Origin of myrmekite as it relates to K-, Na-, and Ca-metasomatism and the metasomatic origin of some granite masses where myrmekite occurs.

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ABSTRACT

Petrologists generally agree that granitic rocks form by crystallization from magma. Some granitic bodies, however, have been modified by alkali metasomatism at subsolvus temperatures (350-550°C). Myrmekite offers a petrographic means to detect the occurrence and conditions of alkali metasomatism in granitic rocks. Because metasomatic minerals inherit the hypidiomorphic textures of the former magmatic rocks, K-, Na-, and Ca-metasomatism may go unrecognized were it not for the additional presence of myrmekite in combination with other textural and mineralogical features, including microfractures, parallel alignments of silicate lattices and feldspar twin planes, quartz sieve-textures in ferromagnesian silicates, and sizes of quartz vermicules. Deformation is a precursor for myrmekite formation, but recrystallization may obliterate the deformation textures, so that myrmekite is the only remnant of prior deformation history. Crystals under stress develop a porosity that opens rocks to hydrous fluid that moves through the crystals and create a large-scale, coupled, dissolution-reprecipitation that totally modifies the rocks’ compositions. The loss of displaced Ca, Mg, Fe, and Al on a plutonic scale causes shrinkages as the residue becomes more granitic. The maximum width of the quartz vermicules in myrmekite correlates with the An content of the primary plagioclase being replaced that was once in the surrounding non-myrmekite-bearing relatively more-mafic rock. Examples of the use of warty and ghost myrmekite as clues for metasomatism are illustrated for the Bonsall Tonalite near Temecula, California, the Papoose Flat pluton in eastern California, the Twentynine Palms pluton north of Palm Springs, California, and the Vrådal pluton in Norway. Rim myrmekite forms in granitic rocks where deformation is at a minimum. Po halos in biotite in granitic plutons are additional clues that the granite has formed by metasomatic processes. Contrasting these textural and mineralogical features that occur during K-
metasomatism is a second type of wartlike myrmekite formation by Na- and Ca-metasomatism in many granitic plutons in Finland, Iran, New York, China, and other localities.

INTRODUCTION

In writing this article the authors recognized that twelve different issues need to be addressed to give guidance to the reader as to the direction the article will go. These twelve issues are listed below.

1. Nature of metasomatism and the problems it causes in making interpretations

When dealing with rocks that have been metasomatized, the investigator is always faced with the problem that the processes of replacement destroy the evidence for rocks or minerals that are alleged to have once been present prior to replacement. As an analogy, if a person only observes the adult butterfly and never the prior stages of the egg, larva, and pupa, then explaining to this person how the butterfly is formed is almost impossible. On that basis, throughout this article the reader must remember various prior stages in a metasomatic process that the reader can no longer see but is still valid evidence for the metasomatism.

2. The old concept of “granitization”

Tuttle and Bowen (1958) brought together all available experimental and petrographic evidence to suggest that granite bodies on a plutonic scale formed from magma. Their work clinched the opinions of most petrologists that the bulk compositions of granitic rocks correspond with thermal minima on the liquidus in the system Ab-Or-Q as well as the system Ab-An-Or-Q. Because their experimental results have been confirmed for innumerable examples in natural rocks, in the minds of most petrologists, Tuttle and Bowen also placed considerable limitations on the scale of possible subsolidus bulk chemical changes in granites. On that basis, it was logical that compositional changes in crystallizing magmas involved local reorganization within and between crystals and that transport of components were relatively limited by fluids circulating within cooling plutons. Therefore, large scale “granitization” by solid-state diffusion (Read 1948) was ruled out. No igneous petrologist could imagine a process in which a rock composed initially of quartz plus a sodic plagioclase can subsequently acquire, by a metasomatic process, exactly the amount of K-feldspar so that the rock’s composition lies on the liquidus minimum.

We agree that old-style “granitization” is untenable and that most granitic plutons are primarily magmatic in origin. We disagree, however, with Tuttle and Bowen’s suggestion (1958) that because the chemical changes that occur in natural rocks formed by magmatic differentiation (e.g., diorite to granodiorite to granite) are duplicated by experimental work that this duplication is sufficient evidence to prove that all large granitic bodies are formed by
maggmatic processes. The lack of proof is partly because chemical studies on granitic rocks formed by metasomatic processes in a replacement sequence of diorite to granodiorite and then to granite give the same chemical trends as those by magmatic differentiation (Collins 1988a). Thus, using experimental studies and chemical analyses of rocks alone are not sufficient to determine the origin of plutonic granitic rocks.

In this article we plan to show that there is another way for mineral compositions to arrive at the thermal minima. A rock system can start with minerals already solidified at high temperatures from magma, but these minerals are unstable as solids at temperatures below the minima. If such minerals in solid rocks are strongly or even minimally deformed to produce an open system, the primary high-temperature minerals, containing abundant Ca, Na, Fe, Mg, and Al, can be modified by the introduction of hydrous fluids that bring in K and Si while subtracting Ca, Fe, Mg, Al, and some Na to convert the primary high-T minerals into minerals that are stable at or near the thermal minima. Thus, Tuttle and Bowen (1958) have not demonstrated that magmatic processes are the only way in which minerals can crystallize at minima T-P conditions on a plutonic scale.

Na-metasomatism on a plutonic scale has been well-documented during the process of fenitization (Winter 2001). Surrounding carbonatite-alkaline silicate complexes are wide Na-metasomatic halos where quartz and feldspars in wall rock gneisses are converted into aegerine (Na-pyroxene), alkali amphiboles (arfvedsonite and riebeckite), nepheline, phlogopite, alkali feldspars, and carbonates. In some places pseudo-syenites are formed that are difficult to distinguish from an igneous rock. Temperatures of fenitization are >450°C but below the solidus (e.g., Pearson and Taylor, 1996). In other geologic terranes, relatively-calcic plagioclase in plutonic igneous bodies is albitized along shear zones by Na-metasomatism and at temperatures below the solidus (e.g., Engvik et al. 2008; Morad et al. 2010; and Plümper and Putnis 2009). In both types of geologic environments the Na-metasomatism is not doubted by the geologic community because the formation of Na-rich minerals is easily recognized as not being present in the metamorphic wall rocks surrounding the carbonatite-alkaline silicate complexes or in unsheared granitic plutons before the metasomatism occurred. Also, the metasomatism occurs across a zone in which gradual changes in mineral compositions are observed or in which veins of albite can be seen to extend into microfractured relatively more-calcic plagioclase.

K has chemical properties that are similar to that of Na, and, therefore, K-metasomatism on a plutonic scale should also be expected. For example, in the temperature range of 25-350°C, sodic and calcic plagioclase feldspars are altered to sericite mica by weathering processes (a coupled dissolution-reprecipitation mechanism) and at temperatures of about 100°C, mobilized K has formed metasomatic authigenic adularia (a low-temperature polymorph of K-feldspar) in Precambrian gneisses and overlying Paleozoic rocks, extending across thousands of square kilometers throughout Missouri, Wisconsin, Iowa, Illinois, Indiana, and Ohio (Liu et al. 2003). However, in the temperature range of 350-550°C (below the solidus), only a few examples of large-scale K-metasomatism are reported in the refereed geologic literature. Pitcher and Berger (1972) and Pitcher et al. (1987) described K-feldspar megacrysts as having been formed by K-

3. Origin of myrmekite and whether myrmekite has any geologic significance

The history of the study of myrmekite needs to be reviewed. Currently, little research is being done on its origin because its presence does not seem to have any important geologic significance. Granite petrologists and geochemists are obviously more interested in doing electron microprobe studies, isotopic analyses, and age dating of granitic rocks and determining how magmatic bodies are emplaced. Consequently, few thin sections are made, and the petrography of the rocks is barely reported. If myrmekite is present, it may go unnoticed or not be mentioned because its origin is so enigmatic and seemingly of no value for the purpose of the article that is being presented. Because the authors of this article believe that myrmekite has high geologic importance, a section on the origin of myrmekite is presented first.

4. Is calculating mass-balance transfers of elements helpful?

It is logical that where plutonic igneous rocks have been modified on a large scale by the introduction of K and Si and the subtraction of Fe, Mg, Ca, and Al so that the chemical composition (and mineralogy) of a diorite would progressively be changed into granodiorite and then into granite, an investigator of these changes might calculate mass-balance quantities of the elements that have been added and subtracted. In that progression from diorite to granite, the volumes of hornblende and biotite would become less, the plagioclase composition would become less calcic and more sodic, and both K-feldspar and quartz would appear and increase in abundance. Such calculations to be accurate would also have to take into account density and volume changes because the newly formed minerals in the more granitic rocks have lower densities and different volumes from that existing in the original rock. Some assumptions also have to be made, i.e., whether to balance the chemistry assuming that at least one element is immobile (Grant 1986). Would the results of such mass-balance calculations distinguish whether a granite is metasomatic or magmatic in origin? The answer is “No!” In the process of magmatic differentiation, the settling of heavy ferromagnesian silicates, calcic plagioclase, and iron oxides to the bottom of a magma chamber progressively enriches the melt residue in K, Na, and Si, so that “magmatic” mass transfers of elements that occur during this settling cause different kinds of plutonic rocks to crystallize. Significantly, these element transfers mimic the same kinds of element transfers that can occur during metasomatic processes. On that basis, chemical analyses of rocks and minerals that are formed by metasomatic processes and mass-
balance calculations are really not helpful in establishing the origin of a rock. Therefore, in this article in which origins of rocks are being studied, no chemical analyses are presented, no mass-balance calculations are made, and only the general sense of what elements have moved in or out is reported.

5. **Can large-scale metasomatism to form granitic masses of plutonic size be demonstrated?**

   Generally, granite petrologists accept K- and Si-metasomatism to form granitic rocks on a scale of a few cubic centimeters, perhaps a cubic meter, but not cubic kilometers. The authors believe that large-scale K- and Si-metasomatism on scales greater than 1 cubic kilometer can be demonstrated, and evidence to support such a concept is given.

6. **What is the nature and source of fluids that could produce large-scale metasomatism?**

   Unfortunately, this question is not easily answered, and only speculations can be provided. Deep unknown sources are likely the only possibility (such as the dehydration of a subducting slab), but, of course, using the unknown is never a satisfactory answer.

7. **Where do removed elements go?**

   Answer to this question is also difficult to resolve. Because of erosion of overlying rocks, in only few places can sites be found where dissolved elements could have moved. Because Fe and Ti are relatively immobile, they are primarily the only elements that reveal movements and concentrations during metasomatic processes.

8. **Is there another bit of evidence that supports large-scale metasomatism, such as Po halos in biotite?**

   As it turns out, Po halos in biotite in granite plutons also provide strong evidence that the chemical and mineralogical composition of the pluton has been modified extensively by metasomatic fluids. Po halos are commonly found in granitic plutons coexisting with myrmekite.

9. **Need for more research on isotopic and trace element studies**

   The various plutonic rocks that are suggested to be modified by metasomatic processes open up many areas of research possibilities that need to be done. Isotopic and trace element studies could provide additional evidence that metasomatism did or did not occur in some rocks.

10. **Scales of images used in the article**
Only a few images of thin sections petrography in this article have a scale-bar, but they are all photographed at low magnification (40x). Therefore, their scales are comparable, and the widest part of each rectangular image (left to right) that lacks a scale bar in most figures is about 4.5 to 5.0 mm prior to enlargements to fill a 2-column width of this article. Unfortunately, many of the thin sections were photographed when the technology for including scales was not easily possible, and it was not recognized at the time that such scales would be expected to be present. Now, re-photographing the thin sections is not possible.

Sixty-two images of mineral textures have been selected from more than thirty-one different geologic terranes and are used to illustrate the differences in styles for the two kinds of metasomatism: (a) K-metasomatism and (b) Na- and Ca-metasomatism. Other images can be seen in website articles listed at http://www.csun.edu/~vcgeo005/index.html.

11. Use of non-peer reviewed articles and the safe guarding of electronic website articles for future referencing

The senior author’s website, which is called Myrmekite, is an electronic journal that contains many articles all of which are non-peer reviewed. Nevertheless, they are repositories of data and images that are not available in published refereed journals. On that basis, they are freely referenced and used throughout our article. Readers can still form their own judgments of the usefulness of the evidence in the images that they provide without prior peer review. See: http://www.csun.edu/~vcgeo005/index.html. Myrmekite has been approved by the Library of Congress with the designation as ISSN 1526-5757 just as other refereed electronic journals have similar ISSN designations. Because reviewers would be concerned that the Myrmekite articles would always be available for future reference use, the Oviatt Library at California State University Northridge has placed all electronic articles on the website into a site called ScholarWorks for permanent storage.

12. Modern studies of mineral-water interface reactions need to be utilized

During the 40+ years in which the various studies on the Myrmekite website were published, no experimental evidence or modern microscopic techniques were available for looking at crystals on a nano-scale, and, therefore, large scale replacements by K- and Si-metasomatism could only be deduced from field and thin section evidence. On that basis, presenting these website articles for publication in refereed journals was always rejected. Modern studies, however, show that mineral-water interface reactions occur that cause minerals to develop a porosity so that fluids can flow through a granitic rock and produce large scale replacements (Putnis and Putnis 2007; Putnis A. 2009; Putnis and John 2010; Putnis and Austrheim 2010; Putnis C. and Ruiz-Agudo 2013). Undoubtedly, tectonically created fracturing plays a major role in permeability development and introduction of fluid into rocks (Jamtveit and Yardley 1997), but replacement reactions by dissolution-precipitation require that fluid infiltrates
every part of a rock and that these fluids move through pores created in the crystals as primary elemental components are replaced. In previous studies, how fluids move through rocks were generally restricted to hydraulic fractures and grain boundaries (e.g., Kostenko et al. 2002), but the production of pores in crystals by a reactive fluid greatly increases the number of possible fluid pathways and the rock’s permeability. As long as there is sufficient fluid and mass transport through the created pores, granitic plutons can be reequilibrated on a large scale. Literally, the fluid can react its way through a granite body by-passing some minerals with which there may be no reaction, or with which a reaction generates a non-porous product which effectively seals the mineral off from the fluid (Putnis A. 2009).

On the basis of this new modern understanding of how fluids can move through rocks and of how an interface-coupled dissolution-precipitation process simultaneously allows K- and Si-metasomatism to occur on a large scale, the geologic interpretations of several articles published in the Myrmekite website are reexamined from this viewpoint.

I. ORIGIN OF MYRMEKITE AND WHETHER MYRMEKITE HAS ANY GEOLOGIC SIGNIFICANCE

In the 1940s to 1970s many different models were proposed to explain the origin of myrmekite, including (1) simultaneous or direct crystallization from a melt, (2) replacement of K-feldspar by plagioclase, (3) replacement of plagioclase by K-feldspar, (4) replacement of plagioclase by quartz, (5) growth of blastic plagioclase around residual, recrystallized quartz, (6) exsolution of myrmekite from high-temperature K-feldspar, and (7) several miscellaneous theories that include combinations of the above (Phillips 1973a; Shelley 1964, 1973a; Smith 1974). Myrmekitic textures are also not a unique indicator of an origin or mechanism. For example, such textures can form in the breakdown of metal solutions, but in rocks that are at subsolidus temperatures, they most definitely indicate fluid infiltration. Numbers (2), (3), and (6) in recent years are generally the accepted possible models and are discussed below.

A. Replacement of K-feldspar by plagioclase

Historically, attempts to explain the direction of replacement have utilized balanced equations that equate mass-for-mass (Becke 1908).

\[
\begin{align*}
\text{KAlSi}_3\text{O}_8 + \text{Na}^{+1} & \rightarrow \text{NaAlSi}_3\text{O}_8 + \text{K}^{+1}. \quad (1) \\
2\text{KAlSi}_3\text{O}_8 + \text{Ca}^{+2} & \rightarrow \text{CaAl}_2\text{Si}_2\text{O}_8 + 4\text{SiO}_2 + 2\text{K}^{+1}. \quad (2)
\end{align*}
\]

potassium feldspar myrmekite

These equations in early years were discredited because there was no evidence that Na\(^{+1}\) and Ca\(^{+2}\) bearing fluids had moved through granitic rocks to replace K-feldspar. Otherwise, whole K-feldspar grains should have become myrmekitic instead of just isolated scattered patches, and myrmekite should have been concentrated near channels through which the fluids
were introduced instead of having a uniform distribution throughout a granitic rock. Nevertheless, some authors continued to support the idea that Na\(^{+1}\) and Ca\(^{+2}\) bearing fluids replaced K-feldspar to form myrmekite (e.g., Narasimaa Murthy and Sadashivaiah 1961; Didwal 1970; Gupta 1970; Ramaswamy and Murty 1972; and Beasley 1983). The senior author has since found places where whole K-feldspar megacrysts have nearly been replaced by myrmekite and that in early stages, the myrmekite is formed along fracture-channels that cut through the K-feldspar.

Part of the problem in early discussions about the origin of myrmekite is that no one realized that some myrmekite had subtle textural differences, and everyone was trying to convince other investigators that their models were the only correct model. These differences in myrmekite type that are formed by replacement of K-feldspar by plagioclase are illustrated and explained in the section: **MYRMEKITE AND TEXTURES FORMED BY Ca- AND Na-METASOMATISM (section X)**

**B. Exsolution to form myrmekite**

Schwantke and other investigators favored a different equation, with variations, in which Schwantke’s molecule Ca(AlSi\(_3\)O\(_8\))\(_2\) is supposedly exsolved from a high-temperature K-feldspar (Schwantke 1909; Spencer 1945; Surya Narayana 1956; Carstens 1967; Ramaswamy and Murty 1972; E. R. Phillips 1964, 1973a, 1974).

\[
\text{KAlSi}_3\text{O}_8/\text{NaAlSi}_3\text{O}_8/\text{Ca(AlSi}_3\text{O}_8)\_2 \rightarrow \text{KAlSi}_3\text{O}_8 + \text{NaAlSi}_3\text{O}_8/\text{CaAl}_2\text{Si}_2\text{O}_8 + \text{SiO}_2
\]

high-T K-feldspar K-feldspar plagioclase quartz myrmekite

The problem with this equation, however, is that (a) no one has ever detected Schwantke’s molecule or produced it in the laboratory, (b) some volumes of quartz vermicules in myrmekite require that the original high-temperature K-feldspar contain more than 40 percent Schwantke’s molecule, (c) some rocks have greater volumes of myrmekite than could possibly have exsolved from the available K-feldspar (Figure 1), and (d) in most places the K-feldspar adjacent to the myrmekite shows no evidence of being depleted in Ca and Na relative to the amounts in the plagioclase of the myrmekite.

Figure 1 comes from a sample of diorite which grades to myrmekite-bearing granodiorite at Monterey, California (Collins 2001a). In this place wartlike myrmekite projects into K-feldspar and the volume of myrmekite exceeds or is equal to the volume of the adjacent K-feldspar. Of course, the third dimension cannot be seen in this image, but because other sites in the same thin section reveal the same disproportionate relationship, the myrmekite cannot have formed by Ca- and Na-exsolution from the small volume of K-feldspar.
Figure 1. Wartlike myrmekite (center, top and bottom) on borders of albite-twinned plagioclase (whitish gray and gray) and projecting into microcline (black). Microcline penetrates and replaces plagioclase along cracks where diorite is in early stages of being replaced by K-feldspar to convert the rock into granodiorite. The volume of the myrmekite is greater than the volume of the microcline. Source: Collins (2001a, Figure 14).

C. Replacement of plagioclase by K-feldspar

Many investigators have noted that myrmekite is common where microcline replaces plagioclase (Bugge 1943; Drescher-Kaden 1948, 1969; Edelman 1949; Seitsaari 1951; Steven 1957; Taubeneck 1957; Sarma and Raja 1958, 1959; Phillips 1964; Shelley 1964; Burwash and Krupicka 1969, 1970; Bhattacharyya 1971; Byerly and Vogel 1973; and Beasley 1983). This association occurs, in part, because plagioclase is particularly susceptible to structural deformation and to the opening of avenues through which fluids can enter and facilitate replacement (Voll 1960; Augustithis 1962; Richter and Simmons 1977; Shirey et al. 1980; and Putnis et al. 2007). The equation is written below as if it balances mass-for-mass, but that is not the case.
\[
\text{CaAl}_2\text{Si}_2\text{O}_8/\text{NaAlSi}_3\text{O}_8 + \text{K}^+ \rightarrow \text{KAlSi}_3\text{O}_8 + \text{NaAlSi}_3\text{O}_8/ \text{CaAl}_2\text{Si}_2\text{O}_8 + \text{Si}_2\text{O}_2 + \text{Ca}^{+2} + \text{Al}^{+3}
\]

plagioclase  K-feldspar  more-sodic plagioclase  quartz  

myrmekite

In this article it will be shown in the next section dealing with Temecula rocks that the equation should be written as being balanced volume-for-volume. Nevertheless, these early investigators made valid observations without knowing that the production of the myrmekite had a far greater significance than just something that was formed in a local area. The fact that myrmekite generally composed less than 0.5 volume percent of the rock just gave the impression that it had no real geologic importance. Even so, many investigators also noted that such myrmekite was commonly formed where deformation occurred and presented various models suggesting that locally-applied stresses caused the myrmekite to be formed (La Tour 1987; Moore 1987; Hippert 1998; Simpson and Wintsch 1989; Vernon 1991; Tsurumi et al. 2003; Menegon et al. 2006; and Yuguchi and Nishiyama 2008).

II. TEMECULA – KEY TO THE ORIGIN OF MYRMEKITE

The senior author found a place where all stages of myrmekite formation can be seen, whereas all other investigators have been forced to come up with explanations for the origin of myrmekite based on seeing only the final product (the “adult butterfly”). The location is in the Peninsular Range of southern California where Cretaceous gabbro, diorite, tonalite, and granodiorite plutons occur along with probable Paleozoic metavolcanic and metasedimentary rocks (the Julian Schist); Tan and Kennedy (2000). The locality occurs in polished “table-top smooth” bed rock in the Santa Margarita stream canyon floor southwest of Temecula, California (Collins 1988b). Figures 2, 3, and 4 show the general locations of the samples from which thin sections and photomicrographs were obtained and which are described in Figures 6 to 15.
Figure 2. Location maps for myrmekite-bearing granite extending from the Woodson Mountain Granodiorite into Bonsall Tonalite near Temecula, California, and locations of outcrops in Figures 3 and 4 (pink rectangle, right side); from Collins (1988b).
Figure 3. Simplified geologic map of rocks in the Temecula area. Bonsall Tonalite (Dior.) and Woodson Mountain Granodiorite are also shown on Figure 4 (left side; arrow pointing to small rectangle labeled Fig. 3 as published in Collins 1988b). Woodson Mountain Granodiorite on Figure 3 also includes a myrmekite-bearing granite facies that occurs on Figure 4.
Figure 4. Schematic outcrop map (not to scale), showing field relationships of different rock types: Woodson Mountain Granite and Granodiorite (grd), Bonsall Tonalite (diorite), and amphibolite dike. Elongate black blebs are xenoliths of amphibolite. Pink outlined areas are photo locations in Collins (1988b), but only photo number 7 (center of figure) is shown in this article as Figure 5. A single off-set line (arrow) cutting across photo number 4 (left side) is a remnant of 2-3 cm-wide Bonsall Tonalite band in the granite.
Figure 5. Bonsall Tonalite (diorite) encloses a dark amphibolite dike on left and at one time also enclosed a thick part of the dike on the right (image is rotated 90 degrees from its position in Figure 4). A thin barely-visible remnant-rind (2 to 5 cm wide) of this tonalite encloses the amphibolite on both sides of the amphibolite (on upper and lower surfaces). Pointed projections (dark gray) of the amphibolite extend into the tonalite but not into the granite. A thin granite (aplite) vein extends through the amphibolite (center of photo). Myrmekite-bearing granite occurs in the upper and lower center and also extends between the thin and thick parts of the amphibolite dike parallel to the tonalite-granite contact (left of photo center). Arrows point to two diamond-saw cuts where a sample (15 cm long, 2.5 cm wide, and 2 cm depth) was broken out. Continuous thin sections and polished surfaces for scanning electron analyses were made from this sample, and all gradational transition stages between the tonalite and granite were observed. Photomicrographs of these transition stages are shown in Figures 7 to 15.

The amphibolite dike has the first appearance of being younger than the granite because the amphibolite dike cuts through the granite (Figure 4). Closer examination, however, shows that the amphibolite must be older because the granite extends through the amphibolite (photo 7 in Figure 4; Figure 5). Therefore, the geologic relationships shown in the outcrop are not as enigmatic as a first examination suggests. If the granite was younger than the amphibolite dike and was formed by magma extending from the Woodson Mountain Granodiorite (Figures 3 and 4), how could a hot, viscous, granitic liquid penetrate the Bonsall Tonalite to form the granite dike without disrupting, breaking, or displacing the amphibolite dike? Forming the granite by replacement processes is at least an alternative explanation.

Prior to selection of crystals for chemical microprobe analyses and to be certain of what was happening in the rocks across the transition from the myrmekite-bearing granite into the granite in the 15-cm-long sample (Figure 5), each of eight circular polished discs (2.5 cm in diameter) in a continuous sequence were photographed at 40x magnification in overlapping images. For each disc, these images (more than 100 per disc) were assembled to form a mosaic (25 cm in diameter). Then, tracing paper was placed on each mosaic, and all minerals grains (hornblende, biotite, K-feldspar, plagioclase, quartz, and myrmekite) were mapped, locating each
mineral type by different pencil colors. These maps were not done to establish modal changes across the transition from diorite to granite because changes in modes were established by using a Wentworth stage on the various thin sections to make measurements of the modes (Collins 1988). Instead the maps were necessary because mineral changes could not be memorized in narrow fields of view in the microscope across a thin section or disc and because these maps provided a means of seeing what happened to the rocks across the 15-cm transition zone from the tonalite to the granite. The maps provided the best way to locate which mineral grains should be analyzed with an electron microprobe. Hornblende in the remnant tonalite could be seen to be progressively replaced by quartz, and most biotite could also be seen to be replaced by quartz although some remnant biotite remained in the myrmekite-bearing granite. The first appearance of K-feldspar and myrmekite could also be identified. For comparison with the changes in the transition zone, a 25-cm-diameter map of minerals grains in unaltered Bonsall Tonalite (as seen in a thin section and polished disc) and also a 25-cm-diameter map of mineral grains in the Woodson Mountain Granite, 10 meters away from the transition zone, were also prepared. From each map, representative mineral grains were selected for 80 electron-microprobe analyses to determine what chemical changes occurred within centers, intermediate areas, and rims of individual grains.

Figures 6 to 15 are representative images of data reported in the mineral grains observed from the above thin section and microprobe studies as well as examples of cathodoluminescence studies of several altered grains, and scanning-electron analyses (Collins, 2002).

A modern interpretation of these figures relies on the recognition that these Temecula rocks have been subjected to an applied stress, and this stress promotes dissolution reactions where the solutions are under pressure. This same stress drives the reactions in the transition zone as illustrated in Figures 7 to 15. These reactions progress because disequilibrium occurs between the introduced fluid as it infiltrates. The composition of this fluid is out of equilibrium with plagioclase or the assemblage plagioclase + biotite + hornblende, and the stress applied to the rock enhances their dissolution. During the process interface-coupled dissolution-reprecipitation reactions occur. The added fluid phase both contributes to (K and Si) and removes (Ca, Na, and some Al) from the solid but deformed diorite while simultaneously generating a porosity that allows the reactions to occur (Putnis 2009).

We begin the dissolution reaction process by starting in relatively undeformed and unaltered biotite-hornblende tonalite several meters from the transition zone (Figure 5). Here the primary zoned plagioclase has cores of An$_{39}$ and rims of An$_{18}$ (Figure 6).
Figure 6. Photomicrograph of Bonsall Tonalite with unaltered hornblende (green); biotite (brown); zoned plagioclase (rims, An$_{20}$, dark gray and cores, An$_{39}$, light gray); albite twinned plagioclase (black, and light gray; left side); quartz (white and mottled gray). Source: Collins (2002, Figure 5).

In the early stages of replacement of diorite in the rind adjacent to the amphibolite dike (Figure 5) both biotite and hornblende show increasing quantities of quartz blebs in their interiors (Figure 7). Eventually, total replacement of hornblende by quartz occurs so that thereafter, there are no clues in the rock of its former existence. Remnants of biotite, however still remain.
Figure 7. **Left side.** Hornblende (brown and olive green) replaced in the interior by tiny grains of quartz (white). Plagioclase (white, cream, gray). **Right side.** Biotite (brown) replaced in interior by tiny quartz grains. Source: Collins (2002, Figure 10).

As the hornblende and biotite is being replaced by quartz, the first stage of alteration of primary plagioclase crystals begins where cores with compositions of An$_{39}$ are chemically altered to become cores of An$_{18}$ to An$_{20}$ while the rims remain the same, An$_{18}$ to An$_{20}$ (Figure 8). Therefore, Ca and Al have been lost from the cores. Microfractures in the plagioclase are not visible under cross-polarized light.
**Figure 8.** **Left side.** Normally zoned plagioclase (with outer sodic rim, An$_{18}$ to An$_{20}$; light gray) in which the core (darker gray, mottled) is microfractured and in the process of losing Ca. **Right side.** Rims and cores have equal An-content of An$_{18}$ to An$_{20}$. Bright birefringent minerals are sericite alterations. Source: Collins (2002, Figure 8).

Under cathodoluminescence chemical alterations, microfractures, and the first appearance of K-feldspar in the core along microfractures or porosity tunnels that cannot be seen under cross-polarized light images become visible (Figure 9).
Figure 9. This cathodoluminescence image has two plagioclase crystals with remnant zoning. The inner plagioclase cores were once rectangular in outline and were relatively calcic (greenish white) and graded to an outer more sodic rim (blue). Microfractures (dark blue veins) extend from the edge of the crystals to the relatively calcic cores. The dark blue indicates that Ca has been removed, leaving Na in the crystal structure. Locally, K-feldspar islands (faint light blue) are inside the core area where loss of Ca is most advanced and where both Ca and Na have been subtracted. Source: Collins (1997b, Figure 3). Karl Ramseyer provided the image.

In a more-advanced stage of replacement K-feldspar has started to replace the plagioclase in the interior of the former primary plagioclase crystal (Figure 10).
Figure 10. **Left side.** Plagioclase crystal (center; tan) replaced in its core by microcline (gray; grid-twinned). One twin plane of the microcline grid-twinning is parallel to the albite twinning of the plagioclase, but this albite twinning is barely noticeable in this orientation. **Right side.** The plagioclase grain (whitish gray) has lost its normal zoning and albite twinning. Plagioclase (light gray) is replaced in the interior by irregular islands of microcline (dark gray). Most nearby plagioclase grains lack these microcline islands even though they may be equally microfractured. The absence of visible microcline is just a matter of chance for K-bearing fluids to reach these other grains but as the metasomatism progresses, they also would be expected to be modified. Hornblende and biotite have locally disappeared and were replaced by quartz (white). Source: Collins (2002, Figure 11).

Other nearby crystals have greater degrees of microfracturing and chemical alteration. Cores and rims of many grains have equal An contents An$_{18-20}$ (Figure 11).
Figure 11. **Left side.** Plagioclase grains (shades of gray), many of which have lost their albite twinning and normal zoning. Cores are An$_{25-30}$; rims are An$_{18-20}$. Quartz (cream, white). **Right side.** Plagioclase crystals (whitish gray) are speckled. Some have cores An$_{18-20}$; rims An$_{18-20}$. Others have reversed zoning with cores An$_{15}$; rims An$_{18-20}$. Some have tiny islands of K-feldspar which cannot be seen in thin section in plane or cross-polarized light but can be seen in a cathodoluminescence image in Figure 9. Hornblende is gone, but some biotite (dark brown; upper right side) remains. Source: Collins (2002, Figure 12).

Eventually, across the zone of deformation, reverse zoning in the plagioclase is established with Na-rich cores of An$_5$ and rims of An$_{18-20}$. Where this occurs, pores in the silicate crystal structure are revealed as is seen in Figure 12 – a scanning-electron microphotograph (magnified 1,600x).
Figure 12. Scanning-electron microphotograph (magnified 1,600x). Plagioclase (pl, dark gray) is full of pores (black) with angular boundaries parallel to the plagioclase crystal structure. Many pores are elongate and parallel to the plagioclase crystal structure and are bordered by concentrations of K (lighter gray, incipient K-feldspar, kf) which are beginning to replace the plagioclase crystal structure along the walls of the pores. Locally, Ba is concentrated in celsian (barium feldspar; ce, white), and in some places the celsian is on pore walls adjacent to the K concentrations. Because the Ba +2 ion is about the same size as the K +1 ion, introduced Ba commonly substitutes in the crystal structure in the same places where K is replacing the Na and Ca. The identification of element concentrations is confirmed by microprobe studies.

In more advanced pore-formation in altered plagioclase crystals, where pores are larger and where K-introduction produces visible K-feldspar, rosettes of tiny hematite crystals line pores in the K-feldspar to make the K-feldspar pink (Putnis et al. 2007). The iron in the hematite is consistent with its coming from the iron that is released as quartz forms sieve-textures in nearby hornblende and biotite (Figure 7), which is simultaneous with the plagioclase alterations.

A key to understanding how myrmekite is formed is revealed where myrmekite first appears (Figure 13). Speckled crystals (sericite alterations) on the left side have reversed zoning with Na-rich cores of An\textsubscript{5} and rims of An\textsubscript{18-20}. Toward the right side a few of the grains of the same size are beginning to recrystallize to form myrmekite with tiny quartz vermicules (first fuzzy in appearance but then more distinct). Farther along the transition, quartz vermicules are more distinct and obvious (Figure 14).
Figure 13. Left side. Microcline (large light gray grain) has now completely replaced a former plagioclase crystal as a continuation of process observed in Figure 10. Speckled granules of plagioclase surround the microcline. The speckling is sericite and clay alteration because of the ease at which these grains are weathered. Two arrows (left side and right top center) point to early-stage myrmekite formation. The same image is on the right side but the image is at a different optic orientation. The microcline (black) in the right image is the same as microcline (dark gray) in the left image. Source: Collins (2002, Figure 13).

Because the pores in the crystal structure (Figure 12) make many crystals structurally weaker, the crystals are easily broken during shearing to form a cataclastic aggregate of small grains. At both places (indicated by two arrows) the reversed-zoned plagioclase crystal is beginning to form myrmekite. As recrystallization occurs, the quartz vermicules are formed because there is more silica in the reversed-zoned plagioclase crystal structure than can combine with residual proportions of Ca, Na, and Al to form only plagioclase, so silica is left over to form quartz vermicules. Thus, it is clear here that the myrmekite did not form by Ca- and Na-metasomatism of primary K-feldspar or by exsolution of Ca, Na, Al, and Si from primary K-feldspar.

Plümper and Putnis (2009) point out that relatively calcic cores of zoned plagioclase are commonly sericitized forming an alteration product of fine mica + albite while the more sodic rims are unaltered. This sericitic alteration would occur where fluids moved through pores into the interiors of the zoned plagioclase crystals at temperatures and pressures less than that which occurred in the Temecula transition rocks for Figures 12 and 13 and where later in the sequence similar altered grains become (a) myrmekite, (b) engulfed to become ghost myrmekite, (c) become albite-twinned plagioclase An$_{12-15}$, or (d) microcline, depending upon the kinds of replacements that occurred in different places (Figures 14 and 15). Therefore, the sericitization of the altered grains (Figure 13) is a later low TP modification that did not occur during the time in which myrmekite and other alternative replacements were in process.
On that basis, with continued deformation and with the addition of more K-bearing fluids, locally some of the tiny myrmekite grains are engulfed and much of their plagioclase content is replaced to form ghost myrmekite (Figure 14). “Ghost myrmekite” is a remnant of former myrmekite that has lost most or all of its former plagioclase, leaving only quartz blebs instead of vermicules in the K-feldspar.

**Figure 14.** Ghost myrmekite in microcline (light gray; right half of both images). Images on the left and right are the same areas but in a slightly different optic orientation. Most granules (left side) are now myrmekite although the tiny quartz vermicules are difficult to see, but they have lost the speckled appearance that is seen in Figure 13 because the residual crystal structure has recrystallized without any pores and is less easily weathered under lower TP conditions. Source: Collins (2002, Figure 14).

In other parts of the transition zone, former plagioclase crystals are not so thoroughly granulated, but the coarser crystals still exhibit reversed zoning with Na-rich cores of An$_5$ and rims of An$_{18-20}$. In some places a part of an altered reverse-zoned crystal can become wartlike myrmekite that projects into adjacent microcline while being attached to non-quartz-bearing plagioclase, as in Figure 15.
Figure 15. Wartlike myrmekite projecting into microcline (dark gray; grid-twinned). Source: Collins (2002, Figure 19).

In most places where K ions come into altered (reverse-zoned) plagioclase crystal structures from one side to form K-feldspar, the interior Na and Ca ions, displaced by the K ions, can readily escape out the other side. However, in some places a fluid pathway route is prevented either because the adjacent plagioclase forms a crystal seal against the K-feldspar which is not broken by later fracturing or because Na ions came into an altered plagioclase crystal from the opposite direction and filled the fluid pathways. When either condition occurs, the residual amounts of Na, Ca, Al, and Si ions in the altered plagioclase crystal structure may not be in the proper ratios to recrystallize only as more-sodic plagioclase, and silica is left over to form quartz vermicules. The way in which the myrmekite is formed results in the wartlike myrmekite projecting into the microcline on one side while being attached to the non-quartz-bearing, recrystallized sodic plagioclase on the other side. Because fluid pathways are rarely blocked, generally no more than 0.5 vol. % of the newly formed granite is myrmekite. In some geologic terranes, myrmekite may range up to 2 vol. %, but the abundance of myrmekite is not important. Its presence should be noted because it is a clue that significant chemical and mineralogical changes have occurred in the rock.

Another way to look at myrmekite is to realize that it is a non-porous product through which fluids cannot continue to move whereas the adjacent K-feldspar and plagioclase represent
places where fluids once moved progressively through connected pores in both feldspars while exchanging ions (Putnis A. 2009).

Note that the albite twinning in the myrmekite with tiny quartz vermicules (white) is continuous with albite twinning in the non-quartz-bearing plagioclase and is also parallel to one of the twin planes of the grid-twinning in the microcline. As suggested for Figure 15, this parallelism strongly suggests that the microcline is not primary orthoclase that inverted to microcline but is secondary replacement-microcline that has inherited its crystal structure from a former altered plagioclase crystal structure as in the scanning-electron image (Figure 12); see (Collins, 1998a). Support for this possibility is provided by Niedermeir et al. (2009). These investigators have done studies using an aqueous solution of KCL to show that natural albite crystals can be replaced by K-feldspar at 600°C and 2 kbars pressure. Subsequent transmission electron microscopy (TEM) diffraction contrast and X-ray powder diffraction (XRD) examination of the replacement results show that the K-feldspar has a very high defect concentration and a disordered Al, Si distribution, compared to the parent albite. On that basis, the recrystallization of such disordered Al, Si distribution would logically recrystallize directly as microcline rather than the more ordered Al, Si distribution that occurs in orthoclase.

Other many small reverse-zoned plagioclase crystals, instead of being replaced by K-feldspar, are recrystallized as plagioclase without pores. This recrystallization happens because Na that has been displaced from other sites by introduced K has entered these crystals and converted them into secondary plagioclase crystals with albite-twinning (upper left corner of Figure 15; white and black), and these crystals have compositions of An$_{12-15}$. Moreover, the volume of secondary quartz has also increased (from 5 to 25 vol. %) because of the simultaneous Si-metasomatism of much of the former altered plagioclase crystal structure as in the Bonsall Tonalite to form quartz sieve-textures (Figure 7). Eventually, the quartz blebs coalesce to form scattered anhedral quartz grains. By looking at the granite in thin section, one would not recognize that this quartz was mostly secondary and not primary or that the plagioclase (An$_{12-15}$) was also secondary. The hypidiomorphic texture looks as if it formed by crystallization from magma because the K-feldspar inherits its crystal outline from the primary plagioclase crystals.

Similar examples occur in some metamorphic rocks. For example, in Precambrian anorthositic granulite of the Bergen Arcs in western Norway primary crystals inherit a texture where fluids moving through fractures have caused the formation of eclogite. In these places the eclogite is not just a function of temperature and pressure but the infiltration of fluids that have caused replacements to occur (Austrheim 1987).

Several thin sections of the myrmekite-bearing Woodson Mountain Granite (10 to 100 meters away from the transition zone in nearby outcrops look no different from the granite adjacent to the transition zone (Figure 5), and in some places tiny islands (1-2 cm wide) of partially replaced tonalite lacking myrmekite occur. These islands grade into the granite without a sharp boundary and contain zoned plagioclase and 3-5 % biotite similar to Figure 6. Such islands cannot logically be considered as xenoliths (or enclaves) caught up in magma but give support that all the Woodson Mountain Granite in this area has also resulted from the metasomatic replacement of the Bonsall Tonalite on a plutonic scale. Nevertheless, these
observations do not necessarily mean that all of the Woodson Mountain Granodiorite was modified by metasomatism, because the main body of the pluton lacks myrmekite and likely still preserves its original unaltered magmatic composition when it was emplaced as an intrusive mass. At any rate, in those places that have been modified most of the Fe, Mg, and Ti were removed from the rock; much of the Ca was removed; but most Na remained behind in secondary recrystallized plagioclase as K and Si were introduced to convert the Bonsall Tonalite into the Woodson Mountain Granite. Chemical analyses of all rock types support these elemental additions and losses (Collins 1988a).

In most other granitic rocks in other geologic terranes that contain wartlike myrmekite, all the above sequential alteration stages (Figures 7 to 15) are missing, and what is observed is the final recrystallization product of just wartlike myrmekite and/or ghost myrmekite being present. Therefore, the textural changes that are observed in the Temecula rocks become a clue to what occurred prior to the formation of wartlike myrmekite and ghost myrmekite in these other terranes.

One important aspect of the gradual stages in the metasomatic process in the Temecula rocks which finally results in the formation of myrmekite is the recognition that myrmekite and secondary K-feldspar are created simultaneously. That is, the myrmekite is not a later product that occurs as a result of another time (thousands or millions of years later) in which hydrous fluids are introduced again to cause alterations. The recognition of this simultaneity by a petrologist embarking on thin section studies should firmly establish the real significance of the discovery of wartlike myrmekite. His or her approach to looking at rocks in the field or in thin section textures can suddenly have a new aspect that would not have been considered before. If a petrologist knows in his or her mind that a plutonic igneous rock is entirely formed by magmatic processes, then all that he or she sees are those bits of evidence that support that view and other bits of evidence that are contrary to that view are ignored. As Goethe has said: “We see what we know.”

On the basis of this understanding of how wartlike myrmekite is formed as one of the consequences of K-metasomatism of primary plagioclase in the Temecula rocks, we can now look at the same kinds of textural evidence that occur in these rocks as they might be applied to other geologic terranes.

### III. TEXTURES ASSOCIATED WITH AND FORMED DURING K-METASOMATISM

Three different kinds of textural features shown in the Temecula rocks provide evidence that large-scale K-metasomatism can exist in many plutonic igneous terranes. These are (a) K-feldspar replacements along fractures, (b) parallel alignments of silicate crystal structures and feldspar twin planes, and (c) quartz sieve-textures in ferromagnesian silicates. Each of these textural features is a condition that occurs prior to or during the formation of wartlike myrmekite that is discussed in subsequent sections.

#### A. K-feldspar Replacements Along Fractures
Evidence that the K-metasomatism is in solid rocks at temperatures below the solidus exists in places (a) where K-feldspar projects into microfactures in adjacent primary plagioclase (Figure 16) and (b) where broken plagioclase crystals are surrounded by the K-feldspar so that plagioclase islands are created in parallel optic continuity with the adjacent, larger, unbroken plagioclase crystal (See also Appendix: A. Microfracturing, Figures A, B, C, and D).

In early stage of K-feldspar replacement of plagioclase, the K-feldspar is interstitial as the K-feldspar replaces albite-twinned and zoned plagioclase along broken grain-boundary seals and as veins penetrating fractures that cut the plagioclase crystals. The absence of the primary plagioclase in the space or veins now occupied by K-feldspar gives evidence that K-metasomatism has caused the removal of the plagioclase (Figure 16).

**Figure 16.** K-feldspar (light gray; extending from left to right through center of photo); albite-twinned and zoned plagioclase (black, white, gray) above and below the K-feldspar. Note curved fracture that extends upward through a plagioclase crystal above the top edge of the K-feldspar and K-feldspar penetrating the plagioclase along this fracture. Myrmekite with tiny gray vermicules borders the plagioclase (white, upper right quadrant) against the K-feldspar (narrow gray vein). Source: Collins (2003, Figure 6).

In not every place are visible microfractures preserved where myrmekite is found. That is, a metasomatically formed granitic rock may not show any outward sign of deformation, such as a foliation or gneissic appearance. The microfracturing that allowed metasomatic fluids to
enter may be no more than breaking of grain boundary seals or breakage of crystals without any offset of fractures resulting in granulation. Or, because microfracturing cannot be at increasingly smaller scales to reach all parts of a crystal, mechanisms that increase the porosity allow the metasomatic fluids to reach all parts of a crystal (Putnis A. 2009). Nevertheless, the system can be opened to fluid movements, and, subsequently, if recrystallization is complete, no trace of the deformation may be visible. In spite of this lack of evidence for the existence of micro- or macro-fractures, they are necessary as a prior or concurrent condition along with a development of through-going porosity before wartlike myrmekite can form, and the myrmekite is the only outward clue that K-feldspar has replaced the interiors of microfractured primary plagioclase.

**B. Parallel Alignments of Silicate Crystal structures and Feldspar Twin Planes**

In the Temecula transition zone one plane of the grid-twinning of the microcline is aligned with the albite-twin planes of the plagioclase in the myrmekite which connect to the adjacent non-quartz bearing plagioclase in optical continuity. This happens because where secondary K-feldspar forms by K-metasomatism of primary plagioclase, a coupled dissolution-precipitation mechanism occurs that allows the K to move into the plagioclase silicate crystal structure and displace Ca and Na and some Al to form the K-feldspar (Putnis et al. 2007). This also happens because the connected porosity tends to migrate along the crystal structural planes in the plagioclase. Because the former silicate crystal structure of the plagioclase crystal is still present after Ca, Na, and Al are removed, (a) one twin-plane of the grid-twinning in microcline is inherited from the albite twin-plane of the plagioclase or (b) the Carlsbad twin-plane of orthoclase or microcline is inherited from the Carlsbad twin plane of the plagioclase. Therefore, the replacement-K-feldspar has parallel alignments of one of its twin planes with twin planes of the former plagioclase crystal that is being replaced. Thus, these same kinds of relationships should be expected in other terranes where alterations have produced wartlike myrmekite by K-metasomatism. Figures 17 and 18 are examples typical of fluid-induced mineral replacements along connected pores parallel to the former plagioclase crystal planes, but see also the Appendix B Parallel alignments, Figures E and F). As noted earlier, because in many places microcline commonly forms, it may not first be orthoclase that converts to microcline (Collins and Collins 1998).
**Figure 17.** Microcline (black and white; grid-twinned; lower left side) is in the process of replacing the interior of an albite-twinned plagioclase crystal (tan) toward the right. Microcline (upper right) also is in the upper right. A twin-plane of the microcline parallels the albite twin-planes of the host plagioclase. Myrmekite and cataclastic fragments of remnant plagioclase are along the border between the microcline and the plagioclase. Source: Collins (2003, Figure 20).
Figure 18. A Carlsbad-twinned plagioclase (Pl) with sericitized calcic core is replaced by K-feldspar (Kf). The plagioclase crystal is in the massive, pink Cape Ann granite several meters from the contact with the Salem diorite and is similar in size and shape to Carlsbad-twinned plagioclase crystals in the diorite. At the right end of the plagioclase crystal, its calcic core is truncated (vertical line, TL) and replaced by orthoclase whose Carlsbad twin plane (CbTP) is inherited from the former plagioclase crystal structure that once filled this space. Source: Collins (1997c, Figure 6).

C. Quartz Sieve-Textures in Altered Ferromagnesian Silicates

As in the Temecula site (Figure 7) commonly where relatively-mafic rocks are being microfractured and/or deformed such that K and Si can be introduced to cause K-metasomatism, the coexisting biotite and hornblende are also affected by Si-metasomatism to produce secondary quartz blebs as Fe, Mg, and Al are removed from their crystal structures by through-going metasomatic fluids. Figures 19, 20, 21, and 22 are examples from four different geologic terranes where later in transitions zones myrmekite occurs between K-feldspar and plagioclase.
Figure 19. Quartz sieve-texture in biotite in metasomatized rocks in Iran. Biotite (brown) replaced by quartz (white) along cleavage surfaces. Microcline (light gray) may also have replaced portions of the biotite (right side). Source: Behnia and Collins (1998, Figure 25).
Figure 20. Quartz sieve-texture in hornblende in metasomatized Salem diorite in South Hamilton, Massachusetts. Hornblende crystals (dark brown to black); quartz (center, white). Microcline (gray, grid twinning). Source: Collins (1997c, Figure 5).
Figure 21. Quartz sieve-texture in biotite in granite from Helsinki, Finland. Sieve-texture in biotite (lower right quadrant and upper left quadrant). Biotite (tan) may be partly replaced by K-feldspar (light gray; lower right quadrant) along cleavage planes. Source: Collins (1998a, Figure 7).
In undeformed relatively-mafic rocks in nearby areas adjacent to the above four examples, the hornblende and biotite crystals lack any quartz blebs. Modal studies of thin sections show that the quartz blebs coalesce to totally replace the hornblende and biotite in most places. Commonly, however, some biotite may remain. On that basis, some Fe, Mg, Al, and some Ca from hornblende must leave the system during the Si-metasomatism.

Because quartz sieve textures in ferromagnesian silicates can be formed by other geologic processes, they can be used only in combination with items A and B and the presence of myrmekite as indication that metasomatic processes have affected the rocks.

D. Wartlike myrmekite and its characteristics
In the Temecula transition rocks, the wartlike myrmekite (Figure 15) is seen projecting into K-feldspar so that the area that is myrmekitic is between two pincers of the K-feldspar on either side. The quartz vermicules taper toward the K-feldspar, and the plagioclase An-content of the plagioclase in the myrmekite adjacent to the quartz vermicules becomes more sodic, the narrower the quartz vermicule. In some kinds of wartlike myrmekite in other terranes, the quartz vermicules narrow still further and disappear toward the K-feldspar where a quartz-free (Na-rich) albite rim may exist against the K-feldspar (Figure 23; see also Appendix C. Wartlike myrmekite, Figure G).

Figure 23. Myrmekite (center, with branched vermicules) bordering a K-feldspar megacryst (gray, upper left). Quartz is light gray and white. Biotite is brown. At the outer edge of the myrmekite against the K-feldspar, the quartz vermicules become narrower until they disappear where the plagioclase of the myrmekite forms an outer albite rim. Source: Collins (1997e, Figure 5).

Quartz vermicules in myrmekite are not always oriented so that their long wormy-lengths are visible. In some places the long axis of each vermicule may be oriented perpendicular to the plane of the thin section. Therefore, under cross-polarized light they may appear black or gray because the view is down the c-axis of the quartz. An example is in the Vrådal pluton in Norway (Figure 24).
Figure 24. Non-typical myrmekite is in two places: (a) in the center of the image below the white quartz grain and the grid-twinning microcline (top; black) and (b) to the left of the black wedge of microcline that projects into the plagioclase (gray). Quartz vermicules are not seen because the view is down their long axes. Quartz-free, albite-twinned plagioclase with a speckled core is in the bottom left quadrant, and its albite-twin planes are optically continuous with the albite twinning of the plagioclase in the myrmekite. One plane of the grid-twinning of the microcline is parallel to the albite twin-planes of the plagioclase (light gray). Biotite (tan; lower right). Source: Vrådal pluton in southern Norway. See: Sylvester and Collins (2008).

E. Ghost myrmekite

Like the ghost myrmekite in the Temecula transition rocks (Figure 14), in some places after wartlike myrmekite is formed on the margins of K-feldspar crystals, renewed deformation and microfracturing may cause the system to open again so that the crystal grains under applied stress develop a porosity so that the plagioclase of the myrmekite is replaced by additional K-feldspar, leaving remnant quartz ovals of the former vermicules. This additional replacement produces a trace remnant of the former myrmekite, called “ghost myrmekite.” An example of ghost myrmekite in early stages of being formed is in the Wanup pluton where remnants of the original myrmekite are still present (Figure 25). The K-feldspar can be seen growing around
some of the former coarse quartz vermicules. (See other examples of ghost myrmekite in Appendix D. Ghost myrmekite, Figures H, I, and J)

**Figure 25.** Myrmekite with coarse quartz vermicules (white and cream), enclosed in microcline (grid pattern, light gray). Plagioclase of myrmekite (center, dark) with coarse quartz vermicules (mostly oriented perpendicular to the plane of the thin section) is speckled brown (sericite alteration). Source: Collins (2001b, Figure 18).

Another early stage of forming ghost myrmekite is also seen in Figure 26 where the sizes of the vermicules of the former myrmekite are a bit smaller because the An content of the primary plagioclase is less than that in Figure 25.
Figure 26. Cathodoluminescence image with quartz blebs (black) in microcline (Kf, light blue) in which the quartz blebs are the same size as in the adjacent quartz vermicules in myrmekite (My) bordering plagioclase against the microcline. Source: Hopson and Ramseyer (1990a) and Collins (1997k, Figure 9).

Hopson and Ramseyer (1990a) argued that the myrmekite in Figure 26 results from replacement of primary K-feldspar by Ca- and Na-bearing fluids which react with the K-feldspar to form the quartz vermicules (equations (1-2)). Then, these investigators suggest a reversal in the replacement direction (3-4). That is, K-replacement of the plagioclase in the myrmekite then takes place, so that quartz is left behind as islands (quartz blebs) in the K-feldspar (Figure 26).

Ca- and Na-metasomatism

\[ \text{KAlSi}_3\text{O}_8 + \text{Na}^{+1} \rightarrow \text{NaAlSi}_3\text{O}_8 + \text{K}^{+1}. \] (1)

\[ 2\text{KAlSi}_3\text{O}_8 + \text{Ca}^{+2} \rightarrow \text{CaAl}_2\text{Si}_2\text{O}_8 + 4\text{SiO}_2 + 2\text{K}^{+1}. \] (2)

\text{potassium feldspar} \rightarrow \text{myrmekite}

**Reversed reactions**

\[ \text{NaAlSi}_3\text{O}_8 + \text{K}^{+1} \rightarrow \text{KAlSi}_3\text{O}_8 + \text{Na}^{+1}. \] (3)

\[ \text{CaAl}_2\text{Si}_2\text{O}_8 + 4\text{SiO}_2 + 2\text{K}^{+1} \rightarrow 2\text{KAlSi}_3\text{O}_8 + \text{Ca}^{+2}. \] (4)

\text{plag. in myrmekite} \rightarrow \text{quartz} \rightarrow \text{potassium feldspar}
It is unreasonable for a chemical reaction to reverse itself inside a crystal when no logical changes in chemical gradients, potentials, or independent variables (T, pH₂O) would be expected to exist there to cause a reversal. Once the first Ca- and Na-replacement occurs, how can K be re-introduced to drive the reaction in the opposite direction? If the reactions were reversed and the reactions proceeded in the way proposed by Hopson and Ramseyer (1990a; Collins, 1990) in the balanced mass-for-mass equations, why does quartz remain in the K-feldspar when reversals of the equations should consume it?

The absence of K-feldspar and myrmekite in undeformed Bonsall Tonalite in the Rubidoux area, the gradual appearance of K-feldspar and myrmekite in the Bonsall Tonalite (where the tonalite is first deformed), and the equal volumes of K-feldspar and myrmekite at first deformation sites suggest that both K-feldspar and myrmekite are formed simultaneously. Also, the equal size of quartz vermicules in myrmekite in the Bonsall Tonalite and in myrmekite in the leucogranite supports the hypothesis that myrmekite in both places were produced from primary plagioclase with the same An content.

The seeming appearance of primary K-feldspar replacing plagioclase in Figure 26 is visually misleading. It is actually a one-directional but incomplete K-replacement of primary plagioclase that forms the myrmekite in the first place but renewed deformation opens the system so that continued replacement of some of the plagioclase followed by continued K-replacement that leads to the textures seen in Figures 25 and 26. Ultimately, where replacement of the plagioclase of the myrmekite is more complete, the result is ghost myrmekite that is in Figure 27 where some quartz is left behind as tiny blebs in the secondary K-feldspar because the silica cannot be incorporated into its composition.
Figure 27. Ghost myrmekite in Rubidoux Mountain leucogranite. At the extinction position for microcline (black), clusters of quartz blebs (white) are in faint traces of remnant plagioclase or as isolated oval islands in the microcline. The maximum sizes of the quartz blebs in the microcline match the maximum diameters of quartz vermicules in myrmekite (left side) where plagioclase is white and quartz vermicules are gray. Source: Collins (1997b, Figure 11)

On the basis of these characteristics of myrmekite and ghost myrmekite and the evidence that replacement features occur in granitic bodies on a plutonic scale, this information is applied later to rocks in the Papoose Flat pluton, the Twentynine Palms quartz monzonite, and to the Vrådal pluton on the basis of these same textures that occur in the Temecula area.

F. Rim myrmekite

If the amount of deformation in granitic rocks is very minor and if the amount of available introduced K is minimal, rim myrmekite may form on the outer edges of primary zoned plagioclase crystals (Figure 28).
Where rim myrmekite is found, the amount of K-metasomatism is minimal with little change in bulk chemical composition or results where there is local migration of K and Na on a sub-millimeter scale. However, rocks containing rim myrmekite may grade into rocks that contain wartlike myrmekite. On that basis, the conversion to rocks having only rim myrmekite to rocks containing wartlike myrmekite exists where increased deformation and increased amounts of K are available to form secondary K-feldspar. This secondary K-feldspar can be in addition to any primary K-feldspar that may also be present.

IV. SIZES OF QUARTZ VERMICULES AND THE FIRST APPEARANCE OF K-FELDSPAR

In the course of studying several different geologic terranes where transitions occur from relatively mafic, undeformed plutonic rocks that lack both K-feldspar and myrmekite to recrystallized more-granitic rocks containing K-feldspar and myrmekite, the senior author found a correlation between the maximum thickness of the quartz vermicules in myrmekite with the primary An-content of the plagioclase in the relatively mafic rocks. Moreover, myrmekite always occurs in combination with the first appearance of K-feldspar that coexists with secondary recrystallized plagioclase. Therefore, the myrmekite need not be a later product of hydrothermal Na- and Ca-metasomatism of the K-feldspar (Becke 1908) and is unlikely the product of exsolution from the newly-formed K-feldspar (Phillips 1964; Shelley 1964).

An example of the first and simultaneous appearance of K-feldspar and myrmekite where primary plagioclase is deformed occurs in the transition from undeformed tonalite containing...
primary plagioclase (averaging about An$_{30}$) to myrmekite-bearing granite near Temecula, California (Figures 13 to 15). Here, the myrmekite has relatively tiny quartz vermicules at the first appearance of K-feldspar that coexists with recrystallized plagioclase An$_{12.15}$. (Figure 29; Collins 1997a) Note that the An-content of the parent primary plagioclase (An$_{30}$) is about twice that of the An-content of the recrystallized plagioclase (An$_{12.15}$).

Figure 29. Myrmekite at Temecula, California. Primary plagioclase in former quartz diorite (averaged about An$_{30}$. Source: Collins (1997a, Figure 3). Coexisting, recrystallized, secondary plagioclase (upper center; tan; albite-twinned) is now An$_{15}$. Microcline (gray, black, and white; grid-twinned). Scale: Width of image is about 5 mm prior to enlargement.

The small maximum-sizes of the quartz vermicules in Figure 29 strongly contrast with the very large quartz vermicules in myrmekite where biotite-bearing gabbro containing primary plagioclase An$_{80}$ in the nose of the Split Rock Pond anticline in New Jersey (Sims 1953, 1958; Collins 1997a) was converted into garnet-sillimanite gneiss containing K-feldspar and
myrmekite in which the recrystallized plagioclase is An$_{40}$ (Figure 30). Note also that the An$_{80}$ is twice the An$_{40}$.

**Figure 30.** Myrmekite in deformed biotite-orthopyroxene gabbro layer near Split Rock Pond, New Jersey. Former primary plagioclase is An$_{80}$, but coexisting recrystallized plagioclase is now An$_{40}$. Scale: width of image is about 5 mm prior to enlargement. Source: Collins (1997a, Figure 6).

The image shown in Figure 30 was selected from thin sections of 912 samples collected from more than twelve different layers of igneous rocks (sills) in the Split Rock Pond anticline and across distances of more than 25 km from the nose of the Split Rock Pond where the primary igneous rocks are undeformed and then progressively away from the nose into the limbs where these primary rocks are strongly deformed (sheared), recrystallized, and metasomatically modified. Gradational to the limbs from the fold nose where K-feldspar is absent, both K-feldspar and myrmekite appear, and the quartz vermicules in the myrmekite in each different layer of differing primary igneous compositions have different maximum thicknesses. These thicknesses correlate with the plagioclase An-contents in the nose of the fold. The greater the quartz vermicule thickness, the higher is the An-content in the parent igneous rock. Because this relationship is consistent throughout the 912 samples in each of the different rock layers, the
correlations between quartz vermicule-thicknesses in myrmekite in modified rocks and An-contents in plagioclase in rocks lacking K-feldspar and myrmekite are strongly supported.

Another example of correlations of primary plagioclase An-contents and thicknesses of quartz vermicules in myrmekite exists in a probable, former, layered igneous intrusion in the Wanup pluton in Canada (Lumbers 1975; Collins 2001b). Here, however, no traces of the original undeformed rocks remain. All that exist are the recrystallized, strongly-deformed layers with K-feldspar, myrmekite, and secondary plagioclase. (See Figure 31 for a schematic of this anticline.)

![Figure 31](image-url)

**Figure 31.** Cartoon of isoclinal anticline in the layered Wanup pluton, showing locations for felsic (a), intermediate (bcdefhi), mafic (g), and theoretical ultramafic rocks (below g), from top to bottom. Gradual changes in An-content of plagioclase are shown, ranging from An$_{22}$ at the top to An$_{54}$ at the bottom.

The petrography and mineralogy of the rocks in the nose, limbs, and core of the anticline (Collins 2001b) reveal that the granitic Wanup pluton is a folded layered intrusion, with relatively more-felsic rocks at the top (Figure 31 a), intermediate compositions in the middle (Figure 31 bcddefhi), and more-mafic rocks at the bottom (Figure 31 g).

The relative sizes of the quartz vermicules in myrmekite in relatively more-felsic rocks, intermediate, and more-mafic rocks at the bottom are shown in Figures 32, 33, and 34.
Figure 32. Myrmekite with tiny quartz vermicules (white) surrounded by microcline (gray). Plagioclase (albite-twinned; black and gray); quartz (white); biotite (brown). Tiny quartz vermicules are typical of myrmekite in north end (top, nose of anticline) of the Wanup pluton and of outermost sodic-granitic layers of eastern and western limbs of the anticline where plagioclase is An$_{22-24}$ (Figure 31 a). Scale: width of image prior to enlargement is about 2.2 mm.

Where the megacrystal quartz monzonite and granodiorite contain plagioclase An$_{34-43}$ (Figure 31 i), the myrmekite has quartz vermicules with intermediate maximum thicknesses (Figure 33).
Figure 33. Myrmekite from a megacrystal granitic layer in the Wanup pluton on the south side of Route 69. Microcline (black, grid pattern) contains large and small quartz blebs of ghost myrmekite. Biotite (brown); plagioclase (albite-twinned; gray and white); quartz (white). Coexisting recrystallized plagioclase is An$_{38}$. Scale: width of image is about 5 mm prior to enlargement.

And where the megacrystal quartz monzonite and granodiorite contain plagioclase An$_{43-54}$ (Figure 31 g), the myrmekite has quartz vermicules with maximum thicknesses (Figure 34).
Figure 3. Myrmekite with thick quartz vermicules in megacrystal granitic rock near base of Wanup pluton, south of Route 69 (Figure 31 g). Microcline (black) with scattered quartz blebs (white); some quartz blebs are the same size as those in the myrmekite, and others are tiny in ghost myrmekite. Plagioclase (albite-twinned; light and dark gray); quartz (mottled gray). Plagioclase An$_{45}$. Scale: width of image is about 5 mm prior to enlargement.

Thus, even though no remnants of the primary igneous rocks are present, a direct correlation exists between the An content of the secondary recrystallized plagioclase and the thickness of the quartz vermicules in the myrmekite. As the An-value of plagioclase increases in different layer locations on Figure 31 from An$_{24-30}$ (Figure 32) to An$_{34-38}$ (Figure 33) and then to An$_{45-52}$ (Figure 34), the maximum thickness of the quartz vermicules increases. The thicknesses of the quartz vermicules in myrmekite in Figure 34 are comparable to those shown in Figure 30 where the known correlated parent plagioclase is An$_{80}$.

On the basis of the correlation of plagioclase An-content with sizes of quartz vermicules in the Wanup pluton, it is logical that tiny sizes of quartz vermicules are found in myrmekite in
an S-type granite in Australia where the An-content of the secondary recrystallized plagioclase is less than An$_{25}$ (Figure 35).

![Image](image_url)

**Figure 35.** Tiny white vermicules in myrmekite, Cottonwood Creek pluton. Plagioclase (black) in the upper left and upper right quadrants. Primary plagioclase averaged about An$_{24}$. Scale: width of image is about 5 mm prior to enlargement. Source: Collins (1997a, Figure 2).

None of these correlations between primary plagioclase An-contents and maximum thickness of quartz vermicules in myrmekite in recrystallized rocks far from the parent relatively-more-mafic rocks would be predicted by the two generally accepted models for forming myrmekite, either (a) by Na- and Ca-metasomatism of outer rims of primary K-feldspar (Becke 1908) or (b) by exsolution of Na, Ca, Al and Si ions from primary K-feldspar (Phillips 1973; Shelley 1973). The K-metasomatic model for the origin of myrmekite, however, provides an explanation as to how the maximum sizes of the vermicule widths correlate with the An-content of primary plagioclase because the greater the left-over Ca and Al in altered plagioclase crystal structures prior to the recrystallization to form myrmekite (discussed in a later section), the greater is the volume of quartz in the vermicules.

All observations of myrmekite and the associated sizes of quartz vermicules in many different terranes lead to the conclusion that the **An-content of the primary un-recrystallized plagioclase is roughly twice the An-content of the recrystallized (secondary) plagioclase coexisting with K-feldspar** (Collins, 1988a). Consequently, this relationship should incite the field geologist to look for remnants of a former, relatively more-mafic, parent rock from which the myrmekite-bearing relatively more-granitic rocks were formed by recrystallization and replacement.
This concept was applied after Simpson and Wintsch (1989) had described myrmekite with tiny quartz vermicules bordering K-feldspar in the Santa Rosa mylonite zone near Palm Springs, California, and with slightly coarser vermicules in the Hope Valley shear zone between Connecticut and Rhode Island. Simpson and Wintsch (1989) had suggested that during progressive deformation in granitic protoliths (near Santa Rosa, California, and along the border between Rhode Island and Connecticut), myrmekite in mylonitic, ductile shear zones is commonly concentrated in the rims of sigmoidal K-feldspar crystals forming a quarter structure in the shortening directions (Passchier and Trouw 2005).

These authors explain the asymmetric distribution of myrmekite by preferential K-feldspar breakdown-reactions at sites of high-differential stress during retrograde metamorphism (Shelley 1993). Asymmetrically distributed myrmekite is, therefore, suggested to be a product of extra strain energy along high-pressure sites favoring Na- and Ca-metasomatism of the K-feldspar to produce myrmekite.

The senior author does not agree with the assumption that the K-feldspar was primary magmatic and that the myrmekite was formed by deformation-induced Na- and Ca-metasomatism. His sampling beyond the shear zones in both the Santa Rosa (Collins 1997j) and the Connecticut-Rhode Island (Collins 1997o) areas revealed undeformed biotite diorite whose primary plagioclase was being replaced from the inside out by K-feldspar due to K-metasomatism.

The deformations were, therefore, continuous and had, not only affected the shear zones, but also the older plutonic country rocks, thence bringing about metasomatic changes in the mineralogy.

The high-pressure and high-differential stress sites (suggested by Simpson and Wintsch) are illogical places for fluids carrying Na and Ca ions to move because fluids always move to low pressure sites. Instead, these high-pressure sites represent “tight” places that were transformed into myrmekite, because the escape of Na and Ca ions lacking avenues for fluid or ion flow from the altered plagioclase structure was not easily accomplished and, consequently, the K-feldspar only incompletely replaced the former primary plagioclase.

The tiny sizes of the quartz vermicules in myrmekite in both field areas are proportional to the An-content of the primary un-recrystallized oligoclase plagioclase (An$_{25-32}$).

This section highlights that changes in rock compositions are tied to the first appearance of K-feldspar and myrmekite in deformed and recrystallized limbs of anticlinal folds. It further clearly shows that these changes occur on a plutonic scale (>1 km$^2$). The newly formed minerals were created in an open system with free movement of elements rather than by solid-state diffusion leading to granites (old-style granitization) as proposed by Read (1948).
V. CAN LARGE SCALE METASOMATISM TO FORM GRANITIC MASSES OF PLUTONIC SIZE BE DEMONSTRATED?

The observation that sizes of quartz vermicules in myrmekite are controlled by the An content of plagioclase in rocks that do not contain any myrmekite and over large distances in the Split Rock Pond anticline and the Wanup pluton implies that other areas might show this same relationship and be support for large-scale metasomatism. Another example was observed in a steeply plunging anticline that occurs in the Gold Butte area of Nevada in that the An-content of plagioclase in a parent relatively-mafic rock correlates with the maximum thicknesses of quartz vermicules in myrmekite (Figure 36; Collins 1997i). Fryxell et al. (1992) and Volborth (1962) suggested that the mafic and ultramafic igneous rocks in this anticline were younger and intruded as phacoliths into older garnetiferous gneisses, which these investigators thought were metapelites on the basis of their high aluminum contents. After mapping this anticline in greater detail and studying 150 thin sections, the senior author found that the igneous rocks are actually older than the garnetiferous gneisses and are completely gradational to the supposed metapelites (Figure 36). Figure 36 shows the large scale (>1 km²) of the processes that changed the various igneous rocks into metapelite-appearing gneisses that contain K-feldspar and myrmekite. As an aid to understanding the geologic relationships in Figure 36, a schematic of the rock layers in the anticline is shown in Figure 37.
Figure 36. Location of Precambrian Gold Butte area, Nevada, and geologic map of Gold Butte anticline. Mine symbol locates vermiculite mine. Grid pattern from Gold Butte 7 1/2 Minute Quadrangle, Nevada, is 1 km on a side. Symbols A-F refer to units in Figure 37. Sites a, b, and c (outlined in red boxes) are illustrations not discussed in this article as well as various numbers representing sites of 20 chemical analyses (see Collins 1997i). Garnet gneiss includes gneisses that may also contain sillimanite +/- cordierite. Symbol for dark mafic rocks (paired-line patterns) include mafic diorite, gabbro, mafic gabbro, and pyroxenites that have abundant mafic silicates.
Figure 37. An interpretive diagram, showing relative positions of units A-F in the Gold Butte anticline; not at the same scale as in Figure 36. A = felsic diorite; B and D = mafic diorite; C = mafic diorite, gabbro, mafic gabbro, and ultramafic rocks (black areas indicate two layers rich in mafic silicates); E and F = mafic diorite, gabbro, and mafic gabbro. Section lines 1-5 are explained in a later discussion about this figure. Source: Collins (1997i).

In this anticline (Figure 36), multiple sequences of ordered layering occur. More-sodic diorite layers occur at the top of each sequence that progressively becomes more calcic and mafic toward the bottom where the layers are gabbro or ultramafic rocks. In Figure 37 the layers (top left) are undeformed and un-recrystallized whereas the equivalent layers (right side) are increasingly deformed, recrystallized, and modified chemically progressively away from the nose of the fold. In the deformed limb K-feldspar and myrmekite make their first appearance in the igneous layers near the nose of the fold, and their abundances then increase away from the nose. Moreover, the greater thicknesses of vermicules correlate with the higher An-contents of primary plagioclase in the corresponding undeformed layer in the opposite limb. Simultaneously, the ferromagnesian silicates in the undeformed layered rocks are progressively recrystallized to form (a) garnet, (b) garnet + sillimanite, or (c) garnet + sillimanite + cordierite in the strongly deformed limb in direct correlation with the Mg/Fe compositions and abundances of the ferromagnesian silicates and with the higher aluminum-contents of the plagioclase with higher An-contents in the primary igneous rocks. An example of garnet gneiss is shown in Figure 38.
It is quite clear that the igneous rocks are not separate injections of phacolithic bodies into supposed metapelites because each of the igneous rock layers is gradational to the different metapelite-appearing rocks.

The fact that garnet gneiss compositions are controlled by the compositions of the primary mafic igneous rocks is shown in three separate areas (a), (b), and (c) where localized transitions between the two rock types occur (Figure 39). Locations of a, b, and c are outlined in red-lined squares on Figure 36.
Fig. 39. Schematic diagrams of structural relationships in igneous rocks and garnetiferous gneisses.

In (a) mafic diorite grades to garnet-sillimanite gneiss. In (b) mafic diorite grades to garnet-sillimanite gneiss and then into garnet-bearing leucogranite where so much Fe, Mg, Ca, and Al have been subtracted from the rock by through-going metasomatic fluids that mostly only sodic plagioclase, quartz, and a trace of K-feldspar remain. The leucogranite is shown in Figure 40.

Figure 40. Felsic fine-grained leucogranite with tiny garnets. Lens cap (5-cm-wide) for scale.
In (c) of Figure 36 a localized metasomatized and recrystallized shear zone extends across the contact between mafic gabbro and mafic diorite, and across the contact the gneiss composition abruptly changes from garnet-sillimanite-cordierite gneiss to garnet-sillimanite gneiss. The mineralogical changes in the gneisses shown in (a), (b), and (c) clearly show that these changes are controlled by the mineralogical and chemical compositions of the adjacent parent mafic igneous rocks.

What is also significant is that the conversion of the igneous rocks into garnet gneisses is not simply an isochemical change but involves large-scale subtraction and addition of elements in an open system to leave a residue that recrystallizes as garnet gneiss. For example, on the basis of an isocon diagram (after Grant 1986) possible elemental gains and losses occur between a mafic gabbro and adjacent garnet-sillimanite-cordierite gneiss along strike (Figure 41). The gains are K, Na, Al, Ba, Zr, and Rb, and the losses are Ca, Mg, Fe, Mn, Ti, Zn, P, and Sr. During the conversion, SiO$_2$ seems to remain nearly constant for the whole rock, if the volume remains constant (Collins 1997i).
Figure 41. Original element concentrations $C^O$ in mafic gabbro protoliths in comparison to altered element concentrations $C^A$ in replaced rocks that are now garnet-sillimanite-cordierite gneisses (after Grant, 1986). Major oxide abundances are in wt%; trace elements are in ppm. Some values are proportionally scaled. Solid line = constant volume reference frame; short-dashed line = constant mass reference frame. Elements plotting to the left of any reference isocon represent gains; those to the right represent losses. Data points should be connected with a line to the origin (zero point). What is significant for the scaled values is not the distance of the data points from the lines representing constant mass or constant volume nor the distance from the origin but the angles or deviations of the lines to the data points relative to the central dashed or solid lines.
On the basis of understanding the relationships between the compositions of the mafic rocks and the resulting compositions of their conversions to different kinds of garnet gneisses on a small scale (Figure 39 a, b, and c), a further understanding of what has happened in the Gold Butte area on a large scale (Figure 37) is apparent. All layers are folded such that one limb of the fold is bent underneath the other limb to produce an overturned fold (Figure 37). In that folding process, the rock layers in the underlying limb are stretched (sliding past each other) such that microfracturing allows hydrous fluids to come in and change the rocks metasomatically. The degree of sliding is indicated on Figure 37 by the fact that lines 5, 4, 3 intersect the various layers at right angles, but lines 2 and 1 (at the same spacing interval) show that equivalent positions have been shifted from being at right angles to being inclined, so that inner layers have slid relative to outer layers. Thus, rocks in the overlying limb are not stretched and remain unaffected, but are stretched (likely microfracturing the primary igneous crystals) in the underlying limb, opening up the system to metasomatic fluids. The metasomatism in the underlying limb converts each layer into more granitic compositions progressively away from the nose of the fold as shrinkages in volume also occur. The shrinkages are aided because the primary ferromagnesian silicates are recrystallized as andradite garnet \((G = 3.84)\), packing some of the residual Ca, Al, Fe and Mg into smaller volumes. Garnet is absent in each of the layers in the overlying limb. If the garnets had formed by dynamothermo-metamorphism, then they should occur in both limbs, and that is not the case. Because the more felsic layers at the top are affected differently from the more mafic layers at the bottom, contrasting wartlike myrmekite types are formed with the predicted coarser vermicules in the more mafic rocks where the primary plagioclase has higher An content and less coarse vermicules are formed in the more felsic layers where the primary plagioclase has lower An content. The final results of the tight folding and metasomatism cause the rocks in the inner tightest part of the fold to form a massive granitic residue in which much of the original Ca, Al, Mg, and Fe have been extracted in escaping fluids. The conversion to a granitic residue is facilitated because the forces that are tightening the fold are crushing the ends of layers B, C, and D against layer E above layer F (Figure 37). As a consequence several different elements are being extracted in fluids. By leaving the system these fluids allow the rocks to shrink in volume. Apart from its far larger scale the massive granitic residue in Figure 37 compares well with the small leucogranite body in Figure 39 (b). In the latter case field relationships clearly show that shrinkage (volume loss) had taken place as the mafic rocks were altered by metasomatic fluids and recrystallized. Myrmekite again provided the initial clue to look for this large-scale metasomatism.

Probably, most field geologists finding these leucogranitic lenses in the midst of garnet gneisses would interpret them to be magmatic bodies injected from an outside source into the supposed metasedimentary garnet gneiss sequence. But field relationships make it evident that each granite body was not intruded by force; neither do they carry any gneiss enclaves. Instead, each occupies a place that has undergone volume reduction due to compressional forces creating the tight isoclinal fold. It seems logical to assume that the elements Mg, Fe, Al, and Ca after having been extracted from this volume and combined with hot hydrous fluids could possibly create lamprophyre dikes higher in the crust.

On Figure 36 is a mine symbol that indicates the location of an abandoned vermiculite mine. The vermiculite formed locally in biotite-rich diorite where hydrous fluids moved through
microfaults to remove K ions from the biotite to form vermiculite in the strongly deformed limb of the Gold Butte anticline (see generalized equation).

$$K(Mg,Fe)_3(AlSi_3)O_{10}(F,OH)_2 + nH_2O \rightarrow (Mg,Fe)_3(Al,Si)_{4}O_{10}(OH)_2 .4(H_2O) + K^{+1} + HF$$

During the vermiculite forming reaction, K gets released. These places, therefore, represent potential sources of K which consequently moved to other places in the Gold Butte anticline to form K-feldspar and myrmekite.

**VI. APPLICATIONS OF CHARACTERISTICS OF METASOMATISM THAT OCCUR IN THE TEMECULA AREA TO OTHER TERRANES**

On the basis of the progressive changes in textures during metasomatic alterations in the Temecula area (Figures 6 to 15) and the similar textures and data that are observed in other terranes where wartlike myrmekite coexists with these textures (Figures 16 to 41), three different geologic terranes are examined to see how these common characteristics can be applied in making interpretations about origins of granitic rocks in these places. These three terranes include the Papoose Flat pluton, the Twentynine Palms quartz monzonite, and the Vrådal pluton.

**A. Papoose Flat Pluton**

The Papoose Flat (PF) pluton in the Inyo Mountains of California (Figure 42) is one of the most studied plutons in California because of its petrology, interesting contact relationships with the deformed sedimentary wall rocks, and the controversy over how the pluton was emplaced (Dickson and Sabine 1967; Sylvester et al. 1978; Brigham 1984; Nelson 1987; Dickson 1966, 1995ab, 1996; Paterson et al. 1991; de Saint-Blanquat et al. 2001). The contact relationships and emplacement of the Papoose Flat pluton are not discussed in this paper, and only a few of the many papers discussing these topics are cited above. Instead, the subsolidus history of this pluton is emphasized with particular interest on the Ba and K oscillatory zoned orthoclase megacrysts that contain concentric shells of tiny plagioclase, biotite, and quartz inclusions (Dickson 1994). Two hypotheses have been suggested for the origin of the megacrysts. Brigham (1984), de Saint-Blanquat et al. (2001), and Vernon and Paterson (2002) considered them to be of primary igneous origin, whereas Dickson (1996) suggested an isochemical recrystallization origin. However, on the basis of studies of textures in thin sections of samples of both rocks and 80 different isolated megacrysts collected throughout the PF pluton and in the Campito sandstone wall rocks (contributed by Frank Dickson), these two hypotheses are questioned, and a third non-isochemical metasomatic origin for the formation of the megacrysts is proposed in the following sections.
Before starting a discussion of the formation of the orthoclase megacrysts in the Papoose Flat pluton, modern studies (Putnis A. 2009) must be considered that reveal its progressive instability even though it had originally solidified stably from magma. The pluton continued to rise plastically, and during its slow ascent the applied stress-regime subjected the pluton to semi-brittle deformation affecting its mineral content. Under applied stress, the granitic rocks became a virtual sieve as introduced fluids moved through microfractures and broken grain-boundary seals. Consequently, a large-scale porosity developed, so that introduced fluids coming from great depth, moved not only through the microfractures and broken grain-boundary seals, but also through pores in the crystals. These fluids brought in both K and Si which greatly modified the compositions of the primary plagioclase in these rocks in the same way that was discussed earlier in the section describing modified rocks in the Temecula area (Figures 7 to 15).

On that basis, application of the model for wartlike myrmekite being formed by K-replacements of primary plagioclase crystals is appropriate for the concentrically-zoned orthoclase megacrysts in the Papoose Flat pluton in California. The megacrysts in this pluton are thought by most granite petrologists to be phenocrysts. An example of concentrically-arranged small plagioclase inclusions is in an interior corner of a large orthoclase crystal, 2.5 cm long (Figure 43). The regular pattern of aligned small plagioclase crystals is certainly suggestive that they nucleated on faces of a growing orthoclase crystal in magma. But appearances can be deceiving if not looked at carefully.
Figure 43. An inner corner of a zoned orthoclase megacryst has well-developed concentric banding of tiny parallel plagioclase inclusions and a change of orientation of the inclusions at the corner. Many inclusions have sodic rims. At least one plagioclase inclusion coexists with an adjacent tiny plagioclase inclusion (gray and black, arrow; center; about 1 mm above left end of scale line) whose orientation is at right angles to the zoning. Source: Collins and Collins (2002, Figure 11).

The Ba-zonation (Dickson 1994) can be seen in Figure 43 as different shades of gray in the orthoclase and need not be related to magmatic processes because at different stages of K-feldspar replacement of microfractured plagioclase crystals and the growth of the megacrysts, different amounts of Ba ions may come in with K ions during the metasomatic processes as indicated by the deposition of celsian (Figure 12).

Contrary to the expected behavior by magmatic theory, the orthoclase did not nucleate in the Papoose Flat pluton as small euhedral crystals growing to become large euhedral phenocrysts but started as replacements of primary plagioclase along microfractures and between grain boundaries (Figures 44 and 45).
**Figure 44.** Orthoclase (black) with tiny remnant plagioclase inclusions. The orthoclase forms scalloped edges against adjacent plagioclase crystals (light gray; right side and upper left), which would not occur in magma where euhedral rectangular orthoclase would form. Myrmekite with tiny quartz vermicule border three different plagioclase crystals and can barely be seen where fingers of the orthoclase (black; arrows) project into the plagioclase along former fractures or broken grain-boundary seals (right side). The tiny vermicules correlate with the primary sodic plagioclase that has low An-content. Source: Collins and Collins (2002, Figure 5). See Figure 4 in this same website article for another example of early stage replacements.
Figure 45. The growing zoned orthoclase crystal (light gray) is not perfectly rectangular but has irregular growth on different faces of the crystals. In its outer border, the orthoclase encloses tiny inclusions of the ground mass (left side and top edge; arrows). Note that the orthoclase extends beyond its corner (upper right; arrow) to penetrate and enclose rounded inclusions of the ground mass, and that this part of the orthoclase is optically continuous with the main crystal and has not resulted from deformation of the orthoclase to form a "tail." Thus, such K-feldspar megacrysts are subsolvus and post-magma. One elongate plagioclase inclusion (left side) is parallel to the zoning. Source: Collins and Collins (2002, Figure 8). See also Figure 7 in this same article for a similar early replacement stage.

The same kind of granulation of ground mass minerals can be observed in Figure 46. Additionally, shearing planes lined with biotite extend around the right corner of an orthoclase megacryst without traversing nor off-setting the corner.
Shear planes within the ground mass minerals are revealed by biotite crystals (tan) (arrows; top center and lower right) bending around the corner of an orthoclase megacryst (light gray; partly inverted to microcline). A myrmekitic inclusion can be seen in the lower left (arrow; above the scale), but in this optical orientation the quartz vermicules (gray) are hard to discern. On the right edge of the megacryst are quartz groundmass minerals (white, gray, black). Source: Collins and Collins (2002, Figure 18).

As a result of studies of Sierran plutons Moore and Sisson (2007) and Johnson and Glazner (2009) have suggested that K-feldspar crystals continue growing even after 50% of the magma batch has crystallized. This is due to textural coarsening transferring material from small crystals to large crystals during thermal cycling. Similar coarsening of K-feldspar crystals was proposed by Higgins (1999) as Ostwald ripening. Images provided by Johnson and Glazner (2009) show tails on the corners of K-feldspar megacrysts that are similar to those illustrated in Figures 45 and 46. However, such megacrysts in the Sierran granitic rocks do not show any myrmekite in the interiors, intermediate zones, or rims of the megacrysts or any evidence of cataclasis that would be consistent with the megacrysts being formed by K-metasomatism in already solidified granitic rocks. Therefore, the growth of tails on corners of K-feldspar megacrysts can be brought about by Ostwald ripening or metasomatically by tectonically-triggered replacement reactions – two fundamentally different processes.

An examination of the orientation of ground mass crystals relative to the orientation of crystal inclusions in an orthoclase megacryst is instructive because it reveals the degree of replacement of ground mass minerals that exist in a megacryst as inclusions (Figure 47). Many plagioclase inclusions in the outer edge of the orthoclase crystal have scalloped borders (arrows
Scalloping certainly suggests that these inclusions are being replaced and that they have not nucleated on the face of a growing megacryst crystallizing in magma. Five elongate plagioclase crystals (gray, white, black) in the groundmass (right side) have crystallized against each other in parallel alignment (synneusis) and are parallel to the zoning in the megacryst, but some are inclined. It is logical that if the megacryst had continued to grow by K-replacement processes, those crystals in the ground mass aligned parallel to a possible orthoclase face might have been enclosed without replacement and those that were inclined might be replaced. Thus, the concentric zoning of plagioclase inclusions can be explained. Remnants of inclined grains are commonly scalloped in Figures 46 and 47 and in many other images (Collins and Collins 2002). Note that the outer face of the orthoclase megacryst has projections of K-feldspar that extend around individual quartz grains (gray, white, cream) of the ground mass and that continued replacement of these quartz grains would cause them to become smaller rounded remnants like the small oval quartz (bright cream) remnant in the upper left quadrant.

**Figure 47.** Outer edge of a large orthoclase megacryst (left side). Five elongate plagioclase crystals (gray, white, black) in the groundmass (right side) have crystallized against each other in parallel alignment in a magmatic synneusis texture and are parallel to the zoning in the megacryst, but some are inclined. Arrows point to inclined plagioclase inclusions in the megacryst that show scalloped edges. Source: Collins and Collins (2002, Figure 15). See Figure 10 in this same website article for remnant ground mass grains inside a megacryst.

Some particular features of the zoned orthoclase crystals in the Papoose Flat pluton which negate a magmatic origin of the megacrysts are (a) the crystal inclusions that are found at interior corners of the crystals and (b) the occurrence of wartlike myrmekite. For example, a large segment of an aggregate of ground mass minerals is at an interior corner (Figure 48, top center). This aggregate mass of normally-zoned plagioclase crystals have their longer
dimensions parallel to each other and are roughly perpendicular to the inclusions in the megacryst. Two plagioclase grains at the right corner of the aggregate mass adjacent to the myrmekite are parallel to the plagioclase inclusions. The myrmekite has very tiny quartz vermicules (black and white) and is in the right lower corner of the aggregate mass (arrow; left of two white quartz inclusions). Many orthoclase megacrysts contain aggregates of ground mass minerals at interior corners because at corners the potential crystal structure (-201) crystal face there does not provide a surface on which inclusions can be easily aligned. Moreover, if the orthoclase megacrysts were growing in magma, the aggregates of crystals would exist in random places in the crystals as inclusions and not be mostly restricted to interior corners.

![Image](image_url)

**Figure 48.** A large segment of an aggregate of ground mass minerals is at an interior corner (top center) of a zoned megacryst. Quartz (white; right side). Source: Collins and Collins (2002, Figure 17). See Figures 20 and 21 in this same article for other examples of remnant wall rock mineral grains in the corner of a megacryst.

If the tiny plagioclase inclusions in K-feldspar megacrysts had formed by nucleation on outer faces of growing crystals in magma, then everywhere in the pluton, the parallel arrangement of the long dimensions of the inclusions with possible faces should be found. Instead, only where the ground mass crystals exhibit synneusis does this parallel arrangement exist. In that part of the pluton where the ground mass minerals have random orientation, then the plagioclase inclusions also have random orientation with little to no alignments of long dimensions parallel to possible crystal faces (Figure 49). See also Collins and Collins (2002,
Figure 30). This evidence is consistent with the model that the K-feldspar megacrysts grew post-solidification of magma.

![Image](image_url)

**Figure 49.** Interior of Carlsbad-twinned orthoclase megacryst (light gray and dark gray). Disoriented plagioclase inclusions are along Ba-K growth zones that extend from lower right to upper left. None of the inclusions is aligned parallel to the growth zones, and many have borders that are strongly scalloped. This image looks out of focus but is not and has its fuzzy appearance because of the lack of sharp boundaries adjacent to the inclusions. Source: Collins and Collins (2002, Figure 31)

The ground mass surrounding the megacrysts contains 15 to 20 percent quartz. But quartz inclusions in the orthoclase are scarce, and those that are present exist as tiny ovals, as in Figure 38. If the megacrysts had formed by crystallization from magma, this quartz should have formed as crystals with outlines typical of a granophyric texture. This expectation is particularly true when the ground mass surrounding the orthoclase megacrysts contains 15 to 20 vol. % quartz and at the eutectic both K-feldspar and quartz should crystallize simultaneously. An example of such a granophyric texture exists in the Dunan granite, Isle of Skye (Figure 50).
However, at corners of zoned orthoclase crystals large quartz grains are not as easily consumed (Figure 51).

**Figure 50.** Granophyric texture. Source: Collins (1997p, Figure 22)
Figure 51. A large former quartz crystal (yellow) mostly remains at an interior corner as a scalloped crystal. Source: Collins and Collins (2002, Figure 22).

Moreover, let’s look again at the balanced mass-for-mass equations. In the reactions, plagioclase is replaced by potassium feldspar and silica (quartz) is consumed in the transformation to K-feldspar (equation 4).

\[
\begin{align*}
\text{NaAlSi}_3\text{O}_8 + \text{K}^+ & \rightarrow \text{KAlSi}_3\text{O}_8 + \text{Na}^+. \quad \text{(3)} \\
\text{CaAl}_2\text{Si}_2\text{O}_8 + 4\text{SiO}_2 + 2\text{K}^+ & \rightarrow 2\text{KAlSi}_3\text{O}_8 + \text{Ca}^{+2}. \quad \text{(4)}
\end{align*}
\]

<table>
<thead>
<tr>
<th>plag. in ground mass</th>
<th>quartz</th>
<th>potassium feldspar</th>
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This requirement is even more clearly demonstrated when it is understood that the transformation in the rocks in the Papoose Flat pluton is not mass-for-mass but volume-for-volume, the low density of the K-feldspar crystal structure (G = 2.56) in comparison to plagioclase (G = 2.62 to 2.66 for An_{10-30}) also requires more silica for the space in which the K-feldspar forms. Therefore, it is logical that the quartz grains in the ground mass are consumed or nearly so as the K-feldspar grows and replaces the plagioclase in the ground mass (Figure 47).
During the metasomatic replacement of plagioclase by K-feldspar, quartz is consumed. The same process can likewise be seen occurring in myrmekite, where during the replacement of plagioclase by K-feldspar only the narrower parts are affected and likely because the adjacent Na-rich plagioclase is more Si-rich. More voluminous round or oval blebs are left behind to form ghost myrmekite (Figures 14, 25, 26, and 27; see also Appendix: Ghost myrmekite, Figures I and J).

Myrmekite can be found lying along the edge of a megacryst (Figure 52) and in an interior inclusion (Figure 48); it also appears during earliest stages of K-metasomatism (Figure 46). Contrary to orthoclase that commonly crystallizes from magma, a magmatic origin for the myrmekite is ruled out because from melts, quartz-plagioclase intergrowths can form only as granophyric or micrographic textures (Figure 50). The presence of myrmekite along edges of orthoclase megacrysts explains why orthoclase, formed originally as K-replacement product of disoriented, primary plagioclase crystals in the ground mass, eventually managed to engulf myrmekite and enclose it in its crystal structure (Figures 46 and 48).

Figure 52. Myrmekite with very tiny quartz vermicules along edge of orthoclase megacryst (gray). Source: Collins and Collins (2002, Figure 24). Another example of myrmekite existing in the interior of megacrysts can be found in Figure 26 of this same article.
Finally, the Campito sandstone (arkose), an adjacent wall rock of the Papoose Flat pluton, contains isolated, zoned orthoclase megacrysts concentrically enclosing tiny plagioclase crystals, very similar to the concentric pattern that is in the orthoclase megacrysts in the pluton (Figure 53). In one such megacryst tiny biotite inclusions are in the corner, and these inclusions match similar-sized biotite flakes in the adjacent ground mass of the sandstone. In other places plagioclase inclusions match ground mass mineral sizes and appearances. Therefore, it seems unlikely, that during the deformation of the sandstone these megacrysts were sheared off granitic pegmatite dikes extending into the wall rocks (Vernon and Paterson, 2002; Vernon, 1991) or by crystallization in igneous dikes (Brigham, 1984). The megacrysts had to grow in place and at sites completely isolated from granite magma in the pluton. That is, the same K-bearing fluids that moved through microfractures inside the Papoose Flat pluton also moved through fracture openings in the Campito sandstone outside the pluton and produced the same kinds of metasomatic alterations of plagioclase crystals. The idea that the orthoclase megacrysts in the Campito sandstone were formed by metasomatic processes and not from magmatic fluids was originally proposed by Frank Dickson (Dickson, 1966; Collins and Collins, 2002).

Figure 53. Orthoclase megacryst (light gray) in Campito sandstone contains remnant biotite grains of the sandstone enclosed in the border of the megacryst. Tiny plagioclase inclusions aligned in the megacryst are the same size as plagioclase grains in the sandstone. Some inclusions are inclined to the Ba-K concentric layers in the megacryst and would not be expected to nucleate in that position on the face of an orthoclase crystal growing in magma. Source: Collins and Collins (2002, Figure 44).
A final point that needs to be made is that the K-feldspar megacrysts in the Campito sandstone are not the result of contact metamorphism from the heat of the crystallizing pluton because they form only locally and not along all portions adjacent to the pluton contact. This relationship is consistent with the observation that the K-feldspar megacrysts formed well after the solidification of the pluton.

The observations of the K-replacement textures in the Papoose Flat pluton are not meant to suggest that all orthoclase megacrysts in other geologic terranes, which have concentric zoned inclusions, are formed by K-replacement processes, but the evidence in the Papoose Flat pluton and adjacent wall rocks certainly supports such a conclusion. See Collins and Collins (2002) with additional illustrations supporting a K-replacement origin for the orthoclase megacrysts.

**B. Twentynine Palms**

The same modern interpretation for the development of a porosity in solidified plutonic rocks in which introduced fluids allow K-feldspar megacrysts to form can also be applied to the megacrysts in the Twentynine Palms quartz monzonite as in the Papoose Flat pluton. However, we should first look at previous interpretations. There are two schools of thought regarding the origin of K-feldspar megacrysts in granitic rocks such as occur in the Twentynine Palms quartz monzonite. One school advocates their formation from a crystallizing magma, either having the K-feldspar slowly increasing in size (Hibbard 1965; Vernon 1986) or nucleating late and then growing rapidly to large size (Swanson 1977; Winkler and Schultes 1982). The second school favors their formation by K-metasomatism (Schwermerhorn 1956, 1966; Stone and Austin 1961; Cannon 1964). Mehnert and Busch 1985) thought both replacement and magmatic origins are possible in different terranes, but Vernon (1986) thought that all megacrystal rocks have a magmatic origin even when K-feldspar occurs in augen gneisses. In magmatic models, it is argued that K-feldspar must crystallize first as orthoclase, and that after crystallization, it may revert to microcline where rocks undergo stress (Laves 1950; Goldsmith and Laves 1962). Those favoring a K-replacement origin do not believe that formation of orthoclase is necessary prior to forming microcline (Schwermerhorn 1956, 1966).

A magmatic origin for the K-feldspar megacryst in the Twentynine Palms quartz monzonite in and near Joshua Tree National Park, southeastern California, USA, (Figures 54 and 56) is advocated by Miller (1938), Rogers (1955), Brand and Anderson (1982), Trent (1984), Brand (1985), and Hopson (1996).
Figure 54. Geologic map showing five locations (ab, upper left side; cde, lower right corner) where felsic diorite outcrops occur in areas less than mapping scale and where the diorite has gradational transition stages to the megacrystal Twentynine Palms quartz monzonite. This felsic diorite is older than the Twentynine Palms quartz monzonite but of unknown age. Locations are on file with the museum at Joshua Tree National Park.
Figure 55. Topographic and geologic map in Bartlett Mountains northwest of Joshua Tree, California (Hopson 1996), showing a sixth location of remnant felsic diorite interlayered with megacrystal quartz monzonite. (Joshua Tree North, 7 and 1/2 Minute Quadrangle, California, T1N, R6E, Section 21, center of SE quarter, four aligned red dots)

Microscopic textures, suggesting that fluids depositing K-feldspar have corroded and replaced plagioclase, were noted by Rogers (1955). He decided against K-metasomatism, however, because he did not find plagioclase crystals in the original rock as large as the largest megacrysts (up to 16 cm long). In the field, the Twentynine Palms quartz monzonite (Figures 54 and 55) appears magmatic because it contains large megacrysts of K-feldspar and lacks megascopic evidence for deformation (Figure 56).
While agreeing with Mehnert and Busch (1985) that both replacement and magmatic origins are possible for the formation of K-feldspar megacrysts, we present evidence that K-feldspar megacrysts in the Twentynine Palms quartz monzonite in and near the Joshua Tree National Park are not magmatic but are formed from K-metasomatism of diorite.

The megacrystal Twentynine Palms quartz monzonite contains pink to gray, subhedral to euhedral crystals of microcline (25-50 vol. %), which are 1-5 cm long in most places (Figure 56), but range up to 16 cm long. The megacrysts generally contain inclusions of groundmass minerals (10-20 vol. %), but in some places, inclusions are >50 vol. %. The inclusions occur in clusters or veinlike stringers and rarely show preferred orientations. Borders of megacrysts are generally quite irregular and enclose groundmass minerals, although megascopically appearing to have sharp edges. The megacrysts commonly contain plagioclase (An-15) as rod perthite (up to 20 vol. %). Locally, megacrysts have planar parallel orientations, but random orientations are more common (Figure 56).

The groundmass is dominantly plagioclase (45-65 vol. %) and hornblende (10-20 vol. %), but may also contain biotite (1-5 vol. %). Accessory minerals include apatite, allanite, sphene, quartz, magnetite, zircon, myrmekite, and epidote. Microcline (<10 vol. %) may be found in the groundmass, but in some places, only in megacrysts. Plagioclase generally contains few inclusions, is normally zoned and/or albite-, pericline-, and/or Carlsbad-twinned, has irregular borders, and ranges from 2-10 mm long.

The Twentynine Palms quartz monzonite contains K-feldspar megacrysts as much as 16 cm long, and like the megacrysts in the Papoose Flat pluton, the Twentynine Palms megacrysts also contain inclusions of granulated ground-mass grains in many places. In order to form these
megacrysts by K-metasomatic processes, the original host diorite had to undergo cataclastic deformation. Once the megacrysts had started growing, the diorite slowly transformed to a megacrystal quartz monzonite. This deformation was observed across 50 meters in a transition zone from the diorite into the megacrystal granite (Collins, 1997). Collins (1997) suggested that this diorite was the Gold Park diorite, but the Gold Park diorite is Jurassic; younger in age than the Twentynine Palms quartz monzonite which is Triassic (Trent 1998). Therefore, the Twentynine Palms quartz monzonite cannot replace it by metasomatic processes. On that basis, an older diorite (Figure 57) of undetermined age is suggested to be replaced by the Twentynine Palms quartz monzonite. The earliest stage of introduction of K-feldspar in microfractured diorite is illustrated in Figure 58. Here, the felsic hornblende diorite has a cataclastic texture, and the broken groundmass matrix consists of hornblende, quartz, and plagioclase. In early stages of replacement microcline engulfs the ground mass fragments along irregular borders and contains minor wartlike myrmekite with tiny quartz vermicules. Microcline composes 1 to 5 vol. % of the rock. Normal, undeformed, felsic diorite has crystals of hornblende and plagioclase all about the same size.

The remnant, older, fine- to medium-grained mafic diorite contains >20 vol. % hornblende ± biotite ± clinopyroxene, but also present are remnants of a medium-grained, felsic facies (Figure 57), containing <20 vol. % hornblende ± biotite. The felsic and mafic facies occur as small remnants in several locations (Figure 54, locations a to e).
Figure 57. Felsic diorite remnant layer (~0.5 m thick; left side under knife and at bottom left) in megacrystal Twentynine Palms quartz monzonite in Bartlett Mountains northwest of Joshua Tree, California (extending along the area parallel to the four red dots in Figure 55). Contact with the quartz monzonite is about 10 cm above the end of the knife. On right side of image, loose K-feldspar megacrysts have slid down an eroded slope of quartz monzonite (more than 100 m high) to cover portions of the felsic diorite layer.
**Figure 58.** Felsic hornblende diorite with cataclastic texture. Hornblende (brown and green), quartz (clear-white and gray), and plagioclase (gray and dark gray). Two microcline crystals (dark gray; center and left of center) with irregular borders against the granulated groundmass. Myrmekite is on microcline borders. Source: Collins (1997g, Figure 5).

Where cataclasis of the diorite was first modified by metasomatic fluids, the ferromagnesian silicate minerals (mostly hornblende but minor biotite) were replaced by quartz to form a quartz sieve-texture (Figure 59). In nearby diorite that is undeformed the hornblende lacks any quartz inclusions. Just like in the Temecula area (Figures 4 and 24), where hornblende in diorite is being replaced by quartz, a subtraction of Mg, Fe, Ti, and Ca happened to the diorite in its early conversion to the Twentynine Palms quartz monzonite.
**Figure 59.** Felsic diorite with quartz sieve-texture in hornblende (brown and green). Microcline (gray; less than 1 vol. %). Quartz (white, gray, and cream). Albite-twinneed plagioclase (white and gray; upper left). Source: Collins (1997g, Figure 6).

In the transition to where the megacrysts become 3-5 cm long, the K-feldspar megacrysts first begin to make their appearance as 1-cm-sized crystals. Wartlike myrmekite commonly forms here (Figure 60).
Figure 60. Microcline (black) encloses, penetrates in veins, and replaces plagioclase (cream-gray, light gray). Large plagioclase islands are optically continuous with adjacent tiny islands (blebs) of plagioclase in microcline. Large plagioclase islands are locally myrmekitic, containing elongate-oval cross-sections of quartz vermicules (black to gray). Source: Collins (1997g, Figure 10).

Convincing evidence that the K-feldspar megacrysts in the quartz monzonite pluton were formed by K-replacement processes is the fact that many megacrysts contain more than 50 percent granulated ground mass minerals (Figure 61). In some places the scattered inclusions are so abundant that the K-feldspar of the megacrysts is difficult to see. In the field, however, these abundant inclusions cannot be seen on the exterior faces of the megacrysts. In other places the inclusions are less abundant (5 to 20 vol. %), and they are not in concentric zones of aligned inclusions as are in megacrysts in the Papoose Flat pluton. However, tiny euhedral titanite crystals that are in the ground mass remain in the megacrysts because they are not easily replaced.
Figure 61. The entire field of view is a small portion of a microcline megacryst, 2 to 3 cm long with more than 50 vol. % ground mass inclusions, which in outcrop looks like a megacryst in Figure 56. Quartz and plagioclase (gray and white). Microcline (gray) can hardly be seen. Tiny sphene (titanite) crystals (brown). Hornblende is absent and presumably replaced by quartz and/or microcline. Source: Collins (1997g, Figure 11).

In one place, a thin section of a megacryst has a band of granulated ground-mass mafic grains that extends from outside the megacryst on one side and then through the megacryst where it continues out the opposite side into the granulated ground mass without any offset of the megacryst or the band. This relationship is too large to photograph at low power (40x) under the microscope but is readily apparent in the thin section.

Because of the above evidence, it is clear that the K-feldspar megacrysts have grown below solidus conditions in cataclastic broken rock that provided an open system through which metasomatic fluids carrying large volumes of K ions could have moved freely and converted the older hornblende diorite into the Twentynine Palms hornblende quartz monzonite on a plutonic scale. The ground mass surrounding the megacrysts is dominantly plagioclase (45-65 vol. %) and hornblende (10-20 vol. %), but the amount of hornblende in the diorite is generally greater than 20 vol. %.

Most plutonic igneous rocks that contain such large K-feldspar megacrysts in great abundance (25 to 50 volume percent) would be expected to be biotite rich on the basis of (a) the large amounts of K needed to produce such large volumes of K-feldspar and (b) the richness of Fe and Mg in the diorite which would be expected to react with K to make abundant biotite. But that is not the case here. Generally, only 1 to 5 vol. % biotite is present, and in most places it is only 1 vol. %. The absence of much biotite in the Twentynine Palms quartz monzonite that replaced the diorite is logically explained by the absence of much biotite in the original
magnatic diorite. Therefore, the K-metasomatic origin of the K-feldspar megacrysts in the Twentynine Palms quartz monzonite is a logical conclusion. Nevertheless, the production of such large volumes of K-feldspar megacrysts must require the addition of much K from an outside source in order for the K-metasomatism to occur. Locally, the pluton has xenoliths of biotite schists in which the tiny biotite flakes are as much as 90 vol. %. Therefore, such schists (carried up from metasedimentary rocks below) suggest that a deep source of K is present.

**C. Vrådal pluton**

Another example where myrmekite that is associated with K-feldspar megacrysts is a clue to large scale metasomatism is in the Vrådal pluton in Norway. The Vrådal pluton (967 ± 4 ma) (Figure 62) is one of several relatively undeformed Sveconorwegian granitic plutons that intruded the South Telemark Gneiss (ca. 1210 ma) in southern Norway (Sylvestre 1964; Sylvestre 1998; Andersen et al. 2001; Andersen et al. 2007).
Figure 62. Vrådal pluton location in southern Norway (Vr; tiny red oval near center of figure in white area).

It is a nearly circular diapiric pluton, 6 km in diameter, with a central core, about 4 km in diameter (Figure 63).
Figure 63. Model of the Vrådal pluton.

The pluton surface-exposure (Figure 64) consists of white and pink granodiorite surrounded originally by a 750-1000 m-wide shell of mafic granodiorite. K-feldspar megacrysts, 2-5 cm long and strongly oriented concentric to the pluton margin, grew metasomatically in the mafic granodiorite, converting it into megacrystic granite (Figure 64, yellow; Figure 65, and Figure 66).
Figure 64. Geologic map of Vrådal pluton, Telemark, south Norway (Sylvester 1998). Gneiss (gray); schistose amphibolite (blue); quartzite (green); diabase (dark pink), core granite (pale pink); border facies megacrystic granite (yellow). Diameter of diapir is about 6 km.
Figure 65. Megacrystic granite in Vrådal pluton

Figure 66. Schematic diagram of how matchbox-shaped K-feldspar megacrysts define a foliation in the border facies of the Vrådal pluton. Preferred orientation of K-feldspar megacrysts with ac axes parallel to contact and pluton margin.

Sylvester (1964) originally thought the K-feldspar megacrysts were rapakivi type and magmatic in origin. However, because the senior author found myrmekite (Figures 24 and 67) in his thin section collection, it provided a clue that original primary plagioclase in the pluton
could have been deformed and replaced to form K-the megacrysts. Subsequently, evidence described in this section supported this model.

Figure 67. Myrmekite in Vrådal pluton. Myrmekite (lower left quadrant) projects into microcline (grid twinned). One white squiggly-vermicule is seen, but five other vermicules (ovoid) are also present. Albite-twinned plagioclase (altered speckled tan) projects into the microcline and its albite twin-planes (barely visible) are parallel to one of the grid-twins of the microcline. Quartz is white and biotite is reddish tan.

An example of the deformation of plagioclase is shown in Figure 68. Commonly, the primary plagioclase has a relatively calcic core and a sodic oligoclase rim. In some places the oligoclase rims have tiny blebs of quartz in a granophyric texture, commonly formed during last stages of crystallization from magma. (Drawings of thin section textures were done by Arthur Sylvester.)
Figure 68. Bent and microfractured primary plagioclase crystals. Yellow represents quartz.

Where the plagioclase is slightly microfractured, the interior contains islands of microcline, but these islands are not uniformly distributed as is common in antiperthites of magmatic origin (Figure 69). This figure shows early stages of K-metasomatism of the primary plagioclase crystals.

Figure 69. Incipient replacement of plagioclase (light gray) by scattered irregular islands of microcline (black and white). Microcline grid-twinning is parallel to plagioclase twin lamellae. Brown is biotite.

Where plagioclase crystals are strongly microfractured in their outer margins, then the outer sodic oligoclase rim is also replaced (Figure 70).
Figure 70. Replacement of Carlsbad- and albite-twinne
plagioclase (light gray) in core of crystal by K-feldspar along an irregular fracture. Outer
oligoclase rim is replaced by K-feldspar (black) that also penetrates the plagioclase in albite
twin-lamellae (right end of plagioclase core).

Because the relatively-calcic cores of the primary plagioclase crystals are less stable
than the sodic oligoclase rims at temperatures below the solidus, the islands of microcline expand
their replacement until nearly all the core is replaced (Figures 71, 72, and 73).

Figure 71. Calcic core of Carlsbad twinned plagioclase replaced by patches of microcline (gray;
grid-twinne, black and white) out to a sodic oligoclase rim (light gray).
Figure 72. Dark gray and white, optically continuous, string-like remnants of albite-twinned plagioclase in Carlsbad-twinned K-feldspar. Almost complete replacement of plagioclase by microcline (whitish gray; light gray).

Figure 73. Epitaxial replacement of calcic core of a Carlsbad/albite twinned plagioclase by Carlsbad twinned microcline (light gray and white). The primary, more sodic oligoclase rim is not replaced. Patchwork island remnants of plagioclase have the same optical orientation as the unreplaced rims.
The epitaxial replacement of the calcic core of a Carlsbad/albite twinned plagioclase by Carlsbad twinned microcline (Figure 73) happens because the crystal structure of the plagioclase and the microcline are closely related and nucleation of the microcline on the plagioclase structure surface is favored because the interfacial energy at this new interface is reduced (see Putnis 1992, p. 338). Thus, epitaxy transfers the crystallographic information from the plagioclase to the microcline even though the plagioclase dissolves. Experiments by Putnis and Mezger (2004) and Putnis et al. (2005) emphasize that the fluid composition at the interface is the most important factor, and that during the replacement this may be different from the fluid composition in the bulk. The composition of a boundary layer of fluid at the interface is likely to be the most important factor which controls the coupling (Putnis A. 2009).

The suggested explanation for the origin of the K-feldspar megacrysts is as follows. As the diapir rose, its cooler nearly solidified outer shell was deformed by cataclastic flow. Early formed plagioclase crystals cracked when the pluton's core continued to rise diapirically relative to its almost solidified outer shell. The microfractures were loci of, and pathways for, pervasive hydrothermal solutions containing Si that wholly replaced hornblende and some biotite to form quartz and K that partially and wholly replaced plagioclase to form K-feldspar. The relatively sodic oligoclase rims on the primary plagioclase crystals are relatively stable at the lower temperatures at which the K-feldspar replacements formed, whereas the more calcic plagioclase cores were unstable. Therefore, in many places the K-replacements of the plagioclase were from the interior outward. Because of this, some megacrysts have secondary oligoclase mantles that were once mantles on the primary plagioclase, and, as a result, the megacrysts appear as if they were primary rapakivi phenocrysts. Where K-replacements were incomplete, locally myrmekite was formed. Thus, K-feldspar progressively replaced plagioclase in the outer-shell mafic granodiorite. In some places ghost plagioclase zoning is preserved in K-feldspar of partly replaced plagioclase crystals. One plane of the microcline grid twinning typically parallels the Carlsbad/albite twinning of the plagioclase. Because the K-feldspar megacrysts are 2-4 times larger than the replaced plagioclase, they subsequently grew beyond the boundaries of the original plagioclase and enclosed remnants of the oligoclase rims and groundmass minerals. They inherited their concentric preferred orientation from the magmatic flow fabric of primary plagioclase in the mafic granodiorite. Although some K that went into the K-feldspar megacrysts came from the breakdown of biotite in the mafic granodiorite, much K must have come from breakdown of older, felsic, metaigneous rocks in the pluton's source region. The quartz mode increased from 15 to 25%, and biotite (5%) remained in the megacrystic granite. Some Ca released from replaced plagioclase went into titanite. Excess Fe was precipitated as secondary magnetite or as hematite crystals in cavities in the pink K-feldspar megacrysts (Putnis et al., 2007). Excess Ca and Mg were subtracted from the system. Significantly, finding out that the Vrådal pluton had a large-scale metasomatic history was initiated because of the detection of relatively abundant myrmekite in a thin section collection.
Because the K-bearing fluids moved through the deformed outer shell of the Vrådal pluton and replaced primary plagioclase crystals there, the adjacent gneissic wall rocks ought to have been deformed and modified by these same K-bearing fluids as well, and this is observed (Figure 74).

Figure 74. Perthitic K-feldspar (light gray; right side) replaced parts of a broken albite-twinned plagioclase crystal (light tannish gray; left side) and enclosed and partially replaced quartz (remnant ovoid grains; white; right side). K-feldspar has penetrated and replaced plagioclase (center grain) along albite-twin lamellae. Sample is from the Telemark gneiss.

The Vrådal pluton became replaced on a large scale because the rising of the plutonic mass after solidification of the intruded magma created applied stress on the crystals in the outer rim of the diapiric circular cross-section. The applied stress opened the system to development of a porosity through which fluids coming up from greater depths could add K and Si while subtracting Na, Ca, Al, Fe, and Mg, completely changing the mineralogy of the rocks as K-feldspar megacrysts grew with their long dimensions parallel to the outer boundaries of the pluton. Similar replacements to produce K-feldspar megacrysts have also occurred in outer rims of rising plutons as in the Ardara pluton in Ireland (Pitcher and Berger 1972; Pitcher et al. 1987; Collins 1997h) and the Cooma pluton in Australia (Collins 1998b).
VII. STRONTIUM AND OXYGEN ISOTOPE STUDIES

$^{87}\text{Sr}/^{86}\text{Sr}$ studies in metasomatically altered rocks have been conducted in the southern Sierra Nevada of California. Here, a biotite hornblende diorite and pyroxene gabbro mass lies adjacent to quartz diorite of the granodiorite Isabella pluton, northeast of the city of Isabella (Figure 75). The pyroxene gabbro facies is next to the Kern River valley that extends north of Isabella and also west of the river (Collins 1988). The plutons described in this section and in other parts of this general area have been re-mapped, re-named, and dated by U-Pb studies (Saleeby et al. 2008). The quartz diorite and western portions of the Isabella pluton granodiorite north of the South Fork of the Kern River are now named the Cretaceous Rabbit Island facies and the L-shaped hornblende diorite and gabbro mass is the Cretaceous Cyrus Flat facies. Both facies are part of the South Fork Intrusive Suite. The proto-Kern Canyon fault extends through these facies in a broad zone of deformation, and the later Kern Canyon fault extends west of these facies along the Kern River valley (Saleeby et al. 2008). Neither of these faults are shown on Figure 75.

Figure 75. Geologic map showing the location of the diorite facies (dark speckled area) of the Isabella Pluton. The “L-shaped” and northward tapering body of this diorite (left side of map) extends roughly parallel to the Kern River that flows past the city of Isabella (left corner).

The southern part of this hornblende diorite mass is traversed by erosion-resistant granite aplite-pegmatite dikes standing up like walls above the surrounding mafic rocks. Outwardly, they appear as if they had been formed by injection of magma into fracture openings. However, no sharp contacts exist. Each dike is bordered by migmatites which then grades into foliated diorite and finally into massive diorite (Collins 2004a; Collins 2004b). See schematic drawing, Figure 76.
Figure 76. Schematic drawing of aplite-pegmatite dike. See Collins (1988; Collins 2004a; Collins 2004b) for further discussion of details 0, 1, 2, and 3.

The aplite-pegmatite masses formed where movements create foliation planes, which are then bent into S-shaped drag folds in a dextral direction. Then, slight dextral sliding along each foliation plane on both sides of the center of the consecutive S-shaped drag folds pulls apart each foliation plane in the center of the drag folds for the whole length of the folded zone. In that way “openings” or relatively low-pressure sites were created parallel to the length of the granitic dike that eventually forms in these low-pressure sites. These “openings” allowed hydrous metasomatic fluids to enter (Figure 77). The dextral movements that created the S-shaped drag folds also caused microfracturing, broke grain boundary seals, and allowed pores in stressed crystals to be produced (Putnis A. 2009). These S-shaped drag folds are similar to S-C mylonites described by Lister and Snoke (1984) except that the Isabella rocks are not deformed mylonites in which no change in mineralogy occurs and are much larger features (>15 m) and lack mica “fish.”

Figure 77. Bent-foliation zones (exhibiting dextral shearing) in which aplite-pegmatite dikes (3-3’) formed in the low-pressure sites in the center of the S-shaped drag folds (between C-C’ and
These same fluids progressively modified the chemical and mineralogical compositions of the deformed and foliated diorite. As a result, a gradational change can be observed between the massive diorite and the dikes along each foliation plane in the S-shaped drag folds because of the replacement of hornblende and biotite by quartz and relatively-calcic plagioclase (An-34 to An-48) by both K-feldspar and more sodic plagioclase. About half way to the dike where K-feldspar first appears, myrmekite also is present. The result of all these replacements are the pegmatite-aplite dikes in the center of the drag folds. Because of the net loss of Fe, Mg, Al, and Ca during this process, shrinkage occurred when the residual rocks recrystallized into masses of more-granitic composition in the pegmatite-aplite dikes (Figure 78). This shrinkage is similar to what occurred in the limbs of the fold in the Gold Butte area and the Split Rock Pond anticline, described earlier.

**Figure 78.** Two alternatives show what happens if (A) the granitic dikes were formed by injection of magma into fractures, resulting in expansion to make room for the magma or if (B) loss of elemental components causes shrinkage to occur.

Because these dikes in the diorite mass increase northward in abundance from only a few to hundreds of them, the combined shrinkages in the diorite was sufficient to cause the mass to taper northward, as shown in Figure 75. Additional shrinkages occur because the former diorite adjacent to these dikes is converted to myrmekite-bearing alaskitic granite in which most of the primary biotite and hornblende (15 to 25 vol. %) are converted to residual magnetite grains (1 to 5 vol. %) plus quartz (15 vol. %). The parallel deformation, granitic dikes, K- and Si-metasomatism, and shrinkages were made possible because the broad proto-Kern Canyon fault extends through this area (Saleeby et al. 2008).

It is logical that the chemical modifications of the deformed diorite mass should also affect the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios and reset the atomic clock. The reason for this is because the various Sr isotopes including $^{86}\text{Sr}$ would leave the system at the same time that Ca is being subtracted from hornblende and plagioclase and because K and $^{87}\text{Rb}$ would remain behind and be concentrated in
the residual biotite and K-feldspar. In the transition zone the gradual change in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios is revealed in Figure 79 as their values (data points) shift in position upward and to the left from the lower, older age-line projection (120 m.y.) to the upper, younger age-line (83 m.y.). Later dating by Saleeby et al. (2008) gives a U-Pb date for the Cyrus Flat diorite-gabbro at 99.6 m.y. instead of 120 m.y. and the Rabbit Island diorite-granodiorite at 97-99 m.y. On that basis, the younger age of 83.3 m.y. for the metasomatically created granitic dikes likely indicate younger movements along the fault zone that opened up the rocks to K- and Si-metasomatism.
Figure 79. Isochron plot of (1) biotite-hornblende diorite and hornblende-pyroxene gabbro wall rocks, (2) transition rocks, and (3) granitic aplite dikes in the Kernville pluton; see text. Wall rock samples 12-16, dated by Rb-Sr methods are 120.1 +/- 5 million years old. The other samples give an average date of 83.3 million years old. Solid circles indicate diorite, sample nos. 14 and 15; open circle indicates sample no. 16 (foliated diorite adjacent to the Isabella pluton). Black squares indicate modified diorite (nos. 33, 40, and 44); squares shaded in the northwest half indicate transition rocks (nos. 34, 38, and 39); and squares shaded in the southeast half indicate migmatized diorite and aplite dike samples in which plagioclase is replaced by microcline and myrmekite (nos. 35, 36, 37, 41, 42, and 43). Numbers adjacent to symbols refer to sample numbers in Tables 1, 2, and 3, and Tables VI and VII in Collins (1988).

On the basis that these changes in $^{87}\text{Sr}/^{86}\text{Sr}$ data points support the model that these rocks in the Isabella area have been modified metasomatically on a large scale, it is logical that similar studies could be made in the Temecula area. We, therefore, propose SHRIMP micron-scale $^{87}\text{Sr}/^{86}\text{Sr}$ analyses across the 10 cm transition zone from hornblende diorite to the myrmekite-bearing granite that could show progressive changes in $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Similarly, ion microprobe work for $\delta^{18}\text{O}$ traverses should be done.
Moreover, ion microprobe work for $\delta^{18}$O studies could also be made on the primary plagioclase and metasomatic K-feldspar megacrystals in the Vrådal pluton to see if different temperatures for the magmatic and metasomatic phases are present. Finally, studies of mobile and immobile trace elements would also be of interest.

VIII. POLONIUM HALOS AS EVIDENCE FOR THE ORIGIN OF GRANITIC ROCKS BY K-METASOMATISM

In addition to myrmekite, Po-halos in biotite mica are other indicators for the metasomatic origin of some granitic rocks of plutonic dimensions (Collins and Collins 2010). The deformation that opens up the system for migration of K and Si and enables K- and Si-metasomatism also opens the system to migration of radon gas released from scattered uranium atoms in the primary plutonic rocks. The isotope $^{222}$Rn, which is a daughter product of uranium $^{238}$U, is the precursor to $^{218}$Po, $^{214}$Po, and $^{210}$Po. Where $^{222}$Rn along with these three isotopes migrate in fluids to low pressure sites in microfractured biotite crystals, three different kinds of Po-halos can form in the biotite crystal structure (Gentry 1988; Collins and Collins 2010, Figure 2). Whether one, two, or three Po-halos are created depends on the distance and time of travel of the radon gas and the polonium atoms from the uranium source. If the distance and time of travel is long, perhaps only $^{210}$Po-halos are formed (Figure 80; Collins and Collins 2010, Figure 3). Because radon is a noble gas and electrochemically neutral, it migrates, therefore, without any flow-resistance to the low pressure sites; polonium, on the contrary, occurs as charged ions which are carried along with the radon in migrating steam.

![Figure 80](image_url). Po-210 halo (black) in biotite along a healed fracture. Alpha particle damage from radioactive $^{210}$Po ions moving along a former microfracture occurs as a dark band.
Po-halos are absent in biotite in most primary magmatic granitoids. When they are found in granitic rocks, they occur together with wartlike myrmekite and are consequently additional clues for the formation of these rocks by metasomatic processes. An exception, however, are Po-halos that form in large biotite crystals (as much as 10 cm wide; Gentry 1988) in pegmatites containing uranium-bearing minerals which are produced during last-stage crystallization from magma (Collins and Collins 2010). Not all metasomatically produced granitic rocks, however, contain Po-halos in biotite because not all primary relatively-mafic igneous rocks converting to metasomatic granite contain an abundant supply of uranium. In rocks lacking abundant uranium, myrmekite is the only clue to their metasomatic origin.

Creationists, such as Robert Gentry, have always argued that because the geologic community did not accept the model that some granite masses on a plutonic scale can be formed by metasomatic processes, his model for forming them by fiat during the third day of the Genesis Week had equal validity. Now, Collins and Collins (2010), Baillieu (2010), and this article provide strong scientific arguments to support the model that Po-halos can be formed by natural processes.

**IX. CONSEQUENCES OF K-METASOMATISM**

One consequence to consider about the process of K- and Si-metasomatism that changes relatively-mafic plutonic igneous rocks into rocks of more granitic composition is that the new crystals that are formed have different volumes and densities as elements (Ca, Na, Al, Fe, and Mg) are dissolved out and carried away. The porosity that is produced in stressed rocks allows introduced fluids to move through the crystals which change their elemental compositions, but the conversions are not mass-for-mass but volume-for-volume, and not even at equal volume. When biotite and hornblende (density 3.0) are replaced by quartz (density 2.65) to form a quartz sieve-texture and eventually only quartz, the space occupied by the quartz is generally less than the space once occupied by the biotite and hornblende. Thus, shrinkages in local volume exist which then put stress (added pressure) on adjacent primary plagioclase crystals, causing them to be susceptible to continued extension of more pores through these crystals. Thus, in the nearby stressed plagioclase crystals, fluids can remove more Ca, Na, and Al and create more pores into which K can be introduced to form K-feldspar. Moreover, even though the density of K-feldspar (G = 2.56) is lower than the density of primary plagioclase (G = 2.61 to 2.66 or higher), the expanded K-feldspar crystal structure that is created by the introduced K need not fill all the space once occupied by the plagioclase, and, again, local shrinkages can occur. Therefore, additional stress is created locally on adjacent primary biotite, hornblende, and plagioclase, causing further development of porosity in these crystals. On that basis, during the replacement processes, the system is constantly being re-opened for more metasomatic fluids to come in so that more elements can move out in escaping fluids. Thus, K- and Si-metasomatism continues and spreads to increase the volume in which metasomatism occurs. Once K- and Si-
Metasomatism begins, it is almost self-perpetuating. Only when sources of K and Si are depleted in introduced fluids is the metasomatism stopped.

We have already discussed the fact that although the volume losses (shrinkages) are small, when considered on a plutonic scale, as former relatively more-mafic rocks become more granitic, the shrinkages may be significant. This was illustrated for the isoclinal structures that occur (a) in the Gold Butte area in Nevada (Figures 36 and 37; Collins 1997h), (b) in the Split Rock Pond area in New York (Collins 1997l; Collins 1997m; Collins, 1988a), and (c) along the proto-Kern Canyon Fault (Collins 2004a; Collins 2004b). Therefore, potentially, shrinkages could happen in igneous rocks in metamorphic core complexes that are modified by K-metasomatism. Such a metamorphic core complex exists in eastern California and western Arizona where igneous rocks are commonly mylonitized and granulated along horizontal thrust sheets (Davis et al. 1980; Anderson et al. 1979). In an unpublished master’s thesis Podruski (1979) described K-metasomatism to form K-feldspar megacrysts (bordered by wartlike myrmekite) in the Whipple Mountains in the Parker Dam granite that is within this core complex. Therefore, not only were the igneous rocks granulated along the contact between upper and lower plates, they also likely shrank in volume from loss of elements in escaping fluids moving through the contact between the plates as these rocks were converted to a more-granitic composition by K-metasomatism.

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Another locality is in the Alps in Switzerland where rock layers have been thrust past each other during plate tectonics and where fracturing likely accompanied these movements. For example, the Bergell granite (Wenk 1982; Schmid, et al. 1996) that occurs there contains megacrysts of K-feldspar bordered by wartlike myrmekite, and, therefore, a zone of shrinkage could have happened as a former igneous pluton was converted into the more-granitic Bergell granite body by K-metasomatism.

In addition to the losses of the major elements, trace elements and metallic elements are also likely removed by the through-going fluids. The metallic elements could form ore deposits at large distances from their transformed source-granitic rocks (Oliver et al. 2004; Clark et al. 2005; Engvik et al. 2008). Uranium ore concentrations can also be associated with rocks
containing both Po halos in biotite and myrmekite-bearing granitic rocks in the Bancroft District in Ontario, Canada, where deformation and an open system allow migration of radioactive uranium, radon and polonium (Wakefield 1988; Collins 1997f). Where the uranium concentrations are associated with syenite, no myrmekite is present because the rocks are affected by Na-metasomatism instead of K-metasomatism.

Because $^{18}$O isotopes are somewhat fixed in the original plagioclase crystal structures and at ratios with $^{16}$O that indicate the high temperatures of crystallization from magma, these ratios may still be inherited so that high temperatures may be indicated for the K-feldspar even though the K-feldspar crystallized at temperatures below the solidus. On the other hand, in some places the ratios may readjust to indicate temperatures of crystallization below the solidus. Of course, if surfaces waters are introduced, then $\delta^{18}$O/$^{16}$O ratios could be reset to temperatures less than 350°C. But likely at the depths in which plutonic rocks are being transformed into rocks of more granitic composition by metasomatic processes, the source of the fluids is at lower depths and not from surface waters.

Because zircons are relatively stable over a large temperature range and likely not be affected by through-going metasomatic fluid, they still may produce U/Pb age dates indicative of the time in which the original magmatic rocks were crystallized even though these rocks have been replaced by younger granitic rocks. In contrast, because Rb could come in with introduced K and because much of all the Sr isotopes will go out simultaneously with displaced Ca, the $^{87}$Rb/$^{87}$Sr clock will likely be reset so that the timing of the transformation of the original rock into a younger rock of a more granitic composition can be expected and determined (Collins 1988).

An additional consequence is the consideration of the problem of where do large volumes of Ca, Na, and Al ions go that are displaced by introduced K that forms secondary K-feldspar? The places where these ions have gone has been almost impossible to find because where the metasomatized granitic rocks can be observed, the overlying rocks into which the escaping fluids could have moved have been eroded away. Perhaps, they have produced metasomatic quartz-bearing anorthosites as is in the Komaki District of southwestern Japan (Takagi et al. 2007). Granitic rocks immediately adjacent to the Komaki District contain secondary K-feldspar and myrmekite. Of course, the connection with deep-seated K-metasomatic granite below the plagioclase-quartz rock which would supply Ca, Na, and Al to the anorthosite cannot be demonstrated without extensive drilling. Nevertheless, the introduced Ca, Na, and Al into anorthosite must come from somewhere, and the circumstantial evidence suggests that the connection to myrmekite-bearing granite as a source could be possible.

Another possibility of where the Ca, Na, and Al go is in the production of concentrations of secondary epidote. For example, in some places in the Cargo Muchacho Mountains in southeastern California (Collins 1988a) and in the megacrystal quartz monzonite at Twentynine
Palms, California (Collins 1997g) epidote is present to a large extent in the host rocks, well exceeding the amount that could possibly be derived through release of Ca and Al from altered primary minerals. A simple relatively-closed-system alteration to produce epidote will usually have a single generation of epidote. A complicated open-system-alteration to produce epidote during metasomatism will have multiple generations of epidote (Barth and Ehlig 1989) because K-metasomatism requires the system to be kept open by repeated episodes of deformation.

The Cargo Muchacho Mountains are situated in California east of the San Andres Fault and extend to the west of Yuma, Arizona. The mountains consist of biotite-hornblende diorite as well as tonalite plutons that have been extensively faulted, fractured and microfractured (Collins 1988a). They belong to terranes where both gold and aluminum have been concentrated by metasomatic processes on a large scale. Originally, some of these metasomatically altered rocks were mapped as metasedimentary rocks in the Tumco Formation (Sweeney and Bradley 1981), but such rocks are gradational along and across strike to myrmekite-bearing quartz monzonite and then to less-deformed and altered diorite and tonalite. Thus, kyanite, tourmaline (schorl), and muscovite (enclosed in phyllites) are secondary and represent residual concentrations of aluminum left over from altered plagioclase, hornblende, and biotite after their replacement in earlier stages by K-feldspar, myrmekite, and quartz. Kyanite occurs in druses where it projects into cavities. Therefore, these crystals cannot result from contact metamorphism but rather are deposited in an open system in which fluids were able to move through microfractures. Furthermore, so much disseminated gold was concentrated in the metasomatically altered myrmekite-bearing quartz monzonite masses that open pit-mining for gold was feasible in the western section of the mountains near the San Andreas Fault and also in the interior. The conclusion is that (in an area larger than 35 square kilometers) both gold and aluminum occur as large-scale products of K-metasomatism in the Cargo Muchacho Mountains.

Where do displaced Mg and Fe go that are released from biotite and hornblende as these minerals are replaced by quartz? Notably, Mg moves out preferentially over Fe, because of the greater solubility of Mg and because Fe tends to remain behind in hematite and magnetite. Both iron oxides have densities (G = 5.2) greater than the densities of ferromagnesian silicates (G = 3.0) that are replaced by quartz, so the formation of residual hematite and/or magnetite contributes to the shrinkage of the rocks during metasomatic processes. If Fe is abundant, concentrations of magnetite may be produced, such as magnetite enrichments in the outer shell of the Vrådal pluton in Norway (http://www.csun.edu/~vcgeo005/Abstract%20Vradal2.htm), and the magnetite iron ore deposits in the limbs of the Split Rock Pond anticline in New York (Collins, 1988a). Because ore deposits are associated with rocks that are being albited (Oliver et al. 2004; Clark et al. 2005; Engvik et al. 2008), it is not unreasonable to expect ore deposits also to be associated with rocks that have been subjected to K-metasomatism. Because soluble Mg moves out in escaping hydrothermal fluids preferentially over Fe, the escaping Mg could be the means for causing overlying limestones to be converted into dolomites. It has always been a problem as to where the Mg comes from in limestones that cause progressive conversions to dolomite in layers buried far below Mg-bearing oceanic waters.
Other possibilities are lamprophyre dikes that have a puzzling association with granitic rocks in some places (Hyndman 1972; Bowes and Wright 1976). Their high Fe- and Mg-contents make no geologic sense if they are derived from granites, and, therefore, a deep mantle source is usually postulated as to where their Fe- and Mg-rich compositions come from. On the other hand, if these dikes represent deposition from hydrous fluids enriched in Fe and Mg that were escaping from rocks undergoing K- and Si-metasomatism, a deep mantle source is not necessary and makes geologic sense (Collins 1988a).

If biotite and hornblende contain small percentages of Cu, Pb, Mo, and/or Zn in their crystal structures, these metallic elements could be released into the hydrothermal fluids to form sulfide ore deposits in wall rocks. Likewise, Au and U could be released to form local ore concentrations. At any rate, mostly where elements go that are extracted from rocks that undergo K- and Si-metasomatism has to be speculation because it is almost impossible to connect any granitic sources containing wartlike myrmekite with the final destination of where escaping fluids go.

Unlike rocks affected by K-metasomatism, rocks affected by Na- and Ca-metasomatism, described in the next section, generally do not have coexisting great losses in volume because Na- and Ca-metasomatism mostly affect small volumes in outer rims of primary K-feldspar or plagioclase in which the replacements cause little to no shrinkage.

X. MYRMEKITE AND TEXTURES FORMED BY Ca- AND Na-METASOMATISM

In contrast to textures that are formed where K-metasomatism modifies primary plagioclase to produce secondary K-feldspar and wartlike myrmekite are textures due to pure Ca- and combined Ca- and Na-metasomatism. They have to be added to textures of pure Na-metasomatism observed: (a) on the perimeter of carbonatite-alkaline silicate complexes (accompanied by the formations of distinct sodic ferromagnesian silicate minerals like aegerine, arfvedsonite, and riebeckite) (Winter 2001) and (b) in sheared rocks (with albite replacements of microfractured, relatively more-calcic plagioclase (e.g., Engvik et al. 2008; Morad et al. 2010; and Plümper and Putnis 2009). Included are therefore: (a) Na-metasomatism of mesoperthites, (b) combined Na- and Ca-metasomatism of mesoperthites, interiors of microfractured orthoclase phenocrysts, exteriors of primary K-feldspar, grain boundaries contacts between two K-feldspar crystals, and (c) Ca-metasomatism of borders of primary more-sodic plagioclase and resulting from reaction rims between Ca-bearing pyroxenes and adjacent plagioclase.

Textures formed in mesoperthites can be observed in the Lyon Mountain granite gneiss in New York where both primary albite-twinned plagioclase and mesoperthite with coarse plagioclase lamellae exist. Na-metasomatism is easily demonstrated where secondary plagioclase replaces the K-feldspar portion of mesoperthite (Figure 81). During this process
remnants of coarse plagioclase lamellae remain in the secondary plagioclase. However, because the introduced metasomatic fluids contain mostly Na, the secondary plagioclase in such places is relatively sodic, and the exchange of Na for K does not produce any left-over silica to form quartz vermicules.

**Figure 81.** Replacement of K-feldspar (gray; right and bottom sides) in the mesoperthite by secondary plagioclase (dark gray; center) leave plagioclase lamellae (cream) of the perthite projecting into the secondary plagioclase or as islands. Albite-twinne primary plagioclase is in the upper left quadrant. Source: Collins (1997m, Figure 21). See also Figure 22 in this same article for a similar example.

On the other hand, where Na- and Ca-bearing fluids interact with the mesoperthite, (a) the added Ca requires two Al atoms for every Ca atom, and Si is left-over to form quartz vermicules in myrmekite, (b) the vermicules do not taper toward the K-feldspar and are totally inside the albite-twinned plagioclase (center) of the myrmekite, and (c) the vermicules commonly are nearly uniform in cross-section (Figure 82).
Figure 82. Myrmekite formed by Na- and Ca-metasomatism of K-feldspar in mesoperthite. Some quartz vermicules are seen as cross-sections which have an ovoid shape. Plagioclase of the myrmekite is albite-twinned. Source: Collins 1997m, Figure 23).

An example of Na- and Ca-metasomatism of interiors of microfractured primary orthoclase phenocrysts occurs near Alastaro, Finland, where megacrystal granite intrudes gabbro. Aggregate mats of myrmekite are in the microfractured orthoclase crystals, 2 cm long (Figure 83).
Figure 83. Photo of part of a mosaic of magnified images of an Alastaro altered K-feldspar megacryst. The mosaic extends for about 2 m long and 1 m wide. The assembled mosaic image is an aggregate of myrmekite grains filling almost the entire K-feldspar crystal (about 2 cm long in its unmagnified form). Some unreplaced remnant inclusions of quartz and biotite are scattered among the myrmekite grains. The photo includes about one sixth of the mosaic area. Source: Collins (1997d, Figure 7b). See also Figures 2 and 3 in this same website article for early stage Ca-metasomatism along fractures in the interiors of megacrysts.

On the other hand, there are some kinds of wartlike myrmekite that project into primary K-feldspar along its outer rim instead of as replacements in the interior along microfractures (Figure 84). In such myrmekite the plagioclase has uniform composition, and the quartz vermicules are uniform in size and do not taper toward the K-feldspar. Moreover, the plagioclase of the myrmekite is not gradational to and in optical continuity with adjacent quartz-free plagioclase as occurs during K-metasomatism. An example of such wartlike myrmekite is in a thin section of monzogranite from the Alvand plutonic complex, Hamedan, Iran (Figure 84).
Figure 84. Wartlike myrmekite from Iran. Microcline (Mic; right side). Albite- and Carlsbad-twinned plagioclase (Pl; left side). Muscovite (Ms; green, blue, and red). Biotite (tan). Quartz (cream white; upper left). Image is from monzogranite sample from the Alvand plutonic complex, Hamedan, Iran. Photo contributed by Ali Sepahi.

In a more advanced stage of Ca- and Na-metasomatism of primary K-feldspar in rocks in Iran, an aggregate mat of myrmekite is formed when all the K-feldspar is totally replaced by myrmekite (Figure 85), like that in Figure 83. This total replacement by myrmekite also differs from isolated wartlike myrmekite that is formed by K-metasomatism of primary plagioclase. In such rocks, not only does the myrmekite occur between K-feldspar and plagioclase (as in K-metasomatism), it also occurs against biotite, muscovite, and quartz (Figures 84 and 85).
Figure 85. Aggregates of myrmekite (left side) formed around the plagioclase crystals (right side; albite twinned; white and gray) in diorite, which may have resulted from Ca-metasomatism of sodic plagioclase. Quartz (cream; white). Biotite (brown). From the Alvand plutonic complex, Hamadan, Iran. Source: Behnia and Collins (1998, Figure 2).

Another kind of wartlike myrmekite that is common where Ca- and Na-metasomatism of primary K-feldspar occurs is the production of “swapped myrmekite,” which exists in granite in parts of China (Figure 86) and elsewhere. Along a boundary between two primary K-feldspar crystals, introduced fluids carrying Ca and Na ions replace the K-feldspar in positions that alternate along both sides of the fracture to form wartlike myrmekite that projects into the K-feldspars.
Figure 86. Swapped coarse myrmekite grains are on the grain boundary between two differently oriented K-feldspar crystals (K₁ [light gray, right side] and K₂ [dark gray, grid-twinned, left side]). K-feldspar grain (light gray) in upper right with perfect cleavage (black parallel lines) has nearly invisible, uniformly-distributed, narrow spindles of perthitic lamellae as do all other K-feldspar crystals. Meso-to-fine myrmekite is locally found at the lower corner of the coarse myrmekite (right side of the center of the photo). Bar scale of 0.5 mm length; lower right. Photomicrograph is from an un-named granite body in Inner Mongolia. Source: Rong (2002, Figure 3a).

Another example of Ca- and Na-metasomatism is in the Velay granite in France (Figure 87). In this granite, replacement of K-feldspar by myrmekite is observed only in the enclaves, not in the host granite. This replacement is thought to be driven by the influx of Na-rich fluids from the host granite in the range of 450-650⁰ C. These Na-rich fluids cause subsequent reequilibration with the Na-poorer feldspars in the enclaves. Here, the Na replaces Ca in plagioclase, and, in turn, the released Ca replaces K-feldspar to form myrmekite. The cathodoluminescence image is provided by Daniel Garcia (Garcia et al. 1996).
Figure 87. Extensive development of myrmekite replacing a K-feldspar megacryst (blue) in a granodiorite enclave of the Velay granite. Source: Collins (1997d, Figure 10; no scale available from original paper).

Additional examples of Ca- and Na-metasomatism exist where wartlike myrmekite replaces plagioclase in deformed anorthosite masses. In such myrmekite the plagioclase of the myrmekite has a higher Ca content (An content) than in the plagioclase of the adjacent anorthosite mass. No image examples are included here but can be found in the published literature. For example, see De Waard et al. (1977), Dymek and Schiffries (1987), Wager and Brown (1967), and Perchuk et al. (1994).

Finally, Ca-metasomatism can also occur to produce myrmekite as a reaction rim between clinopyroxene and adjacent plagioclase (personal communication with Andrew Putnis) and along K-feldspar veins in granulites (Touret and Nijland, 2012).

XI. DISCUSSION

Although Tuttle and Bowen (1958) showed that the final product of crystallization from silicate melts in closed systems can be granite at the thermal minima on the liquidus in the system Ab-Or-Q, a granite composition can also be achieved by metasomatic processes in an open system, as indicated by examples in this article. This possibility was particularly shown to be true for rocks near Temecula, California, where deformed Bonsall Tonalite was converted to the myrmekite-bearing Woodson Mountain Granite. Because the original tonalite contained about 85 percent relatively-calcic zoned plagioclase and 15 percent biotite and hornblende and because the metasomatic changes produced about 33 percent K-feldspar, 33 percent relatively sodic plagioclase, 31 percent quartz, and 4 percent biotite, the former tonalite ends up with a
granite composition (Collins 1988). This conversion does not mean that in every place where K- and Si-metasomatism occurs, granite is the end product. In some places, a granodiorite, tonalite, diorite, or gabbro can be barely modified so that only 1-5 percent K-feldspar is added, the primary plagioclase is bordered by rim myrmekite, and only a few percent quartz is added. In other places, more K-feldspar (bordered by wartlike myrmekite) and quartz may be added to a relatively mafic rock, but the resulting amount of secondary, recrystallized, more-sodic plagioclase can still be more abundant than the K-feldspar, and the new rock composition is granodiorite and not granite. In still other places, the original rock, because of higher pressures and lateral shearing, may be converted into garnet gneiss. In any case, an open system can allow hydrous fluids to introduce K and Si and to subtract some Na, Ca, Fe, Mg, and Al by differing degrees in different geologic terranes.

In the example of the isoclinal anticline at Gold Butte, Nevada, where the mafic igneous rock layers of one limb were squeezed by high pressure under the other limb (Figure 37), so much Ca, Fe, Mg, and Al were removed that these rock layers were converted first into garnet gneisses but finally into leucogranite. Normally, such a large volume of granite (> 1 km$^3$) would be considered by many petrologists to be the result of magmatic injection, just as Fryxell et al. (1992) and Volborth (1962) thought that the mafic layers were formed by injection of magma into supposed metapelites as phacolithic bodies. However, it does not make geologic sense that granite magma can be injected into a volume that is being squeezed to produce rocks (and minerals such as garnet) that are in the direction of having a smaller volume. What makes sense is that so much Ca, Fe, Mg, and Al were being removed from the former mafic rocks in this volume by through-going hydrous metasomatic fluids that all that remained was a K-, Na-, and Si-rich granitic residue. This leucogranite lacks any wartlike myrmekite because the continual fracturing of the mineral crystals as the rocks were pushed laterally and compressed, produced such an open system that no localized places occurred where the remaining Ca, Na, Al, and Si in the altered primary plagioclase crystal structures were out of balance such that residual silica remained to form quartz vermicules. Also, the recrystallized plagioclase formed in the final metasomatism was so sodic that all residual Si would have been consumed in forming albite, as on the outer rim of myrmekite (Figure 23).

In those places where K-metasomatism occurs, both microfractured primary plagioclase and the replacing K-feldspar are porous and full of connected pores through which fluids can move in and out of the crystals (Putnis et al. 2007). If anyone has any doubts that K-feldspar can replace primary plagioclase by an interface-coupled dissolution-precipitation process at temperatures below melting conditions (Purnis 2008), then Figure 88 should provide convincing evidence that it happens. Clearly an already-crystallized primary plagioclase crystal has been replaced by microcline in myrmekite-bearing rocks in the Vrådal pluton in Norway.
Figure 88. Microcline (top; black and grid-twinned) penetrates and replaces primary plagioclase (bottom; light-gray, speckled, faintly albite-twinned) along an irregular contact, which also includes veins (black) into the plagioclase. One twin plane of the microcline grid twinning is aligned parallel to the albite twinning of the plagioclase. Significantly, remnants of the zoning in the plagioclase are preserved in the microcline, which logically would not happen if the two feldspars crystallized simultaneously from a melt. Thin section from the Vrådal pluton, Norway.

Tuttle and Bowen (1958 and other granite petrologists have rightly said that the conversion of a relatively-mafic plutonic igneous into granite cannot be achieved by solid-state diffusion processes through solid undeformed rocks, but when the solid-state diffusion is on a nano-scale at half the distance between walls of micro-pores in altered and deformed feldspar crystals, then whole or parts of whole plutons can be changed into rocks of more granitic composition across distances of kilometers. The extent of the conversion of the rocks into more-granitic composition are functions of the original igneous-rock chemical and mineralogical compositions, the degree of deformation or openness of the system to allow elemental components to move in and out, and the amount of available introduced K and Si to form secondary K-feldspar and quartz.

In all recrystallized and metasomatically altered rocks, the presence of xenoliths (or enclaves), dikes into adjacent wall rocks or into other coexisting granitic facies, and hypidiomorphic textures will still indicate that the plutonic rocks were originally emplaced as
magma. But if wartlike myrmekite occurs in these rocks, the investigating geologist needs to recognize that it as a clue to possible, either minimal or vast, mineralogical and chemical changes on a plutonic size. The large scale K-metasomatism that occurs in exposed areas (>100 square meters) and as much as several square kilometers and the formation of wartlike myrmekite (a) in the biotite-hornblende Bonsall Tonalite that is converted to granite near Temecula, California (Collins, 2002), (b) in the biotite granodiorite in the Papoose Flat pluton in eastern California (Collins and Collins, 2002b), (c) in a hornblende diorite of unknown age associated with the Twentynine Palms quartz monzonite (Collins, 1997g), and (d) in the mafic granodiorite in outer rim of the Vrådal pluton in Norway should be recognized as a guide to looking at other locations where such large scale (>1 km²) metasomatism might exist. Moreover, in each of the three places that we noted in the first part of the article where early investigators had reported K-metasomatism to form K-feldspar megacrysts is where wartlike myrmekite coexists with the megacrysts (as found by the senior author). These places include (1) the Ardara pluton in Ireland (Collins 1997h; Pitcher and Berger 1972; and Pitcher et al. 1987), (2) the Waldoboro granite complex in Maine (Collins, 1997e; Barton and Sidle 1994), and (3) the granitic rocks in the Fallen Leaf Lake area in California (Collins 2003; Loomis 1961, 1983). If these earlier investigators had known that the simultaneous occurrence of myrmekite and K-feldspar indicated large-scale K-metasomatism, they would have also known that the whole granitic plutons had been extensively modified by metasomatism and not just the primary plagioclase that was replaced by K-feldspar. The K-feldspar megacrysts in the Ardara pluton are in the outer rim just like those in the Vrådal pluton. The art of petrographic studies of thin sections and extensive sampling during careful field work need to be revived.

APPENDIX: Additional Image Examples

A. Microfracturing
Figure A. Remnant islands of plagioclase. K-feldspar (light gray). Islands are optically continuous with a large zoned plagioclase crystal (upper right). Note serrate edges of the plagioclase against the K-feldspar (left side and upper right) where the K-feldspar is replacing the plagioclase. Source: Collins (2003, Figure 8).
Figure B. Penetration along broken grain-boundary seals. Microcline (gray; grid-twin; bottom) penetrates plagioclase (white; top) and surrounds tiny remnant islands of plagioclase that are optically continuous with the larger plagioclase grain (white; top). Plagioclase of myrmekite is also in optic continuity. Source: Collins, (2003, Figure 18).
**Figure C. Penetration along fractures in plagioclase.** Microcline (left side, light gray, cross-hatch pattern) penetrating and replacing broken end of plagioclase crystal (tan). Plagioclase contains muscovite crystals (bright colors). Source: Collins (1997l, Figure 12).
Figure D. Microfracturing with parallel alignment of island. Carlsbad-twinned and zoned plagioclase crystal with speckled, sericite-altered core. Remnant end of the plagioclase crystal (center, right) occurs as an island in microcline (gray, right side) and is in optical parallel alignment with larger plagioclase crystal (left side). Myrmekite with tiny quartz vermicules occurs on corners of plagioclase projecting into the microcline. Source: Collins (2001a, Figure 18).
B. Parallel alignment

Figure E. Parallel alignment of microcline grid-twinning with albite twinning in plagioclase. Microcline (dark gray, grid-twinning) in this photomicrograph from myrmekite-bearing granite is in a more advanced stage of replacement of plagioclase. Plagioclase (light tan) in the center and upper left has been completely replaced by K-feldspar (lower right). Note that the albite twinning of the grid-twinning in the microcline is parallel to the albite twinning in the plagioclase. Source: Collins (1997b, Figure 6).

Figure F. Parallel alignment. This slightly deformed and fractured Carlsbad- and albite-twinned plagioclase crystal (cream gray and gray) comes from a felsic diorite. Tiny irregular islands and veinlets of microcline (black-gray) replace portions of both halves of the Carlsbad-twinned plagioclase crystal. Source: Collins (1997g, Figure 8).
C. Wartlike myrmekite

Figure G. Myrmekite  Aggregate grains of myrmekite project into borders of microcline megacrysts (dark gray; left and upper left corner) and white (on right side). Albite-twinned plagioclase (light gray) occurs in the center of the image. Biotite (brown) occurs in the upper part of the image. Quartz (white) is in the upper right corner and bottom left edge of the image. Source: Collins (2001a, Figure 24).
D. Ghost myrmekite

Figure H. Ghost myrmekite. In this photomicrograph, the thin section of granite in the Sierra Nevada is turned so that the microcline is at extinction (black) in order to show the ghost myrmekite. Albite-twinned plagioclase (tan to light gray and white, upper right side) has an extinction position that is optically parallel to extinction positions of islands of incompletely replaced plagioclase containing tiny quartz blebs (white) which occur scattered to the left of the plagioclase crystal in the microcline and to the far left in the microcline. When the thin section is rotated so the microcline grid-twinning becomes visible, these islands become invisible or nearly so, whereas the large plagioclase crystal is still distinct. The inclined, elongate, rounded grain (left side, center) in microcline is myrmekitic. Source: Collins (1997b, Figure 10).
Figure I. Ghost myrmekite. Microcline (black, at extinction position), showing islands of quartz blebs (ghost myrmekite). Myrmekite with coarse quartz vermicules (faint ovals) is at left of the black microcline. Source: Collins (1998d, Figure 16).
Figure J. Ghost myrmekite.  Myrmekite with coarse quartz vermicules in megacystal granitic rock near base of Wanup pluton, south of Route 69.  Microcline (black) with scattered quartz blebs (white); some quartz blebs are the same size as those in the myrmekite, and others are tiny in ghost myrmekite.  Plagioclase (albite-twinned; light and dark gray); quartz (mottled gray).  Source: Collins (2001b, Figure 19).

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