THE STRUCTURE OF A PORTION OF THE SOUTHERN CALIFORNIA BATHOLITH, WESTERN RIVERSIDE COUNTY, CALIFORNIA

by

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I hereby recommend that this dissertation prepared under my direction by William Willis Jenney, Jr.

entitled The Structure of a Portion of the Southern California Batholith. Western Riverside County, California

be accepted as fulfilling the dissertation requirement of the degree of Doctor of Philosophy

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SIGNED: William Willis Jenney

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ABSTRACT

Near Riverside, California, the Upper Jurassic Bedford Canyon Formation is intruded by several plutonic rocks of the complex batholith of Southern California. The Bonsall Tonalite is the most common granitic formation in the batholith. The Riverside-Perris pluton, the focal point of this study, is the largest pluton of this tonalite. The Riverside-Perris pluton was studied structurally and conclusions drawn from this study are applied to the Southern California batholith as a whole.

The plutonic rocks are demonstrably intrusive into the pre-batholithic rocks and each plutonic rock can be seen to have intruded those preceding it. Evidence favoring forceful intrusion is scanty and it is concluded that the mode of intrusion was not forceful.

The internal structure of the granitic rocks includes the parallelism of mineral grains, dark clots, xenoliths, schlieren and clusters of clots. These parallel arrays may be, in part, relic structures inherited from the source rock of the magma and the pluton-wide patterns were disrupted by local movements within the magma. Most of this local differential motion was unrelated to the regional or even to

the pluton-wide stresses. The movement was the result of the non-homogeneous cooling of the magma, which led to variable mobility of adjacent portions of the intruding magma. These movements also caused the development of localized joint sets, although pluton-wide joint patterns developed after the whole body had solidified sufficiently.

Rifting of the continental crust, analogous to the spreading in the oceanic rises, is postulated to explain the permissive intrusion of the batholith.

INTRODUCTION

The batholith of Southern California is composed of many mappable granitic formations which intrude pre-Cretaceous rocks. The pre-Cretaceous rocks are assigned, on evidence presented by others, to the Jurassic except for a few bodies enclosed in the eastern and northern extremities of the batholith. The composite intrusion as a whole, as well as the intruded rocks, has a generally northwesterly trending structure. Within the batholith, the various plutons trend to the northwest as well.

Of the many granitic formations which have been described in the Southern California batholith, the one which underlies the most area is the Bonsall Tonalite. Tonalite also represents the average composition of the entire massif (Larsen, 1948).

The Riverside-Perris Pluton

The pluton which is the subject of this examination is the largest body of the Bonsall Tonalite in the batholith. It is associated with four other plutonic rock types, two of which are major units in the intrusive complex. The Woodson Mountain Granodiorite and the San Marcos Gabbro are among the five most extensive rocks in the batholithic complex.

These rocks all intrude the Jurassic Bedford Canyon Formation in localities which are contiguous with the pluton which is the focal point of this study. I have called this the Riverside-Perris pluton because of its situation between these two towns. Around and within the Riverside-Perris pluton, the granitic rocks all show evidence of intrusion into the Bedford Canyon rocks and each shows an intrusive relationship to those which preceded it. The evidences of intrusion are generally the inclusion of fragments of the older material and the truncation of the older structures by the younger rocks.

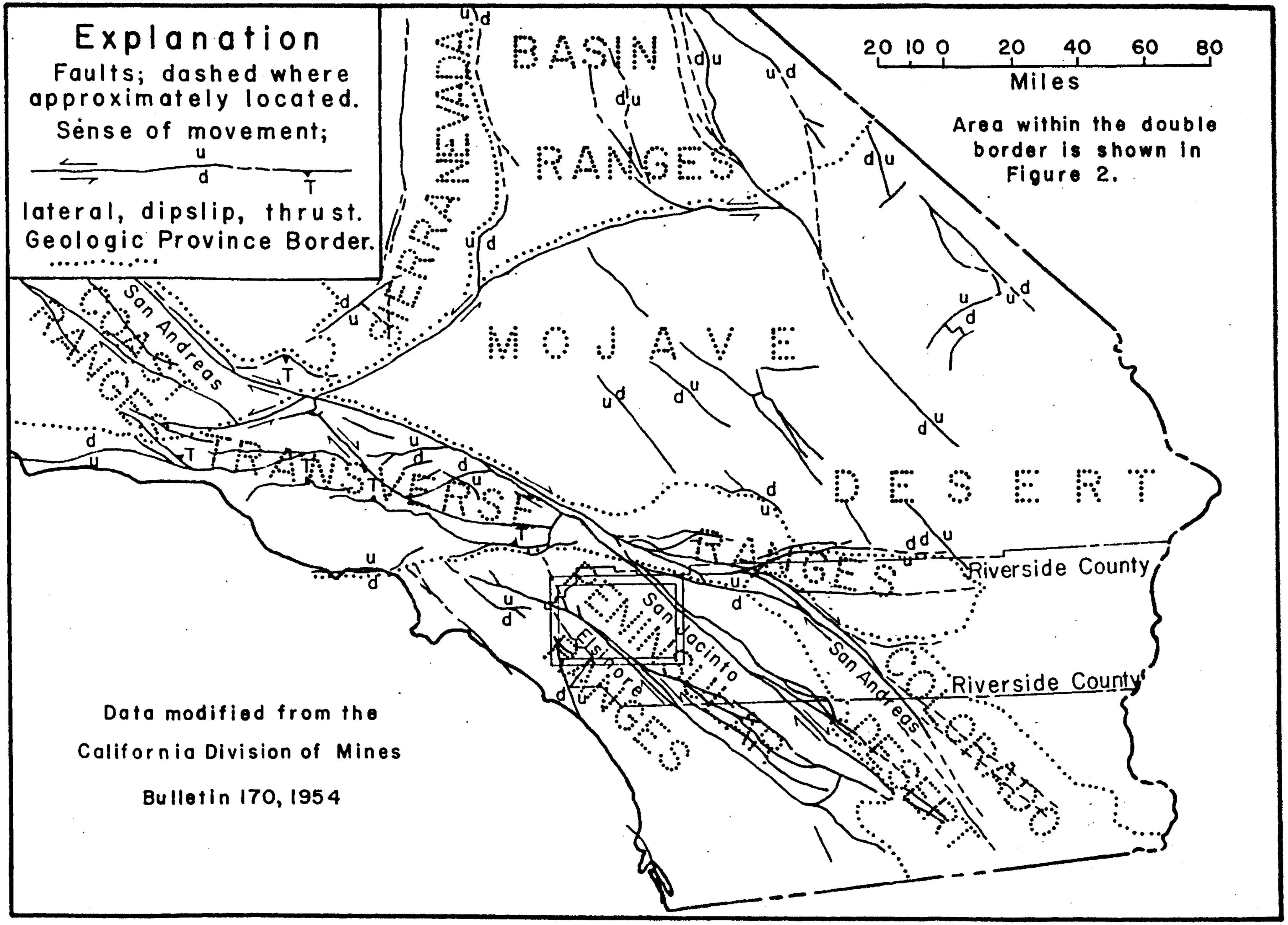
Location and Size of the Area

The irregular north-northwest trending pluton which is the focal point of this study is located in western Riverside County, California, between the cities of Riverside, Perris, and Elsinore (Figures 1 and 2). Latitude 33°50' north and longitude 117°20' west approximately locate the center of this area in southwestern California.

This pluton is about 13 miles long and varies in width from about 4 to about 9 miles (Figure 3). Its area, including the screens of older rock, is approximately 93 square miles.

Purpose of the Investigation

The Riverside-Perris pluton of Bonsall Tonalite has been studied because it represents the whole batholith of



Area within the double border is shown in Figure 2.

Data modified from the California Division of Mines Bulletin 170, 1954

Figure 1. Simplified fault map of Southern California, showing Geologic Provinces.

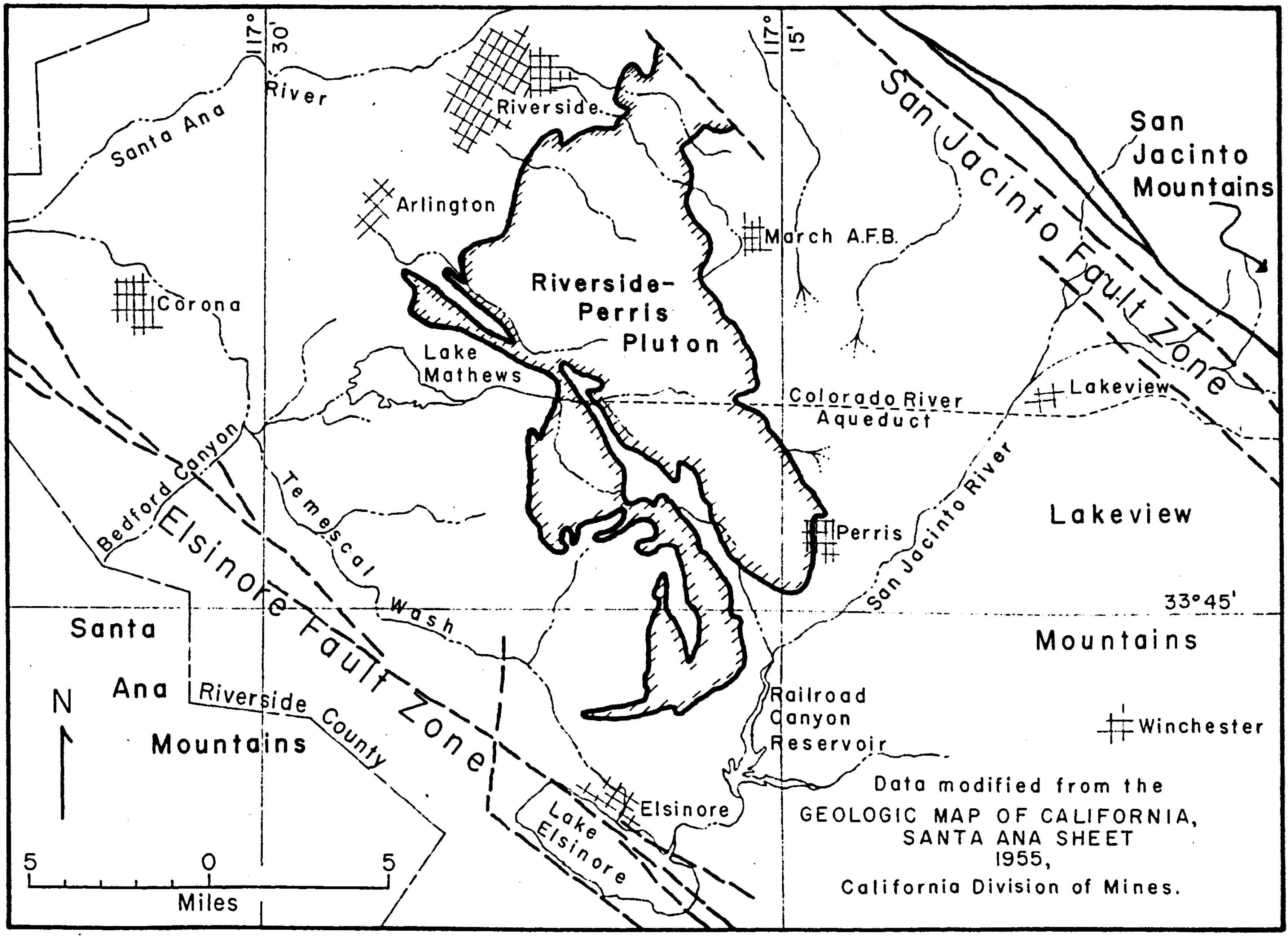


Figure 2. Western Riverside County, California.

Southern California as much as any one body in a complex intrusion can be representative of so large an intrusion. It lies approximately on the axis of the batholith and is bordered by the principal older rocks, as well as by the dominant younger intrusion. The older rocks are the Bedford Canyon Formation and the San Marcos Gabbro; the younger is the Woodson Mountain Granodiorite.

No petrographic work was done by the present author, but the rocks of the batholith have been extensively examined by Larsen (1948). The petrology of the rocks in the area covered by this paper has been discussed by Dudley (1935), Osborne (1939), Engel (1959), and others. The observations of these authors are drawn upon in this study.

The structural study of the Southern California batholith has been neglected, so far as I have been able to discover. The purpose of this investigation is to map the structure in the area of the Riverside-Perris pluton, to add the structural knowledge to the petrologic data already available and to apply this information to the problem of the emplacement of the batholith. This problem includes the mode of emplacement of the granitic rocks, the inter-relationships between the various plutonic rocks, the origin of the ubiquitous dark clots in the Bonsall Tonalite, and the origin of the granitic material itself. In addition, the structural information can be applied to finding the relationships to regional tectonic features which may have been active at the time of emplacement. The Texas lineament and the San Andreas fault, with the associated San Jacinto and Elsinore faults, are such regional features which are near enough to have influenced the development of the northern portion of the Southern California batholith (Figures 1 and 2).

Because of the size and the setting of the Riverside-Perris pluton, as well as its composition, inferences may be drawn which could apply to the batholith as a whole.

Method of Investigation

Mapping was done on a scale of two and one-half inches per mile, using the U. S. Geol. Survey 7-1/2 minute (1:24,000) quadrangle sheets as a base. The rocks were studied in the field to determine the structural elements present, their attitudes, and their inter-relationships. The attitudes were measured to the nearest five degrees, a realistic technique because of the variation in the attitudes of the structures and the difficulty of measuring more precisely with a Brunton compass. Several readings, where possible, were averaged to make one entry in the field notes covering tens or hundreds of square yards. As a rule, measurements in an area of approximately 100 yards square were averaged and the information was plotted on the data map in symbols that cover about 10,000 square yards, on the map. Figures 4 and 5 are

Figure 4. Data map of part of the north-central Riverside-Perris pluton.

The scale and the symbols are the same as on Figure 3.

Figure 5. Data map of part of the central Riverside-Perris pluton, including the septum.

The scale and the symbols are the same as on Figure 3.

copies of two portions of the data map. They are presented to show the kind of information from which the main map (Figure 3) was compiled. In areas of sparse outcrops, the sampling was limited to the number of measurable attitudes. The mapping techniques and philosophy are those of granite tectonics, which were developed by H. Cloos and his colleagues, as outlined by Robert Balk (1937).

GEOGRAPHY AND PHYSIOGRAPHY

In the alluvial plain along the northwestern edge of the pluton, the elevation is approximately one thousand feet above sea level (Figure 3). The highest elevation within the mapped area is 2557 feet at Vayemo Peak, giving a total relief of a little more than 1500 feet for the region.

Surface Forms

The topography of this portion of the Perris block was related by Dudley (1936) to three periods of erosion which resulted in the same number of recognizable surfaces. The oldest, called the "early land form", is expressed only in the slopes of the northwest and east sides of the pluton, where the crystalline rocks rise out of the alluvium. The intermediate, both in elevation and in age, is the Perris surface (Figure 6). This surface, of which the average elevation is over 1600 feet, ranges from about 1400 to 1800 feet above sea level. It comprises, with its monadnocks, the whole of the area mapped (Figures 7, 8, 9 and 10), except for the margins mentioned above. The largest monadnock has an extent of some 25 square miles in the mapped area and extends southwestward to the Elsinore trough (Figure 6). This monadnock is truncated by the Gavilan-Lakeview peneplain at an

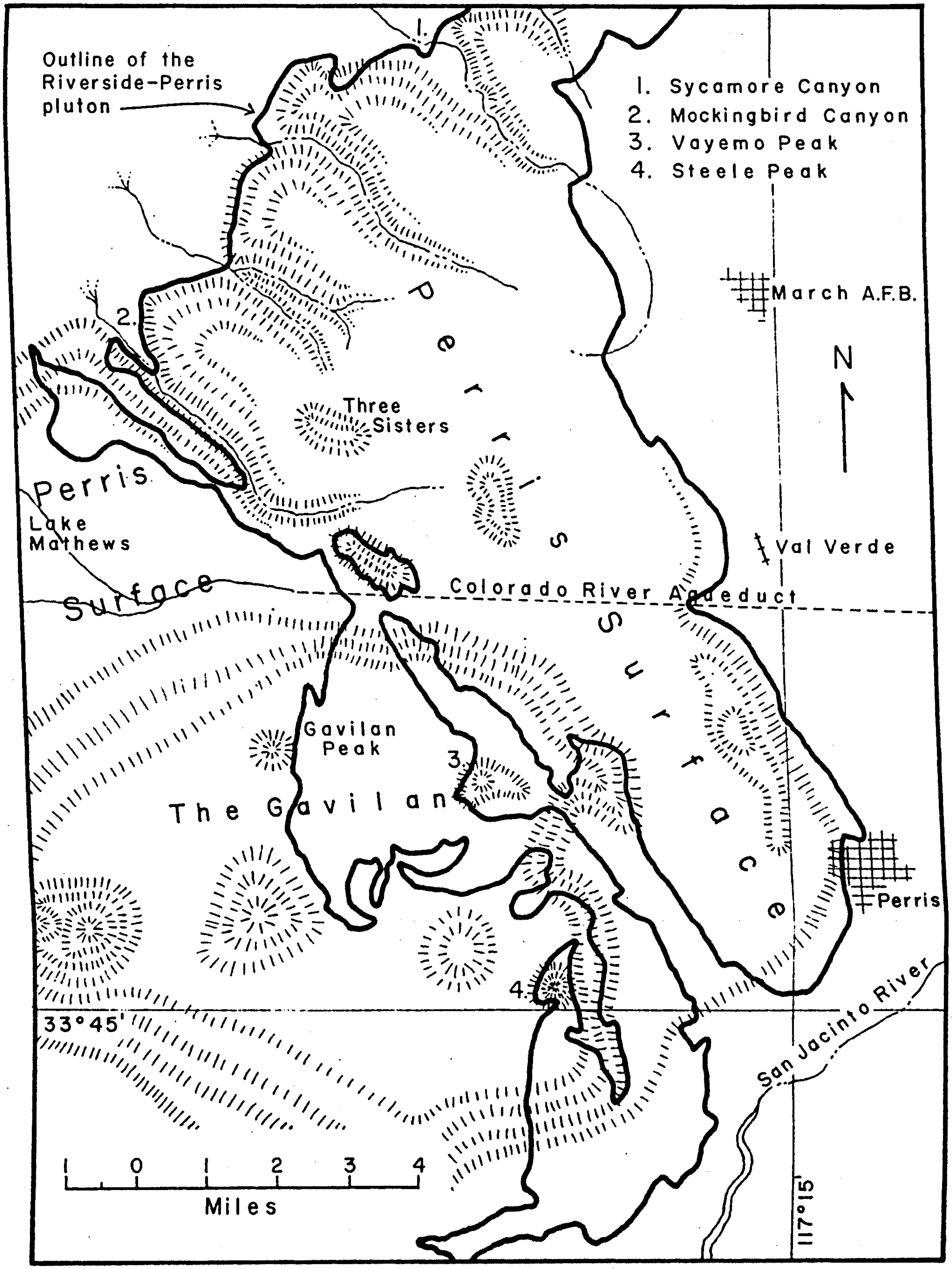


Figure 6. Generalized physiography of the north central Perris block.

Figure 7. The Perris surface west of Val Verde.

The view is east. The rocks exposed are all Bonsall Tonalite.

Figure 8. The Perris surface southwest of Mockingbird Canyon.

The view is north-northwest. The boulders in the middleground are Bonsall Tonalite. Beyond them, and forming the ridge, are rocks of the Bedford Canyon Formation.

Figure 9. The Perris surface near Perris.

The view is southeast. All exposed rocks are Bonsall Tonalite.

Figure 10. Dark clots in the Bonsall Tonalite, showing their resistance to erosion.

The view is northwest, about 2-1/2 miles south-southwest of Val Verde. The Perris surface is in the background.

elevation of about 2100 feet (Figures 11 and 12). This surface is named for the Gavilan (this upland plain) and the Lakeview Mountains, about 16 miles to the east (Figure 2) where the surface is also developed at the same altitude. Briefly, Dudley (1936) visualized a mature surface (early land form) eroded to near base level (Perris surface), then buried and a third surface (Gavilan-Lakeview peneplain) eroded on the fill and the protruding monadnocks. After this, uplift and denudation produced the topography as seen today.

Mann (1955), who worked with the Pliocene and Pleistocene stratigraphy in the Elsinore fault zone described a sequence of events which is considerably different. He would have the "early land form" covered by alluvium in the Paleocene. Later, successive uplifts (or lowering of base level), with pauses for erosion, produced the Lakeview-Gavilan surface followed by the Perris surface (both in the Middle Pleistocene). Mann's timing of the events is at least partly discredited by the discovery of Early Pliocene fossils in an ancient stream channel on the Perris surface (Proctor and Downs, 1962).

The accordance of summits at 2500 to 2600 feet of elevation on the Gavilan-Lakeview surface has been overlooked, apparently, by Mann, Dudley and other authors. The crests of

Figure 11. Edge view of the Gavilan.

The view is southwest. Gavilan Peak is in the right-center. Rocks exposed in the escarpment are Woodson Mountain Granodiorite on the right of the orange grove and Bonsall Tonalite on the left.

Figure 12. Border gneiss of the Bonsall Tonalite.

The view is northwest. The Gavilan-Lakeview peneplain and Gavilan Peak are in the background.

these remnants may well represent the upland portion of the "early land form".

The topography of the Perris and Gavilan-Lakeview surfaces is open and rolling, with scattered erosional remnants standing above them. The dissected northwestern margin of the Perris surface and the almost circular escarpment which surrounds the Gavilan are quite rugged and steep.

Drainage

The northwestern area is cut by several northwest draining canyons or steep-sided valleys. These drainages flow into the Santa Ana River, which flows west-southwest through the large valley lying to the north (Figure 2). The central portion of the area drains west into Temescal Wash, which flows northwest through the Elsinore trough to the Santa Ana River. The flow in the south and southwest is to the south and southwest into the Elsinore trough, either directly or via the San Jacinto River. The San Jacinto River also carries the drainage from a strip, seldom more than a mile wide, along the eastern edge of the pluton. This strip drains eastward and then southward through the Perris Valley to the San Jacinto River. Thus, the whole area eventually feeds the Santa Ana River either directly or through Temescal Wash.

Vegetation

The vegetation is sparse in most of this area (Figure 7, et seq.). This is largely due to man's efforts and the normally semi-arid climate, combined with a dry period which has persisted for many years. The scrub on the hills, which seems to have been burned off fairly regularly (at least partly by intention), is considerably less dense than the chaparral on the more humid Santa Ana Mountains to the west. The flatter areas are harrowed from time to time and grazing is prevalent in all areas not under cultivation or built upon. Persistent drouth has kept new growth to a minimum. Hillside areas which have escaped the periodic denudations are quite thickly overgrown with chest high scrub which can be almost impenetrable (Figure 13).

Exposures

There is almost no alluvium in the regions mapped as crystalline rock. The erosion surfaces mentioned above were cut in bedrock.

The metamorphic rocks are the most resistant to weathering and erosion and therefore generally form the ridges and high ground where they occur (Figures 8, 14 and 15). The batholith has weathered into a grus which underlies practically all of the area mapped as granitic rock. The nature of this weathered material may be seen in road cuts, tunnels and

Figure 13. A boulder field on a brush-covered slope.

The view is north-northwest, about 1-1/2 miles east of Gavilan Peak. All exposed rocks are Bonsall Tonalite.

Figure 14. A ridge of Bedford Canyon Formation rocks in the eastern Gavilan.

The view is north-northwest. A quartz-filled, nearly horizontal joint is evident in the foreground. All the boulders are Bonsall Tonalite.

Figure 15. The septum of metamorphic rocks southeast of Steele Peak.

The view is west-northwest. The light-colored boulders in the foreground and on the hillside are Bonsall Tonalite. The remainder of the exposed rock, including the rubble in the middle-ground, is Bedford Canyon Formation.

Figure 16. The weathered core of an exposed boulder of Bonsall Tonalite.

The view is north, about 1-1/2 miles west-southwest of Val Verde. All exposed rocks are Bonsall Tonalite.

quarries. The masses of undisturbed grus display the attitudes of dark clots, schlieren, joints, and even mineral grains. Rock which is exposed, even slightly above the level of the surroundings, seems hard and sound although it may only have a resistant veneer. If the shell is broken off of some of these exposures, the core is often soft, strongly weathered grus (Figure 16). The sound carapace at the surface is remarkably resistant to erosion since it is, as Warhaftig (1965) discovered in the Sierra Nevada, only wet during the rare rains while the buried rock is constantly damp and subject to weathering. Spheroids of relatively unweathered material remain in the grus. The spheroids, when exposed by erosion, form the typical domal outcrops. The domed outcropping is caused by the partial exposure of a generally spheroidal weathering residual (Figures 17 and 18). These outcroppings are in place, with reliable orientations, as long as erosion of the surroundings has not proceeded too far. As the erosion of the encompassing grus continues, the shape of the erosional remnant becomes pillar-like and the spheroidally weathered core may topple. This process caused the attitudes shown by the boulders to become unreliable over fairly large areas, typically in boulder fields and on the slopes of hills (Figures 13, 14, 15, 19 and 20). Thus, in certain areas, the structural relationships cannot be deciphered.

Figure 17. A domal outcrop of Bonsall Tonalite.

The view is northwest, about 300 yards southeast of Figure 16.

Figure 18. Modified domal outcrop.

The view is north, about 2-1/2 miles northeast of Steele Peak. The shapes of the domes are modified by the joints and foliation within the Bonsall Tonalite.

Figure 19. A boulder field in the Bonsall Tonalite.

The view is north, about 2-1/2 miles east-northeast of Gavilan Peak. The displaced boulders show an unusual concentration of dark clots. There is a rather typical variation in the concentration and in the elongation of the clots over this small area.

Figure 20. A Bonsall Tonalite-Bedford Canyon Formation contact.

The view is south, about three-quarters of a mile east-southeast of Vayemo Peak. The metamorphic rocks (right) strike northeast and dip steeply west. Most joints in the metamorphic rocks are N25°E, 30°SE.

Man's works are continuing to destroy many exposures as he makes fuller use of the land. Many square miles have been blasted, bulldozed, graded, filled, and planted with orchards or covered with houses and streets since they were mapped in this study.

The largest exposures are measured in only a few thousands of square feet (Figure 9). A quarter section may have only a few hundred square feet of exposure (Figure 8, et seq.). A rocky ridge is likely to have only a few usable outcrops on its crest, the attitudes of the rest of the rock being unreliable because of the toppling explained above.

Absence of continuous outcrops of the batholithic and pre-batholithic rocks leaves the recognition of suspected faults to geomorphic evidence. Contact zones are particularly difficult because of the problem of distinguishing between weakness resulting from the shearing, fracturing, and production of gougey material in a fault contact and the weakness produced by similar effects in a dynamic intrusive contact. I mention weakness because the contact zones, where they have topographic expression, tend to be lows rather than highs (Figure 20).

REGIONAL TECTONIC FRAMEWORK

The Riverside-Perris pluton lies in the northern portion of the batholith of Southern California, close to the northernmost extremity of the generally north-northwest to northwest trending structures which make up the Peninsular Range Province of Southern California (Figure 1). Approximately 15 miles to the north lie the foothills of the San Bernardino Mountains, a major part of the westerly trending Transverse Range Province of Southern California. In this 15 mile interval lies a broad valley which is considered an extension of the Perris fault block (Dudley, 1936). The mapped area is approximately on the mid-line of this elongated fault block which lies between the northwest trending Elsinore and San Jacinto fault zones (Figure 2). These zones, bounding the block, are about 20 miles apart and extend 90 miles or more to the southeast (Rogers, 1965). To the northeast, about 25 miles from the San Jacinto fault zone, is the San Andreas fault zone which also trends northwest. These fault zones bound a series of northwest trending blocks. The Perris block is relatively depressed and graben-like between the Santa Ana Mountains block and the San Jacinto Mountains block. The Santa Ana Mountains rise to more than

5500 feet above sea level to the southwest of the Elsinore fault zone and the San Jacinto Mountains rise above the 10,000 foot level to the northeast of the San Jacinto fault zone.

The rocks of the batholith extend uninterruptedly for more than 350 miles southward from Riverside. They then crop out intermittently all the way to the tip of Baja, California, giving the batholith a length of about 1,000 miles (Larsen, 1951). Their combined maximum width is 70 miles and Larsen estimated that they underlie 50,000 square miles.

The San Andreas fault crosses the Transverse Ranges Province a few miles north of Riverside. The Transverse Ranges are the location of the Texas lineament in Southern California (Hill, 1902 and 1928 and Ransome, 1915). The importance and antiquity of this transcontinental structural lineament has been discussed by Albritton and Smith (1956) and Schmitt (1966). Albritton and Smith (1956) give evidence of activity along the lineament in Texas in the Precambrian and Schmitt (1966) demonstrates that the lineament acted as a structural hinge for many periods and was particularly active in the Mississippian and Cretaceous Periods. All of these authors have stressed that the general nature of the activity on the Texas lineament has been left lateral shearing, wherever the structure has been well exposed. The

San Andreas fault is another major structural feature of probable transcontinental dimension. Hill and Dibblee (1953) state that the right lateral strike slip movement on the San Andreas fault may be traced back to the Jurassic. These two major structures, and perhaps the Elsinore and San Jacinto fault zones as well, may have been active at the time of emplacement of the Southern California batholith.

ROCKS OF THE REGION

The rocks in this north central portion of the Southern California batholith have been described by Larsen (1948 and 1951) and others in varying detail. Larsen's work is the most comprehensive and useful, although the structural aspects are almost totally ignored. Of the some 20 granitic rock types which he described in the batholith, six are present in the Riverside-Perris area. Three of the rock types he described as antedating the batholith are also present.

To the south and to the west of Sycamore Canyon, in the northern portion of the Riverside-Perris pluton (Figure 3), there are patches of metamorphic rock which Larsen (1948) assigned to the Paleozoic. These metamorphic rocks are mostly quartzite and phyllite, with some marble. He based the age on admittedly tenuous lithologic correlation with rocks in the Winchester area, about 22 miles to the southeast (Figure 2), whose age was based on a fossil of questionable authenticity (see below under <u>The Bedford Canyon Formation</u>). Because of the lack of any distinctive features which indicate that they should be treated separately, these rocks have been included with the rest of the metasedimentary rocks in this study.

The bulk of the pre-batholithic rock, in the Riverside-Perris area and for the Southern California batholith as a whole, is the Bedford Canyon Formation (Larsen, 1948). Recently Imlay (1963 and 1964) has shown this formation to be Middle Jurassic in age. Also considered to be pre-batholithic is the Temescal Wash Quartz Latite Prophyry of Jurassic (?) age which lies in the southwest portion of the mapped area (Figure 3).

The six batholithic, granitic rocks previously mentioned are, chronologically: the San Marcos Gabbro (the oldest), the Bonsall Tonalite, miscellaneous tonalites, the Woodson Mountain Granodiorite, the fine leucogranite of Rubidoux Mountain, and the coarse leucogranite of Rubidoux Mountain (the youngest). These rocks have all been assigned to the Cretaceous Period by Larsen. Later workers have determined radioactive ages for some of them and also for some of the associated rocks which do not appear in this area. Bushee and his associates (1963) have obtained lead-alpha dates for several of the crystalline rocks in the Peninsular Ranges. These dates are stated to be within the error of plus or minus 10 percent. The Bonsall Tonalite was dated as 120 and 125 m.y. in different localities. The Woodson Mountain Granodiorite was dated at 105 m.y. These dates are from localities 30 to 70 miles to the south and southeast of the Riverside-Perris area, but there seems to be little difficulty correlating these units over these distances because of the almost contiguous series of plutons of the same rock and the very uniform composition of a given rock type. Bushee, et al. (1963) also give the age of 155 m.y. for the Santiago Peak Volcanics, which are wall rocks to the Woodson Mountain Granodiorite in an exposure about eleven miles to the west of the southern extremity of the Riverside-Perris pluton. These volcanics also overlie the Bedford Canyon Formation in the Santa Ana Mountains with an angular unconformity (Gray, 1954). These rocks had been assigned to the Jurassic or Jurassic (?) by various authors. Banks and Silver (1961) have dated the coarse leucogranite of Rubidoux Mountain at 103 plus or minus 2 m.y. These dates, as well as other dates by Bushee, et al. (1963) for other batholithic rocks in Southern California not appearing in this area, are consistent with the sequence of intrusion determined in the field by many authors. They are also consistent with the ages which have been assigned on stratigraphic evidence.

Larsen (1951) estimates that in the northern portion of the Southern California batholith, the San Marcos Gabbro underlies 14 percent of the area, the Bonsall Tonalite underlies 38 percent, and the Woodson Mountain Granodiorite underlies 25 percent. For the batholith as a whole, the tonalites comprise about 60 percent of the rocks in the area. The

Bonsall Tonalite (one of the six tonalites described by Larsen, 1954), is the most common rock unit and represents the average composition of the complex.

To the west, in the Santa Ana Mountains, fossiliferous marine Cretaceous strata overlie and contain clasts of the batholithic rocks (Gray, 1961; Engel, 1959). Tertiary marine and non-marine beds lie in and on the flanks of the Elsinore trough (Mann, 1955; Engel, 1959). Lower Pliocene non-marine sediments fill an ancient stream channel on the Perris surface near Lake Mathews (Proctor and Downs, 1962). Quaternary fill occupies the Perris and Santa Ana River Valleys.

The Bedford Canyon Formation

The type locality of the Bedford Canyon Formation is about 10 miles southwest of Lake Mathews, in the Santa Ana Mountains (Figure 2). The first published date for these rocks was based on a few crinoid stems and brachiopods and given as "probably Late Triassic" (Willis, 1912, p. 505, citing Stanton in Mendenhall's communication). In this general area, later workers have found more brachiopods and several ammonoids and pelecypods. As late as 1959 (Engel, 1959) these rocks were still assigned to the Upper (?) Triassic. In a re-examination of some of the older material and in the light of some newly acquired fossils, Silberling, et al. (1961)

tentatively assigned the Bedford Canyon rocks to the Upper Jurassic. Further work by Imlay (1963 and 1964) on two suites of fossils from areas previously dated as Triassic has shown them to be Middle Jurassic. The suite from rocks at Bedford Canyon is Lower Middle Jurassic and the suite from rocks about three miles farther west is Upper Middle Jurassic.

The proximity of the outcrops of the metamorphic rocks, as little as 1-1/2 miles across the Elsinore fault zone, and the general similarity of the rocks themselves have led all authors who have studied the area to correlate them from the Santa Ana block to the Perris block without question.

Only one fossil has been attributed to the metamorphic rocks in the Perris block. This fossil was in a piece of calcite marble on a talus pile in a quarry at Winchester, about 10 miles southeast of Perris (Figure 2), and was dated as Carboniferous (Webb, 1939). The value of this find has been seriously questioned by workers in the area (D. M. Morton, verbal communication). The source bed of the talus is a coarsely crystalline marble with "abundant calc-silicate minerals (diopside, forsterite, etc.)... a re-crystallized siliceous dolomite, and it shows what appears to be flow banding parallel to its contacts." (Schwarcz, 1960, p. 80).

In addition, Schwarcz (1960) concluded that all the rocks assigned to the Paleozoic in the Winchester area overlie the Bedford Canyon Formation and are, in fact, younger than the Bedford Canyon rocks. Thus it seems that the walls of the pluton, and the septa within it, are no older than Middle Jurassic, a conclusion which places a limit on the age of the intrusions.

Extent

About 10 square miles of the mapped area are underlain by septum-like masses of Bedford Canyon Formation rocks within the tonalite. These metamorphic rocks are exposed as wall rock to the pluton for several more square miles.

Petrology

According to Larsen (1948, p. 19) the Bedford Canyon rocks, as exposed in the Santa Ana Mountains, are chiefly black to dark gray argillites and slates which grade into relatively pure quartzites. In addition to this sequence, he noted some conglomerates and a few thin lenses of dark limestone. He described the metamorphism in the Santa Ana Mountains as "very low grade", and that in the Perris block as "somewhat greater", with slates and sericite schists. Schwarcz (1960) described the phyllites and quartzites of the Bedford Canyon Formation in the Santa Ana Mountains as being in the lower greenschist facies. He stated that there is an increase in regional metamorphic grade eastward across the Perris block to the Winchester area, where the mineral suite is representative of the hornblende hornfels facies, and that "coarser grained schists and gneisses occur as septa" (1960, p. 19) at Railroad Canyon (Figure 2). Schwarcz considered the formation to be representative of a typical gradation from eugeosynclinal sediments in the Santa Ana Mountains to miogeosynclinal sediments in the vicinity of the town of Winchester. Larsen (1948, p. 32) said, "Larger bodies of the older sediments show very little contact metamorphism even a few feet from the contacts." Larsen stated further that some of the smaller bodies show complete recrystallization and increase in grain size, including the development of lime-silicate rocks in limey beds. These statements hold true in the vicinity of this pluton. Quartz-sericite phyllites and arkosic quartzites are the most common rock types of this part of the Perris block, with subsidiary slaty argillites, mafic schistose rocks, and rare marble beds.

Structure

The metamorphic rocks of this area have a general northwest strike and dip steeply to the northeast, but the picture is complicated in detail. In certain localities, the attitudes vary considerably from outcrop to outcrop indicating the presence of structures which modify the regional homoclinal dip. In a few areas, small folds have been identified (Figure 21). The detailed structure is usually obscured because the strongly jointed character of the rock reduces the outcrops to shattered rubble. Although the areas underlain by this rock tend to be topographic highs, as contrasted to the granitic rocks, their surface is composed of soil and rubble with only a few small exposures of low relief (Figures 8, 14 and 15). Thus, the recognition of fold relationships and facings of beds is inhibited by very effective camouflage. In some regions, even the recognition of the general trend of the rock structures is difficult. In certain localities the structure seems to be simple, so that for a distance of a mile or more there is little variation in strike or dip. Elsewhere, each outcrop will have a radically different attitude than its neighbor only 10 feet away. It is not uncommon to climb over a red-brown hill (red-brown being the color of the soil produced by the weathering of the metamorphic rocks in this area) for a quarter of a mile or more and find no usable outcrop, or a very few conflicting ones. These difficulties are compounded by the generally monotonous lithology within the Bedford Canyon Formation.

Osborne (1939), working in the area about equidistant between Gavilan Peak and the Three Sisters (Figure 3), found the foliation of the schists to be about parallel to the bedding. Larsen (1948) indicated that for the metamorphic rocks associated with the Southern California batholith as a

Figure 21. A small fold in the Bedford Canyon rocks.

The view is northwest, about three miles southeast of Steele Peak.

Figure 22. Detail of the Temescal Wash Quartz Latite Porphyry-Woodson Mountain Granodiorite contact.

The view is northeast, about two miles N30°W of Steele Peak. A faint parallelism to the border is evident in the granodiorite (right). The pencil parallels the strike of the planar array of the porphyry.

whole, the "schistosity is nearly everywhere parallel to the bedding" (p. 22). This statement holds true throughout the area mapped here, so far as can be determined from the sedimentary structures observed. The symbols on the map (Figure 3) indicate, equally, the bedding and foliation of the metamorphic rocks.

In the country several miles to the southeast of Steele Peak, Dudley (1935) reported isoclinal folds dipping steeply east and west. Other writers, including Gray (1961) in the area west of the Elsinore trough and Schwarcz (1960) in the eastern part of the Perris block, have found no repetition due to folding. Schwarcz (1960, p. 23) concluded that a homoclinal dip in the section from the Santa Ana Mountains to the eastern side of the Perris block would include about 98,000 feet of section, and that this apparent thickness may be due to the "prying apart of the isolated metasedimentary rocks by the interposed batholithic masses" or to unrecognized isoclinal folding.

The structure within the Bedford Canyon walls and septum should be correctly interpreted to completely understand the emplacement of the Riverside-Perris pluton. As indicated above, great difficulties stand in the way of a correct interpretation, because too little of the structure is revealed.

Temescal Wash Quartz Latite Porphyry

Dudley (1935) named a mass of hypabyssal rock the Temescal Wash dacite porphyry for its exposure along the western edge of the Perris block on the flank of Temescal Wash. Larsen (1948) demonstrated that the average composition was quartz latite and so re-named the formation.

This rock was found to invade and contain fragments of the Bedford Canyon Formation (Larsen, 1948; Pampeyan, 1952; Engel, 1959) and clasts of the porphyry appear in the Upper Cretaceous conglomerates of the Santa Ana Mountains (Gray, 1961). These authors considered the age of this porphyry to be probably Jurassic. The porphyry may be an intrusive phase of the Santiago Peak Volcanics of the Santa Ana Mountains, which are largely andesite and dacite but do include quartz latites and rhyolites (Gray, 1961). The porphyry is intruded by andesite in the Temescal Wash area (Engel, 1959). If these rocks do represent one phase of volcanic activity preceding the emplacement of the batholith, the 155 m.y. age (plus or minus 10 percent) attributed to the Santiago Peak Volcanics (Bushee, et al., 1963) would further apply to the Temescal Wash Quartz Latite Porphyry and definitely establish its age as Jurassic.

Extent

The Temescal Wash Quartz Latite Porphyry acts as wall rock to the tonalite for about 2-1/2 miles along the southwestern edge of the Gavilan (Figure 3). It extends to the south and west for from three to five miles, respectively. There is also a body of this rock, about five square miles in area, which lies nearly half way between Lake Mathews and Corona (Pampeyan, 1952). These two bodies represent more than 90 percent of the exposures of this rock (Larsen, 1948).

Petrology

This blue-gray to blue-black porphyry has white or glassy feldspar crystals which usually range from 2 to 5 mm in length. Larsen (1948) reported the average plagioclase as andesine (An_{38}). Typically, the quartz phenocrysts are 1 to 4 mm in diameter and comprise from 10 to 15 percent of the rock. In the west, the groundmass is aphanitic and the phenocrysts average about 2 mm. The grain size of both the groundmass and the phenocrysts increases gradually to the east, as the plutonic rocks are approached (Pampeyan, 1952). Larsen (1948) and Pampeyan (1952) both contended that the increase in grain size and the devitrification of the groundmass are due to the proximity of the intruding batholith. The mafic minerals are hornblende and biotite, both commonly altered to chlorite. The porphyry here, where it is adjacent to the Bonsall Tonalite, is finer grained than it is elsewhere, where it is affected by the younger batholithic rocks. The maximum size, adjacent to the tonalite, is about 3-1/2 mm for the orthoclase, 2-1/2 mm for the plagioclase, 1-1/2 mm for the quartz, and a groundmass of less than one-half mm grains. This may be attributable to failure of the earlier invading tonalite to raise the temperature of its walls to a high value, as compared to the temperature attained by the walls at the time of intrusion of the later Woodson Mountain Granodiorite.

Structure

Larsen (1948) described the quartz latite porphyry as having an internal orientation parallel to the contacts, with a steep linear parallelism following the dip. Within the body of the intrusion, the strike of the planar structure is generally east. About five miles southwest of the mouth of Mockingbird Canyon, Pampeyan (1952) was able to establish the dip of the Woodson Mountain Granodiorite-Temescal Wash Quartz Latite Porphyry contact as being about 30 degrees west because of the increasing parallelism of the structural elements within the porphyry as the contact was approached. Larsen (1948) also felt that the heat from the later intrusive rocks was enough to re-crystallize the quartz latite and cause movement within it.

The porphyry along the contact with the tonalite in the western part of the Gavilan (Figure 3) is somewhat re-crystallized, as noted above, but the development of its internal organization is negligible. There is a parallelism, only weakly defined, in the biotite-chlorite grains and in the feldspars. The ill-defined strike is northwesterly with an undetermined, but steep, dip. The adjacent Bonsall Tonalite is unusually gneissic in the outcrops nearest the porphyry (Figure 12) as though the tonalite magma were more viscous in this locality, next to the cooler wall rocks. Two and one-half miles southeast, where the tonalite, porphyry, and Woodson Mountain Granodiorite all crop out together (Figure 3) the younger granodiorite shows only a weak parallelism next to the porphyry (Figure 22). These observations also lead to the conclusion that the wall rocks were hotter and therefore did not cause a stiffening of the magma at a later stage of the emplacement of the batholith.

The San Marcos Gabbro

The name San Marcos Gabbro was applied to a group of mafic plutonic rocks in the San Luis Rey quadrangle, 35 miles to the south, by F. S. Miller (1937). This suite varies widely in composition and texture, but exhibits an almost complete gradation between types as well as an intimate spatial association. Miller included rocks in this group which Hurlbut (1935) had named the San Marcos Mountain gabbro. In 1948, Larsen extended the use of the name to gabbros in much of the Southern California batholith, including the area of the Perris block.

Because the gabbro cuts the Jurassic Santiago Peak Volcanics and is in turn cut by the various tonalites and granodiorites, Larsen (1948) considered it to be the oldest rock of the Cretaceous batholith. It seems to be the mafic forerunner of the intrusive complex, as well as wall rock of the tonalite pluton.

Extent

Larsen showed about 100 bodies of the gabbro on his map (1948, Plate 1). These bodies range in size from a few acres to many square miles. They are distributed over the whole extent of the batholith and he (1954) estimated that they made up about seven percent of its area. In the northern portion of the complex, he (1951) estimated that the San Marcos Gabbro underlies about 14 percent of the area. The several small gabbro bodies in the Riverside-Perris area comprise about eight square miles of outcrop.

Petrology

Miller (1937) described the San Marcos Gabbro as being a medium-grained dark gray norite, in its most abundant and average type. The range of composition, however, runs through a broad spectrum of rock types in the gabbro - norite field. The sequence, olivine norite - norite - quartz biotite norite, is paralleled by the sequence, olivine gabbro - hypersthene gabbro - quartz biotite gabbro. These series are paralleled, and in about equal quantities, by a sequence of hornblende-rich gabbros which also grade into a series of noritic hornblende gabbros. The features common to all of these varieties were: the similar relationships to the other rocks in the batholith (especially the age relationships), the similarity in the character of the outcrops, the susceptibility to hornblende enrichment and a "noteworthy similarity in mineralogy and systematic changes in mineral content and chemical composition throughout the group." (Miller, 1937, p. 1399).

Plagioclase compositions vary through the range An_{45} to An_{90} within the various bodies mapped as San Marcos Gabbro (Hurlbut, 1935; Miller, 1937).

Structure

The San Marcos Gabbro is essentially without internal parallelism in the vicinity of Perris and Riverside. The exceptions that do occur generally lie at or near the borders with younger rocks, where mineral parallelism sometimes develops. Elsewhere, the gabbro is normally without discernible internal organization.

Virtually every measurement of the attitude of the internal structure of the gabbro has been plotted on the map (Figure 3). The internal organization of the gabbro is probably stronger than is evident from the map or from the first impression in the field, but the planar structure is localized at best.

The rock is strongly jointed and this fact leads to the difficulty of discovering the primary structures. Weathering along the joints has reduced the outcroppings to the boulder heaps that are the typical exposure, but these boulders may show a recognizable order in the parallelism of the mineral grains. Yet the disarray of the boulders makes it impossible to plot the attitudes.

Contacts

The contacts of the gabbro with the younger and older rocks are not exposed. Some inclusions of the older rocks have been identified in the gabbro, but these inclusions are small and rare and confined to the areas very close to the contacts.

Some parallelism of the internal structure of the gabbro with the contacts has been recognized. This parallelism is more prevalent adjacent to the younger rocks than it is adjacent to the older ones, suggesting that these structures were imposed on the gabbro by the later tonalite. This idea will be more fully developed in the DISCUSSION.

Pegmatite float is commonly found in the border areas near contacts with older rocks, but bodies or veins of this material have not been exposed. Contacts with the Bonsall Tonalite are gradational. This will be more thoroughly discussed later under Contacts in the section on the Bonsall Tonalite. This gradational zone is not present near the border of the Woodson Mountain Granodiorite.

Bonsall Tonalite

The Bonsall Tonalite was named by Hurlbut (1935) for its proximity to Bonsall, California, a town about 35 miles south of Perris. The name was extended by Larsen (1948) to correlative rocks over much of the Southern California batholith, including the Riverside-Perris area. The Riverside-Perris pluton had been previously called the Perris quartz diorite (Dudley, 1935) and the Val Verde tonalite (Osborne, 1939).

The tonalite invaded and contains fragments of the San Marcos Gabbro and the pre-batholithic rocks, and had been considered to be younger than the gabbro by all the authors who have concerned themselves with the batholith, except Schwarcz (1960). His analysis was based on the structural relationship of the tonalite to a small body of gabbro, about one square mile in areal extent, two miles northeast of Winchester (Figure 2). He stated that the contact is perpendicular to the "smeared out" inclusions in the tonalite. A short visit to the area in question showed this relationship to be valid, but the contact is not exposed and the nearest outcrop of tonalite, whose attitudes I judged to be reliable, is at least 300 yards from the presumed contact. The structure of the tonalite there is the regional structure of that tonalite body. At several sites in the Riverside-Perris pluton, the structures of the tonalite show changes from a local to a regional trend over a few hundred yards or less, near the contacts. I believe that this is the situation at Winchester and this belief was confirmed in discussions with D. M. Morton, a geologist with the California Division of Mines and Geology, whose doctoral dissertation on that area is in preparation for the University of California at Los Angeles.

Bushee, et al. (1963) made age determinations of the Bonsall Tonalite in two localities and reported that their findings lie within the error of plus or minus 10 percent of their stated age. Their lead-alpha measurements for zircons in nine samples of the tonalite from the type locality gave an age of 125 m.y. Two analyses of zircons from the tonalite, in a locality about 70 miles south-southeast of Perris, gave an age of 120 m.y. Thus this pluton seems to be of Aptian, or possibly Albian, age (Kulp, 1961).

Extent

Outcrops of the Bonsall Tonalite represent about 28 percent of the entire area of the batholith (Larsen, 1954) and about 38 percent of the northern portion (Larsen, 1951). It is, as noted above, the most common granitic formation. The Riverside-Perris pluton is the largest single body of this rock and it underlies more than 80 square miles. Three to seven miles north and northwest of Riverside, in the Crestmore-Jurupa Mountains area, bodies of this tonalite represent the northernmost extremity of the Southern California batholith (Burnham, 1954; Mackevett, 1951).

Petrology

Hurlbut (1935, p. 615) described the modal composition of typical Bonsall Tonalite as being more than 50 percent plagioclase (An_{40}), 20 to 25 percent quartz, less than five percent orthoclase (but rarely as much as 15 percent) and 10 percent each of hornblende and biotite. The average composition of this tonalite, as mentioned before, approximates the average composition of the entire batholith (Larsen, 1948). Osborne (1939) investigated the petrology of the tonalite in the Colorado River Aqueduct tunnel from a point about 2-1/2 miles south of the Three Sisters (Figure 3) eastward for about seven miles to the alluvium. He found (p. 926) that the composition along the tunnel varied only slightly, and gave the following content as typical: 57 percent plagioclase (An_{40}), 26 percent quartz, 12 percent biotite, 4 percent hornblende and 1 percent orthoclase.

The most striking single characteristic of this rock is the presence of dark clots. The term "dark clot" is used to describe those ubiquitous dark bodies of unknown origin which lie in the more coarsely crystalline gray mass of the tonalite itself. The terms inclusion, autolith, xenolith, remnant, etc., with their appropriate adjectives, are reserved for phenomena whose origin is not open to question. Although the presence of the clots is a characteristic of the rock, their distribution is not uniform. Visual estimates show that they compose from one to five or 10 percent of the rock in normal circumstances. There are several areas, confined to less than 100 yards square, where the dark clots make up more than 30 percent of the rock (Figure 23). These areas of concentration of dark clots have been plotted on an outline map of the Riverside-Perris pluton (Figure 25). There are also localities, plotted on Figure 26,

Figure 23. A cluster of dark clots.

The view is about west. This exposure is in the tonalite terrain, about three miles north of Three Sisters, on the Perris surface.

Figure 24. A zone of mixed tonalite and gabbro.

The view is northeast, about one mile south-southwest of Vayemo Peak.

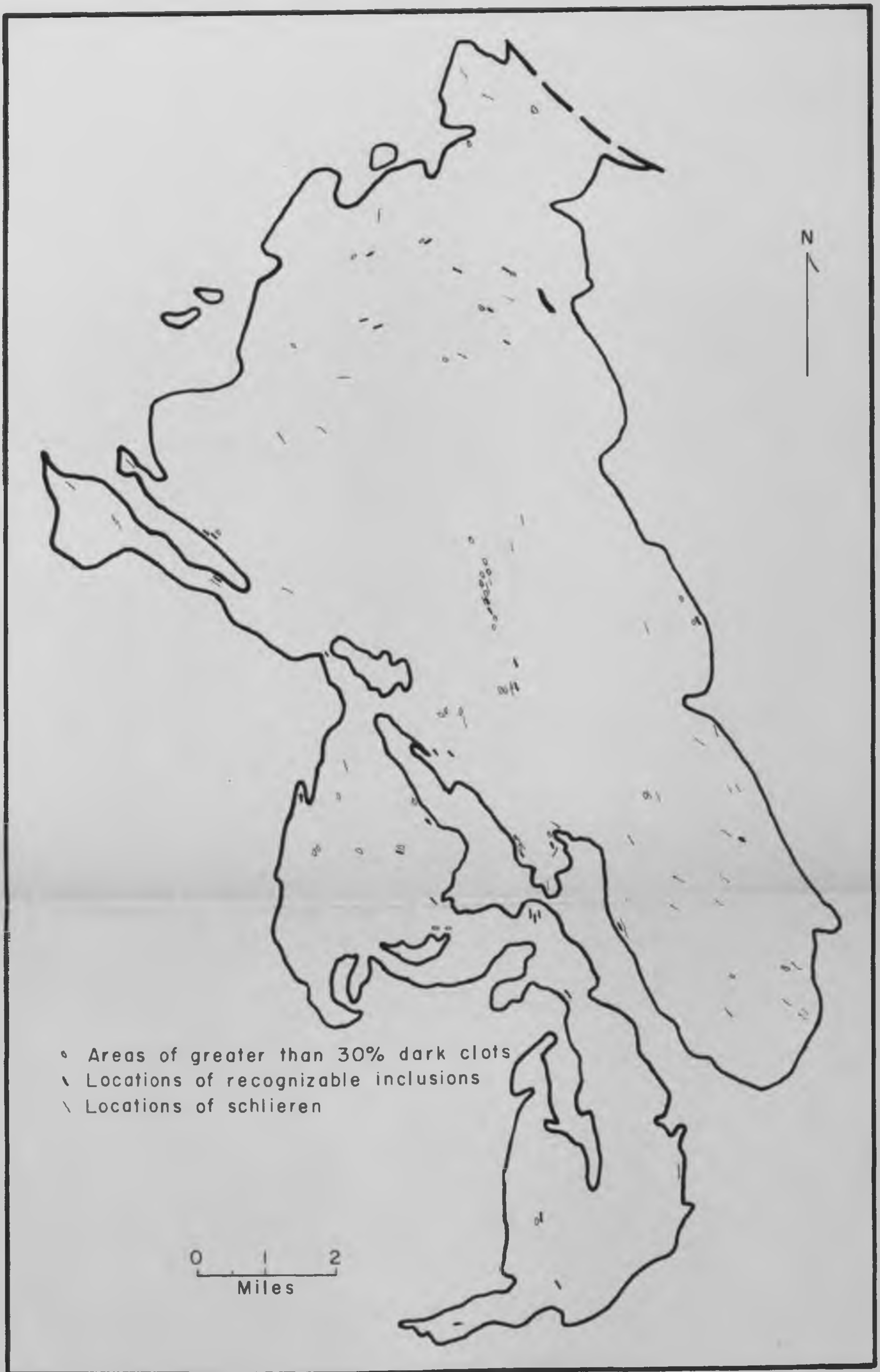


Figure 25. Clots, schlieren and inclusions in the tonalite pluton.

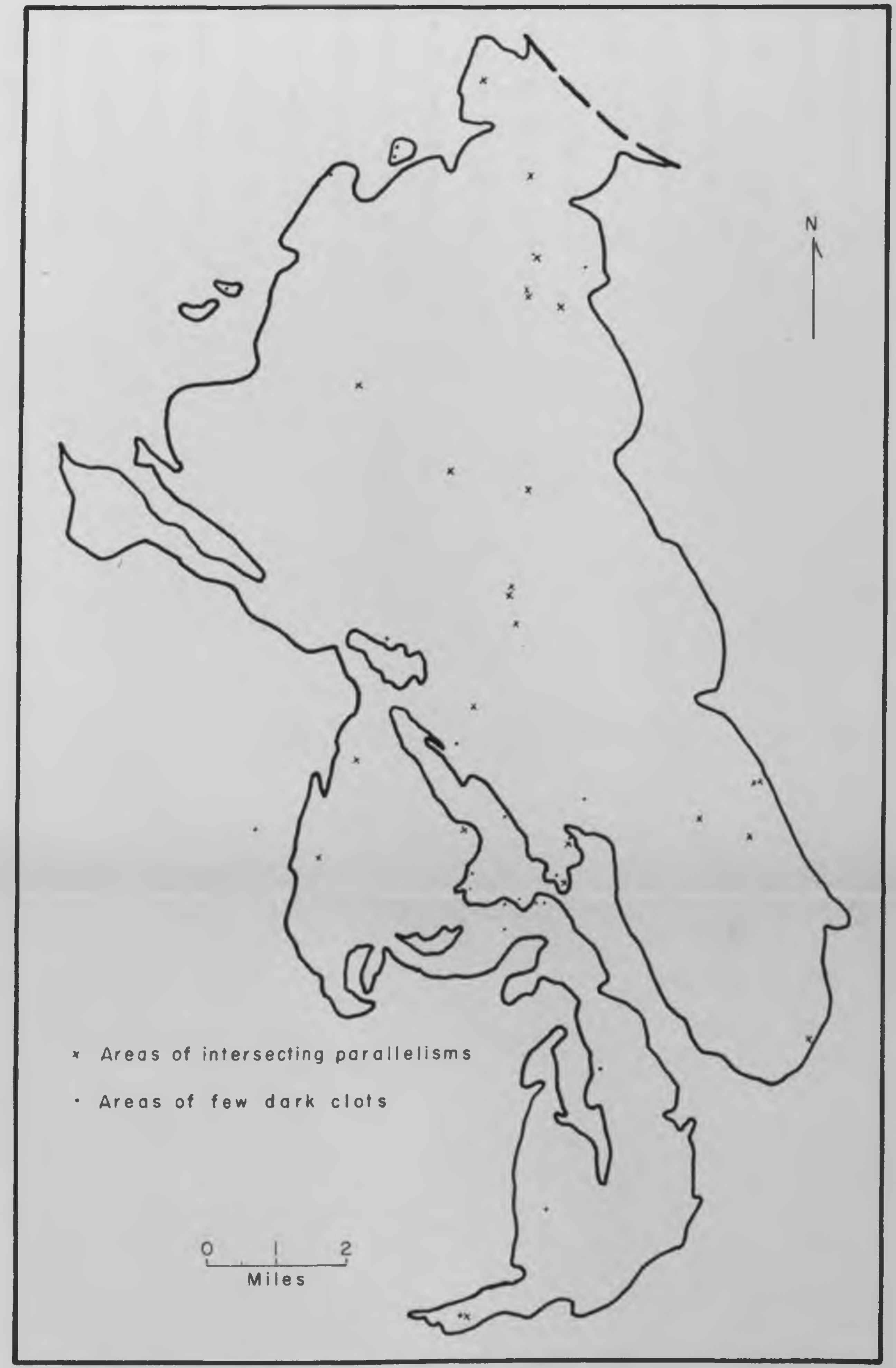


Figure 26. Locations of areas of few dark clots and of intersecting parallelisms in the Bonsall Tonalite.

The dots within the septum and outside of the pluton represent Bonsall Tonalite dikes with few dark clots.

where there are no dark clots over hundreds of thousands of square feet of exposure. The clots range in size from extremes of less than one inch to more than 20 feet long. They normally are between four inches and one foot in length. Many of the illustrations noted above, as well as Figures 24, 27 and 28, show some of the variation in the appearance and density of the clots. The weathering characteristics of these features may be different from those of the mass of the tonalite. Where there is a difference, it will be greater resistance of the clots (Figures 9, 10 and 16).

The composition of the clots, as given by Hurlbut (1935), is 60 percent plagioclase (An_{45}), 20 percent (up to as high as 40 percent) hornblende, 16 percent biotite, and 3 percent quartz plus orthoclase. Osborne (1939) found the average composition of the clots to be 50 percent plagioclase, 10 percent quartz and 40 percent hornblende plus biotite, hornblende being more common. He also showed that many of the clots were derived from the schist, because their plagioclase is more sodic than either the gabbro or the tonalite itself (oligoclase in the clots and andesine and more calcic feldspar in the tonalite and gabbro). He also stated that other clots have plagioclase crystals in the An_{60} to An_{80} range. Osborne concluded that these clots were derived from the gabbro, because of the intermediate composition of the plagioclase in the clots, as compared to the tonalite and the gabbro, and the transitional nature

Figure 27. Unusually drawn-out dark clots in the Bonsall Tonalite.

The view is west, about 300 yards from a gabbro contact 1-1/2 miles S35°W of Vayemo Peak.

Figure 28. Nearly equant dark clots in the Bonsall Tonalite.

The view is south, near the extreme northwest end of the septum. The irregular, horizontal joints are evident in this exposure, which is about 100 yards from the contact with the metamorphic rocks.

of the bulk composition of the clots, which lie between those of the tonalite and the gabbro. Larsen (1948) agreed that it is possible that they are gabbroic, but that they are in almost perfect equilibrium and the main movement and mixing took place before the softening of the inclusions by reaction. He concluded (p. 162) that they came from depth, "as they are abundant in all parts of large bodies and show no relationship to the neighboring older rocks." Larsen's statements seem to be at odds with the situation in the Riverside-Perris pluton, as shown in the next sentences. The reported compositional range of plagioclase in the dark clots in this pluton is oligoclase-bytownite, which seems unlikely to be at equilibrium with the andesine of the tonalite. There is a relationship between the clots and the neighboring older rocks. The wide distribution of the clots is explained by "mixing", which seems unsatisfactory for so large a body and especially so when the Bonsall Tonalite in the rest of the batholith is considered.

In the Riverside-Perris pluton, the clots can be seen frozen in the process of evolution from the schists (Figure 29) and the gabbros (Figure 24). These examples are restricted to the margins of the tonalite body. Away from the contacts of the pluton, the distinctiveness of the inclusions disappears. Even within a few feet of the contact with the slates or arkosic quartzites, very few of the clots

Figure 29. Reaction zone between the Bedford Canyon Formation and the Bonsall Tonalite.

About one mile northeast of Steele Peak, this float boulder lies in the contact zone with the Bonsall Tonalite. An apparent mobilization of the metamorphic rock is seen here. This boulder may also show a possible source of dark clots.

Figure 30. A viscous-shear plane, disclosed by the offset in a dark clot in the Bonsall Tonalite.

The view is southeast, about one mile northeast of Vayemo Peak. The orientation of the shear and the sense of movement are appropriate for a marginal thrust.

can be identified as to their origin. Recognizable xenoliths of earlier rocks within the Bonsall Tonalite have been plotted on an outline map of the pluton (Figure 25). These inclusions are quartzite, for the most part, but slates, phyllites, porphyry and gabbro have been recognized. Inclusions which lie within 200 yards of the contacts with older rocks have not been plotted on Figure 25. While the local origin of some of the clots is certain, the general abundance and distribution of all the clots indicates that local origin is only a partial solution of the problem. Some appear to be locally derived, some from gabbro and some from metamorphosed sediments, and some (if not the great majority) from the site of the formation of the magma. Further comments about this problem are found in the DISCUSSION.

Structure

In the discussions in this paper, magma is taken to mean rock which is mobilized, either by the presence of some melt, whatever the percentage and regardless of the process which formed it, or by virtue of a rheidity in a plutonic rock greater than that of its surroundings. These two possibilities are not mutually exclusive. Granitization is restricted to those processes which make a rock more like a granite without imparting mobility. By these usages, a granitized rock may become a magma by addition of heat, by change in pressure, by increase in volatile content, or by some combination.

The structure of the Bonsall Tonalite is most conveniently discussed in relation to the intrusive and cooling sequence of the rock, as is commonly the case in granite tectonic studies. This sort of treatment presupposes the mobility of the magma and the intrusive nature of the rock. Local development of a border gneiss zone, apophyses of tonalite in the older rocks, conformity of the structure of the tonalite to the shape of the walls, the truncation of the structures of the older rocks by the younger rocks and structures, a tendency towards the development of a chilled margin, the presence of autoliths, and the stoping of wall rock are all evidences of the intrusion of the tonalite. The mobility of the magma is shown by the rotation of included fragments of the older rocks and by the border gneiss, apophyses, and conformity of structure mentioned above. Additionally, the development of planar and linear internal parallelisms, discussed in the succeeding paragraph, is evidence for the mobility of the magma. Planar structures, by themselves, are not unequivocally evidence of flowage. They may be relic structures in a granitized body, or they may be imposed by re-crystallization under stress, as in metamorphic rocks. Linear parallelisms, while not unequivocal indicators of flow where they are distributed over a whole rock body, are strongly suggestive of flow where they are localized and develop to the exclusion or suppression of its planar structure.

Structural Elements of the Flow Phase. The internal structure of the Bonsall Tonalite, while remarkably persistent, is not uniformly developed everywhere. In fact there are exceptional localities where the organization apparently is nil. The internal organization is normally obvious and evidenced by the parallelism of the several structural elements of the flow phase present in the rock. These structural elements are mineral grains, dark clots, xenoliths and (more rarely) clusters of clots or grains, as well as a few schlieren.

The dark clots are the most important and obvious of the structural elements in the pluton. The typical shape of these clots is that of a triaxial ellipsoid, though many may be categorized as disk-shaped or spindle-shaped. Very few can be regarded as angular and fewer still as spheroidal. The axial ratios range from 1:1 to 1:50 and more, although in the higher ranges they begin to lose their clot-like character and become more schlieren-like until they eventually grade into schlieren. The intermediate axis may vary independently through the same range. The normal range of axial ratios is from about 1:1.5 to 1:4.

Figure 31 is an outline map of the Riverside-Perris pluton showing a sampling of the elongation of the dark clots. Some of the elongation may be the result of the original shape of the clots and some of the elongation is caused by stretching. The amount produced by either cause is not known. The sampling is taken in three rather broad bands, which trend northeast across the pluton. The elongations were measured in the plane of the flow foliation, parallel to the strike and to the dip. The absence of symbols on the map does not mean that there is no elongation present, but that no data were recorded for those areas. Localities of no elongation are represented by equant symbols. Some clots are elongated parallel to the strike, with little dip length; others still are equant, and there are transitions. Elongations of greater than 15:1 have not been plotted because they tend to represent maxima, rather than averages and they are very localized. The spread of axial ratios about the average is not uniform. The range of elongations is much greater on the high side of the average. For example, in a rock whose ratios range from 1:1 to 1:10, the average would be about 1:3.

Structural Arrangements of the Flow Phase. The strongest and most widely distributed kind of parallelism in these rocks is the planar array of the structural elements listed above. Even where the elongation of the dark clots

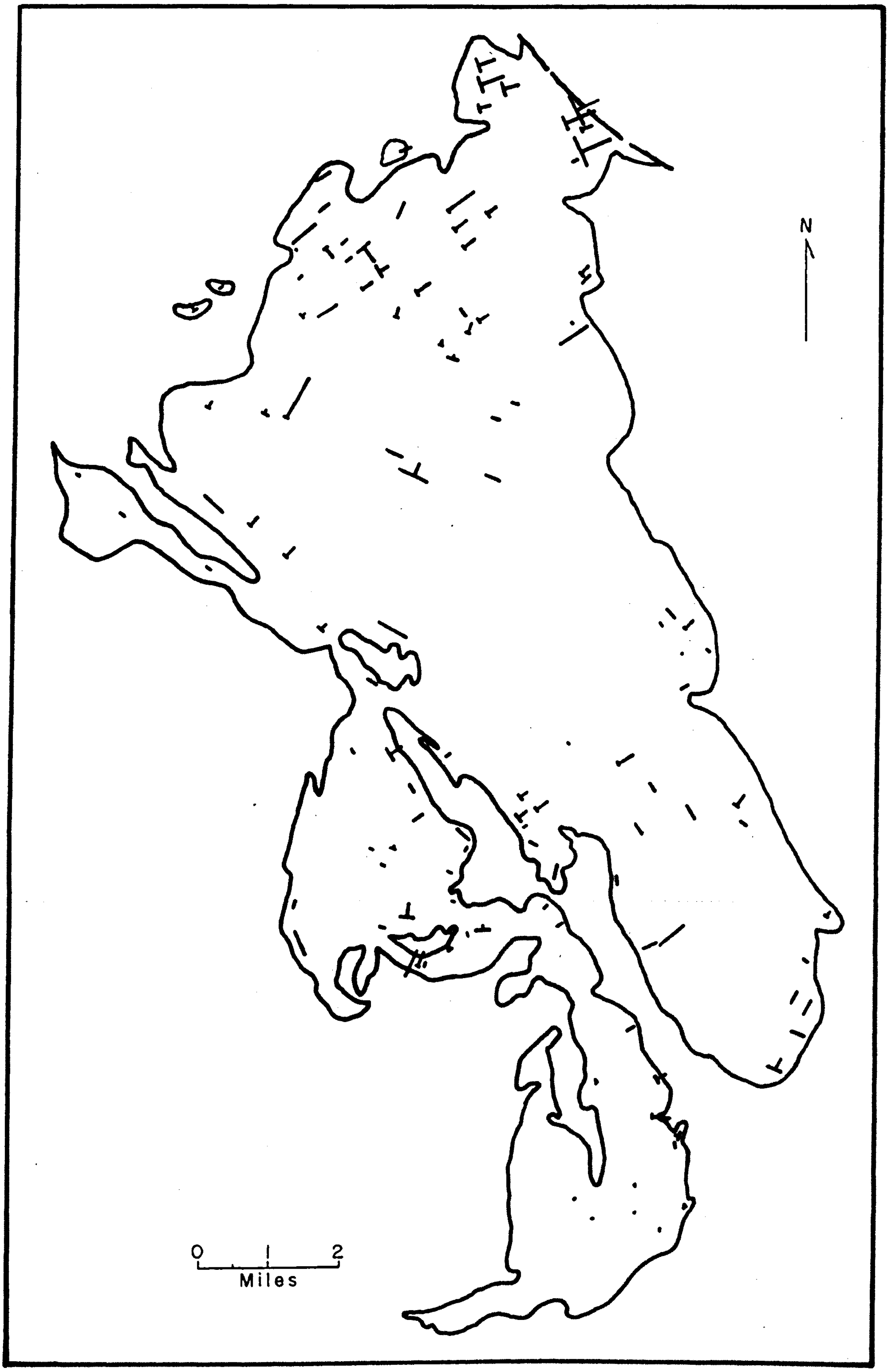


Figure 31. Elongations of the dark clots in the tonalite pluton.

The elongations are shown graphically as a ratio of strike length to dip length.

is very small, less than 1:1.5, the parallelism may be quite evident. Rigorous orientation of nearly equant clots gives a discernible parallelism and the presence of other structural elements may disclose the parallelism when the clots are weakly aligned. The linear parallelism of these elements is also apparent over much of the pluton, either with or without the planar array and to either a greater or lesser degree. Locally, this linear structure may become so prevalent as to completely overshadow the planar structure, if the latter is present at all.

The general trend of the planar structure is north-northwest, with a regional northeastward dip of about 65 degrees. An important exception to so steep a dip is noted below. This is approximately the same as the average attitude of all major structural features of the northern end of the batholith, shown by Larsen (1951) to be N38°W and dipping about 60 degrees to the northeast.

In the Colorado River Aqueduct tunnel the foliation dips most steeply at the screen of Jurassic rock, about 2-1/2 miles south of the Three Sisters (Figure 3), and gradually flattens until it is nearly horizontal where the tunnel enters the valley fill about 5-1/2 miles east (Osborne, 1939). This tendency is not so obvious on the surface because as the dip flattens, the attitudes become more and more difficult to determine on the commonly horizontal exposures. In general, however, there is an easterly shallowing of the dip across the pluton, as shown on the map (Figure 3).

Exceptions to this generalization are present locally. Changes are also present in the regional strike. This regional strike is about the same in the tonalite as in the metamorphic rocks, as well as in the trends of nearly all of the topographic features of this locality, the Perris block as a whole and, indeed, the whole Peninsular Range Province. Over rather large areas, however, the trends in the tonalite are oblique to the contacts. At the contacts with older rock, the planar structure is, where observation is possible, rigorously parallel to the boundary. This will be discussed further in the section on Contacts.

In the northwestern part of the pluton, near the dikes of Rubidoux Mountain leucogranite (Figure 3), there is an area in which the strike of the planar structure of the tonalite is at right angles to the regional structure. The zone lying between the area of northwesterly strikes and the area of east-northeasterly strikes is devoid of useful outcrops. In the eastern portion of this locality, the northwest strikes seem to swing to the north and east to parallel these northeast structures. At the western end of this digression there seems to be a westerly swing in the northwest trending structures. A clue to the origin of the digression may be the strikes of the metamorphic rocks in the area. If these patches of rock are exposures of an underlying mass, or the remnants of a once more extensive pendant, the flow in the intrusive rock could have conformed to the shape of that mass. The only other marked changes of regional strike, except where obviously influenced by border effects, are in regions of relatively shallow dips. Where the dips are shallow, minor changes in attitude will cause major changes in strike.

In the southwestern arc of the pluton, attitudes are found in the planar structure that are nearly perpendicular to the border with the Bedford Canyon rocks. In the 400 to 500 yards that separate these exposures from the metamorphic rocks there are no other outcrops. Based on better exposed areas elsewhere in the pluton, the few hundred yards between the exposures and the walls are enough distance for the structures to swing into concordance with the wall.

During compilation of the field measurements, the recorded trends of the preferred orientation in any given area of outcrop, say 100 yards square, were averaged and plotted on the data map as one symbol. This process was repeated for all exposed areas over the pluton. Figures 4 and 5 are examples, chosen from the data map. In some cases, trends represented on the data map were averaged again in constructing the larger symbols shown on the main map .

(Figure 3). Figure 32 has been prepared to show the spread of trends in the individual outcrop areas throughout the pluton. Places in which the variation is less than 10 degrees are not represented.

In some places the tonalite exhibits intersecting or crossing parallelisms (Figure 26). In these exposures the structure elements are definitely oriented to more than one preferred direction. Some of these exposures are found where the prevailing trend shows little variation (Figure 32). The crossed trends are not to be confused with a single weak or variable trend, because a single variable trend can be seen to swing in one outcrop, or the trend differs in different near-by exposures with no more than one direction of parallelism present in any one exposure. The crossed trends, on the other hand, can actually be seen to intersect or cross.

The measurement of the attitudes of these structures is very difficult. In some areas the differing trends are represented by an interwoven fabric of mineral grains or clots. In other places, the differing trends are evident in domains where one or another of the trends is dominant. Even in boulders where data could be obtained for strike and dip, the simultaneous solution of two or more strike directions with two or more dip directions led to more than one possibility. A choice between the possible solutions could not be

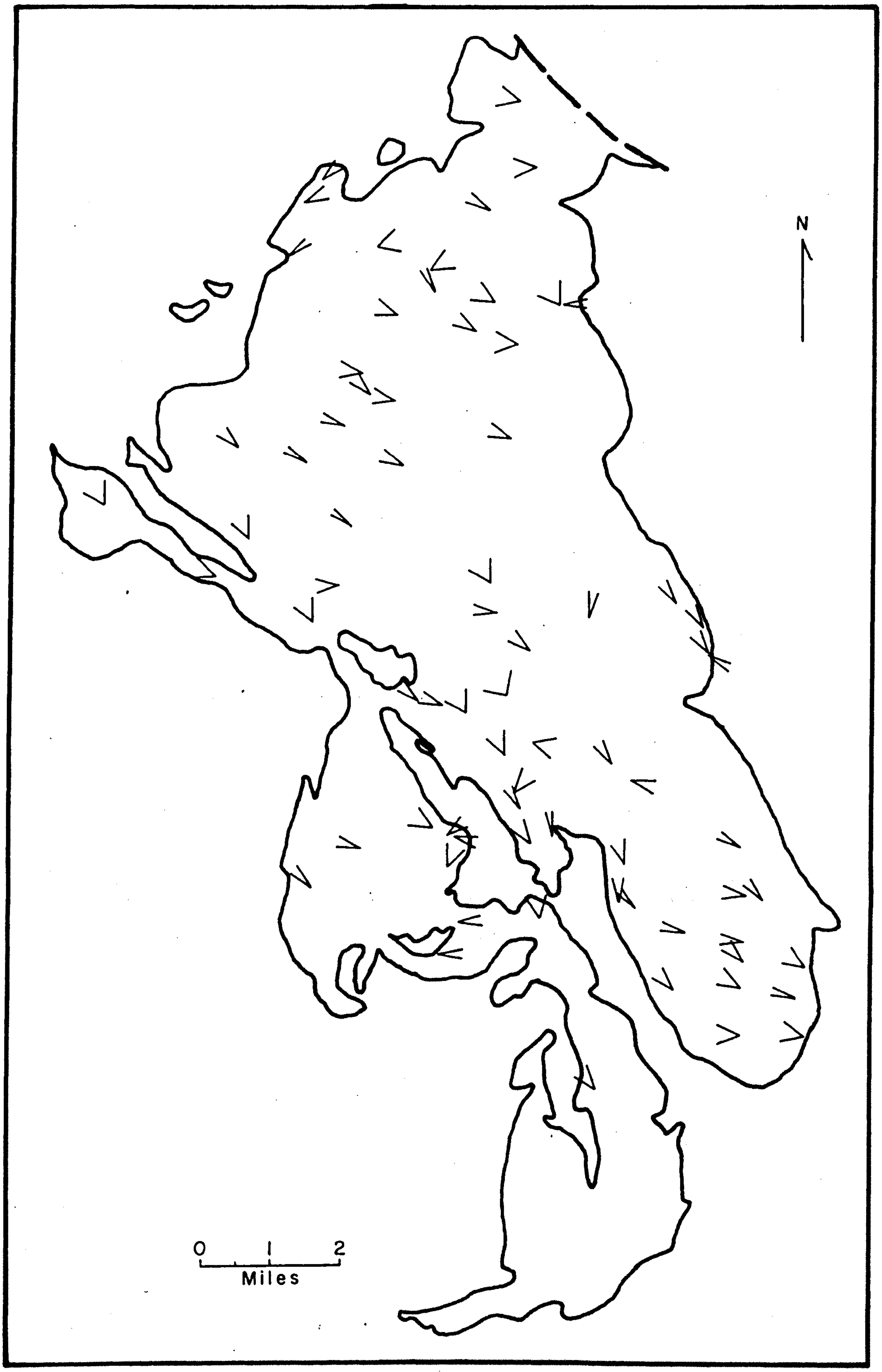


Figure 32. Variability of the trends of the parallelisms in the Bonsall Tonalite.

Variations of ten degrees or less are not shown.

made by comparing them to nearby exposures. Data on the crossing structures have not been presented, other than the localities where they are present.

The possible significance of the crossed or intersecting parallelisms will be discussed in the section on Emplacement of the Riverside-Perris Pluton under DISCUSSION.

Structures of the Intermediate Phase, Viscous-Shear. In a few places, relics of a stage of activity intermediate between fracture and flow have been observed. These vestiges may be more common than they seem, because they are not very obvious. This obscurity is especially pronounced when they lie parallel, or almost so, to the trace of the parallelism on the face of an outcrop or where the rock surface is weathered. The general impression is one of a healed joint with considerable dragging of the mineral grains and clots on either side of the fracture plane (Figures 30, 33, 34 and 35). The presence of dragging indicates a certain mobility of the tonalite at the time of the formation of these planes. Any one of these viscous-shear planes cannot be traced more than a few feet and the drag effects are confined to a very few inches adjacent to the plane. The shear "plane" is a zone, about one-half inch wide, of fine-grained tonalite (about 0.05 inch grain diameter) whose grains are oriented in a planar array parallel to the plane

Figure 33. Close-up of the viscous-sheared clot, shown in Figure 30.

The drag caused by the thrust movement in the viscous-shear plane is evident in the clot and in the mineral grains.

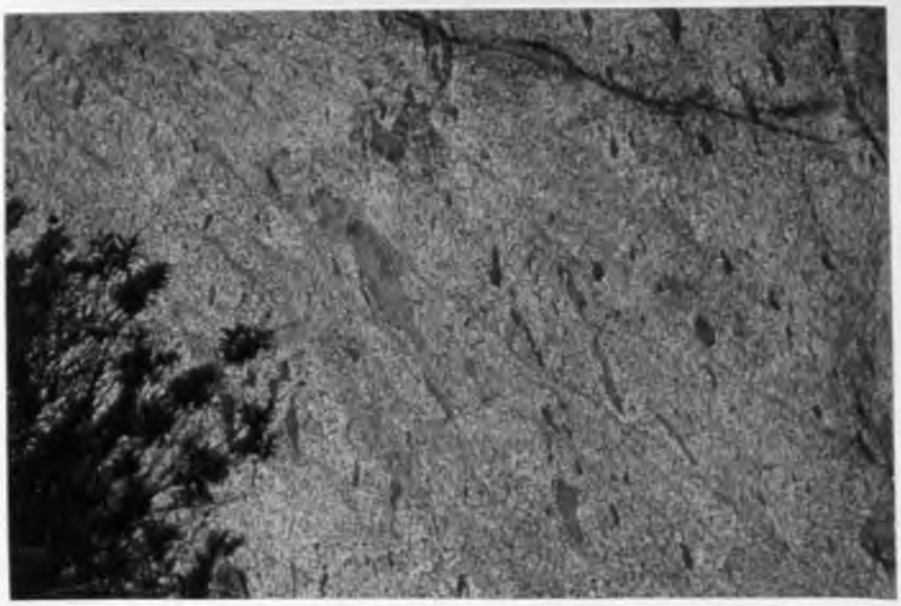


Figure 34. A more evident expression of viscous-shear in the Bonsall Tonalite.

The view is northwest, about one mile northwest of Vayemo Peak.

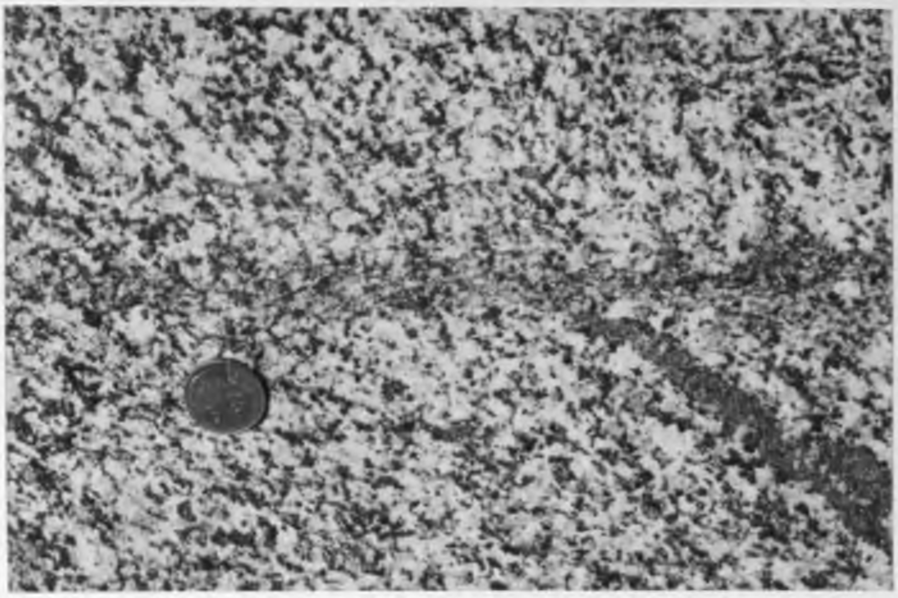


Figure 35. Close-up of a viscous-shear plane.

The locality is the same as in Figure 34. These planes show drag related to reverse motion, but they dip towards the contact and are not marginal thrusts.

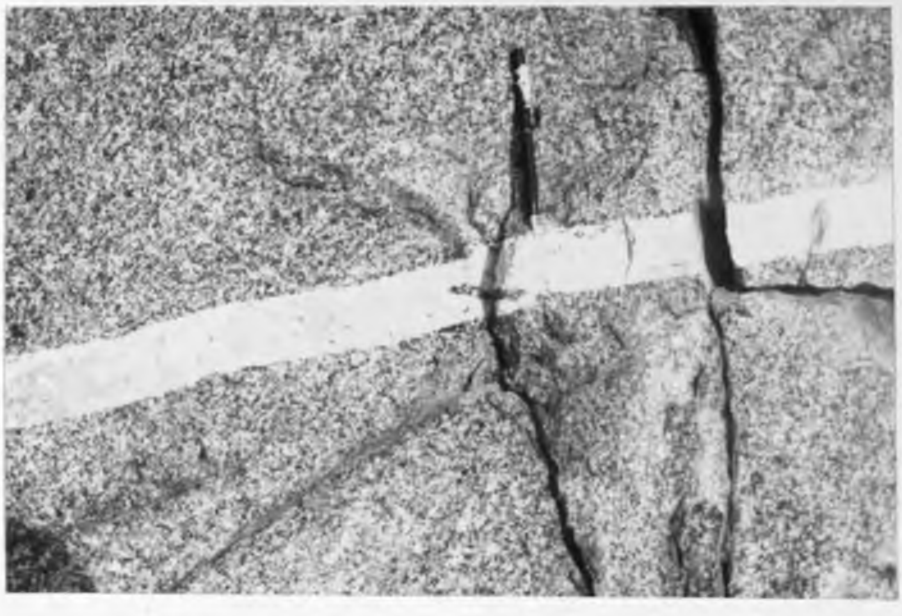


Figure 36. Inclusion of Bonsall Tonalite in a granite dike.

The view is east, about two miles east-northeast of Gavilan Peak. The wall fragment is rotated in place, showing the dilational-intrusive nature of the dike.

(Figures 33 and 35). These structures may or may not be parallel to joints in the locality. They have not been observed parallel to the general planar arrangement of the tonalite, perhaps because the indistinctness of the phenomenon requires it to be cross-cutting in order that it be apparent.

Figures 37 and 38 show features that are possibly related to the viscous-shear phenomenon, or perhaps to the crossed foliation. In Figure 37, the shape of the clusters of dark clots is not concordant with the trends of the mineral parallelism, or with the trends of most of the individual clots. Most individual clots are parallel to the general planar structure, but some of them, and the long axes of the clusters, are normal to the general parallelism. This discordance between the trends of the clots and the shape of the cluster has been seen several times in the tonalite pluton. The remote end of the large cluster in Figure 23 is abruptly truncated across the trend of the cluster and the individual clots. Some controlling structures, possibly shears, trending east-northeast and northwest probably caused the configuration seen in Figure 37. Possible mechanisms are numerous, but the controlling structures are not evident.

Figure 38, as well as Figures 34 and 35, show the development of schlieren as a result of the viscous-shear

Figure 37. Discordant clusters of dark clots in the Bonsall Tonalite, about 2-1/4 miles south of Vayemo Peak.

Dark clots, mineral grains, joints, and schlieren are shown.

Figure 38. Drag effects along schlieren in the Bonsall Tonalite, about 1-1/4 miles east of Vayemo Peak.

Dark clots, mineral grains and schlieren are shown.

process. It seems evident, in Figure 38, that the material in the northeastern segment of the exposure continued to flow long enough to cause its structure elements to be aligned with the margin of the block. The dark clots and mineral grains in the central block seem to be dragged from east-southeast to east-northeast trends by the shearing action of the motion in the northeastern block.

The significance of the structures illustrated in Figures 30, 33, 34, 35 and 38 is that the magma had a differential mobility during its cooling. The orientation of the shears and the directions of movement are not of great importance, because they are a result of local stresses, reoriented at the end of the flow phase. The reason for this reorientation seems to be the inhomogeneous stiffening of the magma in adjacent areas. These new stress orientations produced new flow directions, drag or rupture, depending on the condition of the material. In areas of appropriate viscosity, the result was the dragging and rupturing of the rock and the granulation of the crystals. The viscous-shear phenomenon represents a transitional phase between fluid or rheid flow and brittle rupture.

Structures of the Fracture Phase. Although the Bonsall Tonalite has apparently been well jointed throughout, few joints can be measured in many areas. This scarcity of measurable joints is caused by the nature of the weathering, which seeks out joint planes to start its reduction of the rock to spherical residuals. Even in some areas of many outcrops the attitudes of the joints cannot be determined. Their spacing is revealed by the size of the residual boulders.

Because of the wide range of measured attitudes in the fracture pattern of the Bonsall Tonalite pluton, the breaks have been plotted on equal area projections (Figure 3). In these diagrams, the poles of the planes of the ruptures have been plotted on the lower hemisphere. Two diagrams have been prepared for the pluton as a whole. These are the projections showing all fractures measured in the pluton and showing all measured veins and dikes in the pluton. The other diagrams (numbered 1 through 5 on Figure 3) have been placed on the map in the vicinity from which their data were compiled. Diagram 1 represents the fractures in the area of shallow dips in the northeast of the pluton. The data in Diagram 3 show the fracture distribution in the area of shallow dips in the east-central part of the pluton. The fracture attitudes in the Gavilan are shown in Diagram 4. Diagram 5 represents the fractures in the southwestern arc of the pluton. The area represented in Diagram 2 is greater than that in any of the other local diagrams. Diagram 2 has all of the fractures west of the septum, that lie north of the aqueduct tunnel, and those east of the septum, except for the areas of shallow dips of Diagrams 1 and 3. The number of fractures plotted in each diagram follows: 202 in Diagram 1, 769 in Diagram 2, 256 in Diagram 3, 169 in Diagram 4, 219 in Diagram 5, 246 in all veins and dikes, and 1615 in all fractures of the pluton.

Three groups of joints may be identified almost everywhere in the pluton. Two of these groups may be properly called sets. One set strikes east and is vertical. (See the joint diagrams on Figure 3.) The dip may vary 15 degrees to the north or south, as may the strike, but this set is present in almost every outcrop that reveals joints. The second set strikes north to north-northwest and dips 90 degrees. This set is present almost everywhere that joints are evident. A northwesterly group of joints is nearly parallel to the prevailing planar flow structure. These northwesterly trending joints, whose strike is parallel to the flow structure, are not a true set because their dips are not parallel. The horizontal parallelism of these joints with the planar flow of the tonalite is often a clue to the orientation of the foliation when it is not well developed. Many of these ruptures are parallel to the dip of the flow structure as well, but many of them form a fan and dip steeper or shallower than the plane of the flow and many dip in the opposite direction.

The wide range of dip attitudes is shown in the development of a weak girdle, whose axis is northwest, on the equal area projections (Figure 3). Within the girdle there are two maxima. The first and strongest of these coincides with the attitude of the planar flow structure in the area; therefore it represents those joints that are parallel in strike and dip to the flow structure. The other maximum represents a fairly weak set of fractures with a relatively shallow dip. The presence of the set striking northwest and dipping 35 degrees east was not recognized during the mapping and became apparent during the preparation of the joint diagrams. This maximum also represents a large percentage of the dikes and veins in the pluton (see below).

The four groups of fractures, three of them evident in the field, account for essentially all of the joint directions in the pluton that have a wide range of occurrence. The two vertical sets contribute to the vertical belt shown on the joint diagrams and the shallow dipping northwest set contributes to the belt about the northwest horizontal axis. The northwesterly group of joints that is parallel, in strike, to the planar flow structure contributes to both girdles.

Joints which dip less than 45 degrees represent less than 10 percent of the total number of joints. Their greatest concentration is within 200 yards of the contact zones with older rocks. These joints are not everywhere present in the margins and where they do occur, they usually dip into the tonalite (Figures 14, 28 and 30) and therefore represent feather joints, or marginal thrusts, or both. Elsewhere, the low angle joints are often extraordinarily irregular, as though they expressed a rather imperfect sheeting.

Locally, sets develop which have no apparent relationship to the joints which are visible in the surrounding area. That is, the knowledge of a set or sets of joints in one group of outcrops over a distance of a few hundred feet cannot necessarily be applied to a new area a few hundred yards away. These locally developed joints account for the wide range of attitudes on the joint diagrams and the low maxima.

In only a few cases has the sense of movement on the joints been apparent. These joints are so scattered and the attitudes and sense of movement are so variable that they do not give useful information about the stress environment of the pluton and it is possible that they are related to local stresses. It usually cannot be determined what the relationships are because the joint planes are habitually attacked by the weathering and erosional processes to the extent that deductions about movement are impossible. Where the movement, or lack of it, can be determined, the fractures are generally without noticeable displacement.

The dikes that are present over nearly the whole of the tonalite pluton seem to offer little additional information. The plot of their attitudes on the equal area projection shows that they cluster more tightly about a preferred attitude than do the joints. The dikes and veins dip shallowly to the southwest, spreading to steep southwest dips and on to a few steep northeast dipping ones (Figure 3). They do not show the girdle of vertical attitudes that the joints show on the diagrams and the belt about the northwest axis is much reduced. The minor maximum for the southwest dips has become the dominant feature of the pattern.

If the maximum representing the northwest-striking steeply northeast-dipping joints is also taken to represent the attitude of the planar foliation in the tonalite, the nature of the shallowly southwest-dipping joints and dikes becomes apparent. These fractures lie perpendicular to the lineation in the tonalite, which in turn lies in the plane of the foliation and trends somewhat to the east or southeast of the dip. The southwest-dipping ruptures are the "Q"- or cross-joints of granite tectonics. The cross-joints lie perpendicular to the elongation of the granite and probably open as a result of the contraction of the magma on cooling.

The other, much smaller, maximum on the equal area projection for the dikes, represents the direction parallel to the foliation in the tonalite. The volume of material in these dikes is much greater than in the southwest-dipping joints. This direction includes nearly all of the dikes large enough to plot on Figure 3. The opening of this direction is thought to be related to the tensional stress involved in the emplacement of the batholith and will be discussed in the Emplacement of the Riverside-Perris Pluton under DISCUSSION and in the CONCLUSIONS.

These dikes and veins range in size from one-sixteenth of an inch to 250 yards thick. The latter is an exceptionally large example. Only about 20 dikes have been shown on the map (Figure 3) because those are all that exceeded the minimum practical size for plotting, about 50 yards wide. Of these 20, only three or four are wider than 100 yards. Of those not plotted on the map, a dike eight or ten feet thick is exceptional. The great majority are pegmatitic or aplitic and lithologic types commonly grade one into another within the same dike, which may be a few inches or a few feet thick. Quartz veins and epidote veins are also quite common. The quartz veins, commonly with a few percent of feldspar, are inches in width whereas the epidote veinlets are fractions of inches wide. The lengths of these dikes and veins are indeterminable because of the spotty nature of the outcrops, but many dikes of from one to one and one-half feet thick have been traced for 500 feet. Veins less than one-quarter inch in width that can be traced for 50 feet are not unusual. One notable exception to the lack of continuity of outcrop of these dikes lies about one and one-half miles north of Gavilan Peak. A dike of aplite and pegmatite, which varies from three to six feet thick, crops out continuously for a distance of nearly one and one-quarter miles. It dips only three degrees to the northeast, the shallowness of the dip being also exceptional, so that it essentially parallels the contour lines around the northwest trending nose and valley.

These dikes sometimes have inclusions of tonalite, but the inclusions are very rarely seen. In several examples, the tonalite fragments have been rotated relative to the internal structure of the nearby tonalite walls. Chilling is sometimes apparent at the contacts of the dikes, but it is usually so slight that its presence is open to question. Apophyses are very rare. Although the evidence is fairly scanty, all evidence points to the dikes being intrusive and dilational (Figures 36 and 39).

In some composite dikes, the aplite is at the borders with the pegmatite in the center. In the wider dikes,

Figure 39. Wedging in an intrusive dike.

This photo is of the same dike as in Figure 36. The view is northeast. The net thickness of the dike material is maintained in the doubly wedging portion.

Figure 40. A Bedford Canyon Formation-Bonsall Tonalite contact.

The view is south, about 1-1/4 miles north of Steele Peak. The quartzite (right) strikes N20°E and dips steeply west. Joints lie at N75°W, 25°S. Note the parallelism in the tonalite at the border.

however, there may be an alternation of the two phases several times across the width, or a pod-like distribution of either phase within the other.

The contacts of the dikes are generally not exposed and dike material in place is uncommon. The exposure is often only a line of pegmatite rubble in the soil. The dikes are usually much more resistant to weathering and erosion than the tonalite. Thus they stand up with a few inches or feet of relief above the grus. No dikes or veins have been seen to cross a contact.

Contacts

The nature of the tonalite is the same one mile away from a contact as it is, in many cases, only a few feet away. This statement applies to the mineral content, number and size of the dark clots, jointing, internal parallelism, and the overall appearance of the rock. Exceptions will be discussed in the appropriate sections.

In some places there appears to be a selvage of mobilized, partly digested wall rock a few feet or a few inches thick at the contact of the tonalite and the metasediments. Exposures of this phenomenon have not been seen in their actual relationship to the contact, but float and isolated outcrops in the area of the presumed contact have led to this conclusion (Figure 29). Normally, the intrusion seems to have had only a dynamic effect on the Bedford Canyon rocks (Figure 40), in some localities working its way into joints in a small scale stoping action with no other apparent effect on the host. Evidence of stoping on any large scale is quite rare. To the west and southwest of Vayemo Peak, blocks of the Jurassic rock are found as far as one-quarter mile from the main contact. The most remote block measures about 50 by 75 yards on the surface. Three miles northeast of Gavilan Peak, there is a block 50 yards long which lies about 200 yards east of the septum and several smaller blocks which lie closer (Figures 41 and 42). Nearly all these blocks are concordant, in their gross outlines, with the planar structure in the tonalite, but a few of the xenoliths have been found to be discordant. Blocks have been found whose internal structure and external shape are both parallel to the surrounding planar structure of the Bonsall Tonalite. Other xenoliths have only the internal structure parallel. A few have no features parallel to the structure in the surrounding rock. These relationships suggest that the fragments were carried long enough for most of them to be oriented by the flow of the mobile medium. The large block in Figure 39 is concordant, in its outline, to the structure of the enclosing rock. The average strike of the foliation in the block is parallel to the strike of the planar structure in the adjacent tonalite,

Figure 41. A large xenolith in the Bonsall Tonalite.

The view is about northeast, three miles N60°E of Gavilan Peak. The rubbly outcrop of the Bedford Canyon Formation rocks is evident in this view perpendicular to the strike of the planar structure on the Bonsall Tonalite.

Figure 42. A concordant inclusion of metamorphic rock.

The view is southeast, about 50 yards southwest of Figure 41. Note the parallelism between the internal structure of the xenolith, its shape and the planar array in the Bonsall Tonalite.

but the dip of the metamorphic structure is about 20 degrees shallower than the dips in the tonalite. Samples taken from these xenoliths cannot be distinguished from samples taken from the septum. These indications suggest that the blocks foundered or were stoped by the tonalite and were carried with it long enough to orient the fragments in the environment. This evidently took place at a time when the chemical activity of the tonalite was too weak to make noticeable changes in the xenoliths. Recognizable xenoliths have been found a mile or more away from the nearest exposures of the wall rocks (Figure 25), but pieces as large as one foot long are quite rare within 100 feet of the contacts. The mechanical disruption of the walls by the tonalite is a restricted and minor phenomenon. Materials susceptible to assimilation or to conversion to dark clots evidently react very quickly with the magma. Inclusions therefore seem to either rapidly lose their identities or retain them indefinitely.

In three localities, the tonalite intrudes the metasediments in a series of sills which are from five to fifteen yards thick and are generally parallel to the internal structure of both the metamorphics and the tonalite. These sills are separated by screens of the wall rock that are 10 to 30 yards thick. This rather complex contact zone may be 200 yards thick. Within these sills, the intrusion is typically barren of inclusions of the host, while it may be replete with the usual dark clots. These areas may show stoping phenomena frozen in a state of partial development. Continued intrusion would wedge the screens farther and farther into the main body of magma until they would eventually be independent of the walls.

The contacts of the Bonsall Tonalite with the San Marcos Gabbro are quite different than those with the Jurassic rocks. No exposures of the gabbro-tonalite contacts were found and our knowledge of them comes from float boulders and scattered outcrops in the presumed contact zone. Hurlbut (1935) concluded that the marginal assimilation of the gabbro by the tonalite resulted in a transitional zone between the two rocks, and that there was no true border. The contact between the two rocks in the Riverside-Perris area seems also to be transitional. The gabbro does not appear to have been completely solidified before the advent of the tonalite, or else it was partially re-mobilized by the tonalite. The two rocks seem to be mutually interpenetrating, in some localities, and this phenomenon can grade into a few small dikes of the younger rock cutting through the gabbro or a few recognizable inclusions of gabbro in the younger rock. Figure 24 shows an outcrop of zone of mixing of the tonalite and gabbro. This locality is about 200 yards from the nearest outcrop of unaffected gabbro. The streaking, flowing and mixing are typical of many exposures.

In this exposure, the tonalite is the most prevalent phase, but nearly all gradations between gabbro and tonalite may be found. Compare this illustration with Figure 23, which shows a large cluster of dark clots that do not show any tendency to mix with the tonalite. The locality of Figure 23 is about two miles from the nearest exposed older rock. Figures 19 and 27 show rocks whose clots exhibit considerable flowage deformation, but no tendency to mix. Evidently, some inherent difference existed in the material which was destined to become clots that allowed it to come to chemical equilibrium, or at least made it resistant enough to survive the activity at the border of the pluton. Once in the body of the pluton, the clots seem to have been completely stable, although quite variable in their composition. The non-reaction could be a result of an equilibrium state or the lack of sufficient energy to transform the clot material. An explanation of the difference in the chemical activity at the borders and in the interior of the pluton may be due to the concentration of the water at the cooler margins of the pluton (Kennedy, 1955). Differences in the wetness of the tonalite magma adjacent to the metasediments, which might be expected to have a greater water content, and adjacent to the porphyry, which would be expected to be dry, may explain the development of the gneiss by relating the lower freezing range of the dry magma to the dry porphyry.

The planar array of the various structure elements of the flow phase is usually well developed in the contact zones. This structure is parallel to the contact in strike, but the vertical exposure is so limited that the conformity of the dips can only be assumed on any scale greater than a few feet. The contacts in the central portion of the pluton trend to the north-northwest, parallel to the regional structure and, largely, to the detailed structure as well. Because remnants of the Jurassic rock trend to the north-northwest here, a general conformity of contacts and internal structures is expected and is indeed present. In detail, however, the same condition is found to prevail. The conformity of the structure of the tonalite to the rather tortuous convolutions of some of the embayments in the older rocks is remarkable. The structure of the tonalite changes over distances of less than a quarter of a mile from parallel to the contact to parallel to the regional trend, through angles of up to 70 degrees.

Fragments of tonalite, which have a higher mafic content than the enclosing tonalite, have been identified near the contacts with older rocks. These fragments may carry normal dark clots and exhibit planar and linear parallelisms of the clots and mineral grains. Their mafic content is only slightly greater than the surrounding material and is not atypical of the range of Bonsall Tonalite composition. These fragments are almost certainly autoliths of the tonalite. These autoliths are generally discordant with the parallelism in the encompassing tonalite, probably because they were not carried long enough to be oriented by the flow.

Another typical feature of the contact zones is the prevalence of pegmatite veins and simple quartz veins. Usually the only evidence of these veins or dikes is float, but many are exposed through the soil. The presence of these veins indicates a fracturing of the border zone that is not otherwise evident. The pegmatite is normally composed simply of quartz, feldspar, and negligible biotite. The badly weathered state of the biotite may indicate its presence in a greater amount in fresh rock. Garnets as large as one-eighth of an inch were observed in two localities on the contact and schorlite comprises up to 50 percent of the rock over a few square feet in certain areas.

The planar and linear parallelism are usually well developed in the borders of the tonalite mass, but there are regions in which it is very difficult, if not impossible, to find evidence of parallelism. At its strongest, the development of the parallelism is no greater in the margins than it is in many parts of the body which are removed from the contacts by a mile or more. Areas noted to have weak development of parallelism actually cluster along the septa. The data collected on the elongation of the dark clots (Figure

31) show that there is no special development of the elongation near the borders or the septum.

The grain size varies but little from the average as the contact is approached. What possible chilling there is, lies within the range of the normal variation of grain-size of the tonalite and is therefore not conclusive. The distinct impression given by the field evidence, however, is that many border areas have a smaller than average grain-size. The normal range of grain-size is one-eighth to one-quarter inch, with one-quarter inch being closer to the average. Near the pre-batholithic rocks, the grain-size is often about one-eighth inch. Chilling of the tonalite has not been observed in contact with the gabbro, even in small dikes and irregular apophyses a foot or less wide, because of the warm environment next to the earlier intrusion. The contact with the Temescal Wash Quartz Latite Porphyry has not been observed, but the tonalite which crops out closest to the hidden contact is quite gneissic and granulated (Figure 12), indicating an increased viscosity and the likelihood of a chill zone closer to the contact.

The abundance of the dark clots and their size and composition seem to be unrelated to their proximity to the contact or to the nature of the wall rock. These relationships do not apply to the recognizable inclusions of the pre-existing rocks which become more abundant close to their respective sources.

Flow structures are expected to be best developed where there is a relatively high viscosity, coupled with large amounts of relative motion. These conditions are expected to prevail at the borders of a pluton, and against a septum, where the relative movement is thought to be greater and the temperature cooler than in the center of the mass. Evidence does not show any significant increase in the quality or quantity of the flow structures near the older rocks. In fact, evidence suggests an inhibition in the development of parallelisms near the septa.

Weak development of flow structures results from small relative movement and/or low viscosity. Strongly developed structures may be more or less erased by continued thermal or chemical activity, or the imposition of new structures. The weak development of the planar and linear arrays in parts of the border of the tonalite pluton, suggests the possibility that the septum was moving with the tonalite. It is unlikely that the structures were erased, because the localities of crossing structures are not associated with the areas of weak development. A low viscosity phase is unlikely without excess heat or mineralizers. The somewhat smaller grain-size speaks against the erasure by chemical or thermal activity.

The normal development of the structures along most of the borders can be explained by the movement of the magma, whose viscosity is supporting and moving the older rocks. The reduction of the relative rates and amounts of movement, as the septum is carried along, might explain the failure of the strongest structures to develop at the margin. Also, local relatively stagnant areas exist, which show no structural development.

The impingement of the younger plutonic rocks on the Bonsall Tonalite is restricted to relatively short contacts in the west and north. The intrusion by the Woodson Mountain Granodiorite has caused effects in the tonalite similar to those that the tonalite caused in the gabbro (see below in the section on Contacts under the Woodson Mountain Granodiorite).

The intrusion of the Rubidoux Mountain leucogranite has had no noticeable effect on the structure or texture of the tonalite.

Miscellaneous Tonalites

The miscellaneous tonalites, as defined by Larsen (1948), are present in a few dikes within the mass of the Bonsall Tonalite. One of these dikes has a greater volume than the rest of the dikes combined. This dike, previously mentioned in the discussion of the fracture phase of the emplacement of the Bonsall Tonalite, lies about 3-1/2 miles east-southeast of the Three Sisters (Figure 3). The dike trends north-northeast and is a little more than one mile long and about 250 yards wide. The boundary of this body is impossible to define from surface evidence. Its weathering properties are those of the Bonsall Tonalite and its occurrence is in an area of very low relief and scattered outcrops. The surface there is one of grus, but the few erosional remnants show the dike rock to be a medium-grained, gray to buff tonalite with only minor mafic content.

A faint parallelism of the biotite plates indicates a vertical planar flow structure within the dike and a probable vertical dip of the dike as well.

The smaller dikes of this rock range in size from a few inches to a few feet in width. Their occurrence is the same as that of the aplite-pegmatite dikes previously discussed in the fracture phase of the Bonsall Tonalite and they often occur together.

Woodson Mountain Granodiorite

The Woodson Mountain Granodiorite was named by Miller (1937) for its occurrence at Woodson Mountain, about 55 miles south of Perris. Larsen (1948) extended the range of the unit to much of the batholith, including the Riverside area.

He concluded that the Cajalco quartz monzonite (in the area north and west of Gavilan Peak) and the Steele Valley granodiorite (in the area west of Steele Peak), which were both named by Dudley (1935), fell within the range of composition of, and were correlative with, the granodiorite of Woodson Mountain.

Six lead-alpha measurements of zircons from the granodiorite in the type locality and five more from a locality 20 miles farther southeast, give an age of 105 m.y. Three analyses of monazites from the second locality give the age of 115 m.y. These two dates are stated to lie within the error of plus or minus 10 percent (Bushee, et al., 1963). These dates compare favorably with the 105 plus or minus 12 m.y. age of the granodiorites in the Southern California batholith given by Larsen, et al. (1958). This age correlates with the intermediate position of this rock in the intrusive sequence, as determined by field evidence (Larsen, 1948).

The granodiorite cuts and contains fragments of the Bonsall Tonalite and the older rocks in the Riverside-Perris area. Larsen and others have found this to be true throughout the batholith. Elsewhere, this granodiorite is intruded by younger rocks of the batholith (Larsen, 1948; Pampeyan, 1952).

Extent

Larsen (1951 and 1954) stated that this granodiorite covers about one-quarter of the area of the batholith of Southern California and that it also is the second most common rock unit in the batholith after the Bonsall Tonalite. The granodiorite lies along the southwestern edge of the Riverside-Perris pluton and invades it and the older rocks along nearly five miles of that boundary.

Petrology

The variability of the composition of the Woodson Mountain Granodiorite has been recognized by many authors. Larsen (1948) included rocks in the Riverside-Perris area that range from granodiorite, through quartz monzonite, to granite in his Woodson Mountain unit.

The rock is coarse grained, about one-quarter inch grains, and shows less than 10 percent of mafic minerals. It is a white to light gray leucocratic rock which commonly shows pink casts, because of the orthoclase content. With rare exceptions, the major mafic mineral is biotite. Hornblende is present. The quartz content ranges between one-quarter and one-third of the rock. Because of the wide compositional range, the plagioclase-orthoclase ratio varies considerably but their combined percentage is about 65 to 70.

Structure

The internal structure of the Woodson Mountain Granodiorite is very poorly developed away from the intrusive contacts with the older rocks. Adjacent to the older rocks, a well developed envelope of planar structure is normally present in the granodiorite (Figure 43). The structure is evident in the alignment of mineral grains, inclusions, dark clots and schlieren. This envelope is a few tens to a few hundreds of feet thick, although it may disappear within only a few feet perpendicular to the contact. The structural isotropy is evident over the bulk of the exposures.

The northern part of the Gavilan is characterized by the large number of dikes of the Woodson Mountain rocks that intrude the tonalite in that area (Figure 3). These dikes generally parallel the strike of foliation in the tonalite and seem to dip steeply, possibly parallel to the foliation. They are quartz-monzonitic and granitic in composition and typically coarse grained. Mineral parallelism is very weak or lacking in the central part of the dikes.

Contacts

As is normal in this part of the batholith, the contacts with the older rocks are not exposed, but there is a typical increase in the parallelism of the internal components as the contacts are approached. Little chilling is evidenced,

Figure 43. The Woodson Mountain Granodiorite-Bonsall Tonalite contact.

The view is about N15°W, about 2-1/2 miles south of Three Sisters. The steeply east-dipping planar array in the tonalite (left) is cut by the near-vertical contact with the granodiorite, whose envelope of planar structure is shown.

Figure 44. Rubidoux Mountain leucogranite-Bonsall Tonalite contact.

This disoriented boulder is located at the northern edge of the southern leucogranite dike. The faint parallelism in the leucogranite (bottom), the contact and the planar array in the tonalite are all parallel.

even near the Jurassic rocks (Figure 22). The variation in the thickness of the foliated shell was mentioned above. As the contacts are neared, the incidence of inclusions in the rock increases. The nature of the inclusions ranges from clearly recognizable, evidently unaltered fragments of the intruded rocks, to the dark clots whose source is open to question. These inclusions become more common as the margins are approached because of the increasing proximity of their source. These observations apply equally to the dikes of the Woodson Mountain Granodiorite in the tonalite.

Some mixing of the granodiorite and the tonalite is evident in a few areas within the contact zone. No complete spectrum is present, but rocks are found which show gradations from a few inclusions or schlieren of tonalite in the Woodson Mountain rocks to narrow veins and stringers of the granodiorite in the tonalite. Rocks intermediate in this series show the streaking and blending of the two phases at some localities. Included fragments of the tonalite with partially reacted rims are more commonly seen, but it seems that there were areas which had not cooled below the temperature at which the tonalite could be remobilized by the additional heat of the invading granodiorite.

Contacts with younger plutonic rocks are not present in the area investigated here.

Rubidoux Mountain Leucogranites

Larsen (1948) named these leucogranites for their occurrence at Rubidoux Mountain, in the western part of the city of Riverside. There, he was able to identify and name two phases. He reported (1951) that the coarse phase intruded and carried fragments of the fine phase. In the area mapped here (Figure 3), these distinctions cannot be made.

The entire area mapped by Larsen as fine and coarse leucogranite of Rubidoux Mountain is less than two square miles. Less than one-half square mile is present in the area of Figure 3.

Quartz comprises about 30 percent of the rock and the rest of the rock is composed of the combined feldspar, except for traces of some former mafic constituent. The badly weathered state of the rock makes the identification of the feldspars extremely difficult. The grain size of the rock is generally from one-quarter to one-half inch, although Larsen (1951) stated that the fine-grained phase looks aplitic and part of what is mapped here was assigned to the fine-grained phase by Larsen.

The outcrop pattern leads to the conclusion that there are two nearly parallel dikes of this rock. They trend northeasterly and their dip is not known. The parallelism of the constituents is weak, if present, in nearly all of the rock and only near the contacts is there any consistent development of recognizable parallelism. One measurement near the northeastern end of the southern dike shows that the internal parallelism of the dike strikes northeast and has a southerly dip of forty-five degrees, based on the attitude of schlieren and mafic inclusions. In the southwestern end of the dike, several determinations of the attitudes of the dark mineral traces show that the parallelism strikes easterly and dips to the north. Thus the internal organization of the dike seems to strike parallel to the borders of the dike and the dips are in the same direction as the nearby planar array in the Bonsall Tonalite.

Three northwesterly trends have been discovered in the alignment of the mineral grains in the southern dike. Because of the weakness of the parallelism and the nature of the exposure, it could not be determined if the trends represent planar or linear features or what the dips or plunges might be.

The contacts of these dikes are essentially parallel to the planar structure of the nearby tonalite, at least in strike. The vertical exposures are limited, so the conformity of the dips of the dike and tonalite could not be determined. The contacts are not exposed, except in a few boulders at the southern edge of the southern dike. These boulders are apparently toppled as a result of the weathering and erosional processes. They show the relationships at the contact but do not show the attitudes of the structures. The small inclusions and the mineral grains in the leucogranite are parallel to the contact (Figure 44), as are the planar features of the Bonsall Tonalite.

FAULTS

The recognition of faults, as was previously mentioned, is nearly impossible within the Perris block. The nature of the weathering in the granitic rock and the monotonous lithology of the metamorphic rocks makes fault recognition quite difficult. One exception is noted. This is the Val Verde fault, whose probable trace can be followed for about six miles in the west central part of the tonalite pluton (Figure 3).

Just west of the screen of metamorphic rocks, the log of the Val Verde tunnel, in the archives of the Metropolitan Water District of Southern California in Los Angeles, shows 60 feet of crushed rock. Osborne (1939) said that the trace of the fault in the tunnel was a 10- to 15-foot-wide vertical zone of sheared rock and gouge, which strikes N30°W. According to a map in the Metropolitan Water District (Proctor, unpubl. map), this fault shows surface expression in the right lateral offset of a screen of the Bedford Canyon Formation and of a body of the Woodson Mountain Granodiorite of about 700 feet. A check during this investigation indicates an apparent right lateral displacement of 300 to 400 feet for the screen of metamorphic rocks and an inconclusive relationship of the granodiorite.

One and one-half miles northeast of Gavilan Peak, about 100 yards back into an adit in the tonalite, a gouge zone 12 feet wide strikes N80°E and dips 50 degrees south. Mullions in the fault plane and associated drag and joint features in the workings indicate a right lateral movement with a reverse component in the dip slip. In this same adit, there is another fault which strikes N10°E and dips 30 degrees to the west. The sense of movement cannot be determined for this fault. Neither of these faults show any surface expression.

Study of air photos has led to the identification of several linear features, most of which cannot be recognized on the ground. These lineaments have been plotted on Figure 3. The three lineaments northwest of Perris and the one north of Steele Peak are emphasized by the drainages in their trends, but those to the north and west of Vayemo Peak are not identifiable on the ground.

The lineaments may be faults. The vein structures in three abandoned gold mines may also be faults. At least the lineaments and veins are unusually well expressed fractures.

The faults and suspected faults all lie within one mile of the Val Verde Fault, or its possible extension, except for those north of Perris. The faults and suspected faults strike northwest, northeast, and eastward, and evidently dip steeply except for the shallow dipping faults in the adit and the shallow dipping vein northwest of Vayemo Peak.

Northwest trending structures abound in the Perris block and in the Peninsular Ranges and major right-lateral strike-slip movement is common, so it is not surprising that the major fracture in this area should be a northwesterly trending fault with right-lateral displacement.

Northeast trends are quite rare in the Perris block and in the Peninsular Ranges as a whole. Aside from the big dike of miscellaneous tonalite, the parallel dikes of the Rubidoux Mountain leucogranite, the trends in the nearby metamorphic rocks and the foliation in the tonalite in the same vicinity, there are no evident northeast structures in the Province until one gets about 75 miles to the southeast. There, between the Elsinore and San Jacinto faults, there are many northeast striking faults (Dibblee, 1954, p. 28).

West striking structures are more common than the northeast ones, but still quite rare. Larsen (1948, pl. 1) shows a swarm of dikes that trend just north of west, in an intrusion of Woodson Mountain Granodiorite in the Santa Ana Mountains. The northern end of the Elsinore fault zone has a distinct westerly trend (Gray, 1961, pl. 1). Westerly swings are noted in the flow structures of the Bonsall Tonalite (Figure 3), but for any considerable number of westerly trends, one must go several miles north to the Transverse Ranges Province.

The right lateral sense of movement on the Val Verde fault is the same as that shown by the great faults of the Peninsular ranges and the Colorado Desert (Rogers, 1965). This right lateral movement may mean that the faults that strike northeast in the Gavilan are feather fractures. Those northeast faults, just northwest of Perris, cannot be explained as feather fractures because of their remoteness from the Val Verde fault. They may be a complementary set of faults, and so may these be in the Gavilan; in any case, the northwest and northeast fracture directions are obvious.

The east-striking faults may have been formed during a reversal of the motion on the Val Verde fault, making them feather fractures as well. If their dip is moderate or shallow, they may be thrusts like those in the Transverse Ranges to the northwest (Figure 1), but their traces on the ground seem to be too straight to indicate a shallow enough dip. Once again, it seems that the northwest, northeast, and east fault directions and the right-lateral strike-slip on the northwest direction are the only conclusions that can be reasonably drawn from the evidence.

No conclusions can be drawn about the age of these features, except that they are a result of rupture after the fluid phase of the emplacement of the granitic rocks. It is reasonable to suspect that the Val Verde fault may have its origin prior to the emplacement of the batholith. At least its attitude and sense of movement seem to relate it to the ancient, strike-slip faults of Southern California.

DISCUSSION

The geological events leading to the emplacement of the Southern California batholith and the emplacement itself will be covered in this section. Various possibilities for many of the events will be considered, and the unsuitable choices will be discarded.

Pre-Batholithic Events

The Middle and Upper Jurassic Bedford Canyon Formation (Imlay, 1963 and 1964) represents the great bulk of the pre-batholithic rock in the Peninsular Ranges. This formation is composed of a eugeosynclinal suite in the Santa Ana Mountains, which grades eastward into a miogeosynclinal suite in the Winchester area (Schwarcz, 1960). These rocks have been regionally metamorphosed and the grade of metamorphism increases eastward across the Perris block, therefore in the direction of the miogeosyncline. This unusual circumstance has been described by Larsen (1948) and Schwarcz (1960). No information developed during this work leads to any conclusion about this anomaly.

Foliation in the phyllites and more argillic portions of the Bedford Canyon rocks, adjacent to the Riverside-Perris pluton, is parallel to the relic bedding in nearby arenites. Crossbedding has been identified, but the tops are not evident. The bedding-foliation parallelism has been recognized by many authors who have worked in the Santa Ana Mountains and the Perris block. Rock cleavage in the slaty rocks is imperfect and evidently parallel to the regional, and local, northeast-dipping structure. Of the several authors, who have dealt with the Bedford Canyon Formation, only Dudley (1935) has reported any appreciable change from the northeastward homoclinal dip. He found evidence of isoclinal folding in the metamorphic rocks east of Elsinore. In the vicinity of the Riverside-Perris pluton, a few small folds have been identified (Figures 4 and 21) and other changes from the regional attitude of northwest strike and northeast dip have been identified (Figure 3). These variations appear to be minor and only modify the essentially homoclinal dip of the formation.

Jahns (1954) reported that the metamorphic rocks on the east flank of the San Jacinto Mountains, adjacent to the Colorado Desert province, still have the northeastward homoclinal dip and that no major folding has been identified in the pre-batholithic rocks north of northwestern Baja California.

Nothing in the literature has been found to adequately explain the origin of this homoclinal terrain which is 50 miles wide and 100 miles (or more) long. No evidence in the locality of the Riverside-Perris pluton has added any new information. The solution to this problem seems to be critical to the understanding of the geology of the Peninsular Ranges Province. Even when the width of the interposed plutons is removed and allowance is made for some repetition by folding, as yet not generally recognized, the thickness of the sediments in this homocline is tremendous. The great bulk of this material is Middle and Upper Jurassic. The mechanism for batholithic emplacement proposed here (see below under The Stress Environment) requires only the minimum thickness of pre-batholithic accumulation. The suggested spreading of the walls of the batholith would allow the insertion of the plutons between wedges and screens of pre-existing rock, apparently increasing their thickness. The lack of compression in the walls and screens and the absence of the forcing-up of blocks as overburden also minimize the necessary thickness of sediments.

The possibilities for developing this homoclinal structure include upwarping by lateral compression, upwarping by vertical forces, downwarping by lateral tension, downwarping by vertical forces, and dragging of the terrain by the great right-lateral displacement, with some possible vertical component, along the San Andreas and related faults.

A choice among these suggested mechanisms does not seem possible at present.

Because of the parallelism of the metamorphic foliation and bedding, it appears that the deformation and the metamorphism may have been concurrent or at least that the same stress field applied at both times.

The Temescal Wash Quartz Latite Porphyry may represent the intrusive phase of the Santiago Peak Volcanics of the Santa Ana Mountains (Engel, 1959; Gray, 1961). These rocks, then, would represent the volcanic outpouring that is thought to preceed the emplacement of batholiths (Hamilton and Myers, 1967). The volcanics in the Santa Ana Mountains lie in a zone more than 80 miles long and about 10 miles wide (Larsen, 1948) along the western border of the batholith. The porphyry lies along the western edge of the Perris block, between Corona and Elsinore (Figure 2), also within that 10-mile-wide strip at the western edge of the batholith. Deflections in the structures of the metamorphic septum and in the plutonic rocks, between Steele Peak and Vayemo Peak (Figure 3), suggest the presence of a west-northwest-trending structure which may have been present in the pre-batholithic terrain. The intersection of this postulated structure and the major shear zone of the Elsinore fault may have acted as the locus for the porphyry intrusion.

Emplacement of the Batholith

Emplacement of the San Marcos Gabbro

The San Marcos Gabbro is the earliest batholithic rock (Larsen, 1948) and is evidently the mafic forerunner of the batholith. It is restricted to the western portion of the batholith, but some of its exposures are at least 25 miles from the western border.

In the vicinity of the Riverside-Perris pluton, the gabbro crops out along a westerly or west-northwesterly trend in the vicinity of the disruption described above as a possible localizing structure for the porphyry. This zone may be one broad or two narrow trends. The larger mass of gabbro, which extends off of the mapped area (Figure 3) southwest of Mockingbird Canyon, trends to the northwest. A few miles southeast of the Riverside-Perris pluton, a west-northwest band of San Marcos Gabbro plutons extends for at least 25 miles across the Perris block (Larsen, 1948). The various gabbro plutons could have been more extensive before the intrusion of the later rocks, even to the point of filling the entire space now occupied by the later intrusions.

Possible modes of emplacement of the gabbro include: the intrusion of mafic magmas, mafic metasomatism of the earlier rocks, or some combination. If a magma is involved, other possibilities exist; a mantle differentiation, a crustal differentiation, or the mobilization of mafic-rich sediments and igneous rocks. Of the various possibilities, the production of magma by differentiation of either the mantle or crustal material seems to hold the least promise of being the active mechanism. The wide range of gabbroic and noritic composition makes it improbable that they had a source in differentiated materials. The San Marcos Gabbro shows the least amount of evidence of its intrusive nature, yet the internal parallelisms and inclusions of older rocks makes the assumption of intrusion almost a certainty. Perhaps the intrusion resulted from mobilization of heterogeneous, but mafic, source materials.

Emplacement of the Riverside-Perris Pluton

The importance of the regional northeast-dipping homoclinal structure in the pre-batholithic rocks is evident in the shape of the pluton, the attitude of the septum and the trends of the foliation of the pluton (Figure 3). Other structural controls, although less obvious, are represented by trends in and around the pluton. The west to west-northwest direction was discussed in its relationship to the localization of the porphyry and gabbro plutons. Swings of foliation in the tonalite pluton approach this direction. Drainage lines, present and fossil, take this westerly trend. This same west to west-northwesterly trend shows in the long axis of several plutons in the Santa Ana Mountains and in the average strike of the foliation of the Bonsall Tonalite in the San Luis Rey 30' quadrangle, 20 miles south of the area of the Riverside-Perris pluton (Larsen, 1948). An east-northeast structural trend in the northern part of the pluton, involving the metamorphic rocks, the tonalite foliation, and the leucogranite dikes, has been discussed (see Structural Arrangements of the Flow Phase in the section on the Bonsall Tonalite). The southwestern arc of the pluton also seems to be controlled by this east-northeast trend.

A fourth structural direction in the vicinity of the Riverside-Perris pluton is defined by at least five different features. The east-northeast trends in the northern part of the pluton do not reappear east of the alluvial embayment at the mouth of Sycamore Canyon. The northwest trending septum, east of Vayemo Peak, is very nearly cut through by north- and south-trending embayments of tonalite. In Figure 25, a north-trending alignment of the clusters of dark clots and recognizable inclusions is shown in the central part of the pluton. The elongated grouping of inclusions and clots coincides with a topographic high, which can be identified in Figure 3, about 2-1/2 miles east-southeast of Three Sisters. The ridge is also the site of an unusual number of small dikes.

In Figure 26 is a north-trending alignment of areas of crossing structures. In Figure 32, among several possible alignments, is one of fairly wide variations which trends north between the embayed septum and the truncation of the northeast trends at the north of the pluton. There are also a number of deflections of the strike of the planar array in the tonalite as this same northerly trend is crossed (Figure 3).

These several features, discussed in the previous three paragraphs, define one north-trending lineament in the center of the Riverside-Perris pluton. The number of occurrences on this line places their location outside the realm of coincidence. These features are interpreted to indicate the proximity of a septum along that trend. The septum may have been higher than the present surface and have been eroded away, or it may be still buried. Older rocks were not encountered in the portion of the Colorado River aqueduct tunnel that crosses the trend of the lineament (unpublished logs of the Metropolitan Water District of Southern California). The lineament is evidently a zone of displacement, as shown by the truncation of the east-northeast-trending structures at the north end of the pluton. The nature of the displacement is not apparent, but it was pre-plutonic or synplutonic. A post-plutonic rupture of that scale would show surface evidence and would have been identified in the aqueduct tunnel.

The possibility of minor movement on the dislocation during the emplacement of the pluton is good, and may be demonstrated by the deflection of the northwest-trending planar structures as they cross the lineament (Figure 3). Some of the deflections along this line are to the west and some are to the north. These deflections, the area of crossing structures (Figure 26), and the wide variabilities of flow trends along this line (Figure 32) could be caused by the interference of flow patterns near an obstructing septum. A north-trending spur of the Elsinore fault zone also extends into the Perris block (Figure 2).

The four structural directions of the pre-batholithic framework, shown in the vicinity of the Riverside-Perris pluton, are northwest to north-northwest, west-northwest, east-northeast, and north. The control these trends had on the shape and internal structure of the pluton is evident from inspection of Figure 3.

The Bonsall Tonalite was mobile relative to the walls and shows a variable internal mobility as well (see under Structure in the section on the Bonsall Tonalite). Mobile cells within the mass of more viscous material have caused the usually predominant planar structure to become locally overshadowed or obliterated by the usually subsidiary linear structure. In addition to the increase in the linear array, mobile cells are identified by unusual drawing-out of the clots and development of areas of wavy or swirled parallelism. The size of an individual cell varies from a few tens of yards to a few hundreds of yards. No cell has been mapped as a structural entity, because of the scarcity of outcrops and the generally confused patterns developed within the mobile zones.

The viscous-shear phenomenon (see under Structures of the Intermediate Phase in the section on the Bonsall Tonalite) is another evidence of the internal variablity of the mobility of the magma. The continued motion of one area after the surrounding magma has stopped or slowed, leads to the production of schlieren in some cases (Figure 38). In other examples (Figure 30) there is no tendency to develop schlieren in the viscous-shear planes. Other schlieren developed from strictly flowage phenomena, as evidenced by the gradation in the axial ratios of the dark clots from 1:1 to 1:50 and more. In examples of extreme elongation, it is difficult to determine whether the features are clots or schlieren.

The dark clots are derived from various sources. As was discussed under Petrology in the section on the Bonsall Tonalite, Hurlbut (1935) concluded that the source of the clots was the San Marcos Gabbro, while Osborne (1939) found that the most likely source for the clots was the schists but that some clots originated from the gabbro. Hurlbut's work was in an area where most of the wall rock is gabbro.

Osborne's was in an area where most of the wall rock is schist. The evidence leads to the conclusion that any pre-existing rock will be a source of the dark clots if the mafic content is high enough.

The wide distribution of the dark clots and the wide variation of the rock types within one granitic formation, indicate that the magma failed to reach chemical equilibrium. Some gabbroic material probably was not incorporated in the gabbros as they formed, because the pre-existing rocks did not all receive the thermal and chemical activity at the same rate. Moreover, this lack of uniform transformation may have been made inevitable by the inhomogeneity of the parent rock. Thus, in one area, nearly complete melting could set up conditions for fractionation, of varying degrees, of the gabbro, tonalite, and granodiorite. If the gabbro emplacement were accomplished before the complete fractionation, the remaining material would be likely to move with the tonalite, but remain segregated as gabbroic clots. In another environment, the mass could have been only partly mobilized and the higher melting, less reactive material would be carried along as dark clots with the mobilized rock. The clots, then, would represent the more mafic portions of the source rock of the magma and their wide distribution would be the result of the chemistry of the parent rock.

Material picked up by the magma, as it moved to its final site, would be incorporated to the extent of the ability of the magma to react with it. The more mafic fragments would be, of course, the most difficult to incorporate and would be the most likely to remain as dark clots.

Whatever the source, the permanence of the clots, once past the border zone, is remarkable. It seems improbable that the clots could be in equilibrium with their surroundings if they have such a wide range of plagioclase compositions. The ability of the tonalite to react with the clots and included material must have been very small in most parts of the pluton. Kennedy (1955) suggested that the water in a body of magma tends to have the same partial pressure throughout. He further indicated that this condition would be met if the water concentrated in the regions which were cooler and had the lowest confining pressure. The cooler environment at the periphery and the lower confining pressure at the top of the magma body would concentrate the water, and thus much of the chemical activity, in those regions. Consequently, the magma, whether partially or wholly melted, would be more active at the borders and would tend to react with the fragments broken from the walls, whereas within the mass of the tonalite reaction would be much reduced.

The foliation in the Bonsall Tonalite may be at least partly a relic of the structure of the parent rocks.

The production of dark clots by the partial melting of the less refractory portions of the pre-existing rocks would leave the clot material arranged much as it was before melting, in strata or lenses of varying dimensions. The upwelling of the mass as a whole would preserve these features as relic structures, while the internal mobility of the magma would tend to separate the mafic material and distribute it as individual clots. The flattening of the structure to the east, away from the screen of metamorphic rocks, cannot be explained by reference to the "typical" domed intrusion. A viscous upwelling of material with an initial homoclinal dip to the east would lead to the formation of a structural arrangement such as is seen in the Riverside-Perris pluton. The margins would lag, because of friction, and the interior parts would rise more rapidly, causing any single horizon to dip less and less steeply as the area of greatest movement is approached.

Unequivocal criteria for distinguishing between purely flow structures and residual structures are difficult to formulate. Parallelism with the border, where the border is parallel to the regional trend, is not unequivocal evidence. Parallelism to a border at moderate to large angles to the regional trend or the parallelism to the edges of an embayment are criteria for flowage dominated structure. The cells of local mobility, which were mentioned as evidence for the flowage of the magma, are also flow induced structures. The areas of extraordinary concentration of dark clots may represent relics of once larger concentrations of mafic material which were broken-up and spread-out by differential flowage. The areas of crossing structures may be caused by the development of flow, at angles to the relic structure, which was frozen before the older structure could be obliterated, or they may represent two or more generations of flow structures.

During the upwelling of the magma, the intruding rock conformed to the borders while the interior portions remained unaffected by the shape of the boundary. Strike changes of up to 70 degrees in less than one-quarter of a mile were discussed under Contacts in the section on Bonsall Tonalite. The areas in which these swings occur seem to represent the transition between flowage domination and the zone dominated by residual structures.

The lineation, which is varyingly developed in different localities, is roughly parallel to the dip of the planar array. As a general rule, though, it lies a little to the east or southeast of the dip in the plane of the foliation (Figure 3). The lineations may show either the path of entry of the tonalite, the elongation of the pluton, the attitudes of relic linear structures, the presence of local cells of mobility, or they may represent some combination of these possibilities. The general trend of the lineation over the pluton reveals a pluton-wide pattern to the linear structure. An orientation related to the flow from the source of the magma, to the elongation, or to the structures of pre-existing rocks could apply to the pluton-wide lineation pattern. These pluton-wide patterns could all be related to the regional homoclinal structure. If the lineations are related to the path of entry, they would show an up-dip migration of the magma. The same possibility applies to the development of lineation by the elongation of the pluton. The relationship of relic linear structure to the attitudes of the original material is obvious, but there is no evidence in the screens and walls of a lineation in the prebatholithic rocks. In any of these cases, the east and southeast deflection of the lineation, compared to the dip, could be explained by the right-lateral drag from the major northwest-trending faults in the vicinity. Local increases in intensity and changes in orientation of the linear parallelism could be related to the variation in local mobility. Responses to new environments could either be along the earlier formed structures or along newly developed ones.

The rupture pattern in the pluton may be related to the dynamics of the emplacement and cooling sequence, to regional stresses after the solidification of the mass, or to a combination of both. The shallow dipping fractures, which dip into the tonalite, near the contacts of the pluton seem to represent the feather joints of granite tectonics or, in some cases, marginal thrusts. Their location, near the borders, connects them with the dynamics of the intrusion. The shallow, southwest-dipping joint set, which is perpendicular to the lineation, is also associated with the intrusive and cooling sequence of the magma, as is the dike system which followed these joints. The remaining joints are not attributable to the primary structures of the pluton, with the possible exception of those parallel to the foliation. Joints that dip steeply and strike parallel to the planar array in the tonalite are mostly parallel to the dip of the foliation as well. While joints parallel to the foliation in granites are not discussed in Balk's (1937) summary of granite tectonic phenomena, it seems logical that there would be a propensity for a joint set in this orientation.

The evidence compiled during this study does not eliminate the possibility that the joints, except those containing dikes, could have been formed at any time after the consolidation of the batholithic rocks. Lack of evidence of later tectonic activity, except for that associated with the great faults of Southern California, leads to the assumption that the joints are related to the batholithic emplacement.

The possible modes of plutonic emplacement are; as a melt, as a partial melt, as a rheid, by granitization, or by some combination. As was indicated in the section on Structure under the Bonsall Tonalite, structural evidence eliminates the possibility of granitization for the Riverside-Perris pluton. Metasomatic processes are not discarded as possibilities leading to the formation of a molten fraction. Structural criteria for differentiating between magmatic and rheid emplacement are unknown. A possibility exists that there would be a different behavior of the dark clots in a rheid. The clots might be more deformed, like schlieren, under plastic deformation, whereas in a melt or partial melt, the flow would take place in the fluid phase between the clots. The suggestions, above, that the clots may have been formed as a result of incomplete differentiation or partial melting would place limits on the choice between rheid and molten flow. The preservation of the relic structures and "stratigraphic" distribution of the dark clots, discussed above, is possible with either the rheid or fluid flow.

The sequence of the insertion of the aplite and pegmatite dikes and the intrusion of the miscellaneous tonalites is not demonstrated by visible cross-cutting relationships in the Riverside-Perris pluton. The fact that the aplite-pegmatite dikes and quartz veins follow primary fractures suggests that they were more closely associated with the intrusion of the pluton than the dikes of miscellaneous tonalite, whose orientations are parallel to directions not generally associated with primary features. The tensional stress caused by the cooling of the magma is postulated to have opened the cross joints, perpendicular to the direction of elongation, and to have allowed the insertion of the aplite-pegmatite dikes and quartz veins. An easterly oriented tension direction is necessary to explain the miscellaneous tonalite dikes. This easterly oriented tension is not consistent with the primary stresses in the pluton. It is suggested, then, that the miscellaneous tonalite was intruded after the other dikes and veins.

Emplacement of the Woodson Mountain Granodiorite

The Woodson Mountain Granodiorite was emplaced along the western part of the batholith (Larsen, 1948), as were the gabbro and the porphyry. Of the various mechanisms for the emplacement of plutonic rocks, granitization must be eliminated as the active agent for this rock because of the flowage evident in both the dikes and in the main mass of the granodiorite near the Riverside-Perris pluton.

Insertion of the Rubidoux Mountain Leucogranite

The Rubidoux Mountain leucogranite was intruded into the tonalite in the same structural direction as most of the miscellaneous tonalite dikes and the Woodson Mountain

Granodiorite dikes in the Gavilan. They all took the orientation parallel to the foliation in the Bonsall Tonalite. The time of emplacement and the orientation of the stresses were evidently different than during the emplacement of the earlier dikes. A northerly orientation for the tension direction is suggested for the time of emplacement of the leucogranite.

Interactions Between the Plutons and Their Walls

The proximity of the rocks in the intrusive sequence is a measure of their chemical similarity and it is a generality that the proximity in time of two rocks in the intrusive sequence will determine the amount of contact effect developed at their common border. The gabbro is more affected by the intrusion of the tonalite in the western and southern Gavilan than it is by the later intruding granodiorite in the same areas (see under Contacts in the sections on the San Marcos Gabbro and the Bonsall Tonalite). The tonalite is considerably affected, in the northwestern Gavilan, by the Woodson Mountain rocks, but not at all by the leucogranite near Riverside (see under Contacts in the sections on the Bonsall Tonalite and the Woodson Mountain Granodiorite).

In addition to the effect of the proximity in the intrusive sequence, there seems to be an effect by the overall heating of the region as the sequence progresses. The common lack of parallelism within the granodiorite may be attributed to the higher regional temperature, as well as to the rarity of inclusions and the generally leucocratic aspect, which tend to inhibit the recognition of the structure. The higher temperature, later in the sequence, is further evidenced by the differing nature of the contact between the tonalite and the porphyry on the one hand and the granodiorite and the porphyry on the other. In the central Gavilan, about 1-1/2 miles south of Gavilan Peak (Figure 3), the tonalite is granulated and very probably chilled (Figure 12) at the contact with the porphyry. In that locality, the gabbro is more than one mile away. One and one-half miles north and south of the gneissic outcrops, the tonalite does not show the gneissic character near the porphyry. The gabbro had previously invaded these areas, raising the rock temperatures. Two miles northwest of Steele Peak, the Woodson Mountain Granodiorite intrudes the porphyry in a region already occupied by gabbro and the tonalite. The granodiorite shows only a weak foliation adjacent to the porphyry (Figure 22). This weak structure seems to be the result of the warming of the rocks by the succession gabbro, tonalite, and granodiorite.

The imposition of structure on the Temescal Wash Quartz Latite Porphyry by the later intrusion of the plutonic rocks has been suggested by Larsen (1948) and Pampeyan (1952). The imposition of structures on the San Marcos Gabbro, bordering on the Riverside-Perris pluton, is probable. The weakness of the parallelism in the gabbro and the poor quality of the exposures makes proof of the imposition very unlikely. The zone of mixing in the border regions between the tonalite and the gabbro indicates that the gabbro was soft enough to be deformed. The development of stronger parallelisms in the gabbro near the tonalite contacts and the concordant development of these parallelisms with the contacts, indicate that the deformation was imposed.

The Stress Environment During the Emplacement of the Batholith, as Inferred from the Structures in the Riverside-Perris Pluton

The structures in the borders of the granitic plutons in the Riverside-Perris region do not show abundant evidence of forceful intrusion. Deformation of the wall rock, formation of a border gneiss plate, formation of marginal thrusts and feather joints are not evident or are developed only on a very local scale. The presence of areas of very weakly developed parallelisms in the borders of the plutons also militates against the forceful intrusion of the granitic rocks. However, the local development of border gneiss, the imposition of structure on the gabbro and porphyry and the development of a few fractures that represent marginal thrusts indicate that the intrusions were not entirely passive.

Stoping, while present in the tonalite pluton, was not very important as a mechanism. It modified the site of emplacement and, perhaps, the pathway to that site. Space can be made by raising the roof over the pluton, by spreading the walls, or both. These motions may be induced by the magma, its entry may be permissive, or the mechanism may be a combination. A molten, or partially molten body of rock, having the lower density, will be out of gravitational equilibrium with the same rock in an unmolten state and therefore tend to rise. A certain amount of upward advance of the magma can be allowed by the uplift of the overburden by the buoyant effect of the molten rock. This gravitational differential would allow the melt to rise into contact with cooler terrains.

The difficulty of driving the pluton into the cooler regions of the crust is inherent in the physical and chemical conditions which prevail at those levels. The intrusion has lost nearly all of its thermal energy and chemical activity, as shown by the observed lack of contact metamorphic effects. It has lost much of the bouyancy of the gravity differential at the same time that it is getting stiffer and harder to move. The lifting of the cover and the spreading of the walls by the stiffening magma would have a strong expression in the structure of the marginal zones. Under these circumstances, the margins should show evidences of forceful intrusion. Since these evidences are not present, the forceful spreading of the walls by the intrusion is eliminated from the possibilities. If we postulate the lack of a tensional environment, we are forced to explain the lateral compression of up to 50 miles of rocks into only a few miles, or explain the uplift and erosion of tens of thousands of feet of overlying rocks between the intrusion of the last of the batholith and the deposition of the marine Cretaceous rocks in the Santa Ana Mountains. These sediments lie on tonalites and other granitic rocks (Rogers, 1965) and carry clasts of them (Jahns, 1954). Schwarcz (1960) estimated that 98,000 feet of Jurassic sediments would be required in the Santa Ana Mountains - Winchester interval in the absence of repetition by folding or by "prying apart" by the interposed masses of plutonic rocks. This thickness could be made unnecessary by the pulling apart of this portion of the Peninsular Ranges.

In a tensional environment, the plutonic rock will fill space as rapidly as it is created. Its bouyancy will aid its advance against the overburden. The postulated tension would lead to a thinning of the overburden. In fact, the only material to overlie the intruding plutonic rocks would be sediments which would slump or otherwise be deposited into the growing rift. Magmas which reached the surface would add volcanic material to the sediments filling the rift. The overburden could still be inconsequential, compared to that traditionally considered in the emplacement of batholiths.

The tension, regardless of its orientation, would tend to be relieved by the formation of a widening fracture or fracture zone that would form parallel to the existing structural grain in the rocks. In the Peninsular Ranges Province of Southern California, the northwesterly grain was established by Jurassic time (Larsen, 1951). Because the batholith does not pinch noticeably to the north, it is likely that the northern end of the fissure was widened along some transverse structure. The Transverse Ranges, the location of the Texas lineament in Southern California, trend just south of east and would seem to provide just such a structure. Additionally, any easterly orientation of the tensional stress, except just south of east, would also tend to open up the Transverse Range area to batholithic intrusion. The Transverse Ranges have a large volume of plutonic rock, consisting mainly of granites to quartz diorites (Bailey and Jahns, 1954). The dikelike aspect of many batholiths is manifest.

The release of pressure in the tensional environment would assist the melting of large volumes of rock, if the melting is not dependent on the amount of water or other volatile materials, which would escape with the pressure release. This lowered pressure could lead to the partial melting, mentioned above as a possible source of the clots, as well as the increased mobility of already molten rock. The increased mobility would tend to interrupt its fractionation. The relative cooling effect of the lessened pressure would be greatly subdued by the low thermal conductivity of the rock. Large volumes of magma were produced. This structural study has not developed evidence about the generation of the magma beyond the probability that the magmas were developed in a tensional environment, or at least survived the effects of a tensional environment, and were emplaced in the mobile state.

The spreading of the crust and the advance of the granites would be greatly enhanced, once begun, by the bowing down of the M discontinuity in response to the decreased pressure. A phase change at the discontinuity would increase the volume locally and would help make up the loss of volume resulting from the rising magma.

The upwelling of the magma would follow the gradual widening of the pathway so that only in the middle of a pluton would the spreading be evident, and then only if it were without an immediate successor. Each succeeding intrusive body in the complex batholith would face a similar situation and would only have to adapt to a small percentage of the spreading. The succeeding, more fluid body would find it easier to meet the newly given space requirements than the cooling, stiffening body already in place. Local inhomogeneities would cause a certain randomness of the axes of the various plutons, but the general trend would be obviously parallel to the locus of the spreading, that is parallel to the major structural direction of Southern California.

The permissiveness of the environment would be consistent with the lack of development of flow structure adjacent to the walls in certain border zones of the Bonsall Tonalite. A rate of intrusion greater than that permitted by a purely permissive mechanism would cause the formation of some border effects. The possibility that the septum was carried along by the magma (see Contacts in the section on Bonsall Tonalite) cannot be discounted. The upward transport of the septum by the stiffening magma seems to be consistent with the model of permissive emplacement. The lateral displacement of the wall rock screens, postulated by Hamilton and Myers (1967), is not consistent with the evidence in this pluton. A stronger shell of structures would be developed at the screens and perhaps the structures would even show a piling-up on the lee side. Stoped blocks would not drift away from the lee side, and blocks are found on both sides of the septum in the Riverside-Perris pluton.

Origin of the Fault Pattern

The faults and suspected faults in the vicinity of the Riverside-Perris pluton may be related to pre-batholithic structures or they may be superimposed on the rocks by later stresses and be unrelated to pre-existing faults. The structural evidence is so scanty that the resolution of the problem seems impossible. The suggestion that batholiths, once they are formed, act as resistant buttresses and cause the accumulation of structures around their borders by the deflection of structures that would normally pass through them (Hamilton and Myers, 1967), seems to have merit. The Val Verde fault would have been pre-existing with continued activity during and after the emplacement of the batholithic rocks, by analogy to the great faults of Southern California that also strike to the northwest.

CONCLUSIONS

The eugeosynclinal to miogeosynclinal suites of Middle and Upper Jurassic rocks, including the Bedford Canyon Formation, were regionally metamorphosed prior to the introduction of the plutonic rocks. The grade of metamorphism increased eastward across the Santa Ana Mountains and Perris blocks from the greenschist facies on the west to the hornblende hornfels facies on the east. The rocks were deformed into the broad homocline, dipping steeply northeast, during this metamorphism. The evidence available does not allow a definitive choice between the possible modes of formation of the great homocline. The proposed regional tension during the emplacement of the batholith may make a compressional origin of the homocline untenable.

The volcanic forerunners of the batholith are represented in the Riverside-Perris area by the Temescal Wash Quartz Latite Porphyry. The porphyry stock entered the zone of volcanism on the western margin of the site of the batholith where the Elsinore fault zone is intersected by the west-northwesterly disruption that passes through the Gavilan. The earlier extent of the gabbro is not evident, but it could have been much more extensive than at the present. The source of the gabbro is suggested to have been the mafic-rich sedimentary rocks and volcanic rocks of the geosynclinal sequence. The intrusion of the gabbros along the west-northwest trends seems to indicate the presence of permeable structures in that orientation. They may be related to the influence of the nearby Texas lineament, a structure of transcontinental magnitude.

The Bonsall Tonalite was formed by the rheomorphic or anatectic mobilization of the earlier geosynclinal sequence. The dark clots were formed as a result of the incomplete melting of the pre-existing material. The magma was intruded into a tensional environment in which certain localities were the sites of restricted forceful effects brought about by the inhomogeneous mobility of the magma and the non-uniform rate of spreading of the batholithic rift. The site of the Riverside-Perris pluton was influenced by pre-existing structural controls, most of which are of unknown nature. Their trends, however, are evident: northwest to north-northwest, west-northwest, east-northeast, and north. The insertion of the greatest number of veins and dikes was accomplished during the time when the cross joints were opened by the tensional forces parallel to the direction of elongation of the magma. The large dikes of miscellaneous tonalites and Woodson Mountain Granodiorite were emplaced during the dominance of the regional tensional environment.

The dikes of the leucogranite intruded an analogous structural site, parallel to the foliation, but seemingly due to another stress environment, possibly related to a local reorientation about an unexposed metamorphic buttress.

The data collected during this investigation do not lead to any conclusions about the Woodson Mountain Granodiorite, except that it is an intrusive body younger than the Bonsall Tonalite.

When applied to the batholith as a whole, the conclusions drawn from the Riverside-Perris area present a fairly unified picture. Mobilization, at different times, of the pre-batholithic rocks of various chemical composition, perhaps coupled with differentiation of the magmas, lead to the intrusion of the various granitic formations. These magmas, mobilized in a tensional environment, intruded the overlying rocks and were emplaced in the ever-widening interval between the walls of the batholith. As they rose towards the surface, they were overlain by an ever-thinning overburden. The gravitational force was sufficient to move the magmas up into the thermal zone where they were emplaced, possibly aided by the effect of the volume increase if the mantle material went through a phase change in response to the decreasing pressure. Structural evidence points to the possibility of the batholith being emplaced in a continental analogue of the rifts in the oceanic rises. This analogous situation would give a source of heat, and possibly material, for the formation of batholithic magmas and simplify the mechanics of the emplacement of the batholith by eliminating the space problem and opening a free path for intrusion.

The results of this study indicate that further work in this area would be desirable. Particularly fruitful subjects of investigation would be the petrologic studies dealing with the nature and origin of the dark clots and the tonalite in general, with particular emphasis on the presence of relic minerals and fabrics.

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Figure 3.

The Riverside-Perris Pluton
and Associated Rocks,
Western Riverside County, California.

Topography from U.S.G.S. 7.5', 1:24,000 Series
Geology by William W. Jenney, Jr., 1962-1968