mountain.

"Another variety of felsic granodiorite, well foliated and locally very schisotose, ...is composed of quartz, microcline and orthoclase, oligoclase, biotite, muscovite, and scattered accesory minerals. "...Numerous inclusions and wispy remnants of quartzite and other metamorphic rocks are present in many areas, and in places there are all gradations between these pre-batholithic rocks and typical granodiorite.

"It was formed probably in part by migmatization and granitization of schist, quartzite, and associated rocks, and in part by intrusion of granodiorite magma along the contact zone between these metamorphic rocks and the Woodson Mountain granodiorite. Its foliation and schistosity are interpreted as a relict structure in most places, and as a flow structure in others."

Dike Rocks

"Masses of fine-grained lamprophyre cut the gabbroic rocks in many places. "...The constituent minerals include plagioclase in the bytownite-anorthite range, hornblende, and magnetite, with local ilmenite, hypersthene, and augite."

"Aplitic dikes of granodioritic composition are exposed...on a low hill northwest of Hiriart Mountain..." "...Most of the aplitic dikes consist of quartz, potash feldspar, oligoclase, muscovite, biotite, and scattered magnetite, with rare garnet, zircon, tourmaline, and other accessory minerals. The rocks are light-gray to buff, fine-grained, and thinly schistose. They are even-grained in some places, but elsewhere they grade inward from aplitic borders to pegmatitic centers. In a few places they closely resemble the younger pegmatite dikes, but grade along the strike into typical aplitic rocks that are cut by irregular veinlike masses of pegmatite and quartz."

Pegmatites

"At least 400 pegmatite dikes are exposed in an area of about 13 square miles. Most of them trend northward and dip gently to moderately westward. ...They are remarkably persistent, and range from small stringers to large dikes with bulges nearly 100 feet thick. ...The pegmatites occur mainly in gabbroic rocks, and appear to have been emplaced along a well-developed set of fractures. These fractures are independent of the primary structural features of the enclosing rocks, and transect contacts between major crystalline rock units...Graphic granite is the chief constituent of the outermost units. ...Discoidal masses of coarse-grained pegmatite form the innermost zones, or cores, of many dikes. Some cores are composed of quartz, perthite, or an aggregate of these minerals, and others consist of quartz and giant crystals of spodumene. ..."pocket pegmatite" consists of "coarse-grained quartz, albite, orthoclase, microcline, muscovite, lepidolite, and tourmaline." "All the gem tourmaline and beryl, as well as the commercial concentrations of lepidolite, occur in the pocket pegmatite. ...Fine-grained granitoid rocks, composed chiefly of quartz and albite, ...know collectively as 'line rock', are strikingly marked by alternating thin layers of garnet-rich and garnet-poor pegmatite, or of schorl-rich and schorl-poor pegmatite. ...The Pala dikes are believed to have been formed by crystallization of pegmatite liquid that was injected along fractures during the final stages of consolidation of the southern California batholith."

(Discussion of Pegmatites From My Text)

.... These pegmatites generally have sharp contacts against tonalite and gabbro and lack structural and textural features that support a replacement origin (Jahns, 1954; Jahns and Tuttle, 1963). On the basis of oxygen isotope fractionations between minerals, the pegmatites are estimated to have been emplaced at temperatures of about 700 to 730° C. Crystallization began with the bordering, garnet-bearing "line rock" (aplite) and ended with the coarse-grained gem-bearing pockets at temperatures near 565° C (B. E. Taylor and others, 1979). Therefore, the pegmatites must have crystallized from fluids that were above melting temperatures for granite. However, some of the pegmatites in the area can be traced continuously along strike into fine-grained granitic bands (3 to 5 m wide), containing muscovite and biotite but lacking the bordering, garnet-bearing "line rock". Here the granitic bands contain abundant myrmekite and remnant, zoned plagioclase (xenocrysts from the gabbro. This association indicates that the lateral fringes of the pegmatites have formed in place by recrystallization of sheared gabbro and at temperatures below the melting point. Moreover, in a few places adjacent to the aplite "line rock" and pegmatite dikes, granitic veins (3 cm wide) cut the tonalite and gabbro wall rocks. In these granitic veins all transition stages of cataclasis and replacement can be seen that lead to the formation of aplite and pegmatite. From the gabbro into the vein, the ferromagnesian silicates progressively develop quartz sieve-textures and then disappear to form pure quartz. Coexisting plagioclase breaks down to form myrmekite, plagioclase of lower An content, K-feldspar, and muscovite. In the center of the vein, scattered, tiny garnet crystals occur. The preserved cataclastic textures indicate that the former gabbro, now replaced by the granite vein, was also recrystallized at temperatures below the melting point. On this basis it is logical to assume that the pegmatite and aplite dikes were former zones of cataclastically sheared gabbro that were recrystallized initially to form myrmekite-bearing granite. But locally the introduced hydrous fluids were hot enough to melt the granitic residue and eventually cool and crystallize as the gem-bearing pegmatite.

The direct conversion of gabbro and tonalite to aplite and pegmatite dikes containing Al-rich minerals, muscovite, schorl, lepidolite, spodumene, rubellite, and topaz, is consistent with Figures 50 and 51. The density of these Al-rich minerals (2.80 to 3.57) is greater than the density of the parent plagioclase (2.67 to 2.71) in the tonalite and gabbro, and therefore, the formation of these Al-rich minerals would result in a volume loss. The additional destruction of biotite, hornblende, and pyroxenes in the tonalite and gabbro and the replacement of plagioclase by muscovite, K-feldspar, and quartz (Figure 48, and Figures 55 and 56, discussed in a later section) would provide even more volume losses, so that the breakdown and recrystallization of plagioclase and mafic silicates in the tonalite and gabbro may result in volume losses as much as 25 to 40% of the original volume. Such large volume losses could create vugs and cavities in which the large crystals of the pegmatites could grow. The vastly different mineralogies of the recrystallized aplite and pegmatite versus the tonalite and gabbro and the sharp contacts give the aplite and pegmatite the appearance of having been formed by magmatic injection from an outside source, but their derivation may be directly from the recrystallization of the adjacent relatively Al-rich gabbro and tonalite.

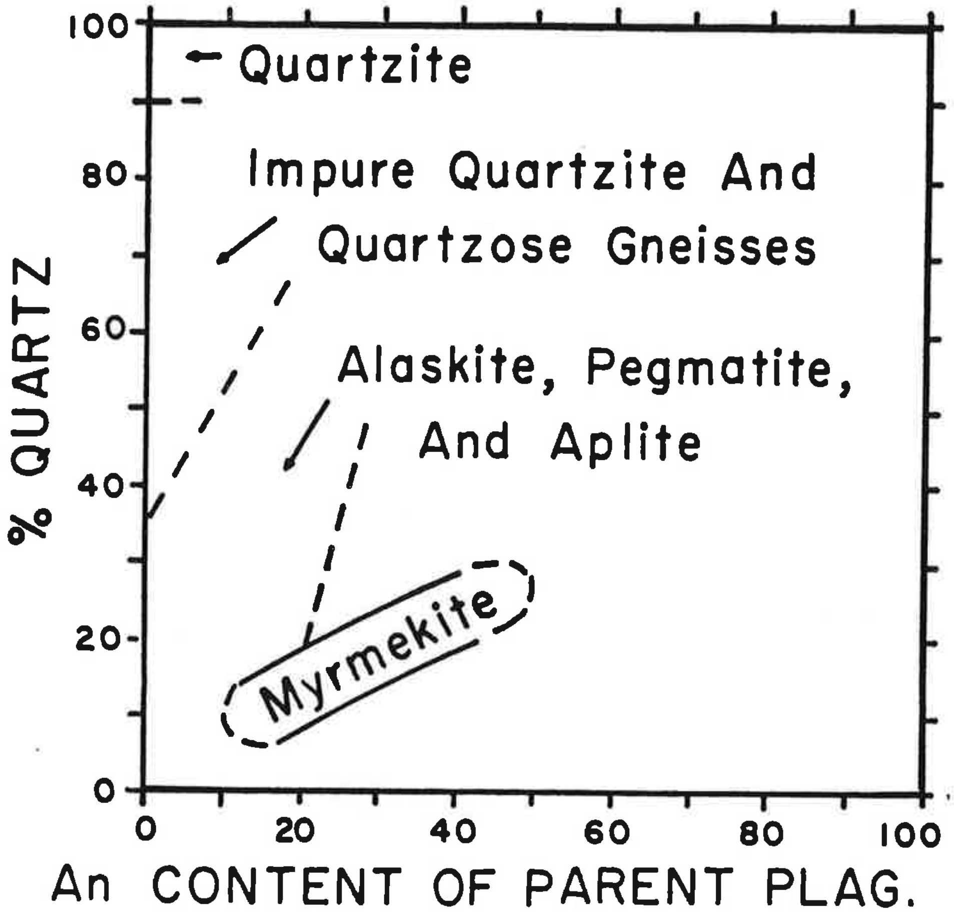


Figure 50. Volume percent quartz plotted against An content of parent plagioclase of original rock showing fields for quartz vermicules and An content of recrystallized plagioclase in myrmekite and for quartz and recrystallized plagioclase in alaskite, pegmatite, aplite, impure feldspathic quartzite, quartzose gneisses, and quartzite.

Metamorphic Rocks --- The Bedford Canyon Formation

In the Pala gem-pegmatite area the Bedford Canyon formation was mapped as a screen between gabbro and granodiorite plutons (Jahns, 1954). This formation consists of quartzites, impure quartzites, and supposed quartzite conglomerates. However, these rocks exhibit severe cataclasis, and in thin section the unsheared portions can be seen to have the same composition as the adjacent pluton (Collins and Cota, study in progress). Across the transition between the plutonic igneous rocks and the "quartzites" the ferromagnesian silicates can be seen to break down to form quartz sieve-textures and then to disappear to form interstitial quartz. Plagioclase in the same places recrystallizes to form myrmekite, K-feldspar, muscovite, and recrystallized plagioclase of lower An content. In the gabbro pluton are granodiorite "dikes" which are schistose (Jahns, 1954). These "dikes" are myrmekitic and have essentially the same composition as some of the metasedimentary-appearing rocks except that they are isolated and look like igneous dikes. The Ca, Al, Fe, and Mg that were subtracted from these schistose dikes may have been transferred to lamprophyre "dikes" which also cut the gabbro.

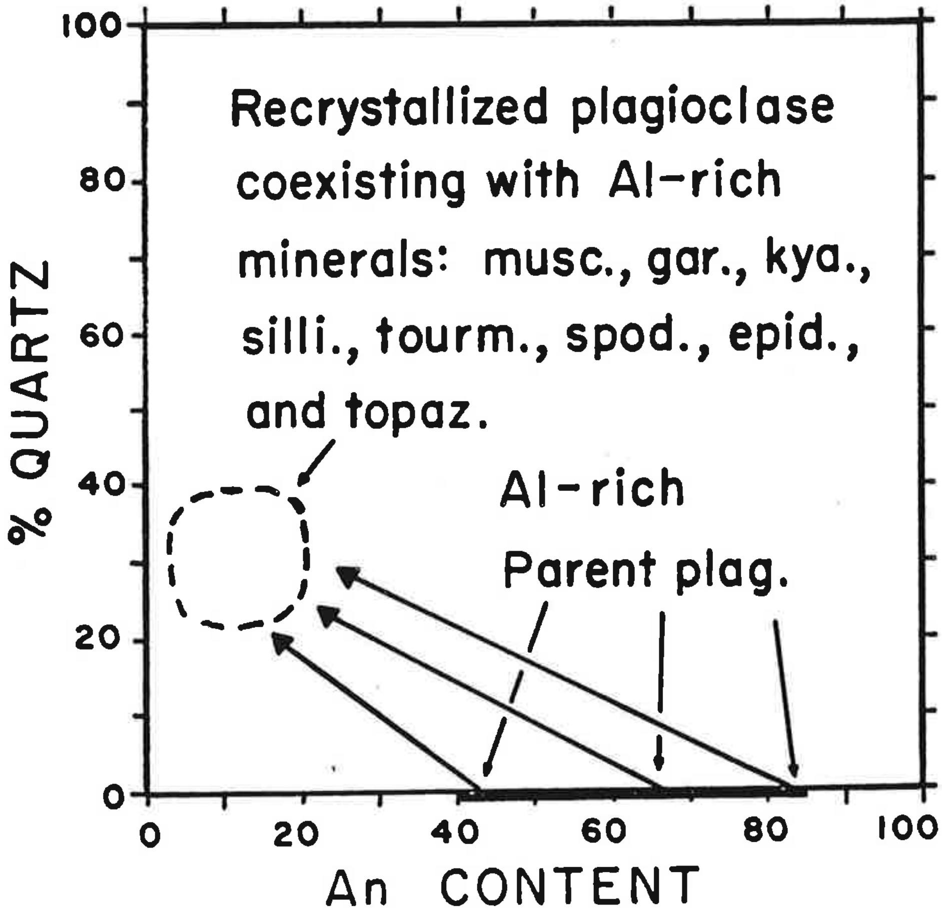


Figure 51. Plot of volume percent quartz associated with recrystallized plagioclase An_{0-20} formed by recrystallization of Al-rich parent plagioclase An_{40-85}. Al-rich minerals, muscovite, garnet, kyanite, sillimanite, tourmaline, spodumene, epidote, and topaz, may coexist.

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